# Geology in the Oceania region: A Preface

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The papers in this issue of *Episodes* are introduced by the Secretary of the 34th International Geological Congress to briefly outline the scope of the forthcoming congress and the geographical region that it emphasises. Of the following 6 framework papers, 5 provide broad outlines of the geology, structure and mineralisation of the major land masses of the Oceania region (Australia, New Zealand, New Guinea and New Caledonia) and one provides historical details of the early geologists in Australia. These papers generally represent several decades of work in the field described and provide thorough, up-to-date reviews of their subject matter. Then there are 21 papers describing aspects of the geology in much more specific regions which some of the planned field trips associated with 34th IGC will visit (Figure 1). These range from discussion of astrobiological aspects seen in the Flinders Ranges of South Australia and Precambrian stromatolites from the Pilbara region of Western Australia to details of modern reef growth along the 2,300 km long Great Barrier Reef. The settings for major Au mineralisation within the Yilgarn Craton of Western Australia (e.g., at Kalgoorlie), base metals in Tasmania and the Mount Isa Inlier (Qld) and iron-oxide Cu-Au mineralisation at Olympic Dam are described. Thus, these papers provide useful backgrounds to many of the field trips associated with the 34th International Geological Congress in Brisbane.

Reviewers from the Oceania region, Europe and North America are individually acknowledged within the various papers. However, we would like to thank the reviewers of earlier drafts of these papers for providing constructive reviews within a tight time frame.

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***Peter Jell*** *is a senior project officer with the Geological Survey of Queensland, coordinating and editing a new edition of the "Geology of Queensland". His background is in paleontology, specialising in trilobites, echinoderms and insects across all ages throughout Australia.*

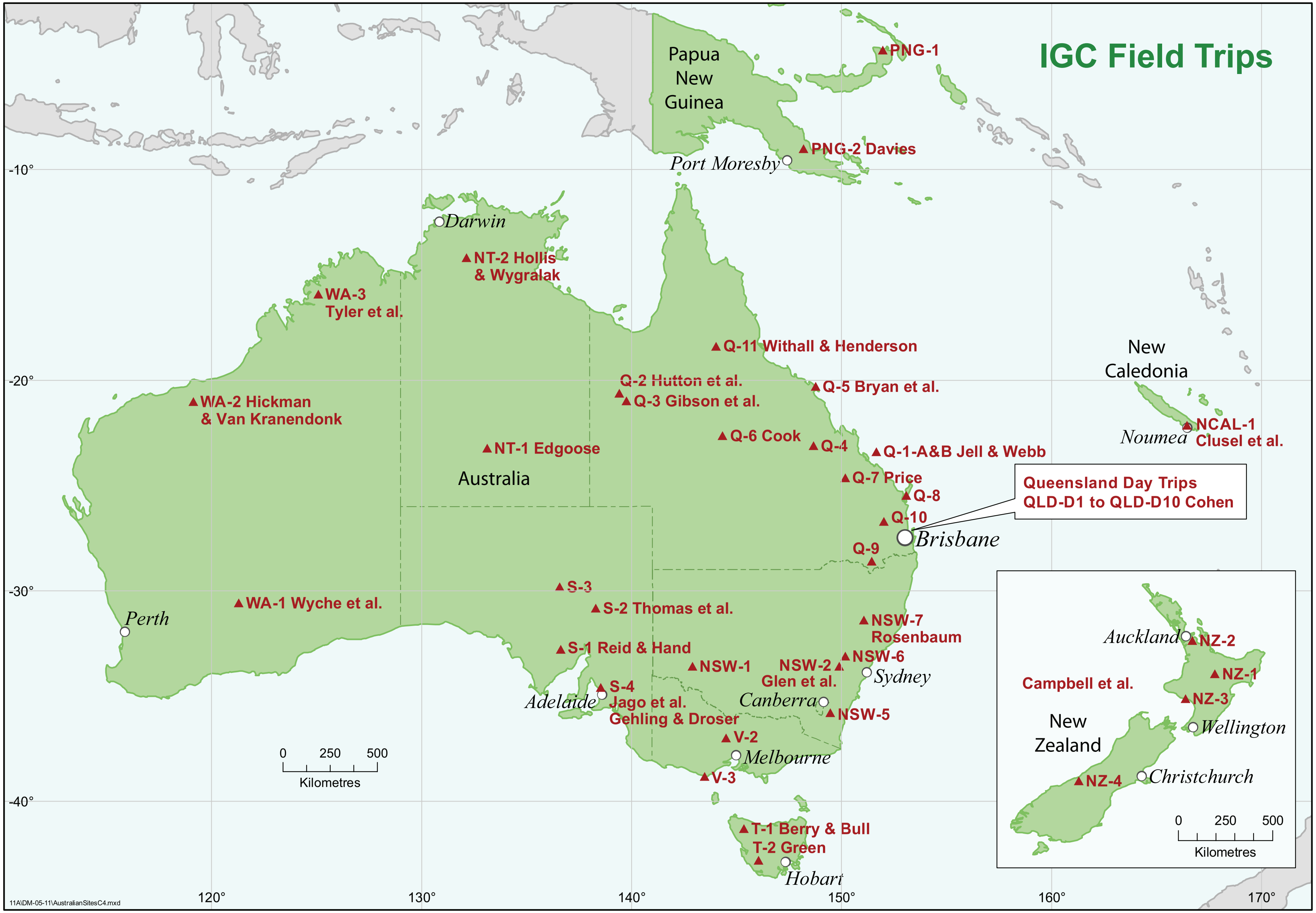


*Figure 1 Locations of 34 IGC field trips (see also www.34igc.org), with background papers within this issue also indicated by authors' names.*

by Ian B. Lambert

# Showcasing the geology, resources and innovative geoscience approaches "Downunder": the 34th International Geological Congress

Geoscience Australia, Secretary General 34th IGC

## Introduction

Since its inception in 1878, the International Geological Congress (IGC) has grown progressively in size and stature to become the largest global geoscience meeting, attracting thousands of delegates from over a hundred countries. The International Union of Geological Sciences (IUGS) is the scientific sponsor of the IGC, which is generally held every four years and is sometimes referred to as the Geoscience Olympics, given that quite a few congresses have been held in the Olympic month and year. That is the case for the 34th IGC, which is being hosted by the Oceania region (Figure 1) in Brisbane, Queensland, Australia, during 5–10 August 2012 (www.34igc.org).

This Special Issue of *Episodes* is designed to stimulate interest in the 34th IGC by describing the geological framework and evolution of the Oceania region and highlighting the geology of many of the areas where field trips are being offered. This article provides an overview of geological highlights, research and teaching in the region, before outlining features of the Congress.

## Geological features of Australia

While Australia is proudly a 'new world' country, it is also known as the oldest continent. As much as the antiquity of its rocks, this is because of the age of its landscapes, with which Aboriginal Australians – the longest surviving continuous culture on Earth – have a powerful connection.

Western Australia is dominated by the Archean Pilbara and Yilgarn cratons. It is home to the oldest dated minerals in the world (c. 4.4 Ga detrital zircons from Jack Hills) in the northern Yilgarn Craton, the oldest identified microfossils and stromatolites (in c. 3.5 Ga chert near Marble Bar in the Pilbara Craton). Western Australia also has evidence of probable Archean meteorite impacts; world class Au and Ni provinces in late Archaean greenstones, diamonds and a magnificently exposed Devonian carbonate reef system in the Kimberley region; the vast Hamersley Iron Formations of the earliest Proterozoic; major bauxite and mineral sand resources in the SW; oil and gas fields on the NW Shelf, currently underpinning our LNG exports; Holocene stromatolites; and pristine modern coral reefs.

The central strip of Australia is dominated by Proterozoic basement terranes and Phanerozoic sedimentary cover. This is a region characterised by high crustal heat flow and "hot" granites; it contains Australia's first pilot hot rock energy project. It has major U, Cu-Au (including the huge Olympic Dam mine) and Pb-Zn-Ag mineralisation (including Mount Isa, Century, Cannington and McArthur River); oil and gas fields in Paleozoic and Mesozoic basins; the Acraman impact horizon; the Wolfe Creek impact structure; Neoproterozoic glacial horizons; ancient fossil macro-organisms in the Flinders Ranges that are the basis for the recognition of the Ediacaran Period; very ancient internal drainage rivers; and several opal fields and other gem and semi precious stone (e.g., sapphires, jade) localities. Tourist icons –including Kakadu in the north, and the world's largest monolith at Uluru (Ayers Rock, in the Uluru-Kata Tjuta National Park) in the arid centre – combine spectacular scenery with ancient Aboriginal culture. The Ranger U mine is an interesting example of effective management of a mine surrounded by the World Heritage listed Kakadu National Park.

The eastern part of Australia is dominated by Phanerozoic rocks. This is where pioneering work was done on the origin of various types of granites and on lithosphere evolution based on mantle xenoliths in Cainozoic volcanic rocks. Queensland has several sites with well preserved fossils of large mammals and dinosaurs, and some large lava caves. Fascinating insights into climatic changes and early human habitation have been found, most notably Mungo man in a lake system in southwestern New South Wales. The Mesozoic–Cainozoic Great Artesian Basin stretches across much of eastern Australia. It hosts groundwaters as old as 2 Myr in what is otherwise an arid continent.

There are vast black coal reserves in Queensland and New South Wales and major brown coal reserves in Victoria. Queensland also has oil and gas fields, including a rapidly growing coal seam methane sector, and large good quality oil shale deposits. The first petroleum giant fields exploited in Australia are in marine environments off Victoria, and there are a series of basins with largely untested petroleum potential along the southern margin of Australia. Onshore Victoria also houses Australia's successful geological carbon storage (CCS) demonstration project, in the Otway Basin.

Australia is unique in having warm ocean currents down both its east and west coasts. Spectacular coastal scenery abounds, from the World Heritage listed Great Barrier Reef which fringes much of the Queensland coast, to the Great Ocean Road of southern Victoria, which is the eastern part of the largest south-facing coast in the world. This wave-dominated coast is home to many endemic marine species. The northern coast in contrast occurs in tropical waters and is dominated by tides. A far lower percentage of endemic species occur here, with many species shared with Australia's northern neighbours.

The island of Tasmania, cut off from the mainland by rising sea-levels c. 6 ka, is characterised by late Proterozoic to Mesozoic geology.

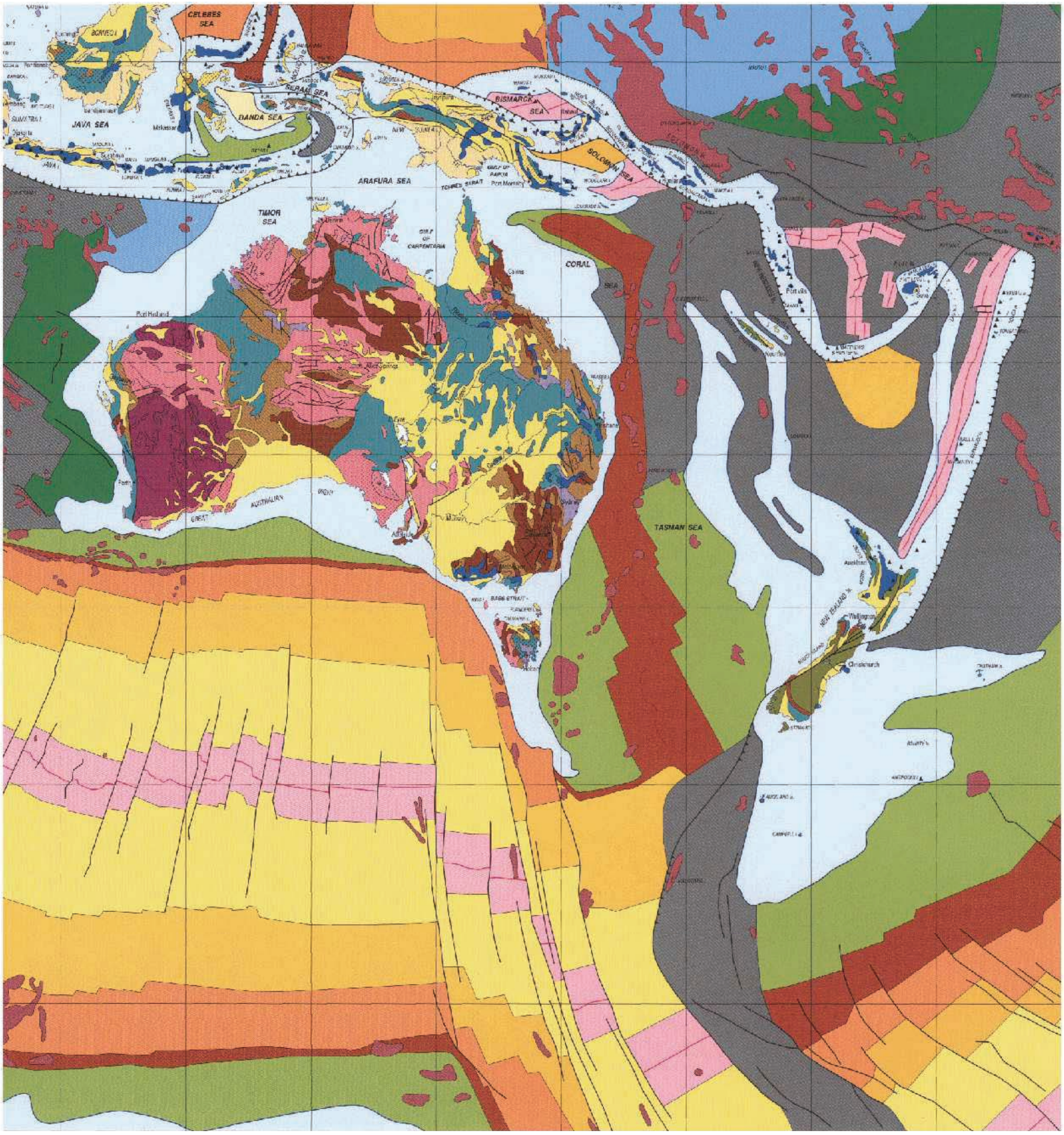


***Figure 1 Oceania and environs. Extract from Geological Map of the World (Commission for the Geological Map of the World).***

Jurassic dolerites, with spectacular landforms, are related to dolerites in Antarctica and South Africa. Scenic wilderness areas abound, including glacially sculpted mountains and lakes and old growth forests.

Australia's unique scenery and its flora and fauna, is complemented by excellent food and produce, including many varieties of world-recognised outstanding wines.

## Geological setting of Oceania

New Zealand and New Guinea and other island countries around the eastern and northern margins of the Australian Plate are located in geologically active settings (Figure 1).

New Zealand straddles the Australian and Pacific plates, whose Mesozoic separation from Australia is detailed in the spreading centre on the floor of the Tasman Sea. The country is noted for its spectacular alpine geology, glaciers, volcanic and geothermal landscapes, strong Maori and British heritages, sheep and some excellent wines. It also has unique fauna and flora. In marked contrast to Australia, it had no mammals other than bats until man came.

New Guinea also exhibits fascinating alpine and volcanic geology in a tropical setting. It features major ophiolite belts, active tectonism and volcanism, uplifted coral terraces, Au and Cu mines. Additional attractions in remote areas include colourful tribes maintaining traditional cultures and customs.

There are many smaller volcanic and coral reef dominated islands dotted through the region from Indonesia to Fiji. New Caledonia, a charming blend of French and Melanesian cultures, is known for its ultramafic rocks which have weathered to form major lateritic nickel deposits, and the ancient floras preserved on its scenic islands.

Australia was long part of supercontinents, such as Rodinia, Nuna and Kenorland. It only became a continent in its own right when it finally broke from the remnants of Gondwana (Antarctica), beginning in the Jurassic. Since then Australia has been part of the fastest moving crustal plate, insulated by extensive areas of oceanic crust from active continent-continent and ocean-continent collision zones, in marked contrast to New Guinea and New Zealand.

The split between Australia and Antarctica was complete by 34 Ma. The geological features of Antarctica include Precambrian gneisses, dry valleys, volcanic activity and the world's largest land glacier. Australian and New Zealand geoscientists have had a long involvement in geoscience research in Antarctica. Recent work has focused on geophysical surveys of the continental shelf, and paleoclimatic trends through detailed studies of cores in ice and offshore sediment.

Through the Cenozoic, intraplate volcanic activity migrated from N–S in at least three parallel tracks down the eastern strip of Australia and beneath the Tasman Sea. During the same period the continental shelf of the Queensland coast became an ideal environment for coral growth leading to massive reef development in the world's largest modern coral reef system, the Great Barrier Reef.

Australia was not subjected to the Cenozoic orogenic activity which occurred in New Guinea and New Zealand. The Australian Continent also escaped the major recent glaciations that scraped surficial materials from wide areas of the northern continents. Its last period of extensive valley and continent-wide glaciation, reflected by striated pavements, U-shaped valleys, and diamictites, occurred during the late Carboniferous and early Permian. From the Mesozoic, there was widespread deposition of riverine, lacustrine and shallow marine strata over older bedrock coupled with climatic desertification due to rapid northward plate motion, culminating in the vast sand dune system in the arid interior of Australia.

The geologically stable setting of Australia has formed and preserved its unusually deeply weathered and topographically subdued landscapes, and the bold reddish hues of the 'outback'. The extensive blanket of weathered rocks and sediments – the regolith – can be several hundred metres thick. The physical and chemical properties of the widespread regolith materials, including groundwaters, can be quite different from surficial materials of other countries typified by shallow cover over fresh bedrock.

The properties of the regolith have to be understood for effective mineral exploration, land/water management and infrastructure planning. Some approaches and technologies used in other developed countries must be modified to take account of the different conditions, and this has been a driver to develop innovative approaches and technologies in Australia. A high proportion of Australia's agricultural production comes from the SE and SW of Australia, where groundwater quality and quantity, and inter-relationships with surface waters, are of concern. Geoscience research is increasingly being brought to bear in efforts to better understand and manage the range of issues involved.

## Geoscience research and teaching

Geoscience in Australia and New Zealand is motivated by several prime factors: continuing to play major roles in meeting global demand for a range of mineral and energy commodities, providing solutions to societal challenges such as groundwater quality and management, satisfying a thirst for new knowledge to understand better the world we live in, making informed land use decisions, building large cities and infrastructure, natural hazards and emergency management.

There is increasing acknowledgement that the outcomes of geoscientific research sustain economies and help safeguard communities. Geoscientists have vital roles to play in multidisciplinary systems approaches to underpin important decisions and policies addressing the major challenges.

Geoscientists are increasingly linked into advances in web technologies and data transfer standards that are making geological and geospatial data more accessible, including for purposes beyond the original incentive for their collection. They are increasingly benefiting from rapid increases in available computing power and in open source processing algorithms which are facilitating analyses and modelling of large and diverse datasets. Combined, these new developments are rendering multidisciplinary approaches much more effective and transparent.

Geoscience research in Australia is carried out within universities, government agencies such as Geoscience Australia and the Commonwealth Scientific and Industrial Research Organisation (CSIRO), industry, Cooperative Research Centres, Special Research Centres, Major National Research Facilities and State and Territory Geological Surveys. In New Zealand, geoscience activities are conducted in Crown Research Institutes, notably GNS Science, and universities. In both countries industry research builds on government funded programs.

A strategic approach, involving separation of funding for research infrastructure from research programs which use the infrastructure, is proving effective and engendering good collaboration in Australia. While the number of geoscience departments in universities has decreased over the past two decades, the standard of teaching and research is high in Australia and New Zealand. Enrolments in Australian tertiary institutions are buoyant, driven in significant part by a very strong resources sector, which is driving the strong national economy. In Australia and New Zealand, geologists from numerous countries are conducting research and exploration projects.

The universities of Papua New Guinea and New Caledonia offer geoscience courses and research.

## Outline of 34$^{th}$ IGC

As a large and prestigious event, the 34$^{th}$ IGC will provide an excellent opportunity to catch up on advances in international geosciences and to meet and network with leading professionals in the global geoscience community. In addition to a very wide ranging and interesting scientific program, many business meetings will be held during the IGC.

Brief descriptions of some of the features of the IGC follow and detailed information is available at the Congress's website: www.34igc.org.

### Organisation

The organising body for the 34$^{th}$ International Geological Congress is the Australian Geoscience Council Incorporated (AGC). The AGC is a Council of Presidents of the major Australian geoscientific societies, and is the peak representative body for the 7000 or so members of the geosciences profession in Australia.

The number of Australians participating in the Brisbane Congress will be maximised by integrating meetings of the major Australian geoscientific societies into the IGC.

The 34th IGC Organising Committee comprises leading Australian geoscientists, representatives of the main geoscience societies and the Chief Executive of New Zealand's GNS Science.

Carillon Conference Management, based in Brisbane, was appointed in 2008 as congress manager.

### The venue

The IGC will occupy the entire Brisbane Convention and Exhibition Centre, an excellent state-of-the-art venue in an attractive cultural and entertainment precinct. This offers ample space for over 30 concurrent scientific symposia, poster displays and a large exhibition under one roof.

### Scientific Programme

Under the overall theme "Unearthing our Past and Future: Resourcing Tomorrow", the 34th IGC will have a scientific programme covering all facets of the geosciences. It has been designed to appeal to academics, government officials and resource industry representatives in equal measure. The full scientific programme encompasses c. two hundred Symposia grouped under 37 Themes. Of particular note are the plenary "hot topic" sessions, which will cover future minerals, energy and water resources; the geological record and climate change; energy in a carbon-constrained world; living with natural hazards; and the geosciences addressing major challenges of the 21st century.

### Workshops and Short Courses

Some thirty professional development workshops and short courses are being offered, covering a wide range of topics.

In addition, three training workshops are being organised for selected participants from low income countries: (i) Sustainable mining in Africa; (ii) Geological sequestration of carbon dioxide; and (iii) Capacity building in risk modelling for natural hazards in the Asia-Pacific region.

### Field trips

The 34th IGC is offering c. 40 field trips to areas of geological interest across Australia, New Zealand, Papua New Guinea and New Caledonia. A bonus for those travelling to the IGC from Europe and Asia is a trip to Langkawi Global Geopark in Malaysia.

### Exhibition

The Congress will feature a major geoscientific exhibition (GeoExpo), which will occupy two exhibition halls and extend into the corridor spaces outside these halls. There will be a wide range of exhibitors – from resource companies, geosurveys and service providers to scientific publishing houses.

Through the GeoExpo and the scientific programme, Russia and China are jointly releasing and promoting the results of a decade of collaborative geological mapping and geophysical surveys in central and eastern Asia. These are expected to be of considerable interest to many delegates.

## Concluding remarks

The 34th IGC offers an unparalleled opportunity to experience the geology and innovative scientific endeavours of a fascinating region, while networking with international leaders in all aspects of the geosciences. Despite the current economic uncertainties, outcomes of geoscientific endeavour will continue to underpin real growth in this part of the world: this is your chance to part of it by attending the 34th IGC. We are extending a warm welcome to geoscientists from around the world to come to Brisbane in August 2012.

## Acknowledgements

The constructive comments of Richard Blewett, Clinton Foster, David Denham and Alan Whitaker are gratefully acknowledged.

*by Brian L. N. Kennett[1] and Richard S. Blewett[2]*

# Lithospheric Framework of Australia

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*The Australian continent comprises an amalgamation of cratonic elements onto which there has been significant Phanerozoic accretion in the east. The result is a complex lithospheric structure with a broad span of ages of material at the surface. The continent is moving rapidly to the north at c. 7 cm/yr, relative to Asia. The collisions with the Eurasian and Pacific plates to the north, coupled to the interaction with the Pacific Plate along the eastern plate boundary through Tonga and New Zealand result in a complex pattern of stresses that is reflected in a moderate rate of intra-plate earthquakes. Extensive geophysical investigations at a continental scale have revealed details of the nature of the lithosphere. The lithosphere is thick (200 km or more) and seismically fast beneath the Precambrian domains of the centre and west of Australia and thins to the east, in a series of steps, to c. 80 km in the Tasman Sea. Large gravity anomalies in the centre of the continent attest to complex deformation in the Phanerozoic that has left a residue of domains with rapid changes in crustal thickness. The development of Australia's generally thick lithosphere has exerted fundamental control on the overall tectonic stability and consequent landscape evolution, the distribution of earthquakes and associated seismic risk, the evolution of sedimentary basins, as well as heat flow and other resource endowment.*

## Introduction

Classical plate tectonic theory has the outer layer of the solid Earth covered by rigid tectonic plates. Globally, there are 14 large and about 40 small tectonic plates, ranging in size from the Pacific Plate, which comprises 20.5% of Earth's surface, to the Manus Microplate, which comprises only 0.016% of the Earth's surface area (DeMetts et al., 2010). These tectonic plates are thought of as nearly rigid blocks of lithosphere, defined by their boundaries and their trajectory across the Earth's surface.

Australia, with an area of 7.69 million km$^2$, is the Earth's largest island and smallest continent. The continental landmass, c. 3,700 km N-S and 4,000 km E-W, occupies a significant part of the Australian Plate, which is currently separating from the Indian Plate in a diffuse zone in the Indian Ocean. Since its separation from Antarctica at c. 80 Ma, Australia has been moving steadily northwards, currently at c. 7 cm/year with respect to a hot spot frame of reference. Australia is the most rapidly moving continent on the globe, and has swept into the southern fringe of Asia with current collision active in Timor and Papua New Guinea (e.g., van Ufford and Cloos, 2005).

The lithosphere comprises two main components: the crust and the thicker lithospheric mantle. The nature of the surface is defined by its geological lineage and has a strong influence on local topography. Information at depth largely comes from geophysical investigations. Airborne magnetics provide strong control on the upper crust, whilst gravity observations are sensitive to deeper structure. Most information on the lower crust and the mantle lithosphere comes from seismological studies. Controlled source studies can provide detailed pictures of crustal structure and architecture, whilst passive seismic studies using distant earthquakes provide information on deeper structures. A relatively new approach exploiting the seismic noise-field is able to provide additional information on crustal structure.

Although the definition of the base of the lithosphere is a difficult problem, the global average continental lithosphere beneath shield areas is c. 200 km thick and beneath tectonically younger areas it is c. 80 km thick (Eaton et al., 2009). The global average crust is c. 40 km thick (Christensen and Mooney, 1995). Despite the average elevation of Australia being only 330 m, the crust shows considerable variation in thickness with substantial areas thicker than 30 km and locally reaching 60 km thick in central Australia. The strong topography on the Moho compared with the surface elevation requires that isostatic compensation occurs at depth, with substantial density contrasts in the mantle lithosphere (e.g., Aitken, 2010).

The best controls on the thickness of the lithosphere come from seismological techniques, particularly surface wave tomography. The large cratonic areas of Australia show substantial portions with relatively thick lithosphere extending to 200 km or more in places. The transition to the shallow lithosphere in the E beneath the Paleozoic fold belts appears to take place in a sequence of discrete steps (Fishwick et al., 2008). The development of Australia's generally thick lithosphere has exerted fundamental control on the overall tectonic stability and consequent landscape evolution, the distribution of earthquakes and associated seismic risk, the evolution of sedimentary basins, as well as heat flow and other resource endowment.

Australia has preserved some of the oldest rocks and the oldest landscapes on the Earth. The continent is also covered by an extensive, but mostly thin, veneer of regolith. The preservation of these features in Australia is a function of the general landscape stability as well as the latitudinal position, which resulted in only minor glaciation during the last ice age. Beneath the regolith and sedimentary cover there lies an amalgam of Precambrian cratons and Paleozoic mobile belts. Understanding the way in which the continental fragments were assembled and the nature of the sutures between them requires the integration of geophysical and geological information on the nature of the whole lithosphere. As a result of major programs at the continental scale, Australia is well endowed with geophysical and geological datasets that allow penetration of not only the shallow cover, but also reveal the nature of the deeper parts of the lithosphere and down to the asthenosphere.

In this paper we provide a summary of the major continent-wide datasets that advance our knowledge of the deeper regions of Australia's lithosphere. With this aim, we make use of potential fields (gravity in particular), basins, radiometrics and heat flow, as well as a variety of both passive and active seismic techniques. We present maps of the main boundaries of the lithosphere, namely: (1) topography/bathymetry, (2) the depth to the Moho, and (3) depth to the asthenosphere. Figure 1 is a map of localities named in the text. Tectonic terms are defined in Table 1 of Huston et al. (2012).

## Physiographic setting of Australia

Australia is an island continent with a landmass defined by a distinctive coastal outline, which has maintained its current shape for the last 6 kyr. At the time of maximum glaciation, the Australian mainland, Tasmania and the island of New Guinea formed a single larger and differently shaped landmass that stretched from the equator to latitude 45°S. With the end of the Ice Age, global temperatures increased, much of the continental ice melted and sea level rose. This caused flooding of the land bridges between Tasmania and the Australian mainland 6 kya, and between Australia and New Guinea 8 kya. The rise in sea level inundated about one seventh of the larger ice-age continent isolating Tasmania, the Australian mainland and New Guinea.

In the absence of major mountain building episodes in the last 200 Myr, much of the present topography of Australia is the result of prolonged erosion by wind and water. Dating of the surface regolith indicates a weathering history stretching back over 300 Myr in some parts of the continent (Pillans, 2008). The landscape was strongly shaped by continent-wide glaciation; large ice caps developed in the Permian when Australia was very near the South Pole, and this glacial influence on the landscape persists to the present day. By the early Cretaceous, Australia was already so topographically flat and of low elevation that a major rise in sea level divided the continent into three landmasses as a shallow sea spread over the land.

Today, the major physiographic features of the Australian continent comprise: (1) a major Western Plateau with localised ranges, (2) the Eastern Uplands with the highest land concentrated along the Great Divide and (3) an intervening zone of Interior Lowlands. The topography is subdued; the highest point on the continent, Mount Kosciuszko, is only 2,228 m above sea level, and the lowest point is Lake Eyre at 15 m below sea level. The average elevation is only 330 metres, the lowest of any of the continents (Figure 2).

Destructive plate margins are generally associated with mountain building, landscape rejuvenation and the formation of large reliable river systems. In contrast, the Australian continent has passive margins on three sides and the fourth is the lower plate of a collision zone in the N. Another important factor shaping Australian physiography is the fact that the continent was already at fairly low latitudes in the Pleistocene, so that glaciation was confined to small areas in SE Australia, with little reworking of the older landscape.

The continental slope on all margins is deeply incised, with steep-sided canyons up to 2 km deep, reflecting either extinct drainage systems or the drowned valleys of current river systems (such as the canyon, to the west of Perth). On the southern part of the eastern margin the continental shelf is rather narrow, with a shelf break closer than 20 km from the coast and the base of the abyssal plain sometimes reaches within 60 km of the coastline (Figure 2). The conjugate southern margins of Australia along the Great Australian Bight and the Antarctic coast, created by the opening of the Southern Ocean, show somewhat wider continental shelves.

The broad continental shelf off Queensland, left behind after the Coral Sea opened, forms a foundation for the Earth's largest single living entity – the Great Barrier Reef (Figure 1) – with 2,900 reefs, 600 continental islands and 300 coral cays created in a mixed siliciclastic-carbonate depositional environment. The wide continental shelf between the coast of northern Australia and Indonesia, Timor and New Guinea represents the drowned remnants of the former single landmass. The extended continental shelf off Western Australia is host to most of Australia's natural gas resources, and is marked by complex embayments and salients left over from peri-Gondwanan fragments now forming basement in Southeast Asia being rifted in the Mesozoic (Huston et al., 2012).

## Tectonic setting of Australia

The Australian Plate is undergoing a complex set of interactions with its neighbours. To the S, an active spreading centre separates the Australian Plate from the Antarctic Plate. This margin developed with the break-up of Gondwana c. 99 Ma, with full separation by 35 Ma. Following a major plate reorganisation in the Pacific Ocean

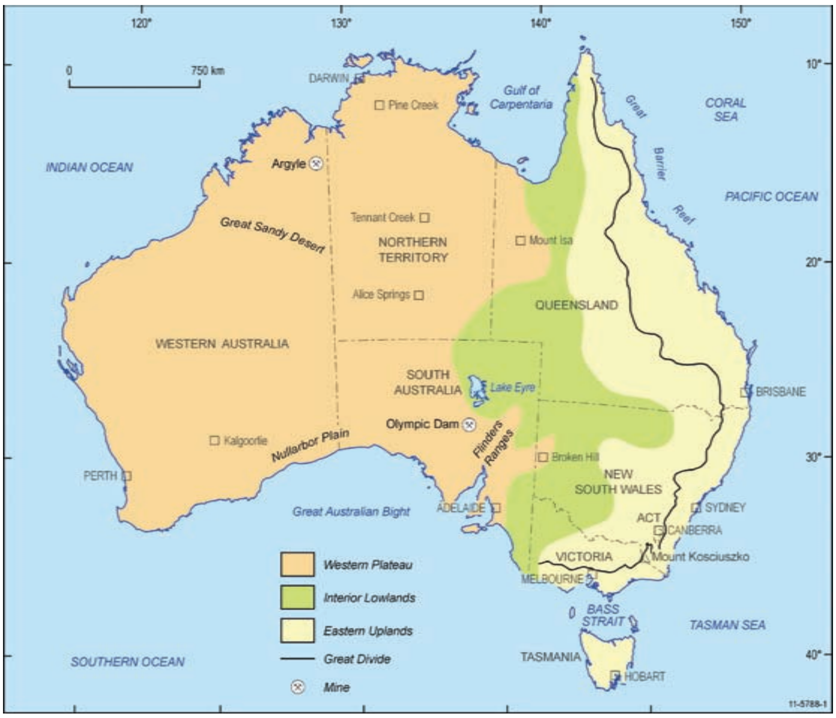

*Figure 1 Map of localities and features mentioned in the text.*

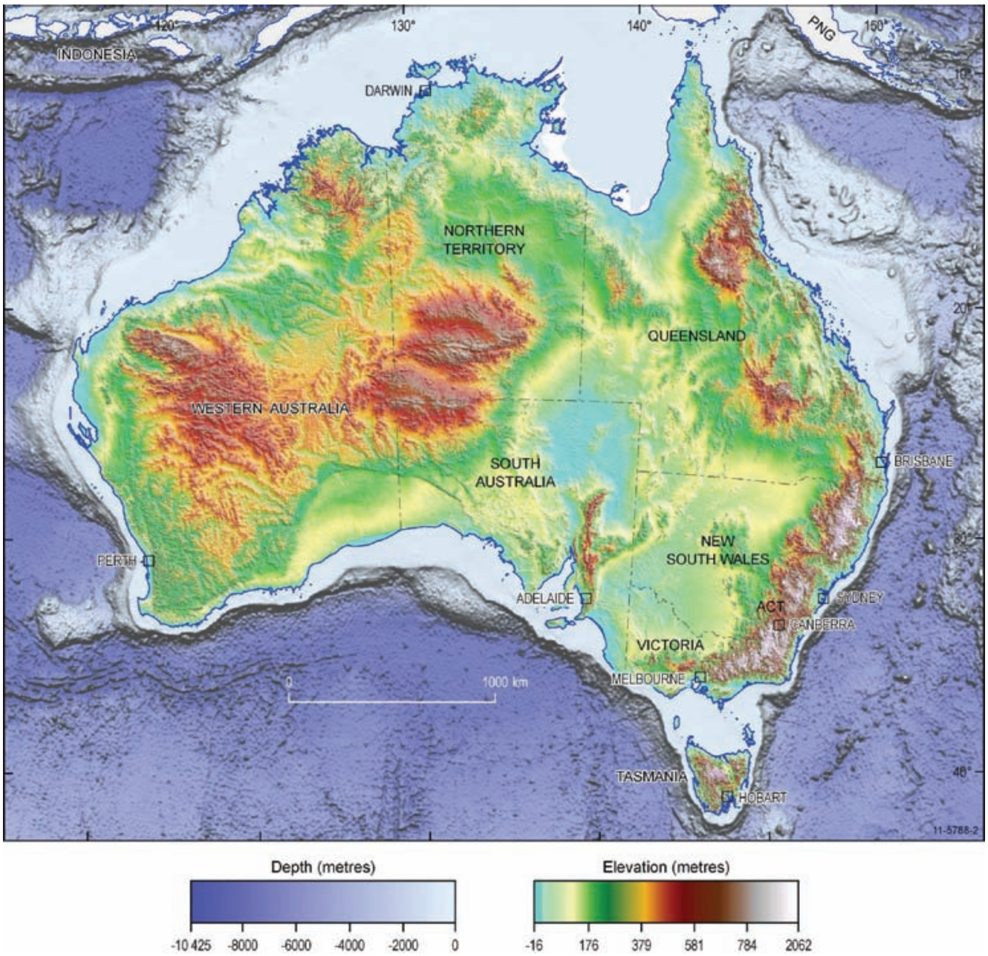


***Figure 2 Topography and bathymetry of Australia (Source: Geoscience Australia).***

to the W the interaction is with the Eurasian Plate, with collision in the Banda Arc region and subduction of the Australian Plate beneath Indonesia at the Java and Sumatra Trenches (Figure 2). Subduction in the Banda Arc region may well already have come to a halt with the arrival of thick buoyant Australian lithosphere that cannot readily descend to depth. The western boundary of the plate is a diffuse zone called the Capricorn Plate, which lies between the Australian Plate and the Indian Plate further to the NW (De Metts et al., 2010).

The boundary forces acting upon the Australian Plate vary from tension in the S and SW to compression in the E and N. The present stress state is largely controlled by compression originating from the three main collisional boundaries located in New Zealand, Indonesia and New Guinea, and the Himalaya (transmitted through the Indian and Capricorn plates). South of latitude 30ºS, the stress trajectories in the Australian continent are oriented E-W to NW-SE. North of 30ºS, the stress trajectories are closer to the present day plate motion, with an orientation between ENE–WSW and NE–SW (e.g., Hillis and Müller, 2003).

at c. 40 Ma, Australia acquired its present N–NE trajectory. Australia has migrated more than 3,000 km along this path at a rate of 6–7 cm/yr over the past 45 Myr, making it the fastest moving continent since the Eocene (Tregoning, 2003).

Following breakup, Australia's initial drift was to the NW. But ridge subduction in the NW Pacific at c. 52 Ma resulted in termination of Tasman Sea spreading, and a change in the Australian Plate vector to its present-day northerly orientation. Sea floor spreading in the Southern Ocean accelerated at c. 45 Ma, and it was not until 35 Ma that full separation of Australia and Antarctica occurred. Australia, as a separate continent, was released from the remnants of Gondwana and it commenced the northward march to meet Asia.

The eastern boundary of the Australian Plate with the Pacific Plate is marked by the Tonga-Kermadec Trench N of New Zealand where the Pacific Plate is being subducted, giving rise to many of the Earth's deep earthquakes. In New Zealand itself, there is oblique subduction along the E coast of the North Island with collision through the South Island moderated by the Alpine Fault. Beneath the Fiordland region of the South Island, the Australian Plate is subducting at the Puysegur Trench. This short subduction zone links to a largely strike-slip boundary passing Macquarie Island and linking to the boundaries of the Antarctic Plate (De Metts et al., 2010).

The Australian Plate is subducting beneath the Pacific Plate along its NE margin at the New Hebrides and Solomon Trenches. The northern boundary of the Australian Plate is complex. Continental Australia is colliding with the Pacific Plate through New Guinea, but

The complex pattern of stress in the continent is expressed in a relative high level of seismicity for what would normally be regarded as a 'stable' intraplate continental region. Recent seismic events with magnitude >6 are not uniformly distributed across Australia but clustered into regions towards the edges of major structural blocks (Figure 4). The rate of occurrence of earthquakes is normally low, but is punctuated by periods of enhanced seismic activity associated with one or more large earthquakes. Since 1901, seventeen earthquakes with magnitude M >6 have been recorded in Australia. The recurrence times for larger earthquakes, such as the 1988 Tennant Creek sequence with three M >6 events in 12 hours, is more than 10 kyr, so that the brief snapshot of seismicity available will certainly be incomplete. Many neotectonic features have been recognised through careful mapping across the continent (Clark et al., 2011). In the Flinders Ranges in South Australia there are clear indications of active faults thrusting Precambrian basement over Quaternary gravels (Quigley et al., 2010). The active intraplate deformation in Australia is likely to be guided by prior tectonic structures and thermal weakening of the lithosphere.

## Geological setting of Australia

Rocks exposed at the surface in Australia span much of the Earth's geological history. The Archean regions include rocks older than 3700 Ma in Western Australia and 3100 Ma in the Gawler Craton of South Australia. The oldest zircon crystals yet found on the Earth, dating

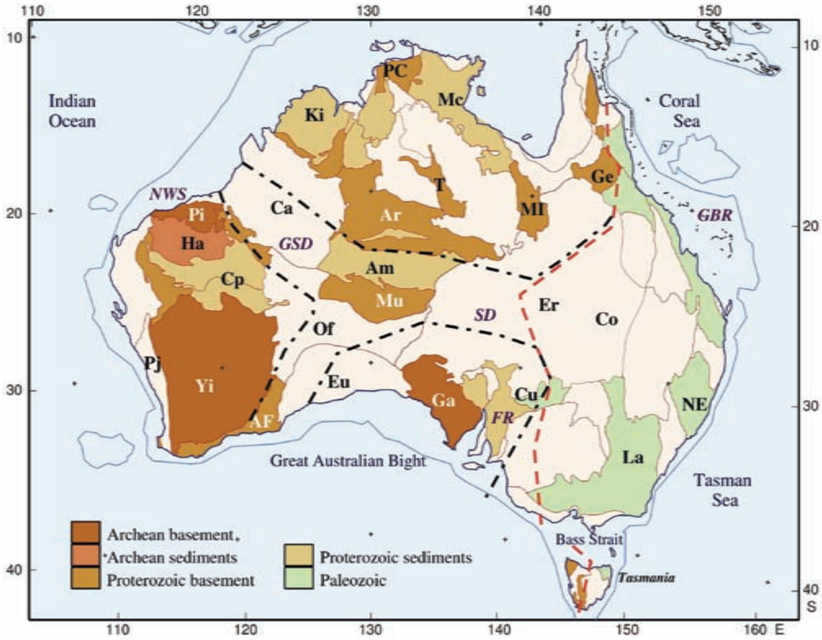


***Figure 3 Simplified representation of the main tectonic features of Australia. The outlines of the major cratons are marked by chain-dotted lines. The approximate location of the boundary between Precambrian outcrop in western and central Australia and the Phanerozoic east (the "Tasman line") based on the reinterpretation by Direen and Crawford (2003) is indicated by a red dashed line. Key to marked features: AF – Albany–Fraser Orogen, Ar - Arunta Block, Am - Amadeus Basin, Ca - Canning Basin, Co – Cooper Basin, Cp - Capricorn Orogen, Cu - Curnamona Craton, Er - Eromanga Basin, Eu - Eucla Basin, Ga - Gawler Craton, Ge - Georgetown Inlier, Ha - Hamersley Basin, Ki - Kimberley block, La - Lachlan Orogen, Mc - MacArthur Basin, MI - Mt Isa Inlier, Mu - Musgrave Orogen, NE - New England Orogen, Of - Officer Basin, PC - Pine Creek Inlier, Pi - Pilbara Craton, Pj - Pinjarra Orogen, T - Tennant Creek Inlier, Yi - Yilgarn Craton; SD - Simpson Desert, GSD - Great Sandy Desert, GBR – Great Barrier Reef, NWS – North West Shelf.***

back to 4400 Ma, occur in Proterozoic conglomerates within the Yilgarn Craton of Western Australia. In contrast, recent volcanic activity in both NE and SE Australia has left distinct volcanic edifices, with the latest eruptions in the SE dated at 4.6 ka.

Much of Australia is covered by extensive Mesozoic and younger sedimentary material (Figure 3). These cover rocks indicate the general tectonic stability of much of the continent from Mesozoic times to the Present. The underlying Australian continental crust was accreted in three major supercontinent cycles, each comprising about one third of the continental area from the Archean cratons in the W to Phanerozoic provinces in the E (Huston et al., 2012). Disparate Archean crustal elements were assembled into three major cratonic zones in the Proterozoic. The West Australian, the North Australian, and the South Australian elements were formed by c. 1830 Ma, and these cratonic elements were joined to the Rodinian supercontinent by 1300–1100 Ma. The fold belt structures of the Phanerozoic Tasman Element comprise the eastern third of Australia, which was accreted onto the eastern margin of the Precambrian cratons (e.g., Collins and Vernon, 1992). The break up of Gondwana, through a series of rifting events from c. 160 Ma, resulted in the formation of the passive margins around Australia, with the formation of the Coral and Tasman seas in the E, the Southern Ocean in the S and the Indian Ocean in the W (Huston et al., 2012). These rift events created the accommodation space for the Mesozoic sedimentary basins that host most of Australia's hydrocarbon resources.

There has been significant subsequent volcanism; in the Mesozoic, Australia was the continental margin of the subducting Pacific Plate and subsequently a chain of hot-spot related volcanism has developed through eastern Australia. The eastern margin of Australia has been influenced by sea-floor spreading in the Tasman Sea from c. 80 Ma and back-arc spreading in the Coral Sea.

The eastern seaboard, including Tasmania, is a patchwork of Paleozoic metamorphic, sedimentary and igneous rocks. These rocks are revealed, as highlands, due to the rift-flank uplift generated by opening of the Tasman and Coral seas. The Flinders Ranges, a Y-shaped region of uplifted Neoproterozoic sedimentary rocks in South Australia, attest to the influence of regional compression across the Australian Plate. Across northern Australia, large areas of mostly Proterozoic metasedimentary rocks occur in the Kimberly, Pine Creek, Macarthur and Mt Isa areas. These basins were filled with vast sandsheets during a time when the Earth's land surface was devoid of the stabilising influence of life, and became the containers for major base metal and uranium mineral systems.

## Penetrating the cover: the lithosphere of Australia

### Potential Fields

Much of the Australian continent is covered by a thick layer of regolith, which can exceed 100 m in many places. This regolith is a product of long term weathering largely unaffected by glacial action. In consequence, potential field geophysics is of major value in providing insight into buried rocks and structures beneath. As a result of the collective efforts of the state and territory geological surveys and Geoscience Australia, most of Australia has been covered with airborne magnetic surveys with high-quality data that provide valuable information on the upper part of the crust. Australia is covered by regional, 10 km spaced, gravity data, but many parts of the continent are sampled on a 1–2 km spaced grid.

The broad features of the structure of Australia appear clearly in the upward continuation of the gravity field across the continent (Figure 5). At an altitude of 25 km the small scale features linked to surficial structure are suppressed, and the major division of the continent can be seen. The western portion has pronounced negative gravity anomalies associated with the thick cratonic roots in the mantle with relatively low density so that they are buoyant. There is a N-S band of positive gravity anomalies near 140ºE, and a zone of weak positive anomalies towards the eastern seaboard. Even in the upward continued image there are considerable contrasts, and a group of alternating bands of E-W trending anomalies in central Australia with higher gravity values is prominent. These features dominate the surface gravity field and are associated with substantial localised changes in crustal thickness in the zone affected by the Devonian Alice Springs Orogeny and the c. 550 Ma Petermann Orogeny in

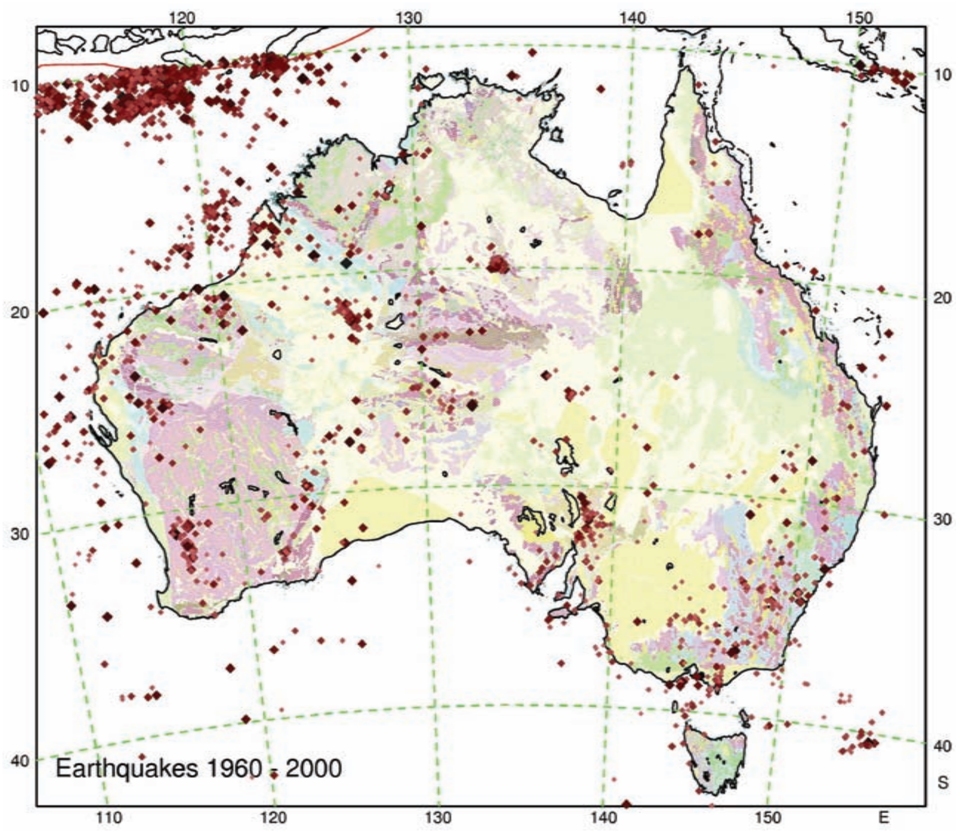


*Figure 4  Earthquake locations for the period from 1960–2000 superimposed on a generalized geological map of Australia. The symbols increase in both size and depth of tone as the magnitude of the event increases.*

the Musgrave Province. Similar bands of alternating N-S trending gravity anomalies in eastern Australia are associated with a series of island arcs and a succession of accretion events onto the evolving eastern margin of the continent. The major mineral province of Mt Isa has strong positive gravity anomalies which trend N-S, a tectonic grain imparted during the Meso-proterozoic. These gravity anomalies associated with Mt Isa are truncated in the S, with a sharp gravity gradient which formed during the latest Proterozoic breakup of the supercontinent Rodinia.

## Sedimentary basins

Figure 6 displays the thickness of Neoproterozoic and Phanerozoic basins across Australia, building on input from reflection seismology, gravity and magnetic surveys. Although a significant part of Australia is covered in sediments, the sequences on the continent are generally less than 7 km thick (Figure 6). Deeper basins (>15 km) occur off-shore particularly in NW Australia; they host most of Australia's oil and gas. Nevertheless the onshore basins are important with major coal deposits in eastern Australia, mostly from the Permian, and gas accumulations in the Cooper Basin in southern Queensland. The majority of these basins are associated with gentle downwarps associated with thermal sagging rather than localized rifting, and many in the western part of the continent lie on thick lithosphere.

## Radiometrics and Heat Flow

A major effort has been made to assemble a continent-scale study combining the results of many different airborne radiometric surveys (Figure 7). The surveys measure the gamma-ray spectrum at a modest altitude (c. 100 m) arising from the decay of the major radioactive elements potassium (K), uranium (U) and thorium (Th) in the top 30 cm of the ground. The energy distribution of the gamma-rays is specific to the decay chain for the particular elements and hence the relative contributions can be measured and extrapolated back to the concentrations of the radioactive elements at the Earth's surface. Corrections need to be made for the background radiation, the height of the aircraft above the ground, and the response and sensitivity of the detector. Further processing is required when many different surveys are combined to ensure that a common scale for element abundance is used across the entire image (Minty et al., 2009).

Figure 7 employs the normal mode of display of radiometric results using a three colour image with the K concentration on the red channel, Th on the green channel and U on the blue channel. This ternary image can be regarded as a chemical map of the near surface distribution of the elements and is strongly correlated with the surface geology and the rate of erosion or deposition in the landscape. Areas low in all three radioactive elements appear as dark hues (ultramafics, quartzites and sandstones) and areas high in all three radioactive elements appear as white hues (felsic volcanics and granites). Weathering, erosional and depositional

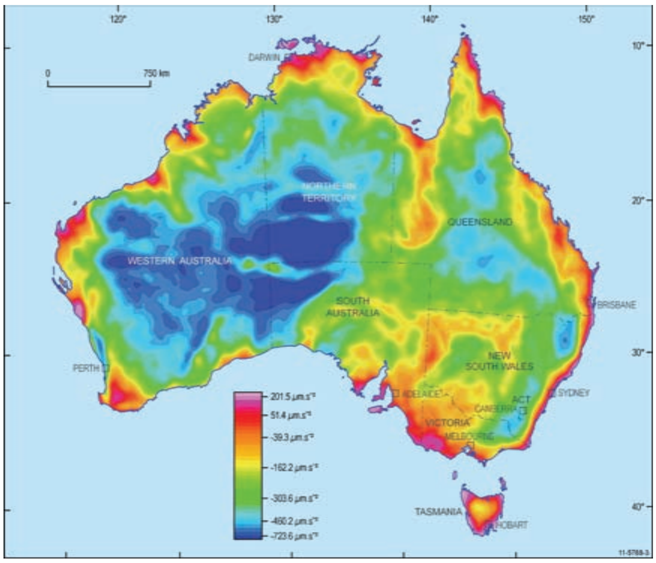


***Figure 5  Gravity field upward continued to 25 km, indicating the contrast between the ancient western core of the continent with negative gravity anomalies (blue) and the east with positive anomalies (yellow to red) (Source: Geoscience Australia).***

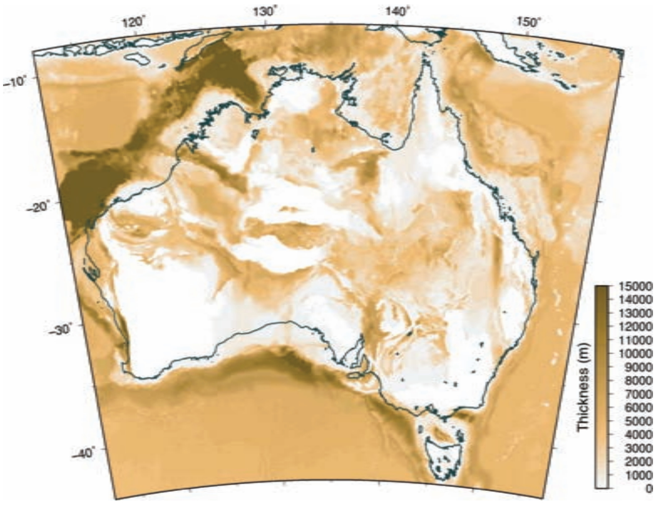


***Figure 6 The distribution of sedimentary basin thickness across the continent based on the OZSeebase study (Frogtech, 2005).***

processes play a large part in forming the radiometric response (Wilford, 2011). Many of the green and green-blue areas, representing high Th but low K, are highly-weathered surfaces rich in Fe.

Localised high concentrations of the radioactive elements are clearly visible in the granites of the Pilbara Craton and the eastern Yilgarn Craton, in central Australia and in the New England region of northern New South Wales.

Many geological features and boundaries are well delineated in the continent scale image, e.g., the Mesozoic shoreline behind the Nullabor Plain with a clear distinction from the Th-rich Yilgarn Craton. The actively eroding fold-belts surrounding the Kimberley region are prominent against the more sombre tones of the Kimberley itself. There is also a clear image of the remote Canning Basin in Western Australia, much of it forming the Great Sandy Desert between the Kimberley and the Pilbara regions. The mineral provinces around Mt Isa and Broken Hill have a distinctive signature. Other concentrations occur in the Flinders Ranges, which is also noted as a zone with enhanced geothermal heat flux, most likely associated with a high concentration of these heat-producing elements in the crust. The granites of the Tasman Element along the eastern margin are bright red and white colours and shades, consistent with their high concentrations in radioelements (Figure 7).

The high concentrations of the radioactive elements in the crust in many parts of Australia leads to high temperatures at depth, particularly where radioactive granites are blanketed with sediments, as in the NE parts of South Australia. Sampling of heat-flow across the continent is rather patchy, but is sufficient for the general pattern of temperature distribution to be determined (Figure 8). There is undoubtedly considerable potential for the exploitation of geothermal energy resources, but many of the best prospects lie far from the major population centres where the energy is needed (Kirby and Gerner, 2010; Jones et al., 2011; Webber et al., 2011).

## Seismological studies of the lithosphere

Australia has made extensive use of seismological methods in the study of the crust and upper mantle using both man-made and natural sources (Figure 9). A number of major refraction experiments were carried out from the 1960s into the 1980s and provide an important control on seismic wavespeeds across the continent. There has been little such work since, except some off-shore/on-shore experiments mostly in Western Australia. Reflection studies of the whole crust have grown from short experimental spreads in the 1960s to large-scale transects. A historical overview and extensive bibliography of the full range of active seismic experiments up to 2006 is given by Finlayson (2010).

A nearly 2000 km long reflection transect with 20 second recording was built up across southern Queensland in the 1980s using explosive sources. Explosive sources continued to be used until 1997, when they were replaced with arrays of powerful vibrator sources. Since 2004 there has been a major national investment in seismic reflection work funded through investment from Geoscience Australia, state and territory geological surveys and, since 2007, the AuScope

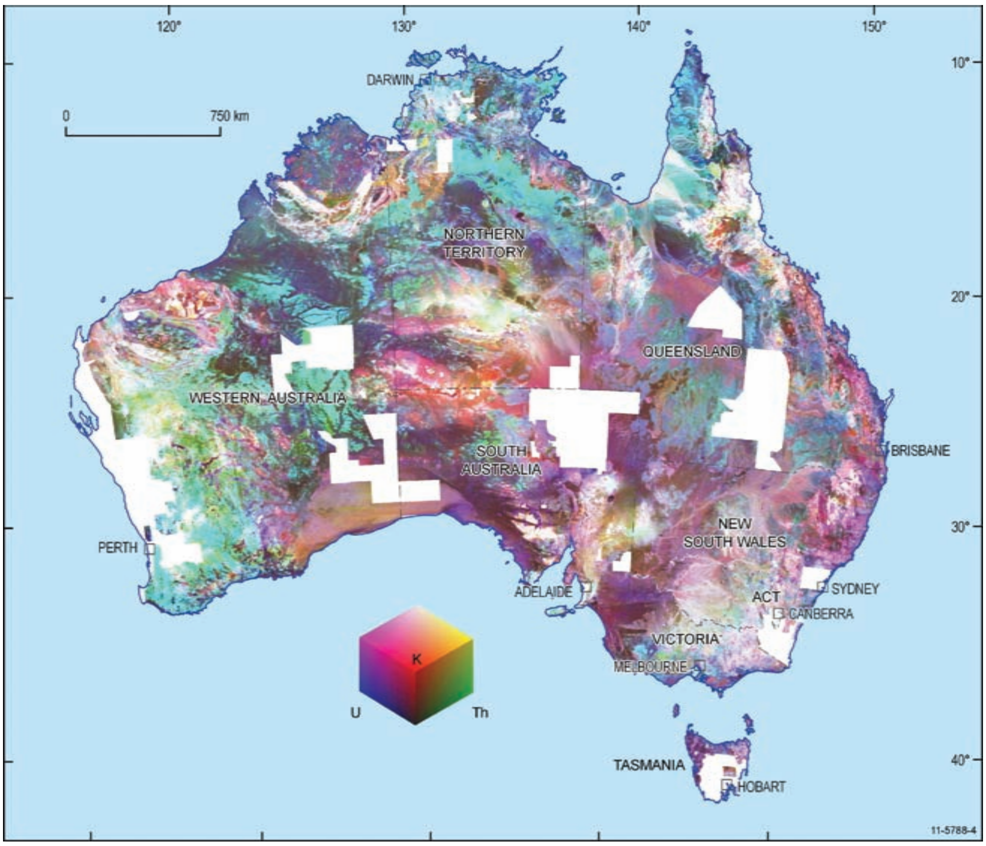


***Figure 7 Gamma-ray spectrometric map for the continent of Australia –K (red), Th (green) and U (blue) (after Minty et al., 2009). The pale grey areas are data acquisition gaps (Source: Geoscience Australia).***

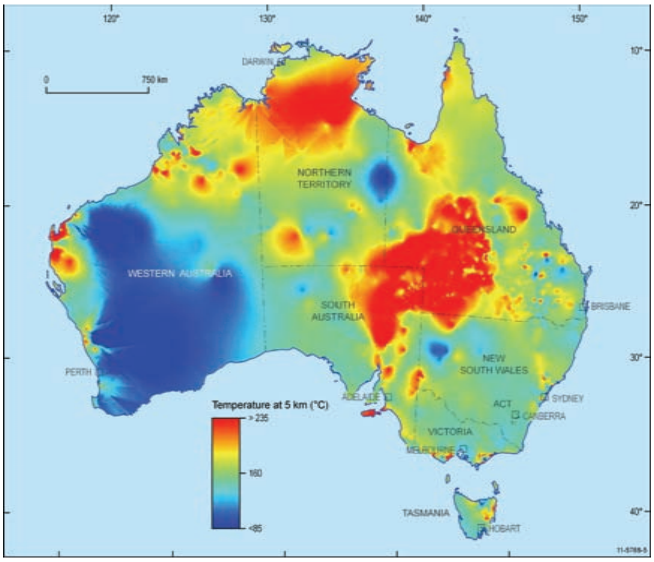


***Figure 8 Distribution of temperature at 5 km across the Australian continent estimated from heat flow studies (Source data: Earth Energy Pty Ltd; AUSTHERM database; Geoscience Australia).***

infrastructure initiative. Over 10,000 km of full crustal reflection profiles have been acquired with recording to 20 seconds or more. This very large effort has provided new insights into crustal structure, architecture and evolution in a number of parts of the continent. The dense sampling provided by the reflection transects has been of considerable value in mapping the character and geometry of the Moho across the continent.

The configuration of earthquake belts around the margins of Australia provides a wealth of events at suitable distances to be used as probes into the seismic structure of the upper mantle. Until recently there have only been a few permanent high-fidelity seismic stations on the continent. Much data collection has therefore been acquired through extensive deployments of portable broadband stations for periods of a few months at each site. This approach was pioneered with the SKIPPY experiment (van der Hilst et al., 1994) where a group of stations were progressively moved across the continent in a sequence of deployments. This style of experiments has inspired many similar efforts world-wide (such as major deployments in South Africa, James et al., 2001, and the US Array, http://*www.usarray.org*).

The combination of a long duration of recording at the permanent stations and the broad spatial coverage of the portable stations provides an excellent resource for studies of the lithosphere. A wide range of techniques, which exploit different aspects of seismograms, can be used to gain information on the 3-D structure in the crust and mantle.

A sequence of dense deployments of 3-component shorter-period instruments across the SE corner of the continent since 1999 has been brought together in the WOMBAT project (Rawlinson et al., 2011). These dense deployments provide detailed information on crustal and mantle structure.

## *Passive seismic*

A wide range of techniques have been employed to exploit the wealth of information available from the extensive coverage of the continent using portable seismic instruments. Most of the studies exploit seismograms from earthquakes, either in the regional belts of events to the N and E of Australia, or from more distant sources. A new technique, which is of particular value for crustal studies, exploits the ambient noise field, by working with the stacked correlation of records at pairs of stations. The result is equivalent to having a virtual source at one of the stations with a receiver at the other. The surface wave energy is the most prominent feature of the stacked correlation and has been exploited by Saygin and Kennett (2010) to map the upper and middle crust.

Large amplitude surface waves can be used in a tomographic inversion to determine the 3-D variations in shear wavespeed in the mantle. These relatively late arrivals in the seismogram travel nearly horizontally from regional events (e.g., Debayle and Kennett, 2000; Fishwick et al., 2005). This approach to surface-wave tomography relies on matching the waveforms on individual paths and then mapping of the path-specific constraints on shear structure into a 3-D model. In contrast, the higher frequency body wave arrivals are refracted back from the variations in structure in the mantle and are particularly sensitive to discontinuities in structure. Observations out to a distance of 3,000 km provide coverage of the structures down through the lithosphere–asthenosphere boundary to the upper mantle transition zone below northern Australia. The combination of short-period and broadband observations has provided detailed information on both P and S wavespeeds and the variation in seismic wave attenuation with depth. Further information on 3-D variations in structure can be extracted from the patterns of travel-time residuals. The combination of these different types of information allows the main features of upper mantle structure to be characterised. There is

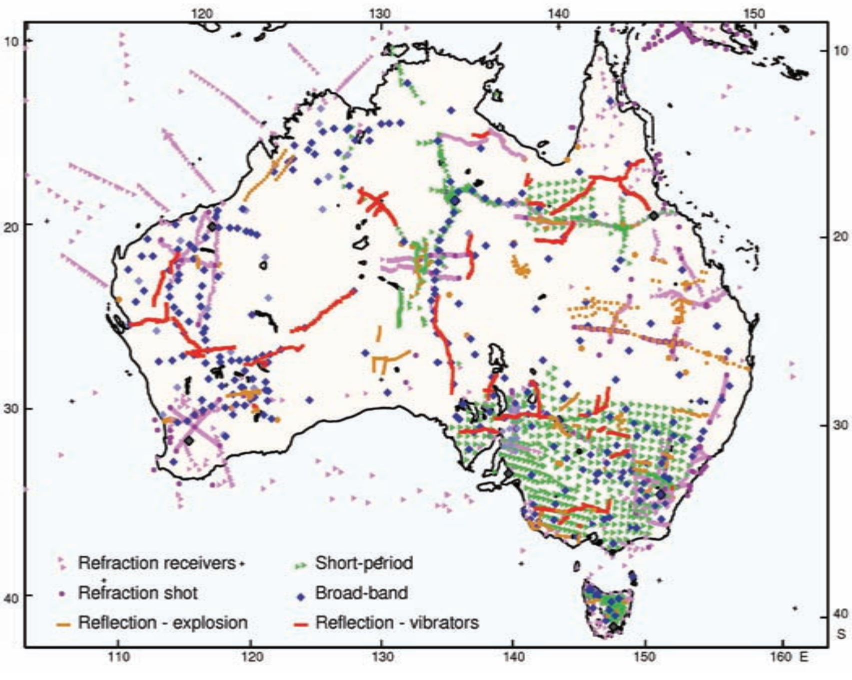


***Figure 9 Locations of seismological studies across Australia showing the locations of active studies: refraction surveys (purple) and shotpoints (circles), reflection lines (orange – explosive sources, red - vibrator sources) and passive studies: broad-band recording stations (diamonds) and short-period stations (triangles), as indicated on the key.***

a dramatic increase in seismic wave attenuation in the asthenosphere and thus such observations are important for mapping out the base of the lithosphere.

## *Receiver functions*

Extensive use has been made of receiver function studies that exploit the conversions and reverberations that follow the onset of the seismic signal. These studies provide important information on crustal and uppermost mantle structure, particularly the depth and nature of discontinuities. The receiver function is constructed from the beginning of a seismogram by extracting the influence of the source; the resulting information on structure near the receiver can then be elucidated with careful modelling.

The receiver function studies provide an important supplement to the limited sampling available from refraction experiments and provide a full continental coverage. Initial work in eastern Australia by Shibutani et al. (1996) was extended by Clitheroe et al. (2000) to cover most of the continent, using portable broad-band stations and the limited number of high-quality permanent stations at the time. A set of receiver function studies have been made of the West Australian cratons as detailed coverage has become available through portable broad-band deployments (Reading et al., 2007).

Despite higher noise levels and more limited bandwidth, valuable receiver function controls on crustal thickness can be extracted from teleseismic records at higher frequencies. For example, stacking and waveform fitting, coupled to local surface-wave dispersion results, has been used to improve the spatial coverage of Moho depths in SE Australia.

## *Body-wave tomography*

The initial application of mapping of the relative delay times of seismic arrivals was to investigate the major gravity anomalies in central Australia (Figure 5) by using linear profiles of portable instruments (Lambeck et al., 1988). These studies revealed the need for substantial, but localised, variations in crustal thickness, to satisfactorily model the seismic results. The crustal thickness variations have since been corroborated by deep seismic reflection profiling across the Arunta Block and, more recently, across the Musgrave Province (Korsch and Kositcin, 2010a).

The arrival times of seismic phases at the various broad-band stations from regional earthquakes have been picked and used in a number of studies (e.g., Kennett and Abdulah, 2011). The distribution of earthquakes means that detailed resolution using seismic waves refracted back from the mantle is confined to the northern part of Australia. Nevertheless these results provide independent corroboration of the presence of fast seismic wavespeed structure extending to 220 km or more beneath the cratons of central and western Australia.

High resolution work has been carried out in SE Australia using the dense networks of short-period stations that can be very clearly seen in Figure 9. Rawlinson et al. (2010) have combined refraction results from a marine survey around the island of Tasmania recorded at land stations with teleseismic data recorded at dense seismic arrays on the island. They have developed a high resolution P-wave tomographic model including a description of the crust–mantle boundary, which is consistent with the limited results from deep reflection studies in Tasmania (e.g., Drummond et al., 2000).

Tomographic inversions using data from distant earthquakes were originally conducted for the individual deployments using the relative arrival times of seismic phases. This approach minimised the mapping of structure from outside the zone for which the inversion is being undertaken. The WOMBAT project (Rawlinson et al., 2011) brings together the results from all the different deployments so that much more comprehensive imaging can be undertaken (Figure 10), revealing considerable structure in the crust and upper mantle. Across such a large region there can be considerable gradients in background structure that are too long wavelength to be imaged by the various deployments. Fortunately, by using information from surface wave tomography, a base model can be constructed from the large scale features in seismic wavespeed. Into this image, more detailed inversions of the arrival times can be incorporated, adding finer scale features to the base model. The results from this combined approach are providing new constraints on the base of the lithosphere in SE Australia.

## *Thickness of crust*

A new model for the depth to the Moho has recently been developed by using the entire suite of available data from refraction, reflection and receiver function studies (Kennett et al., 2011). The coverage is now very good across most of the continent (Figure 11), but there are still a number of remote desert areas with difficult access where sampling is rather sparse.

The inclusion of Moho picks at 20–40 km intervals taken from more than 10,000 km of reflection profiling has been particularly important in constraining the Moho map. Previously isolated seismic

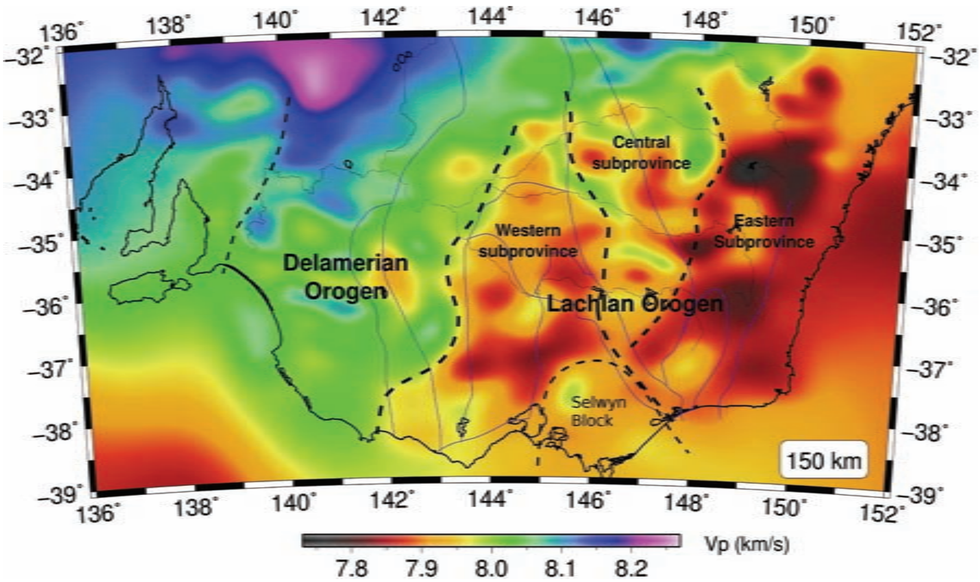


***Figure 10 Seismic tomography for the SE Australian region using data from the WOMBAT array (Rawlinson et al., 2011). The P wavespeed is displayed at a depth of 150 km.***

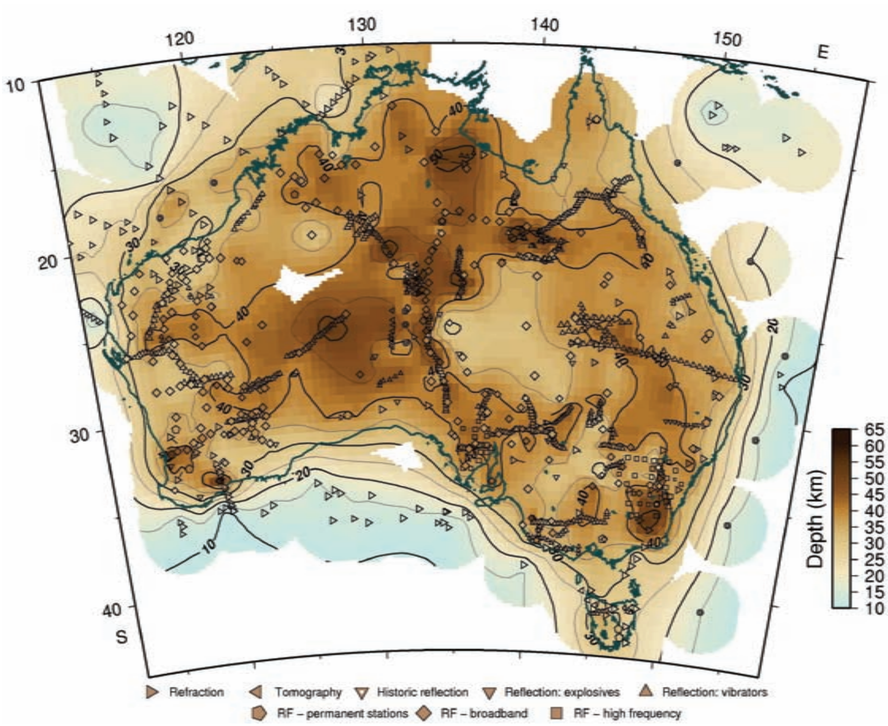


***Figure 11 The depth to the Moho across Australia derived from a combination of seismic refraction, refraction and receiver function studies (Kennett et al., 2011). The values obtained from different classes of observations are indicated by the colours attached to the distinctive symbols for each data type.***

stations can now be linked, and also cross-checks provided on results from other methods. In general, the consistency between techniques is very high and in most places the variation in crustal thickness estimates are less than 3 km. The thickest crust (more than 50 km thick) is mapped beneath the eastern highlands, central Australia and the Mt Isa region. The crust thins towards the continental edge, although it is still over 35 km thick in regions of the North West Shelf (Figure 11). The thinnest crust on the mainland is c. 30 km or less in the NE part of South Australia; a region with coincident high heat-flow anomalies (Figure 8).

There is significant variation in the *seismic character* of the crust–mantle transition. The character varies from a very distinctive and sharp change in reflectivity that may extend laterally for 100s of kilometres, to a more diffuse and difficult to identify boundary. A sharp Moho is particularly characteristic of the main cratonic regions of Archean crust such as the Yilgarn Craton (Goleby et al., 2002), but also the Proterozoic (in the upper crust) regions of NE Queensland (Chopping and Henson, 2009). The more diffuse character Moho occurs in Proterozoic crustal blocks, such as the Curnamona and eastern Gawler cratons (Korsch and Kositcin, 2010b), or northern Gawler Craton to Arunta Block in central Australia (Korsch and Kositcin, 2010a).

The reflection profiling results provide evidence for a number of sharp changes in crustal thickness, notably in central Australia (e.g., Goleby et al., 1989; Clitheroe et al., 2000) where offsets may exceed 20 km. The rate of change of crustal thickness can be rapid, with also small jumps apparently related to crustal sutures as in northern Queensland (Chopping and Henson, 2009). Such features pose a challenge when we try to provide a suitable representation of the Moho surface (Figure 11).

## *The lithospheric mantle*

The seismological results reveal a complex pattern of 3-D structure beneath the Australian Plate. The dominant control on the seismic wavespeeds in the mantle is from temperature, with a decrease in velocity with increasing temperature. As a result the variations in the wavespeed with depth can be interpreted in terms of changes in lithospheric thickness. The cratonic region in the centre and W of Australia is underlain by a thick mantle lithosphere extending to over 200 km depth with fast wavespeeds (especially for shear waves). Beneath the Tasman Element in the E, a region of younger Phanerozoic upper crust, the lithosphere is generally thinner (less than 140 km) and the asthenosphere has a pronounced low velocity zone for shear waves with high attenuation of shear wave energy.

The configuration of regional earthquakes around Australia has been extensively exploited in the analysis of multiply reflected shear waves and surface waves that form the most prominent part of seismograms. A variety of styles of seismic tomography have been employed to build 3-D maps of the shear wavespeed distribution beneath the whole continent (e.g., Debayle and Kennett, 2000; Simons et al., 2002; Fishwick et al., 2005; Yoshizawa and Kennett, 2004; Fichtner et al., 2009, 2010). As a result, structural variations have been imaged on horizontal scales down to 250 km, and significant substructure is apparent within the zones of both elevated and lowered wavespeeds (Figure 12). The centre of Australia has relatively low wavespeeds at 75 km, but there are strong gradients with depth and by 125 km a broad zone of fast wavespeeds is established across the centre and W of Australia that persists to more than 200 km depth. In contrast, in the Tasman Element, the seismic wavespeeds in the mantle are somewhat lower and the lithosphere is relatively thin, with estimates c. 80 km. The Tasman Sea region displays quite low shear wavespeeds (down to 4.2 km/s), probably as a result of residual heat left from failed rifting c. 80 Ma. Fishwick et al. (2008) have presented evidence for the progressive eastward thinning of the lithosphere across Australia, which occurs as a series of discrete steps with quite sharp transitions. This result poses interesting questions as to how the lithospheric steps can be maintained over extended periods of geological time.

Goes et al. (2005) used the patterns of seismic wavespeeds determined from surface-wave tomography to extract information on the temperature in the mantle lithosphere beneath Australia. The mantle lithospheric temperature range is c. 1,000°C, with a broad-scale correlation between temperature and tectonic age. Within each tectonic province there are, however, temperature differences ranging from 200–700°C. There are no significant differences in the temperatures beneath the Archean and Proterozoic regions. In the E, the temperatures approach the moist solidus.

## *Anisotropy*

The path of seismic waves is partly controlled by mantle

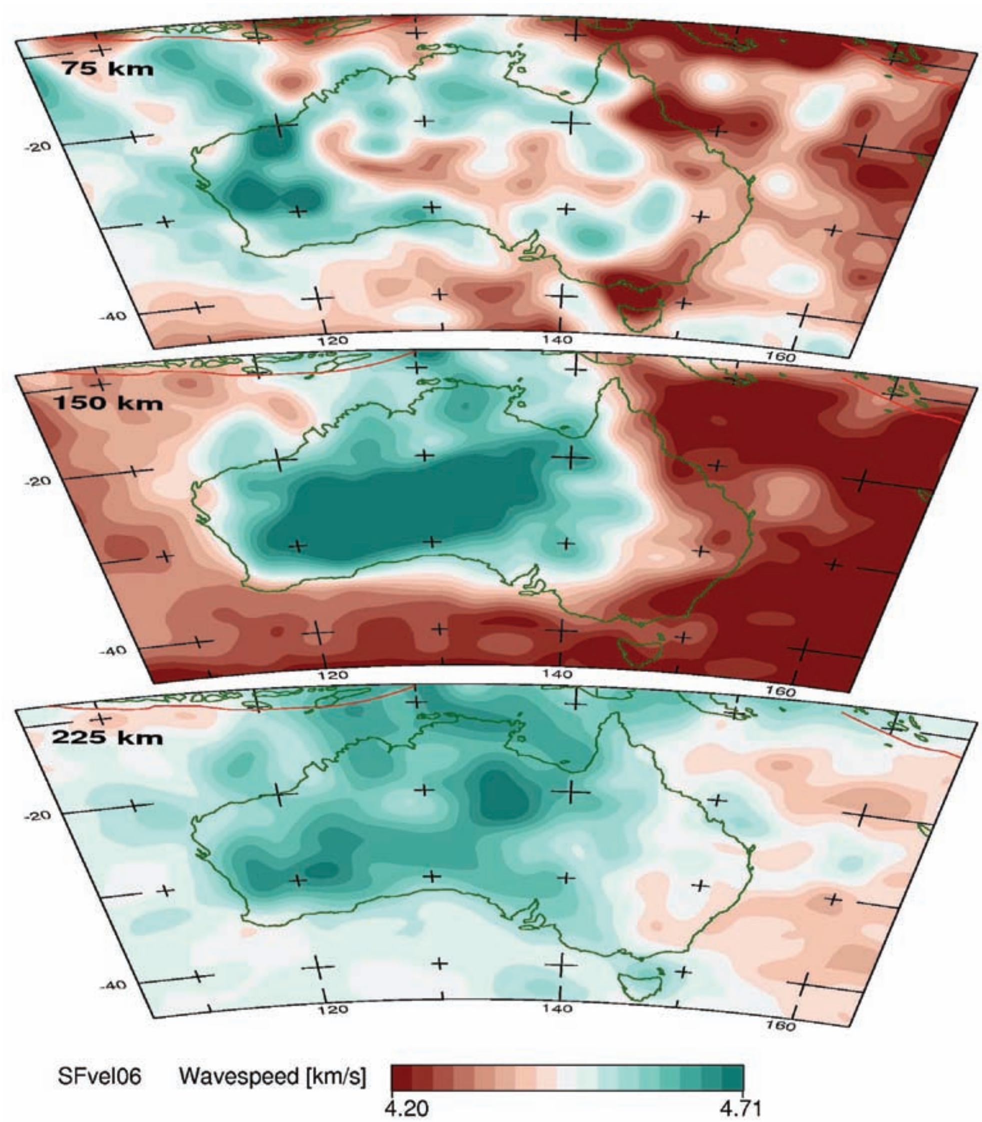


***Figure 12 Map views at depths of 75 km, 150 km and 225 km for the 3-D shear wavespeed distribution below the Australian region, based on the model of Fishwick et al. (2008).***

mineralogy and structure. Olivine, the dominant mantle mineral, responds to stress by aligning its crystallographic structure. It is this structural alignment that influences the transmission of seismic waves. The seismic wavespeeds vary along the crystallographic axes so that waves travelling in different directions have different effective wavespeeds. The effects are most noticeable for shear waves and lead to different travel times for waves with orthogonal polarisation. The difference in propagation time, "shear wave splitting", provides a measure of the strength of anisotropy and hence crystal alignment.

Observations of splitting in the times of arrivals of refracted shear waves on the radial and tangential components of waves returned from the upper mantle were used by Tong et al. (1994) to show the presence of anisotropy in the lithosphere. Subsequently, Heintz and Kennett (2005) made an analysis of shear wave splitting for the core refracted (SKS) waves for stations across the continent using all available stations. The advantage of using such waves is that they are repolarised on exit from the Earth's core so that any shear wave splitting has to be imposed on the wave by its passage through the mantle on the receiver side. There is a weak correlation of splitting with geological features, but many stations show very little splitting of the SKS waves in Australia (Heinz and Kennett, 2005). An alternative approach to anisotropy studies is provided by the analysis of surface waves. Debayle et al. (2005) show that Australia is distinctive among the continents by displaying a twist in the direction of fast propagation of Rayleigh waves from E-W at 100 km depth to near alignment with the absolute plate motion at 200 km depth within the zone of fast seismic wavespeeds. It is therefore likely that the complex patterns of behaviour are induced by the superimposition of effects from multiple anisotropic layers (Fouch and Rondenay, 2006).

## *Transition to the Asthenosphere*

The lithosphere beneath the older parts of the continent can be readily recognised by fast shear wavespeeds (up to 4.7 km/s), but the transition to the asthenosphere is not marked by any sharp transition. Rather there is a gradation from a conductive to a convective regime, most likely linked also to a change in rheology from dislocation to diffusion creep. The presence of the asthenosphere is manifest in enhanced seismic attenuation. The fast lithospheric wavespeeds are accompanied by little loss of seismic energy, enabling high frequency waves to propagate readily from subduction zones into continental Australia (Kennett and Furumura, 2008). These high frequency waves are, however, suddenly lost when the seismic waves penetrate into the asthenosphere, because of its much higher attenuation of shear waves than in the lithosphere (Gudmundsson et al., 1994). Kennett and Abdulah (2011) have used the full range of arrivals at portable stations across Australia to undertake attenuation tomography that confirms the presence of much stronger attenuation below 210 km depth.

Figure 13 shows an estimate of the depth to the base of the lithosphere based on a variety of lines of evidence including the analysis of refracted waves in the mantle, the wavespeeds and gradients deduced from surface-wave tomography and body-wave tomography. The contrast between thick lithosphere in the centre and W and thinner lithosphere in the E is very clear and the nature of the transition is consistent with the analysis of Fishwick et al. (2008). The base of the lithosphere is somewhat irregular and this is likely to impose complex stress patterns from the relative motions of the thick continental lithosphere and the free-flowing asthenosphere. Farrington et al. (2010) have suggested that a southern edge to the cratonic lithosphere has produced edge induced convective flow associated with the northern motion of Australia at c. 7 cm/yr that is

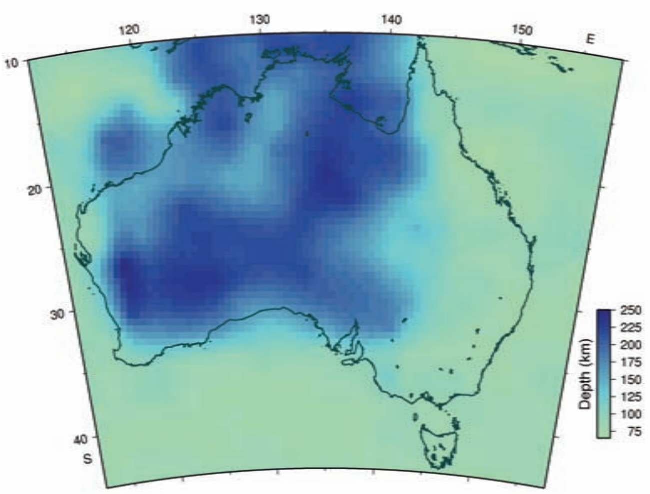


***Figure 13  Estimate of the depth to the base of the lithosphere beneath Australia derived from a variety of seismological data.***

linked to the eruption of basalts of the Newer Volcanic Group in Victoria and South Australia, with the youngest event dated at 4.6 ka.

## *Magnetotelluric Surveys*

In recent years there has been considerable investment in magnetotelluric studies across Australia. Major reflection profiles in Queensland, central Australia and Western Australia have been accompanied by magnetotelluric soundings, with instrument spacing of c. 10 km. Earlier studies using a limited number of instruments have provided insights into crustal lithospheric structure in southern Australia, notably near the major mines at Olympic Dam (Heinson et al., 2006), at Kalgoorlie (Blewett et al., 2010) and across the thrusts belts in central Australia (Selway et al., 2006). Whereas seismic properties depend on the physical state of the material, the electrical conductivity varies over a much larger rather with a strong influence from local chemistry, e.g. the presence of fluids or graphite.

# Discussion

In this summary article we are only able to touch on some of the wide range of studies that have been made of the Australian lithosphere. Many other aspects of lithospheric structure and properties are presented in a set of compilations by Drummond (1991), Braun et al. (1998) and Hillis and Müller (2003).

The processes and timing of lithosphere formation are controversial (Eaton et al., 2009). Old and cold cratonic lithosphere is strong and buoyant, and therefore difficult to destroy and recycle (Lee et al., 2011). Much of Australia is underlain by a thick lithosphere that has undoubtedly influenced the tectonics, topography, seismicity, basin evolution and ultimately the landscape. The deeper lithosphere beneath Australia is imaged well by seismic tomography (Figures 12 and 13). At a deeper level, there are major differences in the thickness of the continental lithosphere between the eastern third of the continent (E of 140ºE) and the western two thirds (Figure 13). The thicker western part of the continent occurs below the known region of Precambrian crust. The thinner eastern part occurs below the mainly Phanerozoic Tasman Element (Figure 3). The lithospheric differences of over 100 km in thickness have a material effect on the way the plate bends under load. Such flexures have predictable consequences and go some way to explain the topographic differences (Stüwe, 1991).

Despite the average flatness of Australia, there are notable differences between the E coast and its hinterland and the rest of the continent (Figure 2). The highest mountains in Australia (up to 2,228 m) occur in the SE, and they are part of a 3,500 km long chain of elevated topography mirroring the geometry of the coast. Although the southern and western coastal hinterlands have similar passive margin tectonic settings with ocean crust of similar age (Müller et al., 1997), they do not show the elevated topography of the E (Figure 2). These topographic differences have influenced the climate and the available topographic head for river networks to form, which in turn influence uplift and erosion rates.

The northward movement of the continent, which started at c. 80 Ma, has brought Australia in the more recent times into collision along the northern plate boundary with Eurasia in the Timor and Papua New Guinea regions (Huston et al., 2012). As Australia has moved northwards it has come under the influence of the Earth's geoid high located beneath Indonesia, which is linked to the gravitational pull from accumulated subducted material at depth. There has been a consequent tilt of the continent towards the N, so that today Miocene sea cliffs on the northern margin of the Eucla Basin are more than 200 m above sea-level (Quigley et al., 2010).

Thick subcontinental lithosphere is difficult to melt, but can be mechanically weakened by stretching. The way in which sedimentary basins evolve through time, from rifting to thermal subsidence is strongly influenced by the thickness and structure of the underlying lithosphere. Heat takes longer to conduct through thick lithosphere, so that the thermal sag sedimentation phase following rifting is delayed and prolonged. Australia's intracontinental basins are widespread, and were filled over a prolonged time interval, as a consequence of the thick lithosphere. Many of these basins, which are typically <5 km thick, comprise several successor basins that range in age from 850 Ma to the Pliocene.

The crustal thickness on the Australian continent ranges from 25–60 km, with much of the crust between 38–42 km thick (Figure 11). The thickest crust occurs beneath the Paleozoic in the E and in parts of central and northern Australia. Deep seismic reflection studies of the Australian Precambrian terranes support the concept of somewhat thickened Proterozoic crust compared to Archean crust, though there is no simple pattern of age progression (Aitken, 2010; Kennett et al., 2011). Drummond and Collins (1986) suggested that lower crustal mafic underplating was the main process responsible for this thickening. Such a mechanism would be consistent with the progressive loss of crustal reflectivity seen in reflection profiling across the regions with thickest crust; since interleaved mantle material reduces contrasts in physical properties. However, the same episodes of upper mantle melting do not appear to have underplated the Archean crust adjacent to the Proterozoic crust in the western part of the continent (Goncharov et al., 1998).

The crust is an integral component of the lithosphere and its thickness also controls the behaviour of the overall lithosphere to deformation (e.g., Behn et al., 2002). Thin crust results in deformation being localised into the mantle and so the width of any rift basin is controlled by the vertical geothermal gradient. Thick crust in contrast, allows stress accumulation in the crust to be greater than in the mantle, so rift basin width becomes dependent on both the vertical geothermal gradient and the rheology of the crust. Rifting tends to exploit former lines of weakness, but does not always succeed; Bass

Strait represents a failed rift in the separation of Australia and Antarctica and this feature is linked to important sedimentary deposits with much of the known oil accumulations. The Centralian Superbasin is also located across the ancient sutures between northern and southern Australia.

Horizontal slices through the tomographic volume reveal additional complexity in Australia's deeper lithosphere. A zone of slower than average shear wavespeeds is imaged below the crust at c. 75 km (Figure 12). This slow zone extends E-W across most of the continent and correlates broadly with the mobile belts that join the three cratonic elements of North Australia, South Australia and the West Australia (Huston et al., 2012). This zone also corresponds with the former extent of the Centralian Superbasin, which formed following rifting during the Neoproterozoic breakup of the supercontinent Rodinia at c. 825 Ma (Li et al., 1999). These same regions have also experienced intense and punctuated contractional deformation that in places has exhumed mid crustal rocks to the surface. The slow wavespeeds at 75 km depth are difficult to explain, and the suitable combination of thermal or compositional structure remains a question of considerable scientific interest. For example, these slow wavespeed regions may have acted as, or continue today to act as, zones of lithospheric weakness. Such zones would favour the formation of sedimentary basins and mountain ranges and be the locus of seismicity.

Finer resolution of more subtle features with seismic tomography is possible where the data density is increased. Figure 9 shows the data network and reveals the much higher density of stations that have been deployed in recent years in the SE quarter of the continent. Tomographic imaging in this data-rich region (Figure 10) is able to resolve the deep boundary between the Neoproterozoic Delamarian and Paleozoic Lachlan orogens (of the Tasman Element) and the various blocks that comprise the latter. The Selwyn Block is thought to be an older crustal fragment of Neoproterozoic to Cambrian age that influenced the younger domains, some of which host the world-class Victorian goldfields. These tomographic models provide constraints on the geometry of the blocks and the location of their boundaries. The slow wavespeed regions in Figure 9 likely correspond to the thermal effects of a mantle hot spot located beneath SE Australia; with the Pliocene and Pleistocene Newer Volcanic Group being the erupted products (Graeber et al., 2002).

Australia has not always been in the largely mid-plate position it finds today. Australia was part of the Earth's earlier (super)continents, with contributions to Vaalbra, Kenorland, Nuna, Rodinia and Pangea/Gondwana (see Huston et al., 2012 and references therein). The evolution of these supercontinents has resulted in different lithospheric blocks first amalgamating and later rifting. As a consequence, the boundaries between these blocks can become pathways for fluids/magmas to metasomatise the mantle. These boundary regions act as a source and pathway for fluids, many of which are implicated in the formation of giant mineral deposits (Hronsky and Groves, 2008). These deep boundaries can be imaged directly or inferred in various places by a number of techniques. Australia has a comprehensive series of deep seismic reflection profiles (Figure 9), which cross many of the continent's major geological provinces and boundaries (Chopping and Henson, 2009; Korsch and Kositcin, 2010a, b). Major structures are interpreted to transect the crust and in places offset the Moho. There are also distinctive differences in seismic character–that is the amplitude and coherence of reflectors–between different crustal blocks in many deep seismic sections. Distinct boundaries between these different 'seismic provinces' are sometimes imaged, and some extend to the surface as known faults or sutures. In other places the boundaries themselves are indistinct and their presence is inferred by the changes in seismic character (Korsch and Kositcin, 2010a, b).

Thick lithosphere also impacts on resources such as diamonds. Diamonds are formed at high pressures (>150 km) and generally require a thick lithosphere and low geothermal gradient to form and be preserved. Australia is major diamond producer, especially from the Argyle deposit in NW Western Australia. The regional extent of this thick lithosphere and the longevity of the landscape should make Australia prospective for diamonds (O'Neill and Moresi, 2003).

A recent compilation of Australia's neotectonic features by Clark et al. (2011) has suggested that the distribution of seismic activity is dependent on whether the crust has been extended or not. In their definition, crust without extension includes cratons, platforms and fold belts, whereas extended crust includes intracontinental rifts of all ages and passive margins no younger than c. 25 Ma. The pattern found was that regions of extended crust tended to be more seismically active. The age of major rifting appears to be important in terms of neotectonic activity level in extended crustal domains so that Paleozoic intra-cratonic rifts and passive margins are less active than those rifted in the Mesozoic, with greater activity in NW than SE Australia (see Figure 6).

The Flinders Ranges in South Australia are accommodating around one third of the seismic strain across the whole the continent, but do not fit into the general pattern of age relations. These ranges are actively uplifting, with thrust faults placing Precambrian rocks onto Quaternary rocks in places (Quigley et al., 2010). This zone of weakness arises from the orientation of the current maximum stress direction perpendicular to basin controlling faults, coupled with the elevated heat flow in the region (Neumann et al., 2000).

## Conclusions

The Australian lithosphere, and its constituent structures, is well mapped by a wide array of continental-scale geophysical datasets using many different techniques including potential fields, radiometry and seismological methods. Such results are of particular importance because extensive sedimentary basins and the draping of thick regolith across the landscape mean that exposed rocks or outcrop is limited.

The Moho, the major boundary in the upper lithosphere, is highly variable in its character and depth below the surface. Depths to Moho across the continent range from 25–60 km, despite an average elevation of only c. 330 m above sea level. There are rapid variations in crustal thickness in central Australia that give rise to major E-W oriented gravity anomalies. The character of the transition from crust to mantle is highly variable with extended gradients accompanying the deepest Moho, probably associated with mafic underplating. There is no simple relationship between depth to Moho and crustal age, though many of the zones of thickest crust are in regions of Proterozoic outcrop.

More than two thirds of the Australian continent is underlain by lithosphere over 200 km thick, which is more than double the global average for lithospheric thickness. This thick lithosphere underlies the Precambrian zones of western and central Australia. It has controlled many features of Australian geology and geography; with a major influence on how the continent has responded to tectonic forces related to breakup and amalgamation. The contrasts in lithospheric thickness provide controls on topography and landscape, seismicity, the evolution of sedimentary basins evolution, heat flow and other resources.

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# Australia through time: a summary of its tectonic and metallogenic evolution

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*The geological evolution of Australia is closely linked to supercontinent cycles that have characterised the tectonic evolution of Earth, with most geological and metallogenic events relating to supercontinent/ supercraton assembly and breakup. Australia mainly grew from W–E, with two major Archean cratons, the Yilgarn and Pilbara cratons, forming the oldest part of the continent in the West Australian Element. The centre of the continent consists of the mainly Paleoproterozoic–Mesoproterozoic North and South Australian elements, whereas the E is dominated by the Neoproterozoic–Mesozoic Tasman Element. The West, North and South Australian elements initially assembled during the Paleoproterozoic amalgamation of Nuna, and the Tasman Element formed mostly as a Paleozoic accretionary margin during the assembly of Gondwana-Pangea. Australia's present position as a relatively stable continent resulted from the breakup of Gondwana. Australia is currently moving northward toward SE Asia, probably reflecting the earliest stages of the assembly of the next supercontinent, Amasia.*

*Australia's resources, both mineral and energy, are linked to its tectonic evolution and the supercontinent cycle. Australia's most important Au province is the product of the assembly of Kenorland, whereas its major Zn-Pb-Ag deposits and iron oxide-Cu-Au deposits formed as Nuna broke up. The diverse metallogeny of the Tasman Element is a product of Pangea-Gondwana assembly and most of Australia's hydrocarbon resources are a consequence of the breakup of this supercontinent.*

## Introduction

Australia severed its last ties to Gondwana, separating from Antarctica at c. 34 Ma (Veevers et al., 1991), largely isolating the Australian continent from active plate margins. Rocks in Australia, however, record a complex, and contentious, geological history that can be traced back to the Eoarchean. In this contribution we present one interpretation of this history, and delve into some of the related controversies. We also link Australia's geological heritage, including its resources, as well as its geography, flora and fauna, to this tectonic narrative. Much of the geology of Australia is the consequence of the amalgamation and break-up of supercontinents and supercratons through geological time, such as Vaalbara, Kenorland, Nuna (Columbia), Rodinia and Pangea (including Gondwana) (Figure 1).

Supercontinent history has not only controlled the distribution of resources, petroleum, coal and minerals, but has surprising influences on processes that have shaped Australia in the past and are shaping Australia now. Australia, for example, shares many floral and faunal affinities with South America and Africa, but few with northern hemisphere continents (e.g., Couper, 1960; Fooden, 1972). The breakup of Pangea isolated Australia from Eurasia and North America. Moreover, when Australia broke away from Antarctica, the resulting seaway opened to allow polar circulation of ocean currents and ultimately the formation of the Antarctic ice cap (Livermore et al. 2005), making Australia, and the world, drier and cooler (Fujioka and Chappell, 2010).

Continental Australia grew predominantly from W–E, with Archean rocks mostly in the W, Proterozoic rocks in the centre, and Phanerozoic rocks in the E. The western two-thirds of the continent consist of three mostly Precambrian elements (see Table 1 for definitions), the West Australian, North Australian and South Australia elements, whereas the eastern one-third is made up of the Tasman Element (Figure 2). These spatial and temporal growth patterns are consistent with supercontinent and supercraton growth, particularly Kenorland, Nuna and Pangea-Gondwana. The distribution of Australia's energy and mineral resources (Figure 2) is also governed by this broad pattern, with each of the four major cratonic elements characterised by distinctive deposit assemblages, both in time and in composition.

The Australian continent evolved in four broad time periods, namely: 3800–2200 Ma, 2200–1300 Ma, 1300–700 Ma, and 700–0 Ma. The first period saw the growth of nuclei about which cratonic elements grew, whereas the latter three periods involved the amalgamation and dispersal of Nuna, Rodinia and Pangea-Gondwana, respectively. In the sections below we present a history of the growth of the present-day Australian continent using this framework, although noting that in many cases there is significant uncertainty and disagreement about specific details and, indeed, about whether some of these processes occurred at all. This geohistory provides context for events that have shaped and changed Australia and the Earth, including the evolution of life and changes in the composition of the atmosphere and hydrosphere. As they have been increasingly linked to geodynamic processes, the evolution of Australia's mineral and

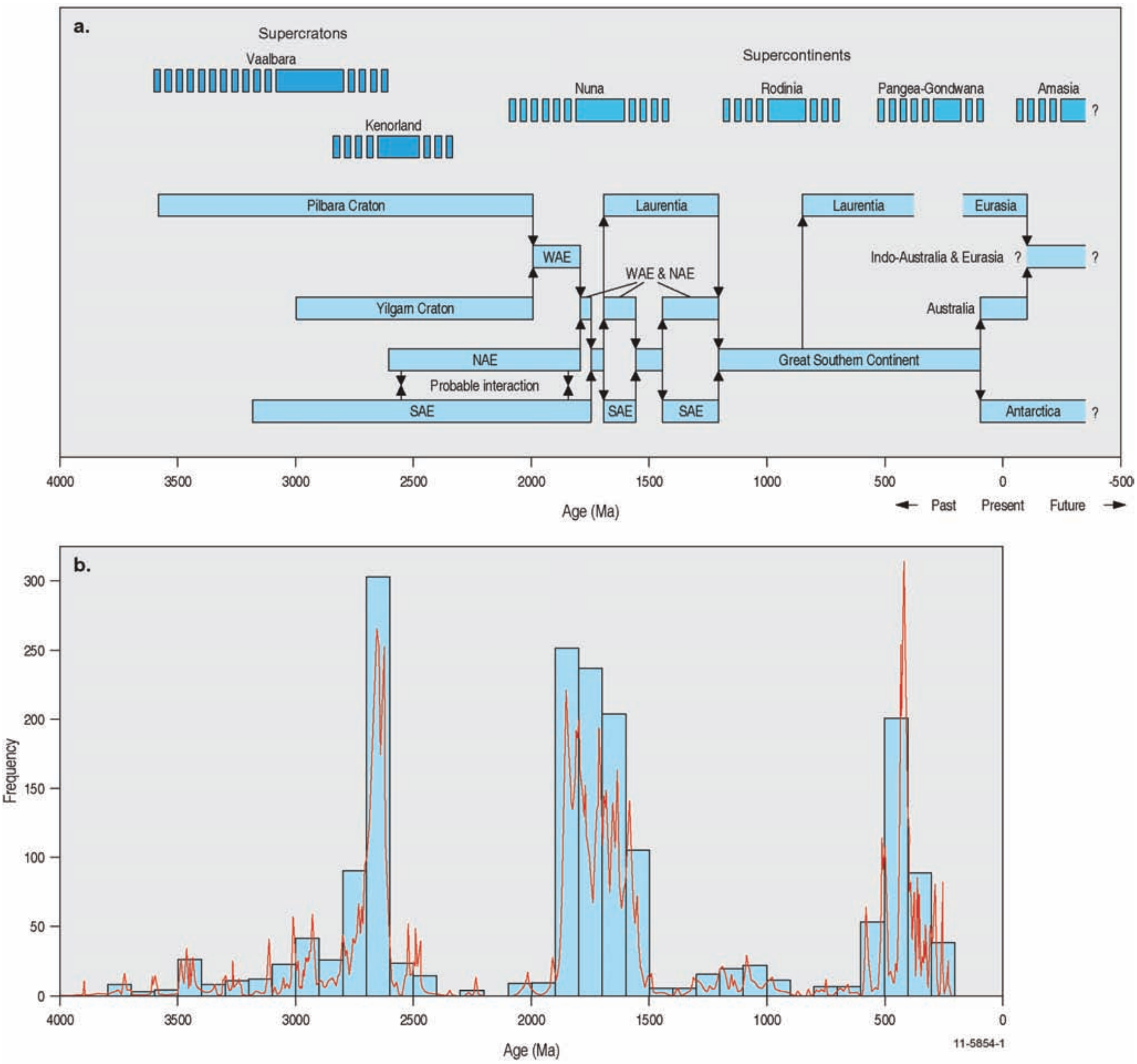



***Figure 1 ( a) Assembly and breakup of cratonic elements of constitute Australia from 3.0 Ga and their relationship to supercontinent cycles. (b) Histogram showing the distribution of the age of igneous rocks from Australia as determined using SHRIMP and other U-Pb analytical techniques (data from Keith Sircombe and OZCHRON database). The acronyms NAE, SAE and WAE refer to the North, South and the West Australian elements, respectively. The upper part of Figure 1a shows the inferred assembly (dashed blue lines to left), stability (solid blue lines) and breakup (dashed blue lines to right) of supercontinents (light) and supercratons (dark) through time.***

petroleum systems, therefore, provides a parallel framework for the evolution of the continent as a whole.

## 3800–2200 Ma: Growth of cratonic nuclei

The oldest rocks in continental Australia, with ages between 3731–3655 Ma, consist of anorthosite and orthogneiss from the Narryer Terrane of the Yilgarn Craton (Kinny et al., 1988; Wilde et al., 2001) and from the Pilbara Craton (as xenoliths in younger plutons; Van Kranendonk et al., 2002). Although these two cratons, which form nuclei to the West Australian Element, are the most extensive exposures of old rocks in Australia, over the past five years, Archean to early Paleoproterozoic rocks have been increasingly identified within the North and South Australian Elements, forming the nuclei onto which these elements were accreted (Fraser et al., 2010; Hollis et al., 2011).

In addition to containing the oldest known rock in continental Australia, the 3731±4 Ma Meeberrie Gneiss (Kinny et al., 1988), the Narryer Terrane contains the oldest mineral known on Earth, which is a c. 4404 Ma detrital zircon from Paleoproterozoic sedimentary rocks at Jack Hills (Wilde et al., 2001). This and other old zircons, which are only 150–350 Myr younger than the age of the Earth, have important implications for the characteristics of the earliest period of the Earth's history. Their very existence indicates that continental crust formed very early, and their heavy oxygen isotope characteristics indicate that oceans were present in the Hadean (Wilde et al., 2001).


***Figure 2 Distribution of Australian resources with respect to major Australian crustal elements and basins.***

The Yilgarn Craton is older in the W than in the E. Rocks of the Narryer, Youanmi and Southwest Gneiss terranes formed between 3730–2900 Ma in the W, and those of the Kalgoorlie, Kurnalpi, Burtville and Yamarna terranes, which constitute the Eastern Goldfields Superterrane, formed between 2940–2660 Ma in the E (Figure 3). Most tectonic reconstructions of the Eastern Goldfields Superterrane envisage arc-related accretion (e.g., Barley et al., 1989; Krapez et al., 2008), although the number of arcs and the polarity of subduction zones vary. Korsch et al. (2011a), for example, infer assembly along a series of E- and W- dipping subduction zones that

*Table 1 Definitions of terms.*

| Term | Definition |
|---|---|
| Supercontinent | A large continent formed by the amalgamation of most or all of Earth's continental land masses. |
| Supercraton | A large, ancestral (largely Archean) land-mass consisting of two or more cratons. |
| Continent | One of Earth's major land masses (or former major land masses). |
| Element | Part of a continent that has some shared broad-scale geological history; often an interpreted proto-continent or collection of such continental fragments (including cratons), that now forms part of an extant continent. |
| Craton | A part of Earth's continental crust that has attained stability and has been little deformed for a prolonged period. |
| Province | A large geological region showing similarities in its geological history, but with a different geological history to adjacent provinces. |
| Superterrane | A collection of two or more terranes. |
| Terrane | A region with essentially similar geology and geological history. |
| Domain | A, usually fault-bounded, region of similar geology. |
| Superbasin | A group of temporally- and genetically-related basins. |
| Basin | A low area at the Earth's surface and of tectonic origin in which sediments have accumulated. |
| Inlier | An exposure of basement rocks completely surrounded by younger basinal rocks. |
| Seismic province | A discrete volume of middle to lower crust, which cannot be traced to the surface, and whose crustal reflectivity is different to that of adjoining provinces, either laterally or vertically. |
| Orogen | An, often linear or arcuate, region that has been subjected to one or more common episodes of deformation and metamorphism (orogenies). |
| Orogeny | Geological event or genetically and temporally closely related events involving rock deformation. |
| Movement | Geological event or genetically and temporally closely related events involving rock deformation. Used where previously defined orogenies (e.g., Alice Springs) are found to include unrelated orogenic events. |
| Event | A temporally- and, commonly, spatially-restricted occurrence of a geological process or related geological processes. This can include magmatism (igneous event), deformation (deformational event, orogeny or movement) or mineralisation (mineralisation event). |

closed between 2780–2655 Ma (Figure 3). Alternatively, Czarnota et al. (2010) inferred the growth of the Eastern Goldfields Superterrane to be related to a long-lived W-dipping subduction zone to the E of the Burtville Terrane. The accretionary processes were likely to have some broad similarities to modern subduction processes, with the formation of backarc basins and major orogenic events as fragments collided (Barley et al., 1989; Figure 3). These processes produced laterally continuous crust-penetrating shear zones which accessed a mantle that was fertilised by subduction. These shear zones were important conduits for Au mineralisation (Blewett et al., 2010), and may be one of the keys to the Au riches of the Eastern Goldfields Superterrane.

The Yilgarn Craton, which includes the Eastern Goldfields Superterrane, has a global resource (production and reserves) in excess of 8,500 tonnes of Au and is one of the two largest global Archean Au provinces. Although most deposits in the Yilgarn Craton, particularly those in the Eastern Goldfields Superterrane, are considered lode Au deposits, the Boddington Au-Cu deposit, which is located in the 2714–2696 Ma Saddleback island arc (Qiu et al., 1997; Korsch et al., 2011a) in the SW part of the craton (Figure 3), is considered to be polygenetic (McCuaig et al., 2001). The earliest phase of mineralisation, at c. 2707 Ma, is interpreted as a porphyry-style, whereas the second phase of mineralisation, at c. 2629 Ma, has a similar age to lode Au mineralisation in the Eastern Goldfields Superterrane (Stein et al., 2001). The first, porphyry-related, stage is one of the earliest examples of arc-related mineralisation known. The Yilgarn Craton is also a major Ni province (Figure 4). Individual Ni deposits are hosted by komatiites, which are high-temperature ultramafic volcanic and shallow intrusive rocks thought to be a product of a hotter Archean Earth (Nisbet et al., 1993).

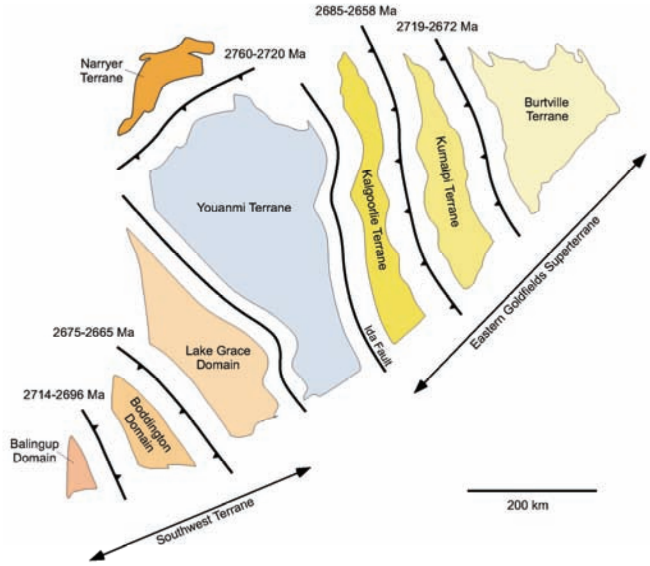



***Figure 3 A model for the assembly of the Yilgarn Craton (based upon Korsch et al., 2011a). The relationships of the Southwest Terrane (Lake Grace Domain) and the Eastern Goldfields Superterrane (Kalgoorlie Terrane) to the Youanmi Terrane are uncertain.***

The Yilgarn Craton appears to have been a constituent of the Kenorland supercraton, which is thought to have also included the Abitibi Subprovince in Canada. These two provinces formed over the same time period, and are the two most richly mineralised Archean provinces known. Kenorland had amalgamated by c. 2660 Ma and began to break up at c. 2480 Ma (Barley et al., 2005).

The 3530–2930 Ma Pilbara Craton is overlain by the 2780–2450 Ma Fortescue and Hamersley basins (Figure 5). Mechanisms by which the oldest (>3200 Ma) part of the Pilbara Craton formed are controversial, ranging from crustal overturn (Van Kranendonk et al., 2002), to formation of an oceanic plateau as the consequence of mantle plume activity (Smithies et al., 2005a), and to tectonic processes analogous to modern plate tectonics (Bickle et al., 1983; Barley et al., 1984; Zegers et al., 1996; Blewett, 2002). By c. 3120 Ma, however, plate-tectonic-like processes must have been active, as the Whundo greenstone belt (Figure 5) is the oldest oceanic arc system known in Australia and one of the earliest known in the world (Smithies et al., 2005b).

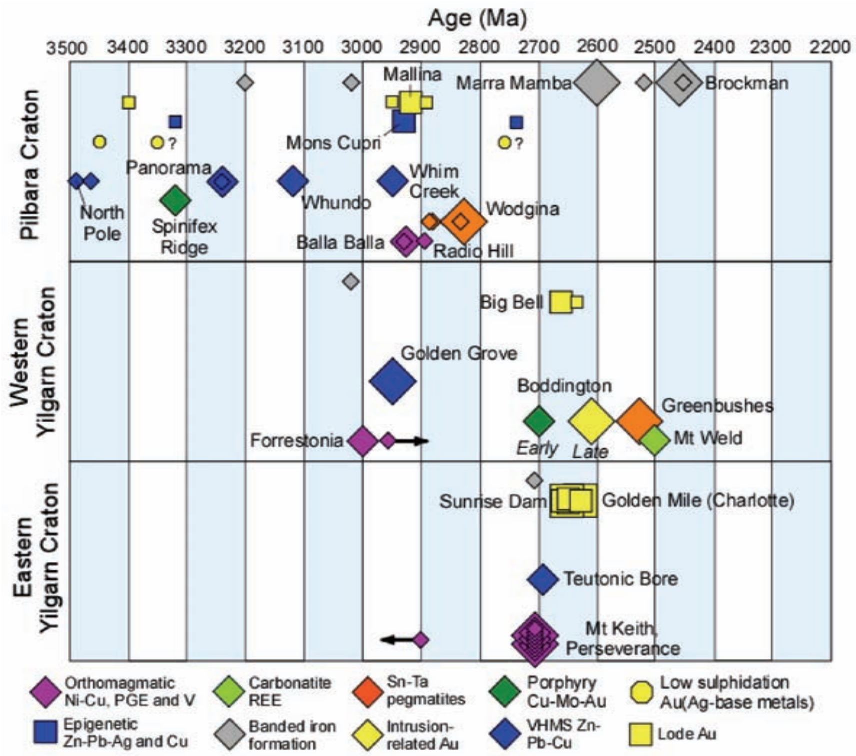



***Figure 4** Major mineral deposits of Australia formed between 3500–2200 Ma (see Figure 2 for localities of named deposits). All deposits of this age occur in the West Australian Element. VHMS = volcanic-hosted massive sulfide. The size of symbols indicates the relative size of the deposits.*

Being one of the best preserved blocks of old crust known on the Earth, the Pilbara Craton contains the earliest evidence of many modern geological processes. For example, the oldest unconformity is preserved by the c. 3430 Ma Strelley Pool Formation (Buick et al., 1995). This highly angular unconformity (Figure 6) overlies the earliest known weathering (regolith) profile. The Pilbara Craton contains the earliest unequivocal evidence of life on the earth, stromatolites hosted by the c. 3490 Ma Dresser Formation (Walter et al., 1980). The Dresser Formation also hosts the oldest ore deposit – the North Pole barite deposit (Figure 4), which produced 129,000 tonnes of barite for use as drilling mud in the NW Shelf petroleum province. The stromatolites are commonly closely associated with hydrothermal barite (Van Kranendonk et al., 2008), which supports the idea that life might have initially evolved in a hydrothermal environment (Baross and Hoffman, 1985).

Although they are generally small and of little economic interest, the Pilbara Craton contains the oldest examples of many types of mineral deposits (Figure 4), including volcanic-hosted massive sulfide, lode Au, epithermal precious metal and porphyry Cu deposits (Huston et al., 2007; Hickman and Van Kranendonk, 2012). These deposits share many features with geologically young examples, indicating that many mineralising processes have persisted throughout geological time. The oldest known hydrocarbons, within ore-related fluid inclusions, are associated with c. 3240 Ma volcanic-hosted massive sulfide deposits in the Panorama district (Rasmussen and Buick, 2000).

The Pilbara Craton was probably a constituent of the oldest supercraton, Vaalbara. This supercraton, which was made up of the Pilbara Craton and the Kaapvaal Craton in southern Africa began to form by 3600 Ma, and began to break up just after 2800 Ma (Zegers et al., 1998; Barley et al., 2005). In comparison with supercontinents, supercratons, which probably were smaller than most modern continents, appear to have been longer lived. The c. 2800 Ma breakup of Vaalbara (Barley et al., 2005) led to the formation of the Fortescue and Hamersley basins, and their equivalents in southern Africa. Rocks in these basins constitute the Earth's earliest preserved passive margin successions (Bradley, 2008). The basin fill is dominated by thick successions of mafic and felsic volcanic rocks and of sedimentary rocks, the most important being the banded-iron formations (Nelson et al., 1999). These 2590–2450 Ma banded-iron formations (Figure 4) formed when reduced $Fe^{2+}$-rich bottom waters were oxidised as they welled up onto the wide passive margin, depositing the iron (e.g., Cloud, 1973). The vast majority of Australia's and the world's banded-iron formations were deposited between c. 2600–1800 Ma, during a period when the Earth's hydrosphere was mostly oxygen-poor (Bekker et al., 2010).

The Yilgarn and Pilbara cratons differ in a number of important ways. Although both cratons have extended geological histories, the Yilgarn Craton is characterised by short period, even catastrophic, crust-forming events. This is shown in Figure 1b, in which a short sharp peak in the ages of igneous rocks corresponds to the final assembly of the Yilgarn Craton between 2720–2655 Ma. In contrast, the period between 3500–2850 Ma is marked by several small indistinct peaks, which may reflect a slower rate of overall crustal growth, particularly in the Pilbara Craton. This pattern is also seen in global data, with the largest peak in juvenile crust between 2700–2600 Ma, and a more diffuse peak at 3000–2800 Ma (Condie, 2005). Hawkesworth et al. (2010), however, noted that these peaks correspond to 'a particular stage in the cooling of the Earth', and possibly to a change in the mode of mantle convection (Korenaga, 2006).

In the North and South Australian elements, recent geochronology has indicated that Archean nuclei of these elements are more widespread and older than previously thought. For example, the recent identification of c. 3150 Ma granites in the South Australian Element extends its geological history by c. 600 Myr (Fraser et al., 2010). Recent dating has also greatly increased the extent of known Archean rocks in the North Australian Element (Hollis et al., 2011). Taken together, these new data indicate a much more significant and prolonged Archean history in both the North and South Australian elements; a history that will become clearer as more data are collected.

## 2200–1300 Ma: Amalgamation and breakup of Nuna

The Paleoproterozoic–Mesoproterozoic evolution of the Australian continent is also controversial. Tectonic models fall into two broad groups, 'fixist' or 'mobilist'. The fixist models (e.g., Etheridge et al., 1987) suggest little lateral movement between crustal blocks, whereas the mobilist models (e.g., Giles et al., 2004; Betts

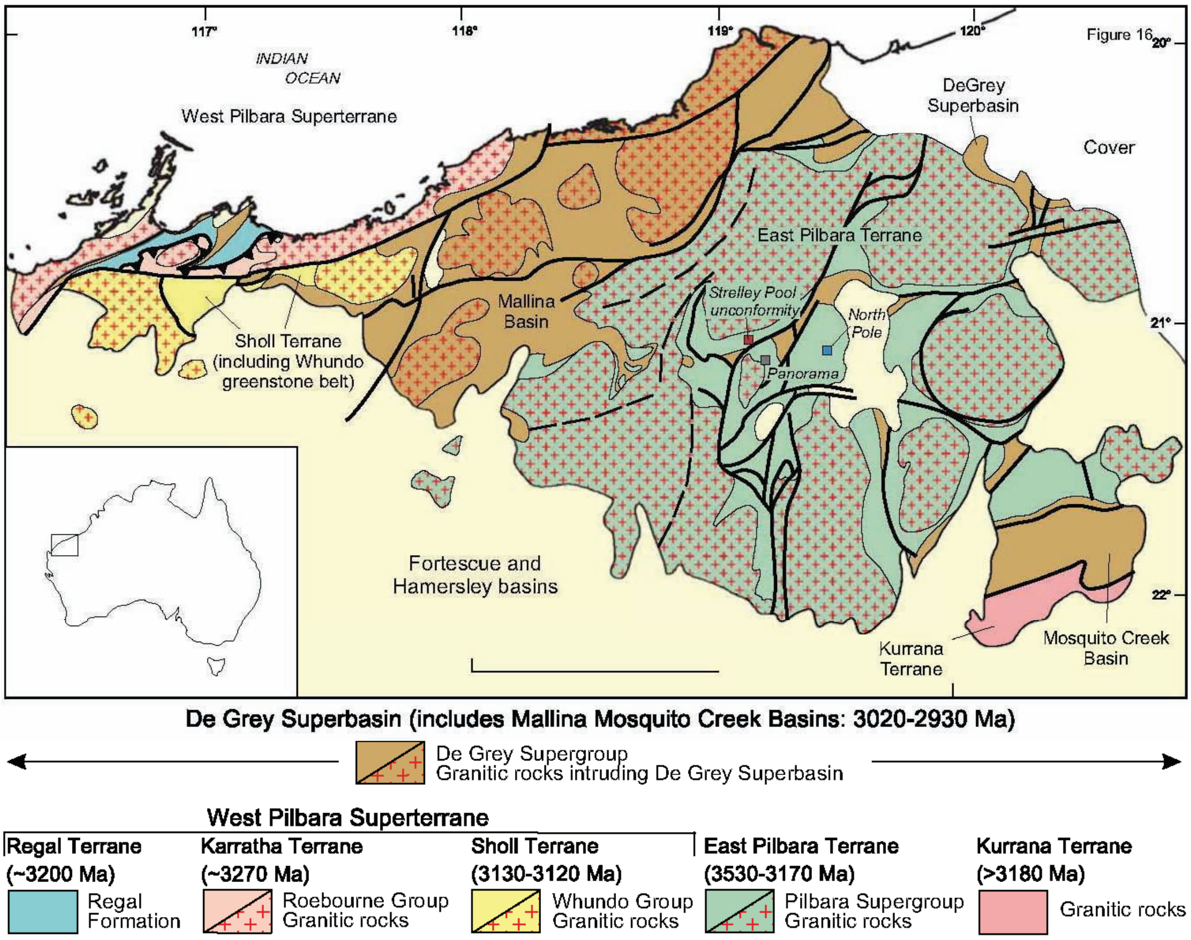



***Figure 5** The geology of the Pilbara Craton (modified after Champion and Smithies, 2007).*

and Giles, 2006; Cawood and Korsch, 2008) suggest large lateral movements occurred between crustal blocks. We have adopted a mobilist model that infers that the three major Precambrian elements of Australia were mostly assembled in the Paleoproterozoic as part of the supercontinent Nuna. However, there remain significant differences of opinion regarding the details of this assembly (Myers et al., 1996; Betts and Giles, 2006; Cawood and Korsch, 2008). We have followed the concept that the southern margin of the North Australian Element was a convergent margin through the late Paleoproterozoic–early Mesoproterozoic (Scott et al., 2000; Giles et al., 2002). Although this is consistent with the broad model of Betts and Giles (2006), alternative models have been proposed (Gibson et al., 2008; Payne et al., 2009). In many cases, the boundaries between provinces are determined in part based upon contrasts in seismic and/or magnetotelluric data.

***Figure 6** The Strelley Pool unconformity, a highly angular unconformity (solid white line) between the 3515–3498 Ma lower Warrawoona Group (formerly Coonterunah Group; dashed white lines show strike) and the 3350–3319 Ma Kelley Group.*

## 2200–1700 Ma: Amalgamation of Nuna

The Pilbara and Yilgarn cratons were amalgamated by a series of tectonic events from 2215–1950 Ma that affected the Capricorn Orogen and amalgamated the West Australian Element (Cawood and Tyler, 2004), which is one of the earliest building blocks of Nuna (Figures 1 and 7). Most of the North Australian Element formed before 1840 Ma as a consequence of the amalgamation of the combined Tanami-Tennant-Isa Province with the combined Kimberley-Pine Creek Province at c. 1840 Ma from the W, the Numil-Abingdon Seismic Province from the E before 1850 Ma, and the Aileron Province from the S before 1840 Ma (Figures 1 and 7; Sheppard et al., 1999; Tyler and Sheppard, 2006;

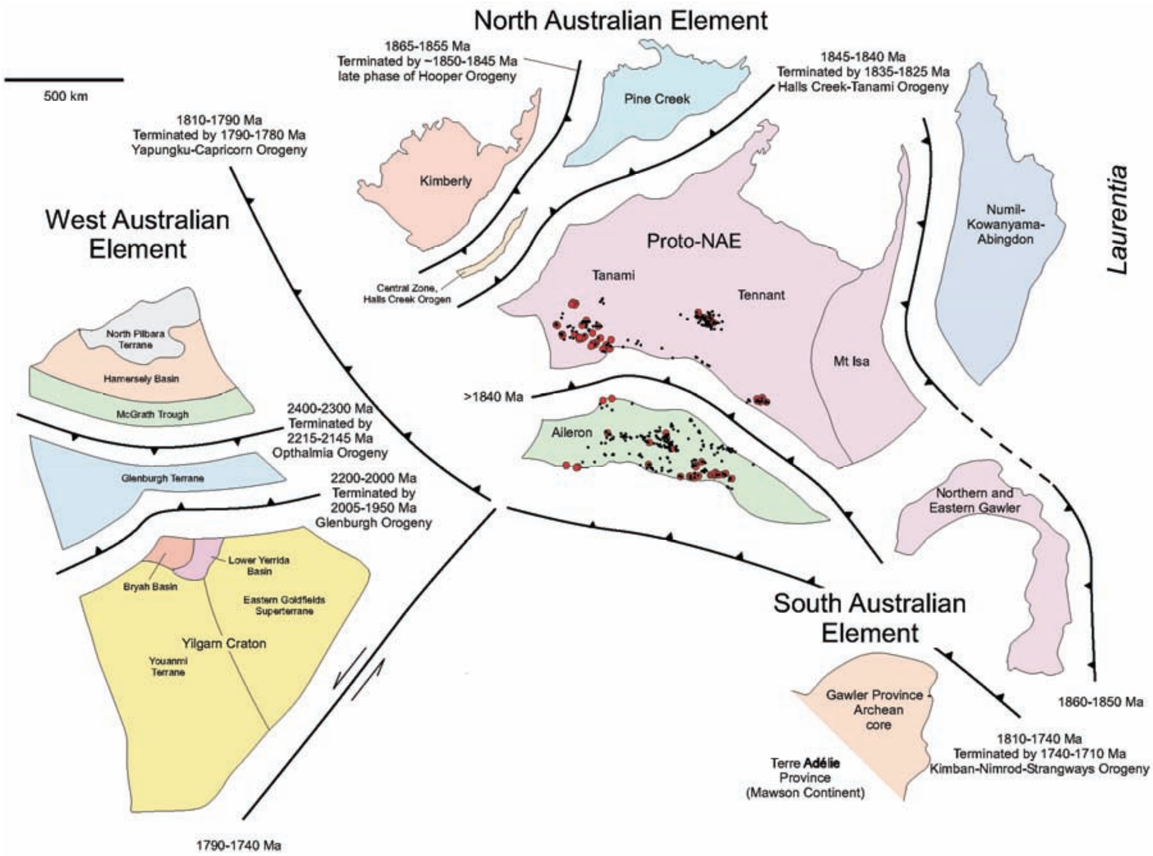



***Figure 7 An interpretation of the assembly of proto-Australia. The circles indicate analyses of 1820–1710 Ma granites from the southern part of the North Australian Element. The larger red circles indicate sodic (≥ 3.5% $Na_2O$) granite analyses, which are also characterised by high Sr, low Y and high Sr/Y ratios typical of subduction-related granites.***

Goleby et al., 2009; Korsch et al., 2012). Along the southern margin of the proto-North Australian Element (Figure 7), c. 1865 Ma turbiditic rocks of the Warramunga Formation (Compston, 1995) and Stubbins Formation (Bagas et al., 2008), may have formed in backarc basins associated with the northward convergence of the Aileron Province.

The northern and eastern parts of Gawler Province, which forms the core of the South Australian Element, grew, at least in part, as a consequence of subduction from the E, and may have been originally part of the North Australian Element (Figures 1 and 7; Payne et al., 2009). This interpretation is consistent with studies by Payne et al. (2006) that indicate the provenance of sediments from the northern and eastern Gawler Province is consistent with a source in the Isa Province, but not the Archean core of the Gawler Province.

By c. 1840 Ma, most of the West and North Australian Elements had been assembled. Although the North and South Australian Elements may have been together intermittently since c. 2500 Ma (Figure 1; Payne et al., 2009), the period between 1810–1750 Ma saw the assembly of two and possibly three Paleoproterozoic–Archean elements into the proto-Australian continent (Figure 7). Beginning at c. 1810 Ma, N- to NE-directed subduction along the southern margin of the North Australian Element resulted in convergence between the North Australian Element and the West Australian Element and the Archean core of the Gawler Province. The West Australian Element was first to dock with the North Australian Element at c. 1790–1780 Ma during the Yapungku-Capricorn Orogeny (Bagas, 2004). After this collision, N-directed subduction continued underneath the the North Australian Element, possibly with a sinistral strike-slip fault developing along the SE margin of the West Australian Element (Figure 7). This period of convergence concluded when the Archean core of the Gawler Province (Figure 7) was accreted onto the combined North and West Australian Element during the Kimban-Nimrod-Strangways Orogeny at 1740–1690 Ma (Betts et al., 2008).

This interpretation is supported by the emplacement of sodic granites and by the formation of mineral deposits characteristic of convergent margins along the southern margin of the North Australian Element. Two discrete belts of sodic granites are present along this margin. An older belt is located in the southern part of the W–NW-trending, 1815–1795 Ma magmatic belt in the Tanami-Tennant-Isa Province (Figure 7). A second, E-W trending belt occurs along the southern margin of the Aileron Province associated with the 1790–1750 Ma granites (i.e., the calc-alkaline-tonalite (CAT) suite of Zhao and McCulloch, 1995). Lode Au deposits (1810–1795 Ma; Cross et al., 2005) and volcanic-hosted massive sulfide (1810–1765 Ma; Hussey et al., 2005), which typically form along convergent margins, developed in the southern part of the North Australian Element. It is likely that Laurentia was joined to the proto-Australian continent on the E until at least 1690 Ma (Betts and Giles, 2006).

## 1700–1300 Ma: Breakup of Nuna

Shortly thereafter, Nuna began to break up. This process, however, was complicated and involved rifting of Laurentia to the E of proto-Australia, but development of a backarc basin system along the southern margin of Proterozoic Australia as subduction stepped to the S (Figure 8; Scott et al., 2000; Giles et al., 2002; Betts et al., 2003). The earliest evidence of extensional processes is the emplacement of 1710–1685 Ma layered mafic-ultramafic intrusions within the southern part of the Aileron Province (Claoué-Long and Hoatson, 2005). The development of basins filled with turbidites and emplacement of tholeiitic mafic rocks along the eastern margins of the South Australian Element in the Curnamona Province and the North Australian Element (Willis et al., 1983; Beardsmore et al., 1988) suggest extension along the eastern margin of Proterozoic Australia beginning at c. 1690 Ma (Betts et al., 2003; Gibson et al., 2008; Figure 8). This extension produced the Calvert and Isa superbasins (Southgate et al., 2000), which, together with the related Curnamona Province, are hosts to the Australian Proterozoic Zn belt. Ultimately it also resulted in the separation of proto-Australia and Laurentia (Betts et al., 2003), with final separation occurring to the E of the Numil-Kowanyama-Abingdon Province in Queensland (Figure 8).

Beginning at c. 1690 Ma, felsic magmatic rocks and minor sedimentary rocks, now orthogneiss and paragneiss, were deposited as the oldest known rocks in the Warumpi Province (Scrimgeour et al., 2005). In addition, Kirkland et al. (2011) inferred convergence and the development of backarc basins along the SE margin of the West Australian Element between 1710–1665 Ma. The inference of N-dipping subduction is supported by 1690–1665 Ma granitic rocks with arc-like geochemical signatures in this region (Eddy Suite; Kirkland et al., 2011) and the Warumpi Province (Argilke Suite; Cawood and Korsch, 2008).

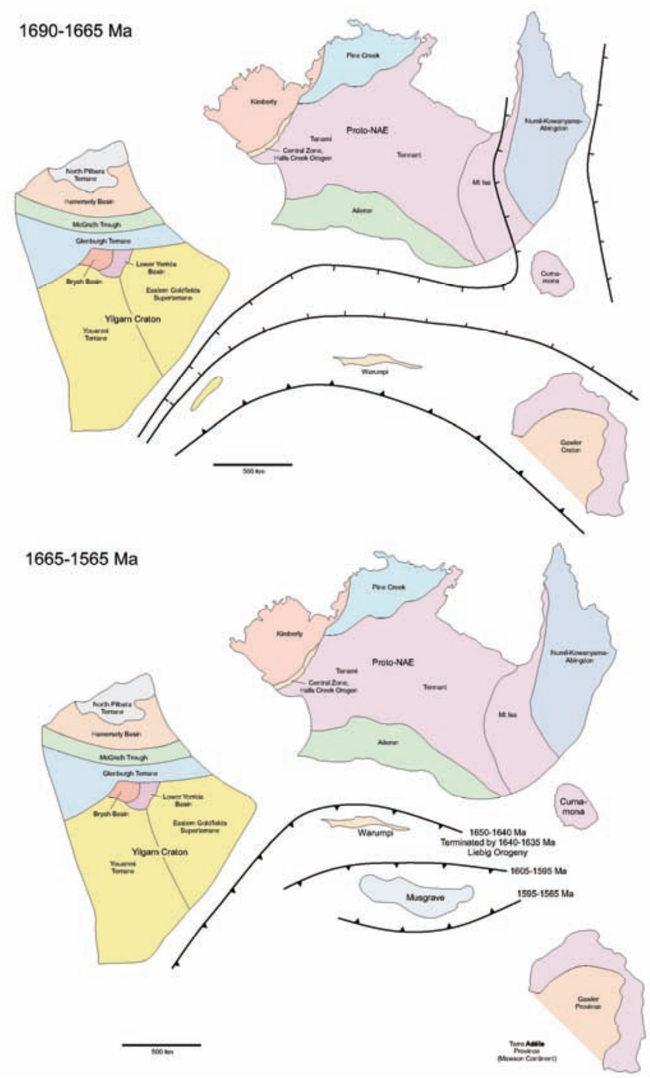



***Figure 8 An interpretation of the evolution of proto-Australia from 1690–1565 Ma. This involved extension along the eastern margin of proto-Australia and separation from Laurentia (1690–1665 Ma) and convergence along the southern margin of the combined North and West Australian elements (1665–1565 Ma).***

At c. 1660 Ma, this continental margin backarc basin system began to close through a series of both S- and N-directed subduction systems that led to the accretion of rocks of the Warumpi Province at c. 1640 Ma (Scrimgeour et al., 2005), followed by those of the Musgrave Province at c. 1590 Ma (Wade et al., 2006), and those of the Gawler Province at c. 1560 Ma (Korsch et al., 2011a). During this period, the Gawler Province was likely the northern extension of the Mawson Continent, which also included the Terre Adélie Craton and large parts of the East Antarctic Shield (e.g., Boger, 2011), although Swain et al. (2008) suggested that the Terre Adélie Craton accreted onto the Gawler Province at c. 1610 Ma, following NE-dipping subduction. At some time after c.1560 Ma, the Mawson Continent (including the Gawler and Curnamona provinces) must have rifted from proto-Australia as these elements converged again during the assembly of Rodinia (Giles et al., 2004).

The period 1640–1500 Ma also involved episodic deformation through much of Proterozoic Australia. The earliest deformational event system, at c. 1640–1635 Ma, affected only the North Australian Element, and involved N-S-directed contraction along the southern margin (Liebig Orogeny; Scrimgeour et al., 2005) and in the E (Riversleigh inversion; Geological Survey of Queensland, 2011). This event may be the result of the accretion of the Warumpi Province (Figure 8; Scrimgeour et al., 2005) and is marked by a U-turn on the North Australian apparent polar wander path (Idnurn, 2000; Figure 9).

Between 1605–1585 Ma, the Isa (early phase of the Isan Orogeny; Geological Survey of Queensland, 2011), Aileron (early part of the Chewings Orogeny; Rubatto et al., 2001) and Curnamona provinces (Olarian Orogeny; Page et al., 2005) were affected by contractional deformation. In the Isa Province, this event was a N-S to NW-SE directed crustal-shortening event (Geological Survey of Queensland, 2011) and may relate to accretion of the Musgrave Province (Figure 8).

Similarly, the northern part of the Gawler Province was affected by high pressure metamorphism at this time (Cutts et al., 2011), although this deformation is not known to extend into the Archean core of the province. Rather, the central part of the Gawler Province is characterised by the emplacement of the 1595–1575 Ma Hiltaba magmatic province, which includes the felsic-dominated Gawler Range Volcanics and coeval Hiltaba granite suite. Betts et al. (2007) interpreted this magmatic event to be the consequence of a migrating hot spot or plume, and the timing of magmatism corresponds to a shift from a contractional to an extensional tectonic environment (Skirrow, 2010). The Chewings Orogeny in the Aileron Province, which is largely a low pressure-high temperature magmatic event continued until 1560 Ma (Rubatto et al., 2001), may be a consequence of backarc extension related to convergence between the Gawler Province and the North Australian Element (Korsch et al., 2011a; Figure 8).

After c. 1570 Ma, thermotectonic activity mostly contracted to

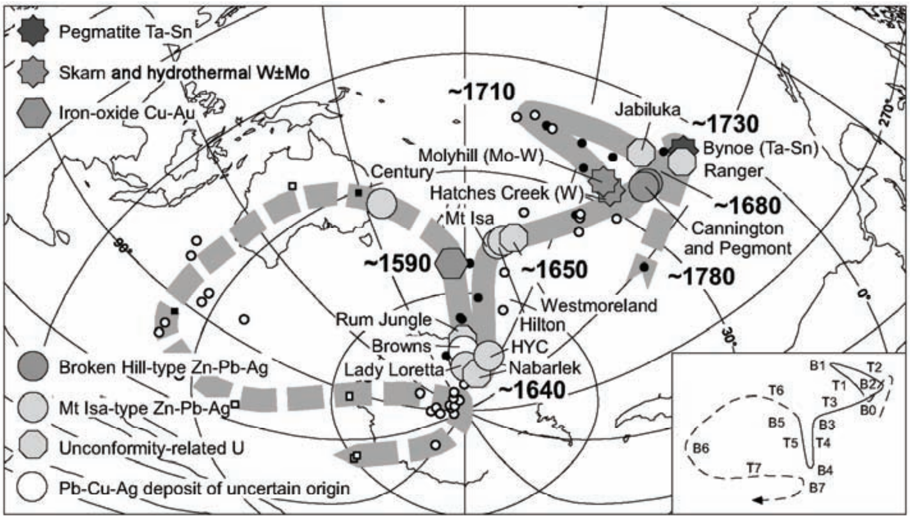



*Figure 9 Variations in the apparent polar wander path for the North Australian Element during the late Paleoproterozoic to the early Mesoproterozoic showing relationships to the timing of major ore deposits (modified after Idnurm et al., 2000). Dashed lines indicate uncertain parts of the wander path.*

eastern Proterozoic Australia, although restricted activity occurred in the Gawler and Curnamona provinces. The 1560–1540 Ma Middle Isan Orogeny (and the Jana Orogeny further to the E), which involved E-W contraction, may relate to convergence between Australia and Laurentia (Geological Survey of Queensland, 2011). Emplacement of the 1545–1500 Ma anorogenic Williams Suite and associated deformation may relate to hot spot movement (Betts et al., 2007).

A large part of Australia's mineral wealth, particularly Fe ore and base metals, formed during the amalgamation and breakup of Nuna (Figure 10). Upgrading of Fe ore deposits in the Hamersley Basin (c. 2008 Ma; Müller et al., 2005), and formation of lode Au deposits in the Pine Creek and Tanami provinces (c. 1810–1790 Ma; Compston and Matthai, 1994; Cross et al., 2005) occurred during amalgamation (Figure 7). In contrast, Zn-Pb-Ag deposits of the Australian Proterozoic Zn belt (1690–1575 Ma; Betts et al., 2003; Leach et al., 2010) formed during breakup (Figure 10), and deposits of the iron oxide-Cu-Au deposits in the Olympic Dam (c. 1575 Ma; Skirrow et al., 2007) and Cloncurry mineral provinces (mostly 1530–1500 Ma; Duncan et al., 2011) may relate to a migrating hot spot or plume (Betts et al., 2007). In the North Australian Element, the timing of formation of many deposits corresponds to bends in the Australian apparent polar wander path (Figure 9; Idnurm, 2000). These bends are responses to changes in plate motion and/or plate reorganisation associated with large-scale tectonic processes, for example at c. 1640–1590 Ma.

The Mesoproterozoic was also a time for the generation of abundant high heat-producing granite magmatism and felsic volcanism rock (Neumann et al., 2000). The resulting rocks contribute to Australia's crustal heat production today, providing a thermal resource for geothermal power as well as a uranium resource for nuclear power. Another curiosity of this time is the development of the world's oldest oil play, although in uneconomic quantities, in the Upper Roper Group of the Mesoproterozoic McArthur Basin, NW of Mount Isa (Jackson et al., 1986).

Iron ore deposits in the Hamersley Basin are unusual relative to other ore deposits in Australia in that they formed as the consequence of three geological events, which were several hundreds of millions to billions of years apart, and occurred in quite different geological environments. Banded-iron formation proto-ore, deposited prior to the Great Oxidation Event (2450–2090 Ma; Farquhar et al., 2010), was upgraded by oxidised, basinal brines at c. 2008 Ma (Müller et al., 2005), which oxidised the Fe and removed Si and carbonate. Then paleo-weathering removed P, leaving high-grade, direct shipping ore (Barley et al., 1999; Taylor et al., 2001).

## 1300–700 Ma: Amalgamation and breakup of Rodinia

Globally, the period between 1300–700 Ma saw the assembly and breakup of the second supercontinent, Rodinia (Figure 11). In Australia, this period was characterised by crustal reworking in the Albany-Fraser-Musgrave belt and the later formation of the intracratonic Central Australian Basin System. New discoveries and more data from existing mineral deposits indicate the presence of important mineralising events during this period, albeit of different character to other times during Australia's history.

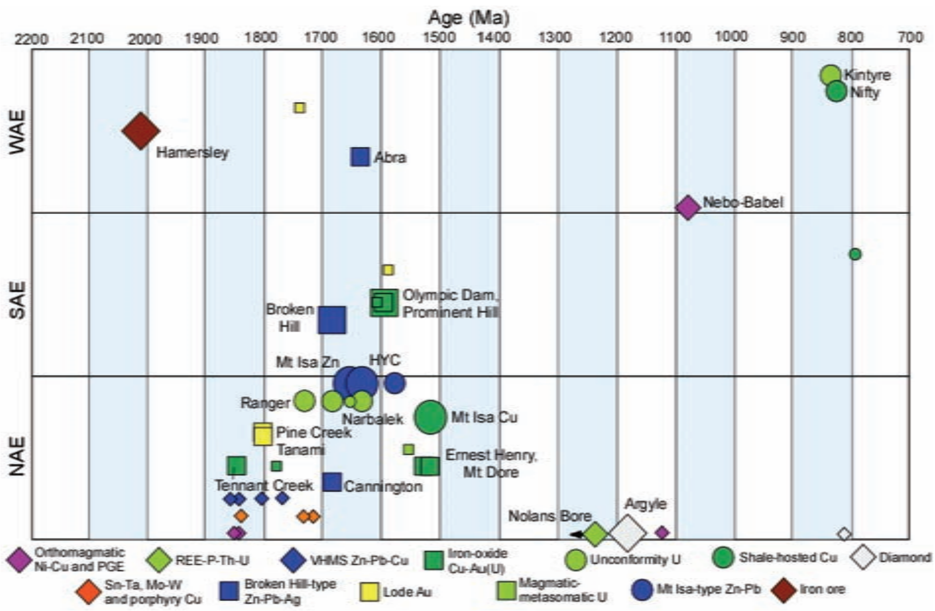



*Figure 10 Major mineral deposits of Australia formed between 2200–700 Ma. See Figure 2 for localities of named deposits and Figure 4 for explanation of acronyms and symbol size.*

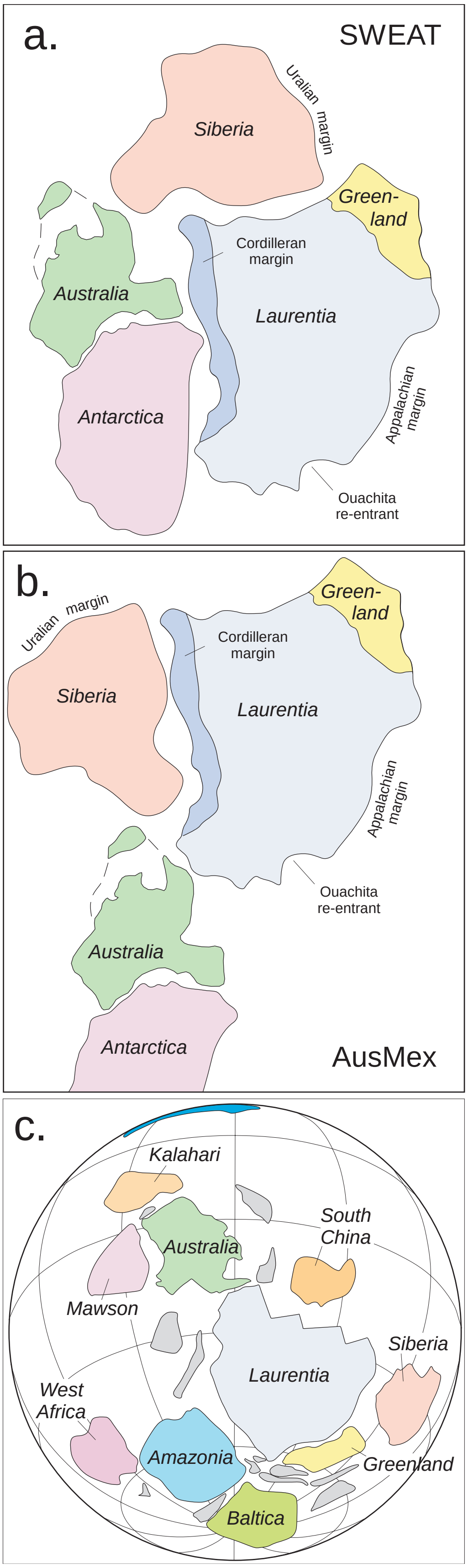



***Figure 11 Possible reconstructions of Rodinia. (a) the SWEAT (South-west (US) - East Antarctic: Hoffman, 1991) reconstruction. (b) the Ausmex (Australia-Mexico: Wingate et al., 2002) reconstruction. (c) the modified AusMex (Pisarevski et al., 2003) model.***

The earliest major events that affected Australia during the 1300–700 Ma period were a series of deformational events that overprinted the Albany-Fraser and Musgrave orogens between 1345–1140 Ma (Myers et al., 1996; Smithies et al., 2010; Kirkland et al., 2011). These events probably relate to the collision of the combined West-North Australian Element with the Mawson Craton, which includes the South Australian Element and the East Antarctica Shield (Boger, 2011). This collision was a prelude to the assembly of Rodinia, which was mainly assembled between 1100–980 Ma (Pisarevsky et al., 2003).

Mineral deposits that formed prior to and during the assembly of Rodinia are quite diverse, and include diamond, REE and orthomagmatic Ni-Cu-PGE deposits (Figure 10). The diamond and REE deposits are associated with alkalic rocks, including the c. 1178 Ma Argyle diamond pipe (Pidgeon et al., 1989), which may be the result of an intracratonic plume event. The volcanic-hosted massive sulfide and lode Au deposits that characterise the assembly of earlier (Kenorland and Nuna) and later (Pangea-Gondwana) supercontinents/supercratons are missing, both in Australia and overseas, which suggests important differences in the geodynamic processes that accompanied growth of Rodinia compared to those that were associated with the growth of Nuna and Pangea-Gondwana.

Rifting leading to the breakup of Rodinia commenced c. 850 Ma with final breakup at c. 750 Ma. One of the earliest manifestations of this process in Australia is the NW-trending c. 830 Ma Gairdner Large Igneous Province (Wingate et al., 1998), which intruded the South and North Australian Elements and extended to the Paterson Orogen in the NW (Hoatson et al., 2008). Extension associated with Rodinia breakup likely led to the formation of some of the Central Australian Basin System. Development of this basin system, which began at c. 850 Ma (Walter et al., 1995), and extended to the Devonian, affected most of central Australia and includes the Officer, Amadeus, Georgina and Yeneena basins as well as the Adelaide Rift System (Figure 2). These basins include the earliest known major salt deposits, in the Bitter Springs Formation of the Amadeus Basin, evidence for several periods of glaciation and the first flowering of multicellular life (see below). Deposition of the Bitter Springs and subsequent salt deposits around the world has lowered the salinity of seawater by at least a factor of 1.5–2 since the Neoproterozoic (Knauth, 2005).

Uranium and Cu deposits in the Yeneena Basin and the Adelaide Rift System formed between 840–790 Ma (Huston et al., 2010); these deposits were related to basin formation and/or inversion. Some reconstructions of Rodinia place Australia adjacent to the Kalahari Craton (Pisarevsky et al., 2003), which contains the highly productive Zambian Copper Belt (Selley et al., 2005). This belt is hosted by a basin of similar age and with similar fill to parts of the Central Australia Basin System; and the ages of ore deposition overlap.

## 700–0 Ma: Amalgamation and breakup of Pangea-Gondwana

The Phanerozoic, particularly the Paleozoic, was a period of extensive tectonic activity in Australia, particularly in the Tasman Element, which makes up the eastern third of the continent (Figure 2). This tectonic activity is related to the assembly and breakup of Pangea, a supercontinent that started to assemble at c. 600 Ma, was assembled by c. 250 Ma, and began to breakup shortly thereafter. Eastern Australia was always along an active margin during Pangea assembly. After c. 180 Ma, Pangea, and then Gondwana, began to break up, culminating with final separation of Australia from Antarctica at 34 Ma (Veevers et al. 1991). Since 65 Ma, Australia has been relatively tectonically stable. Nevertheless, this stability will be relatively short lived in geological terms. Australia is currently moving northward, and is colliding with SE Asia as a prelude to the assembly of Amasia, interpreted by some as the world's next supercontinent (Santosh et al., 2009).

## 700–250 Ma: Amalgamation of Gondwana and Laurasia to form Pangea

Evolution of the Central Australian Basin System continued into the Paleozoic, recording evidence of major global ice ages, from the Yeneena Basin in Western Australia to the Adelaide Rift System in South Australia and King Island in Tasmania (Figure 2). Hoffman et al. (1998) suggested that these glacial events are representative of "Snowball Earth". During the late Neoproterozoic, the complexity of life increased such that towards the end of this era the first evidence of multicellular life appeared, including delicately preserved fossils in uppermost unit of the Pound Subgroup in the Ediacara Hills of the Adelaide Rift System (Glaessner and Wade, 1966). It is notable that these fossils began to flourish after the last of the Neoproterozoic glacial events.

The oldest activity in Australia related to Gondwana amalgamation was in Western Australia (Figure 12). Deformation at 650–600 Ma

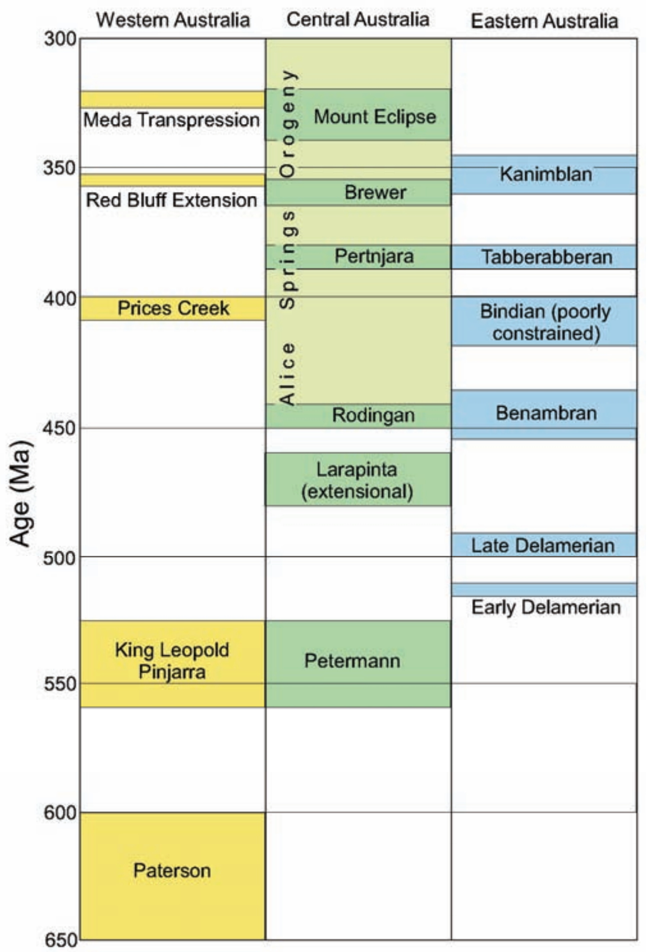


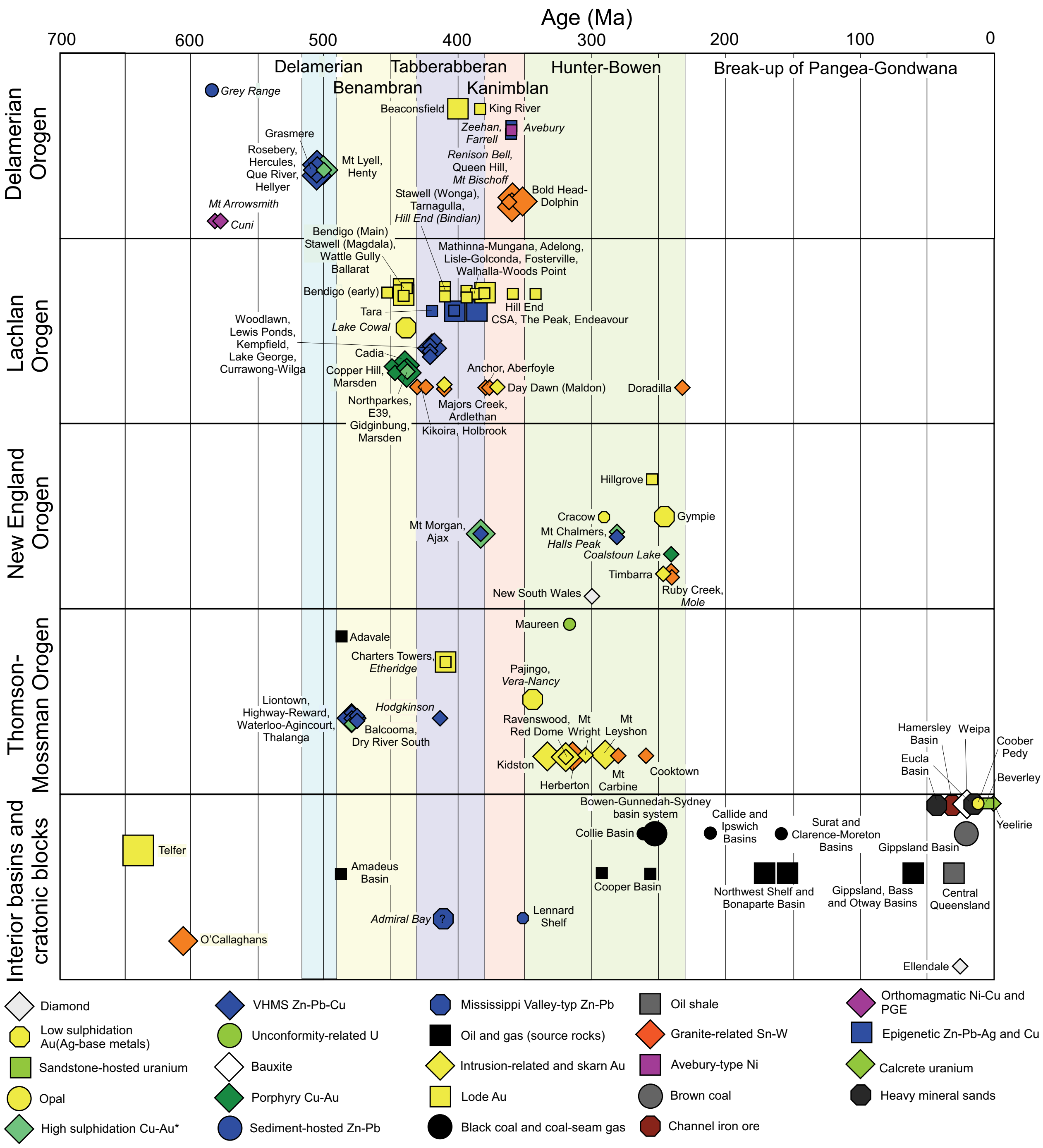

*Figure 12 Space-time diagram showing the relationship of tectonic events in western, central and eastern Australia.*

was accompanied by granite emplacement and related Cu-Au and W-Cu-Zn mineralisation in the Paterson Province (Goellnicht et al., 1989; Huston et al., 2010; Figure 13). More extensive deformation occurred between 560–525 Ma during the coeval King Leopold and Petermann orogenies, which affected and Kimberley and Musgrave provinces, respectively (Shaw et al., 1992; Aitken et al., 2011; and references therein). Zircon U-Pb discordia in the Tanami Province have similar lower intercept ages (Maidment, 2006), suggesting that the linked King Leopold-Petermann Orogeny extended from NW into central Australia (Figure 12). Aitken et al. (2011) suggested that the Petermann Orogeny is an intraplate response to the Kuunga Orogeny, which involved oblique collision between Australia and India during Gondwana assembly.

More direct evidence of Gondwana assembly is present along the southern part of the W coast of Australia, where the Pinjarra Orogen (Pinjarra Element) comprises elements of eastern Gondwana, notably the Australo-Antarctic and Indo-Antarctic domains. These represent distinct continental fragments with different Proterozoic histories that were juxtaposed through oblique collision by c. 522 Ma (Collins, 2003a; Collins and Pisarevsky, 2005). The breakup of Gondwana has since re-separated these domains, although the fragments are still present in the Pinjarra Element (Boger, 2011). Immediately after the King Leopold, Petermann and Pinjarra orogenies, the c. 510 Ma Kalkarindji large igneous province, one of the most extensive known, was emplaced through much of western and north-central Australia, possibly as a response to a mantle plume (Glass and Phillips, 2006).

In eastern Australia, the Tasman Element, which extends from Tasmania to northern Queensland, was built upon a passive margin formed by the breakup of Rodinia (Figure 14). This orogen is part of a global scale orogenic system, the Terra Australis Orogen that exceeded 18,000 km in length and extended along the margin of Gondwana (Cawood, 2005). The Tasman Element developed as the consequence of a sequence of tectonic cycles (Glen, 2005). Each of these cycles, which lasted between 30–130 Myr, initiated with arc magmatism and/or backarc extension, related, in most cases, to W-dipping subduction, and ended with contractional orogenesis as continental slivers and island arcs were accreted back onto the Australian continent (Gray and Foster, 2004; Collins and Richards, 2008). Five such cycles have been recognised (Glen, 2005; Champion et al., 2009): the Delamerian (515–490 Ma), Benambran (490–440 Ma), Tabberabberan (440–380 Ma; includes the Bindian), Kanimblan (380–350 Ma), and Hunter-Bowen (350–220 Ma). Although younger cycles overprint older cycles, particularly in N Queensland, the cycles broadly young eastwards. Collins (2003b) and Collins and Richards (2008) suggest that this cyclicity may relate to switching between long-lived subduction retreat, and associated extension, and short-lived subduction advance, and associated contraction.

There is a progression in the types of mineral deposits formed within individual cycles and an eastward progression in mineral system activity with time (Figure 13). Deposits formed in backarcs (e.g., volcanic-hosted massive sulfide deposits) formed early in cycles, whereas deposits that form mostly during contractional deformation (e.g., lode Au and structurally-hosted base metal deposits) formed late in the cycle. Juxtaposition of these mineral systems has resulted in a highly complex metallogeny through most of the Tasman Element, perhaps exemplified in western Tasmania, which is one of the richest and most diverse metallogenic provinces in the world, with major Zn-Pb-Ag-Cu-Au, Cu-Au, Au, Sn, W and Fe deposits within a few tens of kilometres of each other (Green, 2012). Other important




*Includes high sulphidation epithermal and VHMS deposits
Size of symbol indicates relative size of deposit. Normal text indicates well constrained ages; itallics indicate ages inferred from geological framework

***Figure 13 Major mineral deposits of Australia formed during the period 700–0 Ma, particularly during the assembly and break-up of Pangea-Gondwana. See Figure 2 for localities of named deposits. See Figure 2 for localities of named deposits and Figure 4 for explanation of acronyms and symbol size.***

districts in the Tasman Element include lode Au deposits in the Victorian goldfields and at Charters Towers in N Queensland, and porphyry Cu-Au deposits (e.g., Cadia) in the Macquarie Arc in New South Wales (Figure 2).

Central Australia underwent a series of intraplate deformational events during the Paleozoic. Many are time equivalent to events in the Tasman Element (Figure 12), suggesting that the activity related to convergence on the eastern seaboard also influenced the continent's interior. These central Australian events include development of the Larapinta Seaway and the Alice Springs Orogeny. The Larapinta Seaway linked the Ordovician marine sediments of the Georgina and Amadeus basins through to the Gondwana margin in eastern Australia (Bradshaw, 1993; Figure 14a). These basins host the Larapintine petroleum system, which was sourced from organic-rich rocks deposited in warm and shallow, early Paleozoic seas (Bradshaw, 1993).

The 480–460 Ma Larapinta metamorphic event, which is characterised by granulite facies metamorphism, is an extensional event that is restricted to the Irindina Province (Hand et al., 1999;

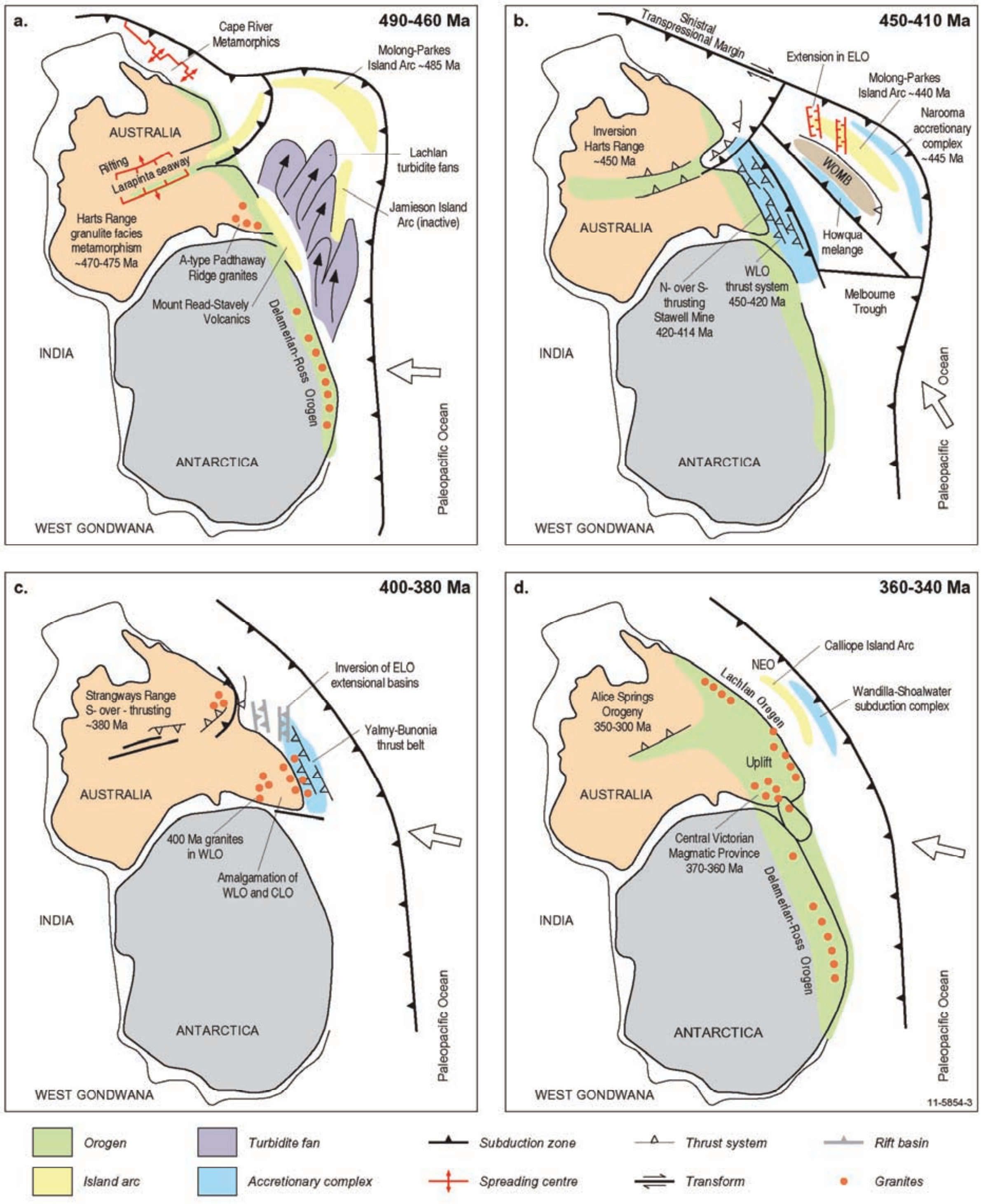



***Figure 14 Schematic diagrams showing the tectonic evolution of Australia between 490–360 Ma (after Gray and Foster, 2004). ELO, CLO, WLO and NEO =Eastern Lachlan, Central Lachlan, Western Lachlan and New England orogens, respectively.***

Maidment, 2005, 2006). This province is interpreted to be a deep (possibly to 30 km), extensional sub-basin within the Larapintine Seaway, with tholeiitic basaltic rocks, which host semi-massive Cu-Co sulfide deposits, present near the base. Hoatson et al. (2005) suggested that these basaltic rocks formed in an extensional environment, and correlated them with the Antrim Plateau Volcanics of the Kalkarindji Large Igneous Province.

The Alice Springs Orogeny was a succession of four contractional events (or movements; Table 1), the 450–440 Ma Rodingan, 390–380 Ma Pertnjara, 365–355 Ma Brewer and 340–320 Ma Mount Eclipse movements, which inverted large parts of the Larapinta Seaway (Gray and Foster, 2004; Scrimgeour and Close, 2011). During orogenesis, deep crustal rocks of the Irindina Province were exhumed from 30 km and juxtaposed with time-equivalent sub-greenschist-grade rocks in the Georgina Basin (Hand et al., 1999; Maidment, 2005; Korsch et al., 2011b). As part of this orogeny, the S-directed Redbank Shear Zone offset the Moho, creating one of the largest gravity anomalies (c. 150 mgal) from continental interiors on the Earth

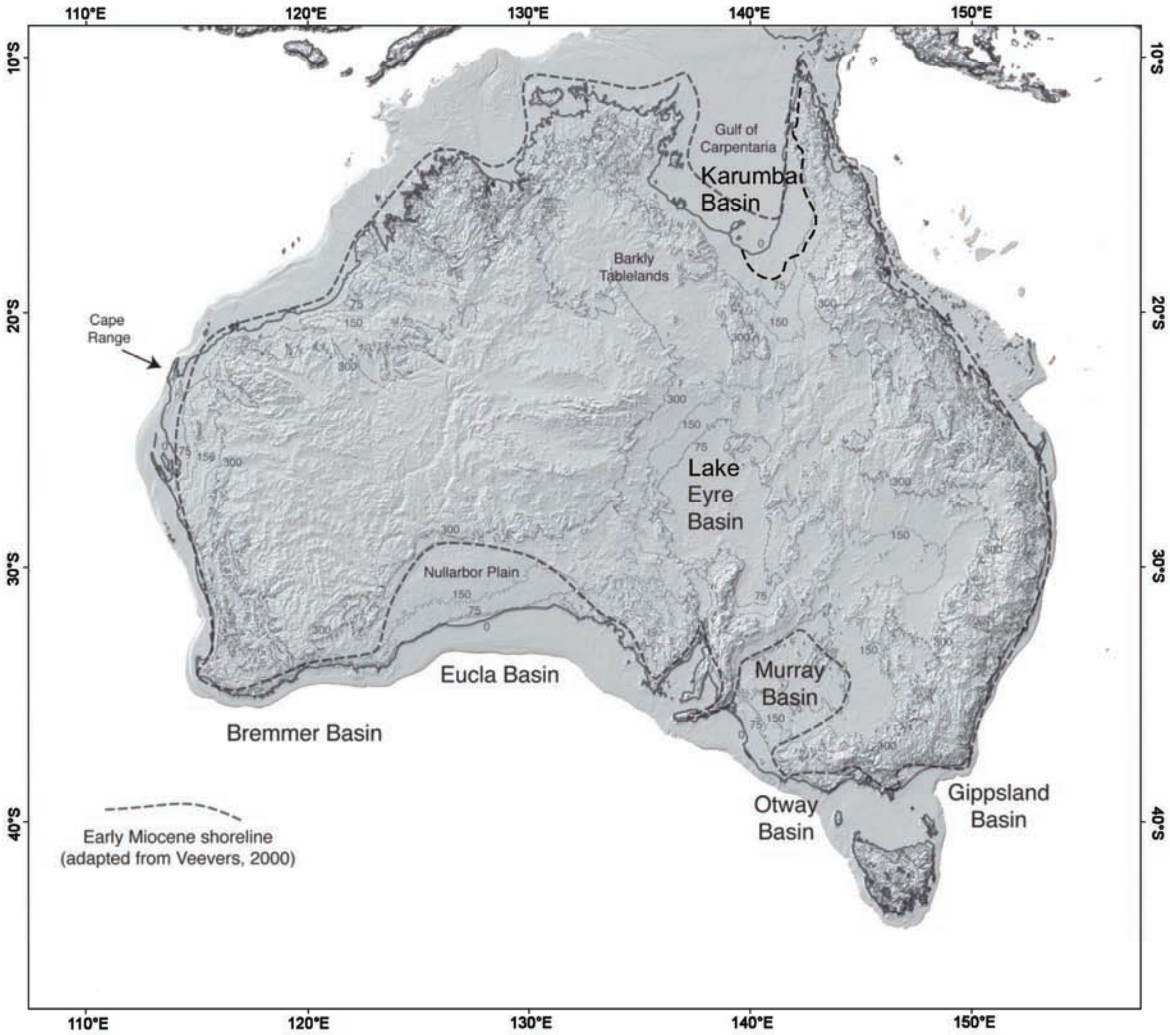



*Figure 15 Distribution of Cenozoic basins in Australia over digital elevation model (modified after Sandiford et al., 2009).*

(Mathur, 1976). Uplift and erosion associated with the Brewer Movement resulted in formation of a foreland basin (Haines et al., 2001) and may have driven hydrothermal flow that formed c. 357 Ma (Christensen et al., 1995) Mississippi Valley-type Pb-Zn deposits in the Canning Basin of Western Australia.

Most of the vast coal deposits of the Sydney-Bowen-Gunnedah basin system in eastern Australia (Figure 2) formed in a Permian foreland basin behind a continental volcanic arc system that developed in the New England Orogen. Counter flow in the mantle wedge above the subduction zone is thought to have created a slowly subsiding basin that included thick coal measures (Waschbusch et al., 2009). The high-latitude position of Australia at this time, an environment similar to the modern Russian taiga, permitted the accumulation of vast amounts of peat from *Glossopetris* and other species (Veevers, 2006). These peat seams, accumulated over hundreds of thousands of years, and were compacted into black coal, with the thickest seams at Blair Athol being more than 31 m thick (Hobday, 1987).

## 250–160 Ma: Stabilisation and breakup of Pangea

After final amalgamation at c. 250 Ma, Pangea remained stable for less than 100 Myr before it started to break apart. This breakup initiated at c. 180 Ma as the Atlantic Ocean and Tethys Sea began to form, separating Gondwana, which included Australia, South America, Africa, India and Antarctica, from Laurasia, which included North America and Eurasia (Dietz and Holden, 1970). The emplacement of voluminous c. 175 Ma tholeiitic dolerite sills (Brauns et al., 2000) in Tasmania, which form part of the Karoo-Ferrar Large Igneous Province, presaged the breakup of Gondwana.

## 160–65 Ma: Breakup of Gondwana and creation of Australia

In the Late Jurassic, a triple point of extension, centred just SW of Perth in Western Australia, began to separate India, Antarctica and Australia (Veevers, 2006). The NW margin of Australia had formed by 155 Ma, when the last of a series of continental slivers (Argo Land, now in Burma) rifted away. Australia separated from Greater India by 120 Ma (Norvick and Smith, 2001; Veevers, 2006) and the W coast faced the Indian Ocean.

Formation of major gas fields of the Westralian petroleum system on the North West Shelf are related directly to the breakup of Gondwana. Late Jurassic rifting along the NW-SE Westralian trend produced deep-marine depocentres into which organic-rich source rocks were deposited (Bradshaw, 1993). Clastic rocks of the Late

Jurassic–Early Cretacoues Barrow delta, which formed as extension continued, buried the source rocks into the hydrocarbon-generating window and provided reservoir rocks. The seal was provided by extensive Cretaceous marine shales that were deposited after continental breakup (Bradshaw, 1993).

Australia separated from Antarctica via an easterly propagating rift system. Rift basins had begun to form as far E as the Polda Basin in South Australia at 155 Ma, and by 145 Ma rifting in the Gippsland Basin in Victoria had occurred (Norvick and Smith, 2001). The submarine Kerguelen Large Igneous Province in the Southern Ocean was emplaced at c. 110 Ma (Duncan, 2002). During the Jurassic and Cretaceous, a large inland sea covered central and eastern Australia, forming the Eromanga and related basins, which are as thick as 1,600 m. These basins formed by dynamic changes to topography possibly caused by corner flow in the asthenosphere below eastern Australia during W-directed subduction (Waschbusch et al., 2009).

At 95 Ma, subduction jumped eastward, stopping the corner flow and triggering uplift of the eastern highlands of Australia (Waschbusch et al., 2009). The Eromanga Basin was inverted and began to erode, and a large delta system was deposited in the offshore Ceduna region, South Australia (Bradshaw, 1993; Norvick and Smith, 2001). By this time, a significant seaway had developed between Australia and Antarctica, although there was still connection through Tasmania. Extension associated with Australia-Antarctica breakup occurred between present-day Tasmania and Victoria to form a rift basin system that included the Otway, Bass and Gippsland basins, which cumulatively host one of Australia's main hydrocarbon provinces (Bradshaw, 1993).

Seafloor spreading along the E coast began c. 84 Ma in the S, and propagated northwards, resulting in the opening of the Tasman Sea and rifting off the Lord Howe Rise and New Zealand. By 56 Ma the Tasman Sea ceased opening (Gaina et al., 1998). Sea floor spreading commenced further N with the Coral Sea opening and the Queensland Plateau rifting from the NE coast of Queensland (Weisel and Watts, 1979).

The long-term and deep weathering of Australia is one of its distinguishing features compared to northern hemisphere continents. Much of the weathering began as Gondwana broke apart. Oxygen isotope data indicate that clays produced during Mesozoic weathering formed in low latitudes as Australia began to move N from Antarctica (Chivas and Atlhopheng, 2010). By the end of the Cretaceous, the fundamental physiography and the paleovalley architecture in today's arid Australia was established (Fujioka and Chappell, 2010). Some of these paleo-land surfaces are as old as the Cambrian, but fission track dating suggests that some of these very ancient landscapes were buried and then more recently exhumed (Pillans, 2007).

## 65–2.6 Ma: Australia girt by sea

The past 65 Ma have left an indelible imprint on the Australian continent, with 80% of the surface geology of Australia made up of Cenozoic deposits. Most of the cover is thin aeolian, lacustrine and fluvial deposits, although many discrete basin entities developed, including the Eucla, Murray, Lake Eyre and the Karumba basins (Figure 15). Sea-floor spreading in the Southern Ocean accelerated at c. 45 Ma, but it was not until 34 Ma that full separation of Australia and Antarctica was achieved (Zachos et al., 2001; Livermore et al., 2005). From c. 45–10 Ma, Cenozoic sediments were deposited in the Lake Eyre, Eucla, Karumba, and Murray basins. The climate was warmer and wetter than today, and rainforests were extensive (Fujioka and Chappell, 2010). During this time, the sea made several incursions onto the land, forming, for example, Eocene (c. 40 Ma) and Miocene (15–5 Ma) shorelines that are preserved along the northern margins of the Eucla and Murray basins (Benbow et al., 1995). These marine incursions have left a legacy of paleoplacer deposits of rutile and other heavy minerals (Hou et al., 2003).

Between c. 45 Ma–5 ka, Australia experienced episodic, mainly mafic magmatism that resulted in a series of volcanic provinces in the Tasman Element (O'Reilly and Zhang, 1995). Many of these volcanic provinces define what are interpreted as a series hot spot tracks that become younger to the S and track Australia's movement northward (Johnson et al., 1989). Others may be the consequence of upwelling of deep mantle (O'Reilly and Zhang, 1995).

Cenozoic basins, regolith, and paleovalleys host other resources, including Fe-ore, coal, oil shale, uranium, and groundwater. A significant proportion of the Fe-ore production from the Pilbara Craton comes from pisolitic iron that formed paleochannels, which were active at c. 30 Ma. These ores, derived originally from Paleo-proterozoic banded-iron formation and Fe ore deposits, have been dated at between 28–5 Ma (Morris and Ramanaidou, 2007).

Brown coal deposits of the La Trobe Valley in SE Victoria, which provide most of that state's electricity, are as much as 130 m thick and formed between 30–20 Ma, when the climate was humid (Fujioka and Chappell, 2010). Similarly, the oil shale resources from pullapart, lacustrine basins near Gladstone and Proserpine in Queensland also formed at this time (Henstridge and Missen, 1982).

By 25 Ma, Australia and Antarctica had separated sufficiently to allow full circulation between the Southern Ocean and the Pacific Ocean, which resulted in major changes to the climate of Australia and Antarctica (Fujioka and Chappell, 2010). The great Antarctic ice sheet began to develop c. 15–10 Ma in the E and c. 10–6 Ma in the W and fundamentally altered the Earth's marine and terrestrial climate and sea levels. The impacts on Australia were large: meridional temperature gradients steepened, the subtropical monsoon contracted northwards, and the continent became drier. The boundaries between climatic zones strengthened, which led to the increased aridification of mid-latitude continental regions in Australia, as well as in Africa and North and South America. Grasslands developed, stimulating the evolution of grazing mammals, and in Australia the rainforests retreated (White, 1994; Fujioka and Chappell, 2010).

Active convergence, which is occurring today along Australia's northern margin, commenced at c. 40 Ma. Timor is now part of the Australian Plate and is moving northward along with Australia (Keep et al., 2002). The consequences of Australia's northward movement are best seen in New Guinea, on the northern margin of the plate. Collision with proto-Indonesia, and related fragments, emplaced the Irian ophiolite and created the present mountainous topography that forms the spine of New Guinea. An active fold-thrust belt has developed in the southern part of New Guinea, with the resulting foreland basins hosting major hydrocarbon accumulations at Tangguh, and Hides (Van Ufford and Cloos, 2005). The collision and underlying structure of the Australian Plate also control the location of Irian Jaya's and Papua New Guinea's giant Au deposits such as Grasberg and Ok Tedi (Hill et al., 2002; Davies, 2012).

Convergence and major dextral strike-slip faulting commenced c. 5 Ma across the Australian Plate boundary in New Zealand. This resulted in far-field shortening in Australia and had an impact on the stress state, causing uplift in the Flinders Ranges, and contributing to Australia's present day earthquake hazard (Clark et al., 2011).

### 2.6–0 Ma: The creation of modern Australia

By 2.6 Ma, the beginning of the Quaternary Period, the current geology of Australia was mainly established. Australia had separated from Gondwana and was moving northward towards SE Asia, setting up the stress field that is prevalent today. Hence, the most significant changes observed in Quaternary Australia are not related to tectonics, but to climate and the arrival of humans. The Quaternary in Australia is characterised by repeated episodes of aridity linked to glacial-interglacial cycles. In Tasmania, the maximum ice extent occurred at c. 1 Ma, with later advances in the past 100 ka being less extensive. These glacial periods were accompanied by periods of aridity: there was an intensification of atmospheric circulation, a reduction in atmospheric moisture and an increase in continentality, thereby leading to more arid conditions (Fujioka and Chappell, 2010). These conditions resulted in deflation and the formation of vast areas of sand dunes. Today, 40% of the Australian arid zone is mantled by aeolian sand dunes and sand plains (Haberlah et al., 2010).

Lake Eyre, in South Australia, reached its maximum water levels at 130–110 ka, with particularly low levels at 95–80 ka and at 65–62 ka. The last glacial maximum saw Lake Eyre completely dry and exposed to the wind, which scoured its base and deepened the lake floor (Hesse, 2010).

Continental Australia was 10°C cooler at 22 ka than now. It was also much drier and windier and sea level was 120 m lower (Fujioka and Chappell, 2010), creating land bridges from the mainland to Papua New Guinea and to Tasmania. The climate warmed, sea levels rose, and the land bridges were swamped by 6 ka. Palynological data record the return of warmer and wetter climatic conditions, which saw the tropical rain forests expand between 8–6 ka and again at 3 ka (Fujioka and Chappell, 2010).

Aboriginal and Torres Strait Islander people have occupied Australia since at least 60–45 ka (Roberts et al., 1990; Pope and Terrell, 2008), and were present during many of the major climate changes that have affected Australia during the latter part of the Quaternary. These changes had major impacts on the Aboriginal people. Changes in climate caused episodic occupation of specific sites as consequences of availability of water and food and changes in sea level. For example, Shark Bay, 800 km N of Perth, was occupied during three distinct periods, between 30–18 ka, between 7–6 ka and since c. 2.3 ka (Bowdler, 1999). A far greater consequence of climate change was the isolation of Tasmanian Aboriginal people. Tasmania was first inhabited at c. 40 ka when a land bridge existed between it and the mainland. This land bridge was severed at c. 8 ka when sea level rose in association with interglacial warming (Pardoe et al., 1991). This isolated Tasmania until the arrival of Europeans just before the turn of the 19th century.

## The future: Toward Amasia

The next supercontinent, Amasia, is expected to be fully formed within a 250 Myr time period (Santosh et al., 2009). The Australian continent, including Papua New Guinea, is currently colliding with the Indonesian archipelago. This convergence has and will continue to produce natural resources, as is currently occurring in Papua New Guinea and Indonesia. Hydrocarbon deposits will continue to form in foreland sedimentary basins related to fold-thrust belts.

## Conclusions

The geological evolution of Australia is a fascinating microcosm of the evolution of Earth. This evolution is closely linked to the supercontinent cycle, with most geological and metallogenic events relating to the assembly and breakup of Vaalbara, Kenorland, Nuna, Rodinia and Pangea-Gondwana. In a broad sense, Australia grew from W to E, with two major Archean cratons, the Yilgarn and Pilbara cratons, forming the oldest part of the continent in the West Australian Element. The centre is dominated by the mainly Paleoproterozoic–Mesoproterozoic North and South Australian elements, whereas the E is dominated by the Phanerozoic Tasman Element. The West, North and South Australian elements initially appear to have been assembled during the Paleoproterozoic amalgamation of Nuna, although the resulting proto-Australia subsequently partially broke-up and was then reassembled during Rodinia amalgamation. The Tasman Element formed mostly during the Paleozoic when the eastern margin of Australia was an active, accretionary margin during the assembly of Gondwana, and then Pangea. Australia's present position as a relatively stable continent, with passive margins on three sides, is the result of its final break-away from Gondwana, when its link to Antarctica was fully severed at 34 Ma. It is presently moving northward toward SE Asia, probably during the earliest stages of the assembly of the next supercontinent, Amasia.

Australia's resources, both mineral and energy, are linked to its tectonic evolution and the supercontinent cycle. Although the mineral and energy resources are known throughout Australia and their formation spans its history, there is a strong clustering of resources both in time and in space and these clusters are associated with the Earth's supercontinent cycles. Australia's most important Au province, the Eastern Goldfields in Western Australia, is the product of the assembly of Kenorland, whereas major Paleoproterozic–Mesoproterozoic Zn-Pb-Ag deposits and the giant Olympic Dam iron oxide-Cu-Au deposits formed as Nuna broke up. The diverse metallogeny of the Tasman Element is a product of Pangea-Gondwana assembly and most of Australia's hydrocarbon resources are a consequence of the breakup of this supercontinent. Very young Au and Cu-Au deposits in Papua New Guinea and Indonesia to the N of Australia may be the very early products of Amasia assembly.

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by David F. Branagan

# Fleshing out the Landscape: Two centuries of Australia's geological heroes

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*Australian Geology, from its beginning, was linked to the development of resources, in a number of distinct colonies, often with little co-operation between them. While the formation of the Commonwealth in 1901 saw the beginning of an Australia-wide approach to geological studies, the former colonies, now states, still guarded certain rights, acting independently, in as far as possible. Separate colonial geological surveys continued as State bodies, focussed largely on resource (including water) studies. Geological education was mainly through a few universities, where research on broader issues was undertaken, with results published in separate, largely colony/state oriented, society journals. The Australasian Association for the Advancement of Science, formed in 1888, was a major influence in bringing geologists into direct contact, and marked the beginning of attempts to rationalise results and focus on Australia-wide aspects. Post World War 11 saw the expansion of geological research, formation of a Commonwealth Survey and the Geological Society of Australia, with the latter providing a medium for promulgation and discussion of ideas. The paper discusses the role of some 180 geologists through two centuries.*

## Introduction

Australia, the island continent, is very large (Figure 1). Its size and physical characteristics offered considerable challenges to the original Aboriginal inhabitants, and the Europeans and Asians who first visited, and then settled, in the country. The Aboriginal peoples had a considerable feeling for landscape, particularly aspects relating to ease of travel, but which also embraced spiritual concepts of the close relation of land to humanity, shown in the 'song lines' linking many parts of 'country'. Knowledge of the location, use and trade of natural materials was both extensive and detailed, numerous 'mining sites' supplying materials for axes, spear-heads, grinding stones, ornamental items and ochre. There was also teaching and identification of 'sacred sites' for initiation and other ceremonies, such sites often inscribed by 'real pictures', time markers or symbolic spiritual images. Although there is little documentation of specific individuals who preserved and passed these matters on to later generations, it is known that Aboriginal people guided European prospectors to outcropping mineralisation during the 19th century. One such European was Ernest Henry, led by Kalkadoon people to the Argylla and Mt. Oxide sites in the Cloncurry region, NW Queensland, 1880–82 (Blainey, 1960).

From European settlement (1788) until 1940, fewer than 1,000 geologists had worked in Australia. About 300 had been members of the Geological Society of London (F.G.S.) formed in 1807. Others belonged to local scientific societies, and, from the 1890s, the Australasian Institute of Mining Engineers (later The Australasian Institute of Mining and Metallurgy). An exponential increase in the number of geologists occurred about 1950, when many graduated and were employed by Commonwealth and State government surveys and research bodies. Projects such as the Snowy Mountains Hydro-Electric Scheme recruited geologists, as did mining and exploration companies. There was also demand for teachers in tertiary institutions. Geologists migrating from Europe and North America added a new mix to an environment which had traditionally been wedded to British practices. Today, approximately 4,000 geologists work in Australia.

Since the last International Geological Congress held in Australia (the 25th, 1976), there has been growing interest in the origins and progress of Australian geology, as summarised in this paper. With several exceptions, only non-living geologists are included in the account, and the story is taken only to the mid 1960s. The paper does not discuss the major theoretical problems of Australian geology, such as the Gondwana concept, ancient glaciations, desert landforms, geomorphology in general or geophysics (for which see Vallance, 1975; Branagan and Townley, 1976; Day, 1966).


***Figure 1 Comparison of Australia and Western Europe.***

# Early European Settlement and Exploration (1788–1850)

## European Geologists, Sydney 1788–1812

The first European settlement in 1788 (the Colony of New South Wales) owed its position to the splendid Sydney Harbour, just N of Cook's Botany Bay. While 'the finest harbour in the world' was a distinct advantage, the rugged basin-like geological structure surrounding Sydney restricted expansion, confining development for almost twenty years. Settlers had to fend for themselves, and while excellent sandstone, and clay (weathered shale) for bricks were readily available, lime was scarce, only sea shells – in accumulations from Aboriginal feasts – offering limited sources of this essential building material. Availability and storage of water was a continuing problem.

The British authorities, perhaps hoping the colony would quickly become self-sufficient, appointed a mineralogist, Adolarious W.H. Humphrey (1782–1829), in 1803. He worked first in Tasmania and moved to Sydney in 1805, but resigned in 1812 to become a magistrate. His contribution to geological knowledge was slight (Vallance, 1981a). In the meantime, coal, discovered in 1791 at Newcastle, 120 km N of Sydney, was already being mined by the Australian Agricultural Company.

## The French

In the early years of British settlement, the French also explored the Pacific region. Their expeditions carried *savants*, who collected minerals, rocks and fossils. The d'Entrecasteaux expedition (1792–1793) made geological observations around Tasmania, mainly though the scientist J. La Billardière (1755–1834), noting spectacular columnar dolerite coastal outcrops, limestone on Bruny Island, and describing coal seams and associated rocks at South Cape Bay, southern Tasmania. That coal (now known to be Triassic) was assumed to be the same age as European coal, later called 'Carboniferous', an assumption marking the beginning of a seventy-year long controversy about coal ages in Australia (Vallance, 1981b).

The Baudin (1754–1803) expedition (1800–1804) included geologically-trained Louis Depuch (1774–1803) and Joseph Charles Bailly (1777–1844) and the zoologist, François Péron (1775–1810). They observed aeolian calcareous coastal sandstones on the Western Australian and South Australian coasts, while Peron's geological observations on Maria Island off Tasmania were perceptive (Banks, 1990). Depuch and Bailly, the first geological visitors to Sydney, recognised the stratigraphic relations between the boldly-outcropping (Hawkesbury) sandstone and the overlying shale, and collected pebbles of granites and metamorphics (as we know them) in the Nepean River, 60 km W of Sydney, carried from the as-then uncrossed (by Europeans) Blue Mountains, suggesting that 'the mountains whence it [the river] takes its rise are themselves of a Primitive nature' (Mayer, 2005). The expedition's collection (796 specimens) was described by the German Leopold von Buch who suggested that rocks lithologically similar to those found in Europe might not necessarily be of the same geological age (Von Buch, 1814). The French-born Francis Barrallier (1773–1853), a British army officer, is now thought to have been the first European to traverse the sandstone barrier SW of Sydney, and he made important discoveries, including fossils (Mayer, 2007).

Louis de Freycinet's expedition (1817–1820) included J.R.C. Quoy (b. 1790), who described the Shark Bay (WA) reefs, and added to the understanding of Sydney region geology. The Duperrey expedition (1822–1825) included R. P. Lesson, whose observations of the igneous rocks (Prospect doleritic intrusion, 45 km W of Sydney, and the Bathurst granite, beyond the Blue Mountains), and his attempt to establish a stratigraphic succession for the region, were noteworthy.

## John Busby, the Second Colonial appointment

Despite Humphrey's relatively unsuccessful tenure, in 1823 the British Government appointed the experienced 'mineral surveyor', John Busby (1765–1857). He reported on the Newcastle coal mines, being worked by the Australian Agricultural Company (Pemberton, 1986). Busby examined iron ore sites, but they were not economic, even in a colony desperate for such material. However, he soon became involved in improving Sydney's water supply, identifying a suitable source 4 km S of Sydney, cutting an inclined tunnel through the intervening sandstone ridge, pumping the water to the southern portal and gravitating the water to the town centre. It took twelve years to complete (Thorpe, 1953).

## Observant Settlers and Visitors

Geological observations by long-term residents began in the 1820s. They included: Phillip Parker King (1791–1856), who, while circumnavigating Australia, made rock collections from many sites that were later described in London by William H. Fitton (1780–1861) (Branagan and Moore, 2008); the Reverend C.P.N Wilton (1795–1859); and the explorer-surveyor (Sir) Thomas Mitchell (1792–1855). The Reverend T.H. Scott (1783–1860) published a widely circulated paper on the geology of New South Wales and Tasmania (Scott, 1824). Another paper by Scott (1831), based on an enforced stay in Western Australia, was one of the earliest about the Swan River settlement.

Wilton, based for years at Newcastle, studied the stratigraphy, and published the first details of the 'Burning Mountain' at Wingen, initially thought to be an active volcano, but soon identified as a burning coal seam (Wilton, 1829; Mayer, 2009). In 1829, the 'Mountain' was also visited by Mitchell who had learnt the skills of military draftsmanship and landscape sketching during the Peninsular Wars. He joined the Geological Society of London and received geology tuition from Fitton. Mitchell made major contributions to Australian geology, mapping in the Sydney region (Figure 2), later in the central-west of New South Wales, and collected important vertebrate fossils from caves, described by European experts (Oldroyd, 2007). His 'eye for country' helped him build roads for the difficult terrain surrounding Sydney, establishing transport corridors that, for the most part, remain essentially unchanged.

In the 1830s the enigmatic Pole, John Lhotsky (1832), the later famous Charles Darwin (1836), Paul Strzelecki (1839) and the Reverend W.B. Clarke (1839) arrived in Australia. Lhotsky (1800–1866) was not popular, so his work, during his six-year stay was (and has been since) undervalued. Lhotsky is remembered for his journey from Sydney to the Australian Alps, early in 1834, which included important mineralogical, geological and geographical observations (Vallance, 1977). In Tasmania (1836–1838), he recorded seven litho-stratigraphic units on an unpublished map of the Port Arthur region.

Darwin's visit, to Tasmania, New South Wales and Western


***Figure 2 T.L. Mitchell's Manuscript Geology of the Sydney Region Map, c. 1830.***

Australia (Nicholas and Nicholas, 1989**),** is well known. While his observations were generally good, his interpretation of the scenery was sometimes wrong. Influenced by Charles Lyell's ideas, he thought the Blue Mountains valleys were formed by marine erosion.

In contrast with Lhotsky, Strzelecki (1797–1873), another Pole, charmed everyone. He had studied geology in France; gained geological and mining experience in the Americas, saw active volcanism in Hawaii, visited Tahiti and New Zealand, arriving in Sydney in April 1839. While controversy continues about his ascent of Australia's highest peak, Mount Kosciuzsko, his geological observations in Tasmania and eastern Australia, entered on a huge map, with numerous cross-sections, were important (Branagan, 1986). His book (Strzelecki, 1845; Paszkovski, 1997), with fossil identifications by British experts, gained considerable acclaim.

Cambridge graduate William Branwhite Clarke (1798–1878), often called the 'Father of Australian Geology', is the best known of early Australian geologists (Figure 3). Arriving in poor health he found the climate suited him. During thirty-five years in the colonies he combined his clerical duties with geology. Although most of his geological work was done in New South Wales (including now southern Queensland), he paid a short visit to Tasmania in 1855. Clarke maintained correspondence with British and continental European geologists and with other colonial workers (Moyal, 2002). He was a major influence on the early development of Australian geology, with his writings and mapping, but was perhaps too wedded to European stratigraphic concepts, tending to 'push' his observations to fit them. His *Researches in the Southern Goldfields* (1860), *The Sedimentary Formations of New South Wales* (1867), and his map of New South Wales (Figure 4) exemplify his work.

***Figure 3 The mature Reverend W.B. Clarke (c. 1860).***

Ludwig Leichhardt (1813–1848?) came to Australia in 1842 with a fine geological background, from two years of studying with the best French geologists. He thought geological events in Australia might not have the same time frame as in Europe. His reputation as an explorer and scientist suffered for years from biased, inaccurate studies, and his important geological papers (Leichhardt, 1855, 1867–68) were largely ignored. Aurousseau (1968) and Roderick (1988) provide a more accurate and balanced picture. His diaries (translated by T.A. Darragh), with their revealing notes and geological sketches, await publication.

Joseph B. Jukes (1811–1869), naturalist on *HMS Fly*, spent three years in the Australian region, meeting Strzelecki and Clarke. He published an important report on the geology of the Great Barrier Reef and summarised the known geology of the continent (Jukes, 1850).

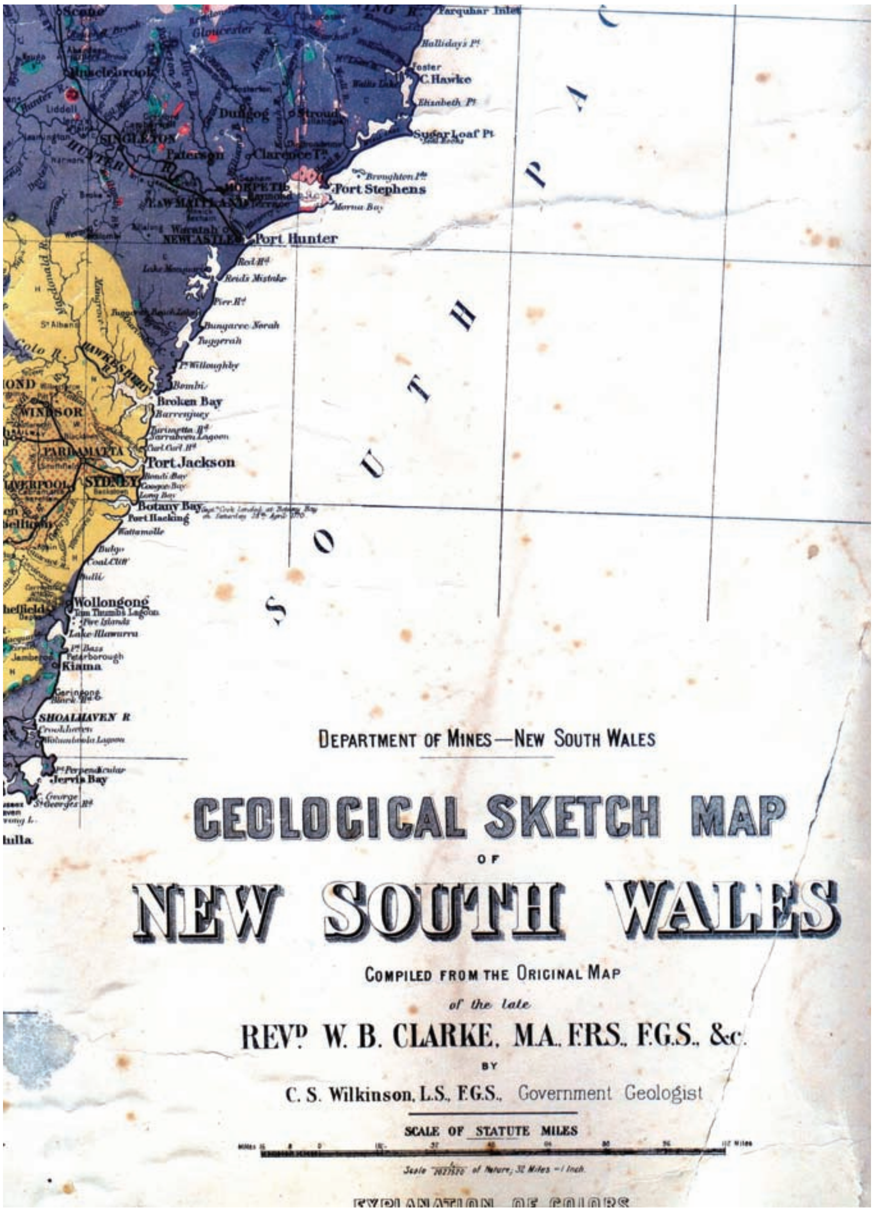


***Figure 4 Portion of W.B. Clarke's Geological Map of New South Wales, edited and published by C.S. Wilkinson, 1880.***

## Colonial Geological Surveys

European settlement began at different times in separate and independent colonies, which maintained competitive independence until the country was united in the Commonwealth of Australia in 1901. Official colonial interest in geology varied considerably over the years. It diminished when Busby retired, but mineral discoveries in several colonies, in the 1840s and early 1850s, revived Government involvement. All the Colonial (later State and Commonwealth) geological surveys established were initially linked to mining and other practical matters, including water supplies.

***Figure 5 Samuel Stutchbury, Mineral Surveyor, NSW, 1850–1855.***

In 1850, New South Wales appointed Samuel Stutchbury (1797–1856) Mineral Surveyor. Stutchbury (Figure 5) had spent three months in New South Wales in 1825–1826, and had extensive mining geology experience in Wales and Cornwall. However, the Government showed little interest in his geological observations, expecting him to be a trained prospector. Stutchbury spent four lonely years working W of Sydney and then N as far as Gladstone (in present Queensland), mapping some 80,000 km$^2$, producing informative quarterly reports, with maps and sections. He suffered from the antagonism of Clarke, who also worked for a time for the Colonial Government on the search for gold. After Stutchbury returned to England in 1855 the Government relied on coal examiners William Keene (1798–1872) and John Mackenzie for geological advice, until 1875, when a Department of Mines and a Geological Survey were established (see below).

## Alfred Selwyn and his Victorian Geological Survey

The most successful of the Colonial Surveys commenced in 1852, following the discovery and rapid exploitation of gold in Victoria, when Alfred R.C. Selwyn (1824–1902) (Figure 6), left the Geological Survey of Great Britain to take up this better paid appointment. Selwyn had gained valuable experience in Britain, mapping Lower Paleozoic rocks in Wales under Andrew Ramsay (1814–1891). He had also mapped coalfields and had an interest in glaciation. An enthusiast for mapping he began fieldwork immediately he arrived in Australia in December 1852, preparing a fine map of the Castlemaine goldfield, NW of Melbourne.

In 1854, Richard Daintree (1832–1878) and Norman Taylor (1834–1894) were appointed to assist Selwyn. Christopher D'Oyly Aplin (c. 1819–1901) joined them in 1856, when an official Geological Survey, with Selwyn as Director, was established. He adopted the British Quarter Sheet system, producing rectangular base maps, six miles by nine miles (scale 1: 31,680). Unlike Britain, there were few surveyed regions, so the geologists had to spend considerable time on topographic mapping as well as recording geological data, so progress was often slow. Nevertheless up to its closure in 1868 Selwyn's Geological Survey prepared and published 65 such maps, covering much of Victoria's mineralised region, and gaining a high reputation internationally (Darragh, 1987). In 1863, the first Geological Map of Victoria was published – the first for any Australian Colony.

***Figure 6 Alfred Selwyn, first Director of the Geological Survey of Victoria, 1852–1868.***

From 1858 Selwyn added George Ulrich (1830–1900), Charles S. Wilkinson (1843–1891), Reginald A.F. Murray (1846–1925), E.J. Dunn (1844–1937), H.Y.L. Brown (1844–1928), the paleontologist Robert Etheridge Jr. (1846–1920), and a chemical laboratory to the Survey. Important petrological work, especially on metamorphic rocks, was accomplished by A.W. Howitt (1830–1908). The Survey was closed in 1868, partly through the machinations of Robert Brough Smyth (1830–1889), Secretary of Mines from 1861 (Darragh, 2000). In consequence, Selwyn moved to Canada to direct the Geological Survey when William Logan (1798–1875) retired the following year. Of Selwyn's staff, Ulrich prospered in New Zealand, whereas the others subsequently had fine careers elsewhere in Australia (see below).

## The New South Wales Survey begins

After closure of the Victorian Survey, Wilkinson was a land surveyor in NSW, but re-established his geological career there, when a Geological Survey, under his charge, was established in 1875. It has continued to the present, and like all the Colonial (later State) Surveys, has had 'ups and downs'. Initially Wilkinson had only Charles Cullen, 'fossil collector', as assistant. They made a large collection for a Geological and Mining Museum, opened in 1876. In 1878, E.F. Pittman (1849–1932) and H.G. Lamont Young (1851–1880), and in 1879 J.E. Carne (1855–1922) joined. Young and his assistant M.K. Schneider disappeared mysteriously in 1880, on a survey of the newly discovered south coast Bermagui Goldfield, and were never found. Young's replacement was H.Y.L Brown, another from Selwyn's Survey. In 1882 Brown moved to South Australia. The same year, T.W. Edgeworth David (1858–1934), the first university graduate employed, was appointed. David's early survey work was in the New England tin fields, which yielded the acclaimed first NSW Survey Memoir. Subsequently, he published memoirs on the Newcastle–Hunter Valley coalfields and adjacent Late Paleozoic glacial beds (Branagan, 2005). Paleontological work was undertaken by Etheridge from 1887. In 1894, he moved to the Australian Museum, and W.S. Dun (1868–1934) continued the paleontology working as a joint appointee of the Museum and Sydney University. Petrological work at the NSW Survey was undertaken by G.W. Card (1865–1943) between 1893 and 1929.

Fine work was done by NSW Survey geologists, including artesian water studies by Pittman, and later by E.J. Kenny (1895–1967) and C. Mulholland (1903–1984) in western NSW. Also notable were Carne's studies (1903, 1908) on coal and kerosene shale, E.C. Andrews' (1870–1948) geomorphology (Andrews, 1910), and geology of Broken Hill (Andrews, 1922), T.L. Willan's (1895–1940) Sydney region geological map (Willan, 1923) and L.J. Jones' (1882–1951) 1930s Newcastle Coalfield mapping. The NSW Survey, after the difficult World War II years, expanded rapidly in the late 1940s under Mulholland (Johns, 1976).

## Tasmania

Official geological work began in 1859, when Charles Gould (1834–1893) was appointed. His extensive mapping, essentially alone, in the difficult western Tasmania terrain, established the island's Lower Paleozoic stratigraphy. As in the other colonies, Gould's scientific work was impeded by the official desire for prospecting, rather than thorough basic geological mapping. His contract was suspended in 1869. Following the discovery (1871) and development of large tin deposits at Mount Bischoff in the NW, and gold in the N, in 1882 Gustav Thureau (1831–1901) was appointed, initially as Inspector of Mines and later Geological Surveyor (until 1889) (McMullen 1996). However, R.M. Johnston (1843–1918), Government Statistician, rather than Thureau, prepared the classic *Geology of Tasmania*. Thureau was followed by Alexander Montgomery (1862–1933)). The Survey became firmly established under W.H. Twelvetrees (1848–1919) between 1899 and 1916, followed by local graduate C. Loftus-Hills (1885–1967) until 1923, who resigned after what he perceived as corrupt practice by his Minister. Work continued with A.H. Reid and P.B. Nye, and, briefly during World War II, with D. E. Thomas and S.W. Carey who made important advances on the geology and tectonics.

## The Victorian Survey resurrected

Murray, Dunn and F.M. Krausé (1841–1918) were re-employed by the Victorian Government in 1870, under Brough Smyth, who directed the geologists' work toward mineral search. Through contact with the other colonies, Smyth compiled the first large geological map of Australia in 1875 showing some details of the inland region (Darragh, 1977). The Survey limped along until the late 1880s when James Stirling (d. 1908), and H. Herman (1875–1962) joined, working particularly in eastern Victoria.

The appointment of J.W. Gregory (1864–1932) as Director, in 1901, saw the Victorian Survey reactivated. A new mapping pattern began. Memoirs and Bulletins were established, and a new State Map was published. However, Gregory's reign was brief, resigning in 1904, E.J. Dunn becoming Director. He opened the Wonthaggi coalfield in Gippsland, and the Geological Museum, which played a useful role until the 1960s. Hyman Herman (Director, 1912–1920), encouraged the graptolite studies of R.A. Keble and H.S. Whitelaw, but Survey activities were restricted during World War I. Brown coal open-cut mining began in the Latrobe Valley in the 1920s, the geological work being undertaken under W. Baragwanath Jr. (1878–1966) (Director 1920–1943). At that time there was interest in possible oil occurrences in East Gippsland, and small discoveries were made near Lakes Entrance.

The Victorian Survey was badly affected by the 1930s depression. There was little fieldwork, except mapping for possible hydro-electricity power stations. Subsequently, D.E. Thomas (1904–1978) (Director 1946–1967) built up a dynamic Survey with twenty staff, embracing a wide range of expertise. In the mid 1960s systematic mapping at a scale of 1:250,000 began, and the first 1:1,000,000 geological map of the State was published in 1963. In this period economic oil resources were discovered offshore, under Bass Strait.

## Queensland

Although Stutchbury and Clarke had worked in what was later the Colony of Queensland, no official survey occurred until gold was discovered at Gympie in 1866. Then separate Surveys were established in north and south Queensland in 1868, Richard Daintree being appointed Government Geologist for the North, and Christopher Aplin for the South. In 1872 Daintree went to London as Commissioner at the London Exhibition of Art and Industry, supervising an exhibition of Queensland rocks, minerals, fossils and photographs. He became Queensland Agent-General and presented a paper and the first geological map of Queensland to the Geological Society of London (1872). Daintree died in London in 1878. Aplin's appointment was short. He was followed, in 1875, by the explorer Augustus Gregory (1819–1905), who worked mainly on coal exploration, resigning in 1879.

In the north, R.L. Jack (1845–1921), previously with the Scottish Survey, replaced Daintree, in 1877, working from Townsville. When Gregory resigned, Jack became Government Geologist for the whole colony, working for the next twenty years, gaining professional acclaim, but with little thanks financially, or support with assistants. His *Handbook of Queensland Geology* (1886) included a new geological map of the colony. Jack, with Etheridge, published *The Geology and Palaeontology of Queensland and New Guinea* (two volumes with a new geological map: 1892). They had earlier (1881) published the detailed, extremely useful *Catalogue of Geological Works on the Australian Continent* which listed publications from many sources (Jack, 2008).

From late 1883 Jack was assisted by W.H. Rands (1861–1914), and, from 1887, by A. Gibb Maitland (1864–1951), who moved to Western Australia in 1896. Jack resigned in 1899 to undertake company work in China, and was succeeded by Rands, who, however, was retrenched in 1902.

Benjamin Dunstan (1864–1933), Assistant Geologist from 1897, from 1902 revived the Survey, with a staff of three. Reports and maps of mineral, gem, and coal fields continued, with a new State geological map (1903, revised 1908). His major publication was the *Queensland Mineral Index* (1913). Renewed interest in oil and natural gas exploration involved H.I. Jensen (1879–1966), formerly geologist for the Northern Territory, working in Queensland between 1917 and 1923. In 1923, Dunstan introduced aerial photography for his Survey's work, improvising equipment and photographing the newly discovered Mount Isa Pb-Zn field (NW Queensland). He also began geophysical investigations for mineral exploration. Dunstan's retirement in 1930 coincided with the Great Depression, and his successor, L.C. Ball (1877–1955), could only assist prospecting by establishing regional offices which remained maintained for many years.

Under A.K. Denmead (1955–1967), specialisation began. State-Commonwealth co-operation commenced with joint mapping for complete coverage of Queensland at 1:250,000 scale. A new state geological map, compiled in association with University of Queensland geologists, was published in 1953.

## Aerial, Geological and Geophysical Survey of Northern Australia

An important development for Australian Geology was the formation in 1935 of the Aerial, Geological and Geophysical Survey of Northern Australia (AGGSNA), with Commonwealth and State financial involvement, seeking ore deposits N of the 20$^{th}$ parallel. The brainchild of the industrialist W.H. Gepp (1877–1954), AGGSNA, with P.B. Nye (from the Tasmanian Survey) as Executive Officer, and J.M. Rayner (1906–1982) (NSW Survey) as Geophysical Consultant, was an important 'training ground' for the geologists and geophysicists who led the resurgence of geology in Australia in the post-World War II years. The earlier Imperial Geophysical Exploration Survey under Broughton Edge (1895–1953), which Gepp helped to develop in 1927, had indicated the likely value of geophysics in mineral exploration. AGGSNA was disbanded in 1940.

## Western Australia

Official geological work in Western Australia began in 1847 with Ferdinand von Sommer (c. 1800–1849), after employment by the short-lived Western Australia Mining Company. Appointed to examine supposed coal occurrences, his contract was extended and he travelled widely, preparing eight reports, some fine maps and sections in less than a year. Glover and Bevan (2010) indicate that he lacked the qualifications he claimed but, in practice, his geological work displayed considerable knowledge and competence.

The Gregory brothers were investigating the geology from 1846. Augustus C. Gregory, (mentioned earlier) is best known. He made surveys in 1846 and 1848, discovering the Irwin River coal measures, and galena on the Murchison River, later the site of the Geraldine Mine. He led an important expedition in the Northern Territory in 1855–1856, before moving to Queensland where he was appointed Surveyor-General. Francis Gregory (1821–1888), also a surveyor, undertook expeditions in 1848 and 1861, and with his brother, Joshua (1815–1850), submitted a paper and *Geological Map of Western Australia* to the Geological Society of London (Gregory, 1848).

The Reverend Charles Nicolay (1815–1897) had a museum of geology at Fremantle from 1861, and acted as geological adviser when there was no official geologist. H.Y.L. Brown, Government Geologist (1870–1872), produced ten reports and three geological maps, but was sacked when the colony got into financial difficulties. The position was re-instated in 1882 when Irishman Edward Hardman (1845–1887) examined mineral resources in the Kimberley region, finding, in 1884–1885, strong indications of gold. Prospectors, following up, discovered the Halls Creek field, and a gold rush ensued. Ironically, the politicians voted not to renew Hardman's position just before the mineral boom began.

As the northern gold development occurred the Government Geologist position was re-established in 1887, and H.P. Woodward (1858–1917), then in South Australia under Brown, was appointed, with two assistants, B.H. Woodward (1846 –1916) and S. Göczel (1856–1918). This was a period of extraordinary prospecting and mining, including the discovery of gold at Coolgardie and Kalgoorlie, (Eastern Goldfields, 600 km E of Perth), and in the Pilbara region, to the N. Such discoveries put the now self-governing colony on a firm financial footing, leading to the formation of a Department of Mines in 1894 (Spillman, 1993).

Woodward resigned in 1895, being replaced by A. Gibb Maitland. He was invited to establish a Survey with assistant geologists, assayer, laboratory staff and a museum, achieving these goals by the end of 1897. Maitland's leadership, until retirement in 1926, saw half of the state systematically mapped at appropriate scales, generally in advance of prospecting. Ninety-one Bulletins were published 'covering almost all aspects of geological investigations known to the science at that time'.

H.W. Talbot (1874–1957), originally a surveyor, was a fine field geologist, despite no formal qualifications, noted for spending more time in the field than any other, 364 days in one particular year! Maitland said Talbot 'contributed more to the originally unknown parts of Western Australia than any other geologist'.

Also notable was E.S. Simpson (1875–1939), mineralogist and chemist, who served from 1897 to 1922 and was later Government Analyst. The survey was, at that time, in advance of most other state surveys in using petrology, notably by R.A. Farquaharson (1883–1959), in the Eastern Goldfields of WA.

J.T. Jutson (1874–1959), initially with few qualifications, was another distinguished worker. His *Outline of the physiographical geology (physiography) of Western Australia*, (1914, later revised and re-issued several times), gained international regard.

Maitland's successors, Torrington Blatchford (1926–1934), Frank Forman (1934–1944), during the Depression, essentially 'marked time', the search for oil having little success. H.A. 'Matt' Ellis (1945–1961) began the revival of geological activities. He was followed by Joe Lord, who recruited geologists from overseas and formed specialist divisions, with a staff of thirty. Despite specialisation, Lord made the preparation of regional geological maps the Survey's first task, with detailed exploration and assessment of underground water. Lord's tenure coincided with the discovery of commercial oil at Barrow Island (1964), the beginning of iron ore exploitation in the Pilbara region, after lifting of the Commonwealth embargo (1960), the result of a belief that Australia had only limited iron resources, and the discovery of large Ni deposits at Kambalda in the Eastern Goldfields of WA.

## South Australia

Significant geology was first done by the knowledgeable, but eccentric, Johannes Menge (1788–1852), who arrived in 1837 under contract to the South Australian Company. His work on Kangaroo Island, and good advice about seeking underground water for the Kingston settlement was ignored. He also noted that the island's limestone contained interesting fossils. Dismissed in 1838, Menge moved to mainland South Australia, searched for signs of mineralisation, collected minerals, and predicted commercial mining, which began with the opening of the Glen Osmond lead mine near Adelaide in 1841 (O'Neil, 1982). In 1844, Menge applied, without success, to be appointed Government Geologist.

The first publication on South Australian geology, *Remarks on the Geology and Mineralogy of South Au*stralia, appeared in 1846, written by Thomas Burr, Deputy Surveyor-General from 1839, then Mineral Surveyor (1846–1847), being replaced by James Trewartha (1847–1850). Benjamin Babbage (1815–1878) was appointed, in 1851, to 'make a Geological and Mineralogical Survey' of the Colony, but apart from a report on the Adelaide water supply (1852) did little geology, moving to other technical activities. From the late 1850s, the Rev J.E. Tenison Woods (1832–1889) carried out extensive work, mainly in the SE of the colony, describing the Mount Gambier volcanic region, metamorphic rocks and correctly identifying the mode of formation of the limestone caves of Naracoorte.

The South Australian Survey, recommended by Ralph Tate (1840–1901) of Adelaide University and the colony's Royal Society, was permanently established in December 1882, with the appointment of H.Y.L. Brown. Brown's previous experience in other colonies, Canada and New Zealand, fitted him admirably for the position, which he held until retirement in 1912, continuing as a Government consultant until his death. He was assisted by H.P. Woodward (1882–1887), then worked alone until 1906, when Henry Basedow (1881–1933) joined him for five years. Brown covered an enormous region, including not only South Australia but also the Northern Territory (Figure 7). A man of few words, Brown, put most of his geological findings on maps. He never complained about working conditions, surviving long periods when his camels travelled without water, and temperatures exceeded 100°F (38°C) for weeks at a time.

Brown published his first Geological Map of South Australia in 1883 (with numerous later editions), and one of the Northern Territory


***Figure 7 Fieldwork regions covered by Henry Yorke Lyell Brown, 1870–1912.***

in 1898. He was followed by L. Keith Ward (1879–1964), a fine geologist who published many papers. He was supported by R. Lockhart Jack (1878–1964), son of R.L. Jack. Like other states, South Australia reinvigorated its Survey in the 1940s when S.B. (Ben) Dickinson became Government Geologist and Director of Mines. South Australia was probably the first State to begin formal mapping at 1:250,000 scale, a practice, instigated by Reg Sprigg (1919–1995), that quickly spread to the other states. Tom Barnes (1957–1970) followed Dickinson and encouraged diversification into the fields of geochemical exploration, seismic surveying, paleontology and engineering geology.

## Northern Territory

Harald I. Jensen was the first appointed Geologist (1911–1916) when the Commonwealth Government took over administration of the Northern Territory in 1911, W.G. Woolnough (1876–1958) joining him the same year. They made a broad regional survey, concerned with the location of mineral resources. There had been earlier geological work in the Territory, by exploring expeditions who made general observations and collected rock and mineral specimens, and particularly by the Gregory Expedition previously mentioned. The first specifically geological report was written by Ralph Tate in 1882, and there were also papers published by Tenison Woods, Etheridge and Brown. After Jensen left, geological work in the Territory became rather *ad hoc*. A. H. Ellis was official Government Geologist briefly (1926–1928), assessing potential metal mines and underground water resources. Important gold production began at Tennant Creek in 1933, an occurrence noted by Brown as early as 1895. Commonwealth geologists were responsible for uranium discoveries in the 1950s, and a Territory Survey was established in 1970.

## Commonwealth Involvement

In 1912, the Commonwealth, as guardian of what became known as Papua New Guinea, sent J.E. Carne of the NSW Geological Survey there to investigate coal deposits. He subsequently examined oil seeps, and noted copper mineralisation. Thus, realizing the potential, the same year the Commonwealth appointed Evan R. Stanley (1885–1924) as Government Geologist. Commonwealth interest in the possibility of oil occurrences began in 1915 when the English expert, Arthur Wade, was employed to assess the oil potential of Papua New Guinea, following which the Anglo-Persian Company was contracted to explore the territory. This was an unsuccessful operation, which ceased in 1929. Wade returned to Australia in 1924 to assess the possibilities within the continent.

The ‘direct line’ of a fully independent Commonwealth geological body began with the appointment of Walter G. Woolnough as Geological Adviser in 1927, assisted initially by Paul Hossfeld (1886–1967) in Canberra, and by the paleontologist, Frederick Chapman (1864–1943), and assistant, Irene Crespin (1896 –1980), in Melbourne. In the meantime Norman Fisher (1909 – 2009**)** became Government Geologist for New Guinea in 1934, joined in 1937 by Lyn Noakes (1914–1990). Then followed the formation of AGGSNNA (see above), the beginning of Commonwealth-state co-operation, although probably grudgingly on behalf of the States, who valued their autonomy.

Harold Raggatt (1900–1968) became assistant to Woolnough in 1940 and replaced him when he retired the following year. In 1942, the Commonwealth Mineral Resources Survey was formed under Raggatt, with P.B. Nye, Assistant Director; J.M. Rayner, Chief Geophysicist; and Fisher, Chief Geologist. This structure remained when the Survey became the Bureau of Mineral Resources, Geology and Geophysics (BMR) in 1946, the name changing to Australian Geological Survey Organization (AGSO) in 1992, (now Geoscience Australia). The range of work carried out by Commonwealth geologists has been considerable. Apart from those named, mention can be made of G.A.M. (Tony) Taylor (1917–1972), awarded the George Medal for his work during the Mount Lamington eruption (21 January, 1951), and the Estonian, Armin A. Öpik (1898–1983), whose paleontological work became legendary, though not everyone supported his contention that the Canberra region had suffered Late Paleozoic glaciation. Exploits of many Commonwealth geologists are discussed by Wilkinson (1996) in his history of Geoscience Australia.

There was other important Commonwealth geological activity. In 1916, the Institute of Science and Industry was established, to undertake research regarded essential for Australia’s prosperity. This included alunite availability, and gold deposits at Bendigo, studied by F.L. Stillwell (1888–1963) from 1916–1919. In the 1920s the Institute was re-organised as the Commonwealth Scientific Industry Research Organisation (CSIRO). Some of its early work was directed towards mineral beneficiation, through Stillwell. In 1935, A.B. Edwards (1909–1960) joined Stillwell, studying ore microscopy, publishing *Textures of the Ore Minerals and their Significance* (1948). Work of international importance was carried out by this group, including the pioneer study by George Baker (1908–1975) of tektites, formed during entry through the Earth’s atmosphere, which attracted the attention of NASA, when investigating satellite re-entry problems. Later geological activities within CSIRO, yet to receive historical evaluation, include coal and hard rock petrology, and marine geology within the Division of Exploration and Mining. The Baas-Becking Laboratory, supported by Commonwealth organizations and mineral industry bodies was established in the early 1950s, doing important bio-geological research.

## Academic Geologists

Australian Universities, from the foundations of Sydney (1852) and Melbourne (1853), have played a continuing role in both the education of Australian geologists and in research.

## University of Sydney

Geology became a formal subject from 1866 under Alexander M. Thomson (1841–1871). He died after a lung infection picked up from fieldwork at Wellington Caves, and was followed by Archibald Liversidge (1846–1927) (Figure 8) with expertise in chemical mineralogy. He created, in 1888, the long-lived Australasian Association for the Advancement of Science (AAAS), later renamed Australian and New Zealand Association for the Advancement of Science (ANZAAS) (MacLeod, 2009). The AAAS brought together Australasia’s geologists (and other scientists) for regular meetings, in different centres, to present their research, become personally acquainted, make friendships (or not), argue about concepts, matters previously only possible by correspondence across the vast distances separating their work places. The Association’s annual reports remain valuable sources of information, particularly the sub-committee reports on glaciation, stratigraphy and structural geology (Vallance and Branagan, 1988).

***Figure 8 Archibald Liversidge, mineralogist and chemist, founder of the Australasian Association for the Advancement of Science (1888).***

Between 1882 and 1890, William J. Stephens (1829–1890) taught regional geology and paleontology. Stephens’ sudden death resulted in the appointment, from the NSW Survey, of Edgeworth David (Figure 9), who was destined to become the doyen of Australian geology over the next forty years. His life was legendary, including Antarctic and military experience, culminating in completion of his *Geological Map of the Commonwealth* (1932) (Figure 10), accompanied by a superb summary of the geology, written in just a few weeks. David’s contribution to Australian geology, through teaching, research, public appearances and popular writing, is unequalled (Branagan, 2005).

***Figure 9 T.W. Edgeworth David, ‘doyen’ of Australian Geologists, by his friend Norman Carter (1909).***

David’s students contributed in almost every region of Australia. They included E.C. Andrews, W.G. Woolnough, Douglas Mawson (1882–1958), W.R. Browne (1884–1975), L.A. Cotton (1880–1963),

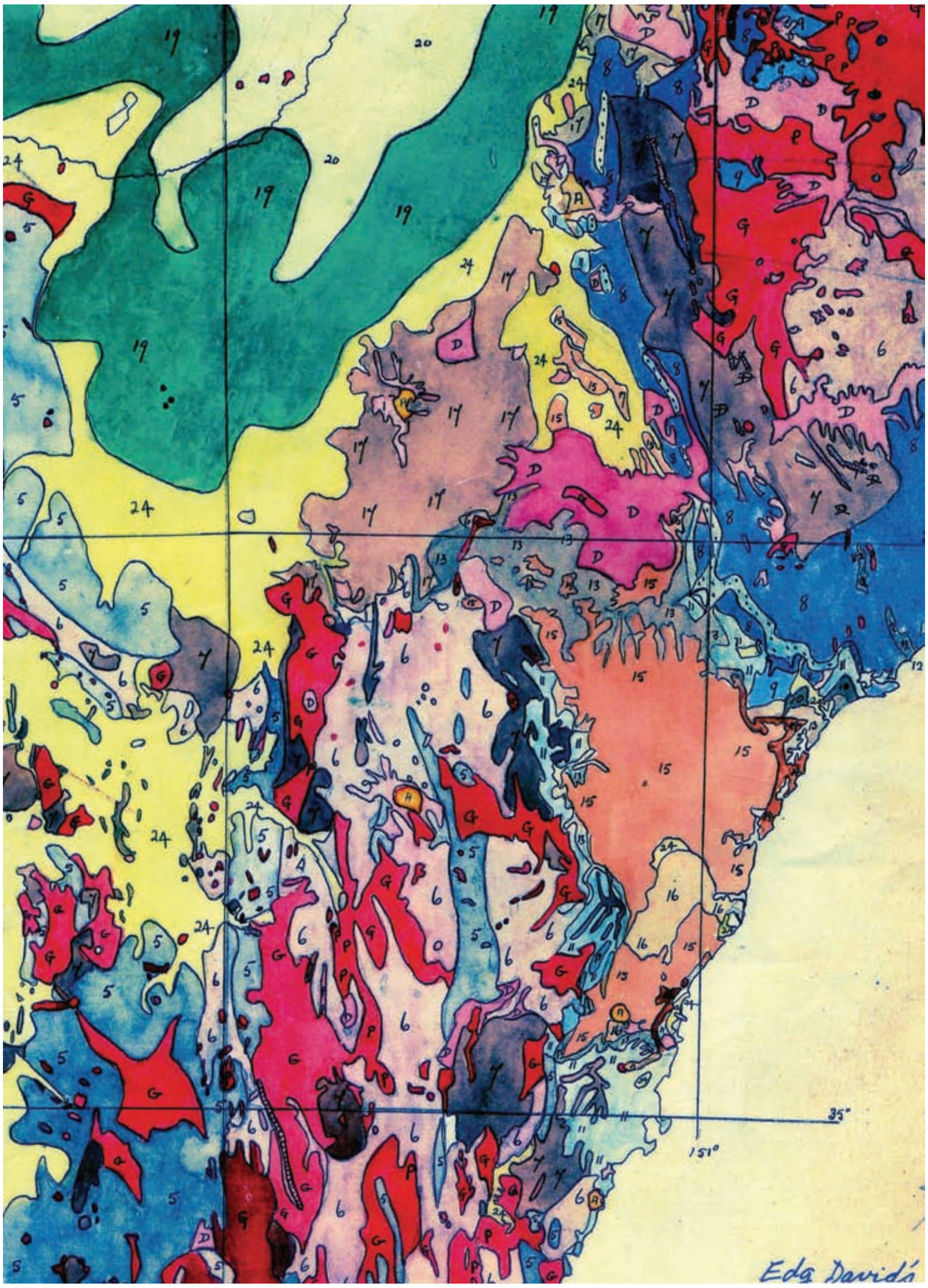


***Figure 10** Portion of Edgeworth David's draft for the Geology of the Commonwealth Map (1929).*

T. Griffith Taylor (1880–1963), Frank Debenham (1883–1965), M. Aurousseau (1891–1983), W.N. Benson (1885–1957), and H.G. Raggatt. When David retired in 1924, Cotton, a geophysicist, was appointed, remaining until his retirement in 1948. The 'David Era' ended with the appointment in 1949 of Englishman Charles E. Marshall (1908 –1992), coal petrologist-cum-engineering geologist.

## Melbourne

In 1853, Irish-born Frederick McCoy (1817–1899), became Foundation Professor of Natural Science, including mineralogy, practical geology and paleontology. McCoy was more interested in museums than teaching, building up a fine collection, and continuing paleontological research, while students complained they never had a geology field excursion. McCoy's contemporary, Ferdinand von Mueller (1825–1896) (at the Melbourne Botanical Gardens), known essentially as a botanist, made important paleobotanical studies, mainly on Tertiary flora. McCoy clashed with Selwyn, the Survey Director, over mining matters. J.W. Gregory (1864–1932) replaced McCoy. It proved a dramatic change, Gregory enthusing students to go into the field, and leading by example. Although Gregory stayed only four years he did more work than most professors undertake in ten or twenty years. He is remembered for his summer journey with students into what he called the 'Dead Heart' of Australia, a term which caught the imagination of ordinary Australians (Branagan and Lim, 1984). Not content with teaching and fieldwork, as mentioned above, Gregory accepted appointment as Director of the Geological Survey of Victoria, but moved to Glasgow in 1904 where his notable reputation was sustained (Leake, 2011).

Ernest Skeats (1875–1953) followed Gregory, building a strong a reputation for petrology, with students such as Stillwell and H.C. Richards (later the first Professor of Geology at Queensland University). Edwin Sherbon Hills (1906–1986), another of Skeats's students, replaced him in 1944, and is remembered for his classic *Outlines of Structural Geology* (translated into many languages). He became Foundation President of the Geological Society of Australia in 1952.

## Adelaide University

At its foundation (1874), among its first appointments (Natural Science) was Ralph Tate, a student of the Royal School of Mines, London. Tate quickly began researching the local geology and established the Royal Society of South Australia. In 1877, his examination of glacial evidence at Hallett Cove, near Adelaide, revived interest in this subject, evidence having been recorded in 1859 by the visiting Alfred Selwyn. Tate's geology ranged from Precambrian to Tertiary (the latter perhaps his greatest contribution to Australian geology). Tate was a 'big' man in many ways, enjoyed student company, and is said to have always travelled with a cask of beer on field trips (Alderman, 1967).

When Tate died the University did not appoint another professor, but two lecturers (probably cheaper!), Walter Howchin (1835–1937) and W.G. Woolnough. The latter returned to Sydney in 1904, and Douglas Mawson was appointed in his place (Ayres, 1999). However, Mawson's long period of Antarctic exploration and military duties saw the teaching largely left to Howchin until

1920, with some short-term appointments. As a Methodist minister, Howchin preached in country places at weekends, and, in conjunction, undertook fieldwork, continuing this practice for ten years following retirement. Howchin's achievements in South Australian geology at least equalled those of H.Y.L. Brown, and included his recognition of Cambrian (now Neoproterozoic) glaciation, which gained him international recognition (Cooper, 2000).

On Howchin's retirement, Mawson became professor and was joined by Cecil T. Madigan (1889–1947) as lecturer. Both physically big men and with strong personalities, they did not always agree, and Mawson made life difficult for his subordinate. The latter, however, gained fame in the early 1930s making the first successful crossing of the formidable Simpson Desert from Northern Territory to Queensland by camel (Twidale et al., 1990).

In 1949 the Geology School was divided, Mawson teaching General Geology, with Eric A. Rudd (1910–1998) appointed for Economic and Mining Geology (the first in Australia). Rudd raised funding through good relations with industry. The brilliant Martin Glaessner (1906–1989) revived South Australian studies in paleontology and stratigraphy when appointed in 1950, culminating in the study of the Ediacara fauna, discovered in 1947 by Reginald (Reg) Sprigg (Weidenbach, 2008). Photos of such fossils are included in Figure 7 of Gehling and Droser (2012).

## Younger Universities

The University of Tasmania (Hobart, 1893), initially had only sporadic geology instruction. Geology became established in 1946 when S. Warren Carey (1916–2004) (Figure 11) was appointed professor. Carey's international reputation came from his controversial ideas on tectonics, involving the notion of Earth expansion.

Geology at Queensland University began in 1911 (Hill, 1981) and, prior to World War II, there were four staff: H.C. Richards (1884–1947), responsible for setting up the Great Barrier Reef (Research) Committee, A.B. Walkom (paleontologist; 1913–1920), W.H. Bryan (1891–1966) from 1920, and F.W. Whitehouse (1900–1973) from 1926. All had fine reputations for research and teaching, covering a wide range of topics. The breadth of research was greatly supplemented by the appointment of Dorothy Hill at the end of the war (see below).

***Figure 11 S.W. Carey 'illustrates' his idea about Earth Expansion, at the ANZAAS Congress, Port Moresby, Papua New Guinea, 1970.***

The University of Western Australia (Perth) opened in 1913, when W.G. Woolnough was appointed and Marcel Aurousseau assisted. Edward de Courcy Clarke (1880–1956) and Rex Prider followed (Glover, 2003; Glover and Bevan, 2010).

New England College of Sydney University, (Armidale, NSW), established 1938, became the independent University of New England in 1954. A.H. Voisey, initially the sole lecturer, continued as Foundation Professor. He was an inspirational teacher, enthusing students to undertake detailed studies of northeastern New South Wales, a theme continued by his successors. He took up the foundation chair in Geology when Macquarie University was opened in 1967, and built up a fine school, including geology, geomorphology and geophysics.

The Australian National University was founded as a research body in 1946, the Research School of Earth Sciences becoming outstanding, with researchers such as Ted Ringwood. Canberra University College, from 1930 an adjunct of the University of Melbourne, introduced science studies in 1958. Strong teaching and research began when undergraduate studies were introduced in the amalgamated National University in 1960 under Foundation Professor D.A. Brown (d. 2010) (Rickard, 2010).

From its beginning the University of New South Wales, initially the NSW University of Technology (1949), was linked to mining education. Early teachers/researchers were L.J. Lawrence from 1951, F.C. Loughnan from 1954, and C.T. McElroy (stratigraphy), all ex-service class-mate graduates from Sydney University. McElroy became Director of the NSW Geological Survey in 1967. Another early teacher was the geochemist, Leo Koch (Lawrence et al., 2003).

Monash University (Melbourne) founded 1958, quickly built up a fine reputation for teaching and research**.** Geology at Newcastle University was 'inherited' from the 1890s School of Mines, then as a college of the then University of Technology, 1952, becoming independent in 1960 (Nashar, 1977).

# The 1920s: Women Geologists become visible

Despite women in Australia achieving the right to vote in 1902, the geological profession remained essentially closed to them for many reasons. As late as 1976, some mining companies refused to allow underground visits by women.

An early would-be geologist was the eccentric, long-lived amateur, Georgina King (1845–1932). She attracted no followers. Whilst Edgeworth David, at Sydney University, from the 1890s, encouraged women students, he was opposed by administrators when making appointments, even at junior teaching levels. One was Fanny Cohen (1887–1975) who published on Broken Hill azurite crystals. She later became a noted headmistress encouraging several generations of students into science. Marie Bentivoglio, scholarship winner (1918), studied mineralogy, and then moved to geographical problems in the newly created Department of Geography under T. Griffith Taylor, where she was, in 1929, acting Professor, before becoming a crystallographer in North America. Also at Sydney University, Ida Brown (1922–1950), a petrologist, worked on the far south coast of New South Wales (1928–1933) before switching to paleontology in 1934, after the death of W.S. Dun. Her fossil studies achieved widespread acknowledgement. Petrologist Germaine Joplin studied contact alteration zones and published well-received textbooks. Despite nominations, she was never elected to the Australian Academy of Sciences. Other Sydney women from the period included Florrie Quodling, crystallographer; Alma Culey, sedimentologist (Branagan and Holland, 1985). Maren Krysko (d. 2010) was a respected teacher at the University of New South Wales.

From Melbourne University came a plethora of paleontologists: Kate Sherrard (1898–1975), graptolite studies, assistant lecturer (early 1920s) and left-wing activist; Isabel Cookson (1893–1973), paleobotanist from 1929, studied Lower Paleozoic vascular land plants and Tertiary fossil plants, and applied paleobotany to oil search. Irene Crespin (1896–1980) (Figure 12), influenced by both Charles Fenner and Frederick Chapman (Commonwealth Paleontologist), worked in the Geological Survey of Victoria, then as Chapman's assistant, succeeding him in 1936, but at half his salary (!). She subsequently moved to Canberra where she worked with W.G. Woolnough (Crispen, 1972). Adelaide University graduate, Nell H. Ludbrook (1907–1995), undertook post-graduate studies in England, returning in 1935 to Canberra, as Assistant to Woolnough and later was with the South Australian Survey (1952–1967). She showed that paleontology could be useful to the resources industry and the *P. Ludbrookiae* zone in the Eromanga and Surat basins honours her work (see Cook, 2012). Mary Wade (1928–2005) contributed to the understanding of the Ediacaran fauna (Turner, 2007).

***Figure 12 Irene Crespin, Commonwealth paleontologist, worked in Melbourne, then Canberra.***

Best remembered of the women geologists, is Dorothy Hill (1907–1997) (Figure 13) of Queensland University, who after post-graduate studies in Cambridge (1930–1938), worked at CSIR to 1943, then at Queensland University where she was later Head of School. Hill gained an FRS mainly for her work on Paleozoic corals, was elected to the Australian Academy of Science, of which she was President in 1970 (Campbell and Jell, 1998).

***Figure 13 Dorothy Hill, FRS, Paleontologist, President of the Australian Academy of Science and the Geological Society of Australia.***

Apart from the South Australian Geological Survey, which employed Maud McBriar in 1948, E.N. Dolling (1950–1952) and Nell Ludbrook from 1952, not until the 1960s did the other State surveys employ female geologists. Final acceptance can be noted by the then relatively early elections of Nell Ludbrook (1968–1969) and Dorothy Hill (1973–1975) to the presidency of the Geological Society of Australia.

## Schools of Mines, Museums and Geological Societies

Technical Colleges and Schools of Mines have contributed to geology, largely in teaching, often being influenced by specific local mining activity. Museums have collected and archived important collections, perhaps most notably in paleontology and mineralogy.

### Schools of Mines

Derived from the various Mechanics Schools of Arts of the 1830s, Schools of Mines and Technical Colleges provided good geological training for potential mine managers and technicians but there was little opportunity for research. Initially located in mining towns, notably in Victoria, later there were important technical colleges in some cities, e.g., Sydney University established a School of Mines in 1892 through the joint efforts of Edgeworth David and Archibald Liversidge (Branagan and Holland, 1985).

The Ballarat School of Mines was established in 1870, under Ferdinand M. Krause (Perry 1984). Others included Bendigo (1873) (Cusack, 1973), Castlemaine, where Thomas Sergeant Hall (1858–1915) made outstanding studies of Ordovician graptolites, including suggesting international correlations. Melbourne Working Men's College included Hall's research colleague G.B. Pritchard (1869–1956), who had 33 hours of lectures per week, plus weekend excursions! In other colonies were Kalgoorlie (Western Australia), Adelaide (1892), Zeehan (Tasmania) (1892), Charters Towers and Gympie (Queensland, from 1901), while technical colleges in New South Wales (from the 1880s) included teachers/researchers such as C.A. Süssmilch (1875–1946) (Newcastle) and J. Milne Curran (1859–1928) (Sydney).

### Museums

Over the years Museums have made important research contributions, perhaps most notably in paleontology, as told in various histories (e.g., Strahan, 1979). Notable staff include the paleontologists, Felix Ratte (d. 1890), Robert Etheridge, A.B. Walkom and Harold Fletcher (1903–1996), the mineralogists, Charles Anderson (1876–1944) and Oliver Chalmers, at the Australian Museum, Sydney; F. Chapman, E.D. Gill and the petrologist, D.J. Mahony, at the National Museum, Melbourne; Charles De Vis (1829–1915) and Heber Longman at the Queensland Museum. Western Australia Museum staff made significant meteorite studies. The South Australian Museum has important mineral collections, while Robert Bedford (1874–1951) at his small museum at isolated Kyancutta, South Australia, did important research on Archaeo-cyathidae. The Tasmanian Museum has an extensive Antarctic collection. Other smaller museums (e.g., Canowindra – fossil fishes) contain important historic and scientific collections.

### The Societies

Learned societies have facilitated discussion amongst geologists and ideas to be debated, as seen above, but space does not allow discussion of the various societies devoted to the geological sciences. For histories of the Geological Society of Australia see Cooper and Branagan (1994); for the Australasian Institute of Mining and Metallurgy (Dew, 1993). Other newer bodies include The Australian Institute of Geoscientists, Australian Society of Exploration Geophysicists, Australian Petroleum Exploration Association and Australian Geomechanics Society.

## Company Geologists and Research

While the history of Australian mining has an increasing literature (Blainey, 1963; Raggatt, 1968; Trengrove, 1976 and the *Journal of the Australasian Mining History Association*), there are few detailed studies of the geologists who have contributed to the search and development of the major ore bodies, except for Glasson and Rattigan (1990). Records of this work are in the papers of various mining companies, and are often difficult to access. Haddon King (1989) illustrates this problem, as of 46 references to his own work 36 are 'unpublished' company records. This field of research offers many rewards, but can also be very frustrating.

In the field of Engineering Geology, for which Australia has a proud record, little history has been recorded. The establishment of the Snowy Mountains Scheme, diverting water from the SE-flowing Snowy River to the westerly flowing tributaries of the Murray River in the late 1940s (Wilkinson, 1996) saw the beginning of modern engineering geology work in Australia, with pioneering conceptual work directed by Dan Moye (1920 – 1975), then being taken up by other later engineering projects in Australia and overseas by Australian engineering geologists.

## Acknowledgements

Particular thanks are due to David Oldroyd and Barry Cooper who pointed out initial inaccuracies and omissions and suggested significant improvements to this paper. Keith Johns, responsible for the idea and then completion, against the odds, of his important compilation of the history of Australian Geological Surveys (1976), was an inspirational source. Keith Scott's advice also greatly improved the paper.

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# New Zealand Geology

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*New Zealand is renowned for its diverse geology and dynamic tectonic setting astride an active segment of the boundary between the Pacific and Australian plates. New Zealand is an emergent fraction (5%) of the largely submerged 'continent' of Zealandia which is half the size of Australia. Zealandia is comprised mainly of continental crust but because it is less than 30 km thick, it is largely below sea level. Zealandia's origins relate to eastern Gondwanaland from which it rifted during the Late Cretaceous to early Cenozoic, with formation of the Tasman Sea floor. Continental Zealandia may be thought of as part of the Australian/Gondwanaland mineral estate, and it is rich in natural resources. However, it was stretched and thinned for 100 Myr, culminating in the Eocene with development of the modern plate boundary. New Zealand largely owes its emergence to plate collision processes within the past 25 Myr.*

## Introduction

This general account of New Zealand geology emphasises only a few aspects of broad current research interest: mineral and petroleum prospectivity, the Alpine Fault and current tectonic activity including rifting, subduction and the Christchurch earthquakes.

Many of the concepts, facts and images presented herein are drawn from a much broader, more detailed reference work on New Zealand geology (Graham, 2008). Concerns and issues relating to natural resources (coal, oil, natural gas, water, and minerals), natural hazards (earthquakes, volcanic eruptions, landslides, tsunami, storms) and natural systems (climate, environment, biosphere, society) dictate direction of research and development in New Zealand earth sciences.

New Zealand is recognised as a small emergent part of a largely submerged 'seventh continent' – Zealandia (Figure 1), that is submergent because of tectonic subsidence caused by Cretaceous-Paleogene rifting, not global rise of sea level. Paleozoic and Mesozoic rocks of New Zealand were forged by inter-plate processes on the margin of Gondwanaland, during the Cambrian to Cretaceous (510–110 Ma).

From Cretaceous–Eocene time (110–50 Ma), extensional tectonics separated Zealandia from Gondwanaland attended by widespread subsidence creating the Tasman Sea and South Pacific oceans. Eocene–Oligocene (50–25 Ma) resumption of subduction and plate collision N of Zealandia developed a new extensional plate boundary configuration through southern Zealandia. During maximum submergence of Zealandia at c. 23 Ma (earliest Miocene) there may not have been any land in the area now occupied by New Zealand (Campbell and Hutching, 2007; Landis et al., 2008).

Since the Miocene New Zealand has been pushed up and Zealandia divided into northern and southern crustal plates so creating the Alpine Fault and subduction zones beneath Fiordland and eastern North Island.

## Extent of New Zealand

The 'Extended Continental Shelf' (ECS) of New Zealand lies within the submarine boundary, recognised by the United Nations, that delineates sovereignty over the seabed spanning from 159°E–166°W, and 23–58°S (Figure 1). The offshore area of New Zealand is c. 24 times its land area.

The land area of 267,707 km$^2$, is about the same size as the United Kingdom or Japan. The highest point above sea level is 3,754 m (Mount Cook/Aoraki) and the lowest point (onshore) is 462 m below sea level (bottom of Lake Hauroko).

New Zealand consists of the North and South islands that are separated by the 20 km wide (minimum) Cook Strait, Stewart Island and subantarctic Auckland, Campbell, Snares, Antipodes and Bounty islands to the S, the Chatham Islands to the E, the Kermadec Islands to the N, and a host of tiny islands proximal to the two main islands. New Zealand is more than 1,600 km long and up to 450 km wide, with a coastline of more than 18,000 km. About 75% of the land is over 200 m above sea level. In the South Island 223 named peaks are more than 2,300 m above sea level.

## New Zealand on a plate

Most of New Zealand lies NE–SW, reflecting the principal 'grain' of the country, parallel to and straddling the active segment of the Pacific–Australian plate boundary. The plate boundary (Figure 2) runs down the eastern side of the North Island, some tens of kilometres offshore, and is defined by the Tonga-Kermadec Trench and Hikurangi Trough. The North Island is on the Australian Plate.

The boundary swings around the SE end of the North Island and cuts through the northern half of the South Island along the Hope Fault (Rattenbury et al., 2006). The Wairau, Awatere, Clarence faults are active sub-parallel faults in this Marlborough region but the Hope Fault is most active and the best proxy for the boundary. The Hope Fault runs along the southern margin of the Seaward Kaikoura Mountains inland towards Hanmer Springs, to cross the Southern Alps where it joins the Alpine Fault near Inchbonnie on the West Coast of the South Island. The Alpine Fault can be traced on land from the entrance of Milford Sound at its southern end, to Tophouse near Lake Rotoiti. It continues to the Cook Strait coast as the Wairau Fault. South of Milford Sound, it skirts around

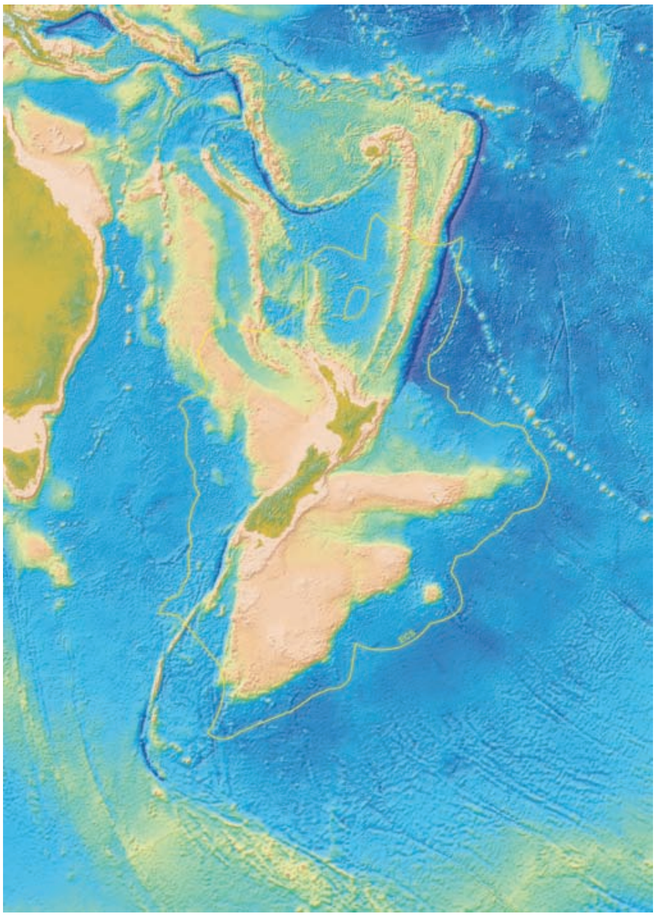

*Figure 1  Zealandia in relation to Australia, New Zealand, New Caledonia, Fiji and the SW Pacific Ocean. This is a gravity–derived bathymetric map. Shading depicts water depth; the darker the colour, the deeper the water. The 2,500 m isobath, located at c. the boundary between green and blue, is taken as a proxy for delimiting the location and extent of the largely submerged Zealandia continent. It is the assumed boundary between oceanic crust and continental crust.*

Fiordland some tens of kilometres offshore, and connects with the Puysegur Trench.

The South Island, S of the Hope Fault and E of the Alpine Fault, is the deforming edge of the Pacific Plate. The Marlborough and Nelson regions and much of the West Coast are on the opposing boundary of the Australian Plate. This plate boundary zone running through New Zealand involves highly-oblique right-lateral collision that varies along the zone causing differences in geology and topography.

In eastern and southern North Island, and NE South Island, continental crust on the Australian Plate is in collision with oceanic crust on the Pacific Plate. The resultant effects relate to normal subduction processes, including ‘Pacific Ring of Fire’ volcanism in the North Island. The Pacific Plate is descending beneath the Australian Plate from E–W at 4–6 cm/year. Subduction-related seismicity can be recognised as far S as Amberley, c. 30 km N of Christchurch.

In general, things are different in the remainder of the South Island from what is happening further N. The collision involves continental crust on the Australian Plate and continental crust on the Pacific Plate.

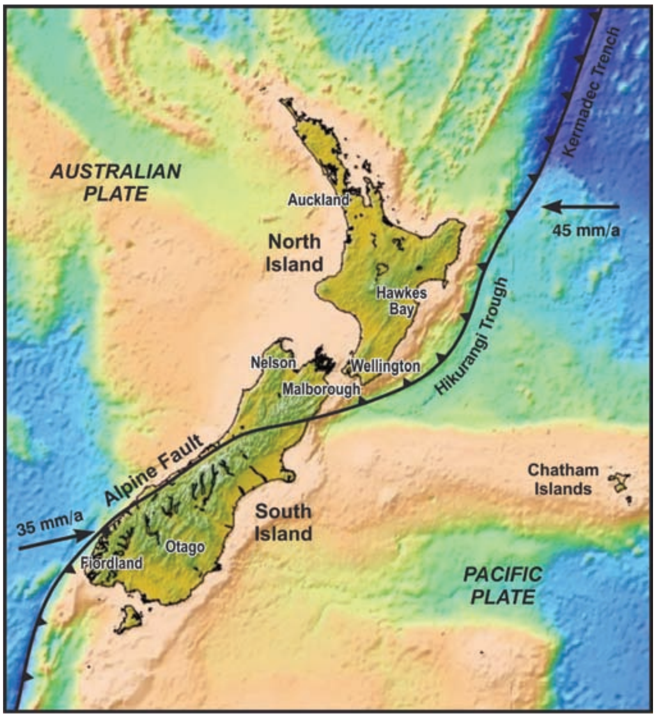


*Figure 2  New Zealand and the modern day plate boundary showing subduction zones and average rates of plate motion.*

The Southern Alps are the result, analogous to the European Alps or the Himalayas.

In the very SW of the South Island, continental crust on the Pacific Plate is in collision with oceanic crust on the Australian Plate. This is the reverse of the situation in the North Island. The Australian Plate is approaching the Pacific Plate from SW to NE at 3–4 cm /year and is obliquely twisted beneath Fiordland. It was motion on this subduction thrust that caused the Mw 7.8 Dusky Sound earthquake of July 2009 – the largest event in New Zealand since the Mw 7.8 Napier earthquake of February 1931 (Beavan et al., 2010a). The nearest volcano to southern South Island is Solander Island, 50 km SW of Te Waewae Bay. Solander last erupted more than 200 kya (Mortimer et al., 2008) and would have been as big as Mount Ruapehu in the North Island.

## The shape of New Zealand

Northland projects to the NW, Mount Taranaki creates a very pronounced boss to the shape of central western North Island, and Cook Strait breaks the country in two. Mount Taranaki is an active, 2,518 m high, subduction-related, stratovolcano that lies, with its predecessors, well W of the active arc, which is characterised by the active White Island, Tongariro and Ruapehu volcanoes. Mount Taranaki may be an expression of residual or remnant volcanism that best relates to a previous orientation of the modern volcanic arc. Eastward roll-back of the arc has occurred within the past few Myr, producing a double or coupled arc; an older arc to the W, the modern arc to the E (Figures 1 and 2). The most recent significant eruption of Mount Taranaki was in 1755. Based on its interpreted history, it erupts every 100–300 years and has done so for at least 120 kyr.

The orientation of Northland relates to active continental rifting (or back-arc rifting) of the Taupo Volcanic Zone (TVZ). If Ruapehu is the fulcrum, the ‘V’ described by Northland and East Cape (eastern

North Island) is a gape resulting from rotation (Figure 2). The rate of E-W rifting of the TVZ at Rotorua is 8–10 mm/year. This motion is being accommodated by active normal faults in the zone, such as the Edgecumbe Fault which last ruptured in 1987.

The Cook Strait region includes the Marlborough Sounds in the NE of the South Island and the South Taranaki Bight to the N. It may be thought of as an artifact of plate collision. In crustal terms, it is a low region developed within continental crust on the Australian Plate that is being drawn down tectonically, as a function of the descending Pacific Plate beneath. The Cook Strait region is where the mechanics of plate collision switches from continent–ocean collision to continent–continent collision. In a sense, the tectonic forces in this region are too weak to raise the crust above sea level. This means that the Marlborough Sounds are a drowned landscape, not just because sea level has risen since the Last Glacial Maximum, 24–18 ka, but because the crust is sinking.

With the advent of lasers, satellites, GPS etc., the manner and rate of deformation of New Zealand is readily determined, and it is interesting to speculate on what New Zealand will look like in the future (Figure 4). New Zealand is subject to 16,000–18,000 recorded earthquakes/year. On average there are 3–6 earthquakes/year that are greater than magnitude 6, a magnitude 7 or greater earthquake occurs about every decade, and a magnitude 8 or greater earthquake occurs about every century.

## Zealandia (83–23 Ma)

New Zealand has an extensive continental shelf referred to historically as the 'New Zealand Submarine Platform' or 'Tasmantis'. With the advent of bathymetric mapping based on satellite-captured gravity data in the 1990s, and other geophysical data, it was realised that New Zealand is part of an extensive area of continental crust (Figure 1) referred to as 'Zealandia' (Luyendyk, 1995) and equal to India or more than half of Australia.

Zealandia rifted away from eastern Gondwanaland with formation and growth of oceanic crust in the Tasman Sea. The oldest basalt on the Tasman Sea floor is c. 83 Ma, and the youngest is c. 53 Ma. These ages, based on paleomagnetic signatures, imply that spreading in the Tasman Sea ceased at least 50 Ma, and that the Tasman Sea floor formed over 30 Myr during Late Cretaceous–Early Eocene.


***Figure 3 Zealandia reassembled: reconstruction of eastern Gondwanaland showing Zealandia, New Zealand and New Caledonia, at 90 Ma. Rifting had commenced at 125 Ma but clean separation was not accomplished until c. 83 Ma. The Tasman Sea spreading ridge formed along the split.***

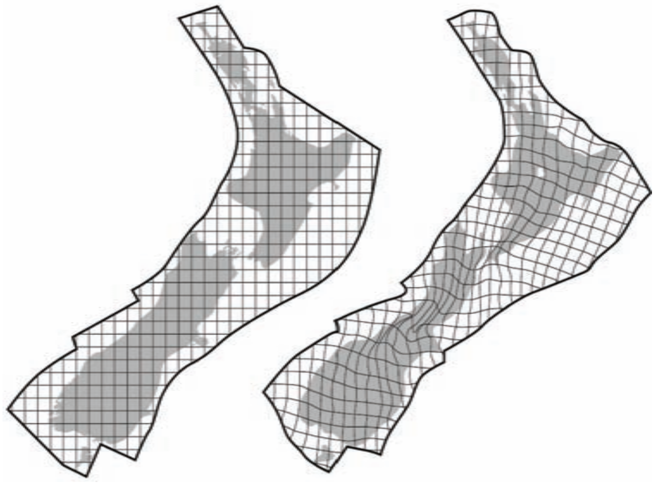

***Figure 4 The deformation of New Zealand as it might appear in 4 million years' time.***

As Zealandia rifted NE away from Gondwanaland, it was stretched and cooled, slowly losing buoyancy and subsiding below sea level. Much of Zealandia is 1,000–2,000 m below sea level. The 2,500 m isobath is a rough proxy for the boundary between continental crust and oceanic crust. The crust of Zealandia is mostly only 10–25 km thick.

Zealandia includes the Campbell Plateau S of New Zealand, the Challenger Plateau to the W, the Chatham Rise to the E, and the Lord Howe Rise, Norfolk Ridge and Three Kings Ridge to the N. The largest emergent areas are New Zealand and New Caledonia.

Rocks attributed to New Zealand's Zealandian geological history prior to the impact of compressional plate boundary tectonism, are the mainly sedimentary formations, with minor volcanics, of Late Cretaceous–earliest Miocene (83–23 Ma) (Figure 5). They represent a transgressive marine sequence relating to the slow sinking of Zealandia and increasingly oceanic conditions. The older Zealandian rocks include widespread Late Cretaceous–Eocene coal measures, whereas the younger Zealandian rocks are latest Oligocene–earliest Miocene, predominantly limestone and greensand.

## New Zealand (23–0 Ma)

A profound tectonic change occurred in the earliest Miocene, although the beginnings may be traced to the Eocene in the far N and S of New Zealand. Movement at the modern-day active plate boundary was very slow during Eocene and Oligocene time in New Zealand, but rapidly accelerated at the beginning of the Miocene (Cande and Stock, 2004).

The Fiordland subduction zone was initially formed in Early Miocene time (Sutherland et al., 2010). However, continental crust of the Australian plate was too buoyant to be subducted, resulting in progressive formation and lengthening of the Alpine Fault along the Eocene continent–ocean transition as oceanic crust was subducted at the Puysegur Trench but continental crust was displaced sideways (Sutherland et al., 2000). An estimated 800+ km of plate motion and 60 km of shortening has taken place since earliest Miocene across the South Island (Sutherland, 1999; Cande and Stock, 2004), resulting

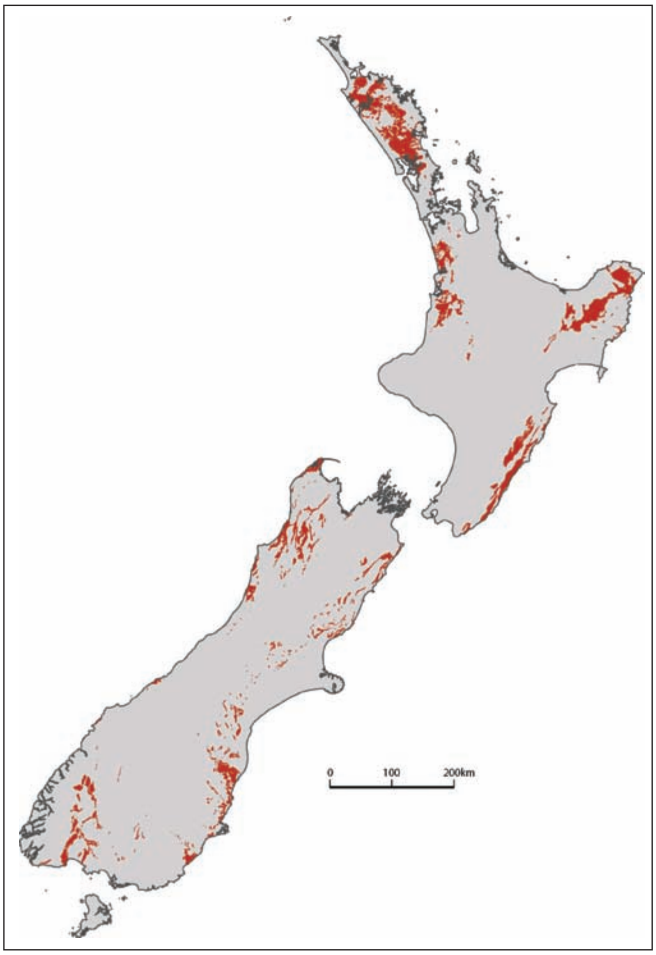


***Figure 5  Distribution of exposed rocks relating to Zealandia, 83–23 Ma.***

in crustal thickening and dramatic mountain building leading to rapid erosion.

This Miocene–Recent crustal collision is responsible for New Zealand's oil, gas and water reserves. The flood of clastic sediment derived from uplift has produced a substantive and widespread regressive sequence. Offshore, this sediment has infilled at least 20 sedimentary basins within New Zealand's EEZ. It buried organic-rich Late Cretaceous–earliest Miocene source rocks so that much of New Zealand's oil and gas is derived from the Cretaceous–Eocene coal. As yet, only the Taranaki Basin, has been explored in detail, and it is the only one being exploited.

Subduction-related volcanism has provided airborne nutrients to improve the fertility of the land for agriculture and forestry. The New Zealand Earthquake Commission (EQC), and the GeoNet Project, the natural hazard surveillance arm of GNS Science, New Zealand communicate hazard mitigation incentives and GeoNet provides information about processes below the surface. GeoNet monitors all active volcanoes in New Zealand, records all earthquakes in real time, and monitors tsunami (see www.geonet.org.nz).

As subduction rolled back to the E during Miocene–Recent, successive volcanic arcs in northern New Zealand have been subject to epithermal mineralisation creating a source of Au and Ag, especially in the Hauraki Goldfields, Coromandel Peninsula. Erosion of rocks derived from late Pleistocene eruptions of Mount Taranaki has produced extensive ironsand (dominated by titanomagnetite) deposits along the W coast of central North Island.

Subduction- related volcanism in the Taupo Volcanic Zone (TVZ) is superimposed on a continental rift (or backarc basin) which includes a string of at least eight rhyolite dominated Pleistocene super-volcanoes (Leonard et al., 2011). The youngest is the Taupo caldera, which last erupted in 233 AD (Taupo Eruption). However, the most recent significant, but relatively minor eruption, the Tarawera Eruption of 10 June 1886, was from the Okataina caldera, near Rotorua, and involved basaltic magma, not rhyolite.

These eruptions can be voluminous and are responsible for substantial blanketing of the central North Island landscape by pyroclastic flows (ignimbrites), pumice and ash deposits. More distal products can be found thousands of kilometres away. These TVZ rhyolite volcanoes are also the source of New Zealand's geothermal energy resources. Recent volcanic eruptions in New Zealand relate to subduction-related andesitic volcanism. The most recent, just 7 minutes long, was from the Crater Lake on Mount Ruapehu on 25 September 2007. New Zealand's most active volcano, White Island, last erupted in July 2000 (Leonard et al., 2011)

Rifting of continental crust in the TVZ (Figure 6) involves stretching of the crust which has made its mark on the edges of the rift zone, including Auckland. The basaltic volcanic field beneath Auckland, may well relate to this process. Some 53 volcanic cones and maars dotting the Auckland landscape are mostly monogenetic, and apparently all erupted within the past 250 kyr, most recently at Rangitoto c. 600 years ago (Edbrooke, 2001).

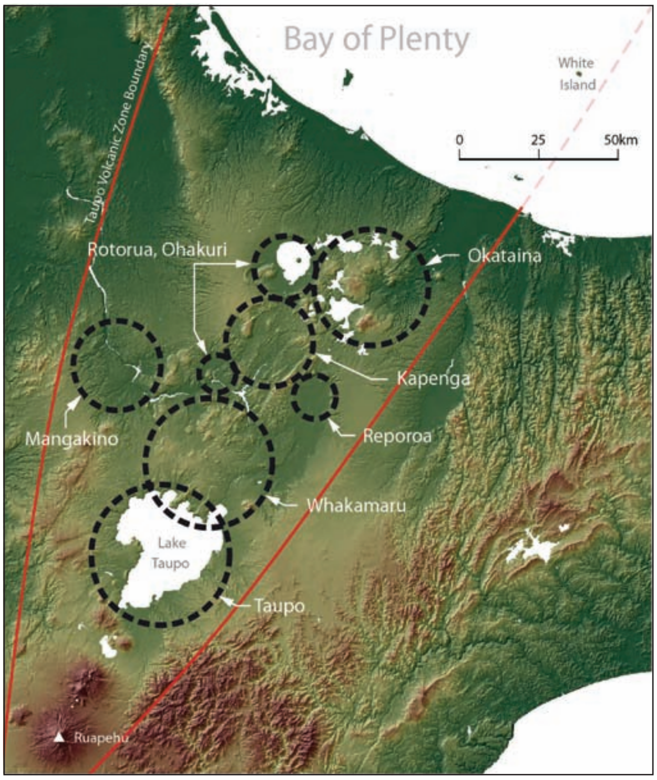


***Figure 6  The Taupo Volcanic Zone (TVZ; defined by the red line) located SW of the Bay of Plenty, central North Island, is an active back-arc rift zone within continental crust (Australian Plate). The TVZ is one of the most geothermally active zones in the world. At least eight rhyolite calderas (named) are recognised. Andesite volcanoes (e.g., Ruapehu, White Island) associated with westward subduction of the Pacific Plate, are superimposed on the rift.***

## Gondwanaland (510–83 Ma)

Uplift of New Zealand and subsequent erosion has revealed the basement geology, especially in the South Island (Figure 7) where Western and Eastern provinces are separated by a long-lived magmatic complex, the Median Batholith (Figure 8). Granites of the latter relate to the Cretaceous rifting of Zealandia from Gondwanaland.

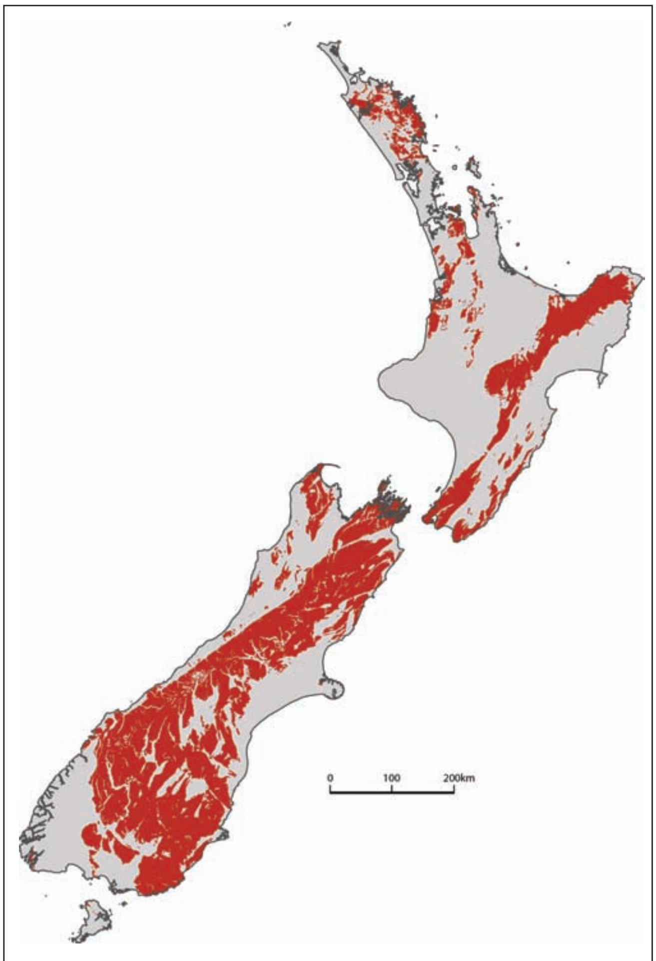


***Figure 7 Distribution of exposed 'basement' rocks inherited from Gondwanaland (red), 510–83 Ma, and all younger rocks and sediments relating to Zealandia and New Zealand (grey).***

The two provinces are dominated by weakly to moderately metamorphosed sedimentary (dominantly marine) and Paleozoic–Mesozoic volcanics. Both provinces are dominated by subduction-related volcanic arc suites, ocean floor associations and accretionary sedimentary wedges. These are organised into elongate belts that have been mapped as tectonostratigraphic units (terranes), generally older to the W, younger to the E. This age progression is compatible with ocean ward growth of the eastern Gondwanaland margin with respect to the western margin of the Panthalassa (or Paleo-Pacific) Ocean.

The Western Province comprises early Paleozoic sedimentary and volcanic rocks and is divided into the Buller and Takaka terranes. Buller Terrane sedimentary rocks are intruded by granite whereas the oldest rocks in New Zealand, early Middle Cambrian (c. 510 Ma), are in the Takaka Terrane being found in the Cobb Valley of NW Nelson (Rattenbury et al., 1998). This terrane also contains an extensive Ordovician record with much poorer Silurian and Devonian records.

The Eastern Province involves the Brook Street, Murihiku, Dun Mountain–Maitai, Caples, Waipapa, Rakaia, Kaweka, Pahau and Waioeka terranes, which range from Carboniferous–Early Cretaceous. The predominant rock type within the Eastern Province is quartzofeldspathic greywacke, largely derived from granitic rocks (Adams et al., 2007, 2011). More than 60% of the New Zealand landmass is greywacke. However, other rock types are present in Eastern Province, including subduction-related volcanogenic sediments (Brook Street, Murihiku, Dun Mountain-Maitai and Caples terranes) and the distinctive Permian Dun Mountain Ophiolite (Rattenbury et al., 1998).

In early 2012, GNS Science completed all 21 geological sheets (1:250,000 scale) within the GIS-based QMap (quarter million) project. This 15-year exercise has generated considerable new geological knowledge and insight, especially in the more remote and mountainous terrains of the Southern Alps, Fiordland, Stewart Island, Northland, central North Island and the Urewera region of eastern North Island (Isaac, 1996; Edbrooke and Brook, 2009; Cox and Barrell, 2007; Turnbull et al., 2010; Leonard et al., 2011; Lee et al., 2011).

Based mainly on detrital zircon age studies the provenance of the greywacke is most probably in the granitic terrains of NE Australia i.e., the E Queensland and NE New South Wales sectors of eastern Gondwanaland.

Much of New Zealand's basement rocks started off as sediment derived from the erosion of mountainous terrain on an active subducting margin of eastern Gondwanaland. This accumulated on the ocean floor and was bulldozed back on to the Gondwanaland margin by accretionary tectonic processes, forming elongate belts (terranes) that built eastwards through time. This lasted from Cambrian–Cretaceous time, and ceased with the onset of continental rifting of Zealandia away from Gondwanaland (Adams et al., 2007).

Investigations of the seafloor around New Zealand have shed light on the submarine depositional environment in which the original greywacke sediments may have accumulated. Mapping of the Cook River and Hokitika River canyons off the Tasman Sea coast of S Westland, suggest that they may be appropriate depositional analogues. Whereas on land these large-volume, rapid-flowing but short, steep-gradient rivers occupy valleys 1–2 km wide, on the sea floor the valleys burgeon outwards and may become 10–20 km wide.

## Mineral resources

New Zealand is relatively well-endowed in natural mineral resources such as Au, coal, fresh water, oil, gas, geothermal energy, ironsand, limestone, clay, zeolite, sand and construction aggregate. The land remains under-explored for mineral deposits and known resources are relatively under-developed (Williams, 1974; Kear, 1989; Brathwaite and Pirajno, 1993; Christie and Brathwaite, 2006).

During 1860–1880 there were Au rushes in Otago, Nelson-Marlborough, West Coast and Coromandel, that boosted the country's population and wealth, leading to a demand for coal for steam-powered machinery and heating, and building stone for public and commercial buildings. Demand for fertilisers for agriculture led to the quarrying of limestone, and, in the central North Island, the mining of geothermal sulfur.

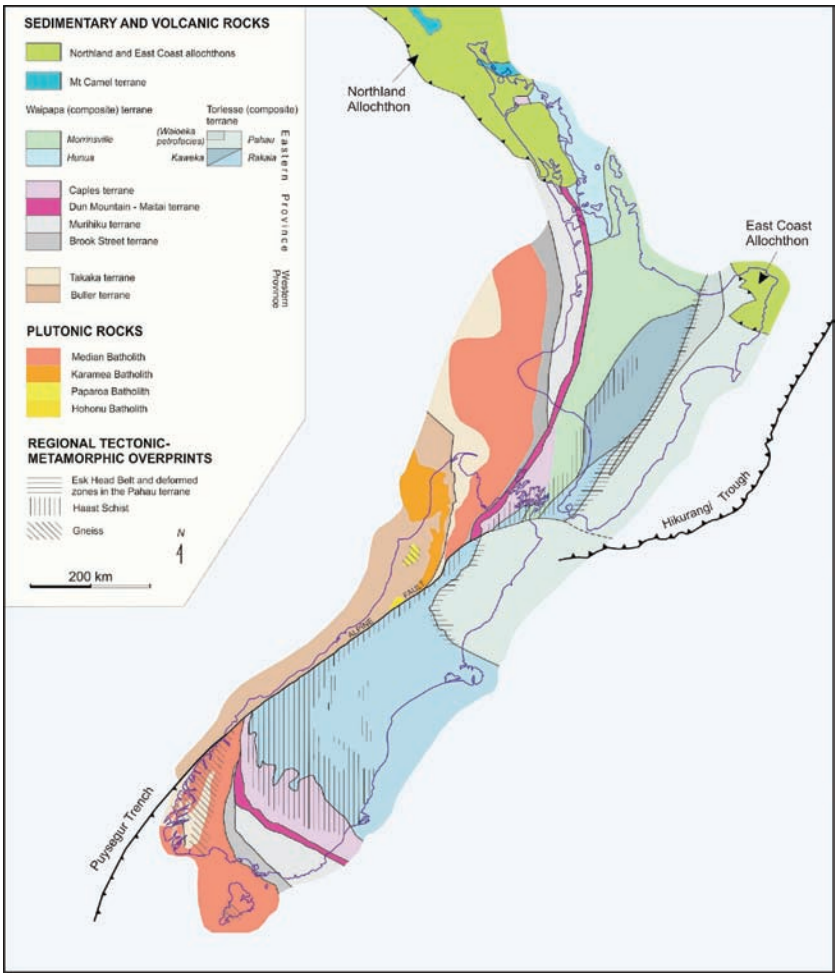


***Figure 8 The distribution of pre-Cenozoic basement rocks in New Zealand, showing tectonostratigraphic provinces, terranes and batholiths. The extent of the Northland and East Coast Allochthons, emplaced in Early Miocene time, is also shown; all other units were assembled by Late Cretaceous time.***

Gold production in New Zealand totalled 900 tonnes (29 M ounces) up to 2005 (Christie and Brathwaite, 2006) and recent increases in the Au price have enabled the mining of low-grade Au deposits such as Macraes. Exploration and research indicate potential for new Au discoveries in Northland, Coromandel, central North Island, Otago and the West Coast.

The black ironsands along the western coast of the North Island were investigated as a source of iron, but the high titanium content defeated traditional blast furnace smelting. This problem has been resolved and ironsand is now being smelted into iron and steel for New Zealand use, and exported (Barakat and Ruddock, 2006; Mauk et al., 2006). Submarine ironsand deposits on the inner continental shelf, mapped in the 1970s, are being re-assessed.

Since the 1940s, exploration for coal has established new coal mines and identified large lignite resources in Southland and Otago (Edbrooke, 1999). Waikato coal is used for electricity generation and is widely used as fuel for the steel, dairy, cement, forestry and meat processing industries, and there are plans to convert the lignite resources into fire briquettes and diesel fuel.

Annual coal production in 2007 exceeded 5 Mt, including high quality coals from the West Coast for export, and lesser quality coals for local power generation.

Geothermal energy has been generated in New Zealand since 1858. A global first, the Wairakei power station near Taupo (central North Island) exploits water-dominated geothermal systems. Geothermal energy from seven power plants now provides 13% (435 megawatts) of New Zealand's electricity needs. There is potential for an additional 1332 megawatts from 11 geothermal systems in central North Island and another 200–400 megawatts from non-traditional sources of geothermal energy such as abandoned deep oil and gas wells in sedimentary basins. Direct usage of geothermal energy for heating and drying in industry, agriculture, horticulture and aquaculture is growing.

Platinum, currently more valuable than Au, has been discovered in plutonic rocks of the Longwood Range in Southland (Christie et al., 2006). Extensive Fe-Mn nodule fields have been mapped along the southern and eastern flanks of the Campbell Plateau in water depths of 4,000–4,500 m (Graham and Wright, 2006). These nodules are a huge potential resource of Fe, Mn, Ni, Cu, Co and rare earth elements.

Extensive phosphate deposits on the Chatham Rise represent a resource of c. 100 Mt of phosphorite (von Rad and Kudrass, 1984).

Recent surveys of the southern part of the Kermadec Arc have located submarine hydrothermal vents with associated massive sulfide deposits containing high concentrations of Ba, Cu, Zn, Pb and Au. These deposits are also associated with unusual microorganisms, which might themselves have value for medicinal or other purposes (Massoth and de Ronde, 2006; de Ronde, 2006).

## Petroleum prospectivity

In 2008, the United Nations Law of the Sea (UNCLOS) accepted New Zealand's legal claim for jurisdiction over a greatly enlarged area of sea floor. New Zealand's Exclusive Economic Zone (EEZ) and Extended Continental Shelf (ECS) are now more than 5.7 million km$^2$ (Figure 1) including 1.5 million km$^2$ of petroleum-prospective sedimentary basins.

Oil was first produced in New Zealand in 1865 from Taranaki, but the first significant discoveries came a century later with two large gas-condensate fields: the onshore Kapuni Field in the late 1950s, and the offshore Maui Field in the late 1960s. They supply natural gas for electricity generation, gas-to-gasoline, methanol, urea fertiliser, and industrial and home heating.

New Zealand's first significant commercial oil field, the McKee Field, was discovered in 1979 in onshore Taranaki. Since 2000, four new fields have been developed in Taranaki: the onshore Pohokura and Kupe gas fields, and offshore Tui and Maari oil fields. By the end of 2007 more than ten oil and gas fields were producing, with two new fields due to come on-stream, but the combined reserves of these new fields are still only c. a quarter of the size of the declining Maui Field.

In 2010, New Zealand's second biggest export commodity in terms of earnings (after dairy produce) was oil. Production is only from Taranaki Basin. A total of c. 2 BBOE (billion barrels of oil equivalent) has been discovered in Taranaki (Uruski et al., 2011), with 19.6 million

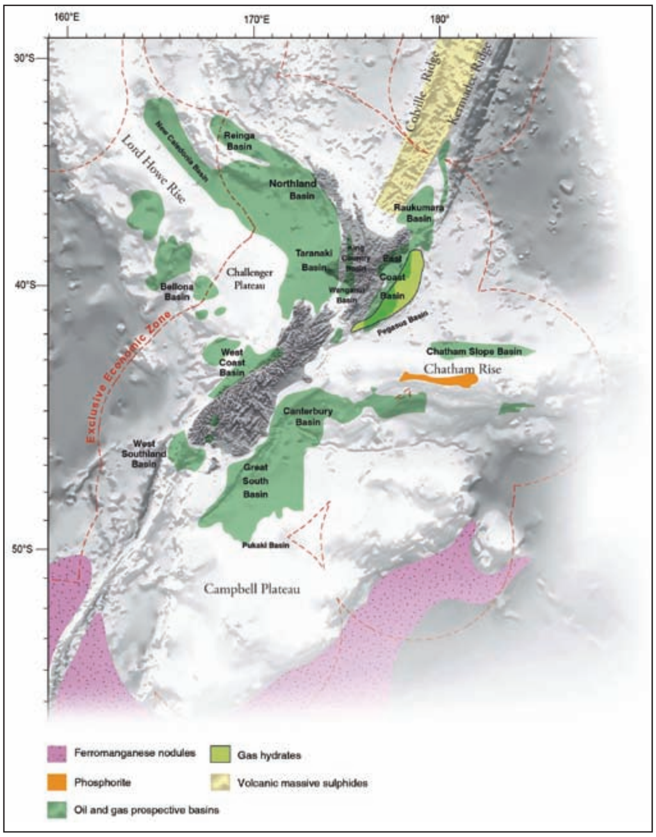


***Figure 9** The economic mineral resources of New Zealand as distributed on land and across Zealandia. Of prime importance are the onshore occurrences of oil, gas, coal, Au, aggregates, groundwater, and geothermal energy, and the offshore potential for oil and gas, metals and methane gas hydrates. The dotted line denotes the limit of New Zealand's EEZ + ECS.*

barrels of oil and 145 BCF (billion cubic feet) gas produced in 2009 (see www.nzpam.govt.nz).

The Ministry for Economic Development has undertaken a number of regional 2D seismic surveys since 2005 in Great South, Pegasus, East Coast, Raukumara, and Reinga basins (Figure 9). Results from these surveys have identified encouraging prospective petroleum plays. Multinational companies ExxonMobil, Anadarko, OMV, Hyundai Hysco, AWE, Mitsui, PTTEP, and Petrobras are evaluating these data, along with locally based New Zealand exploration companies.

Future prospective areas are the Raukumara Basin and the Canterbury Basin. In 2010, Petrobras was awarded a 12,000 km$^2$ exploration block off the East Coast of the North Island in the Raukumara Basin, and is acquiring and interpreting seismic data. Sediments in the basin are as much as 11 km thick, but little else was known about the geology of the region until 2D seismic data was acquired in 2005 and 2007 (Stagpoole et al., 2008). Four major sedimentary megasequences, range between Cretaceous and Recent, with a number of stratigraphic and structural plays (Figure 10). One of the megasequences recognised is a submarine slope failure tentatively correlated with the East Coast Allochthon. No wells have been drilled in the Raukumara Basin.

In the inner Canterbury Basin, eastern South Island Cretaceous source rocks have been recognised in a NE-trending graben, in close association with reservoir sandstones of comparable age. This is the Carrack-Caravel prospect located 50–80 km offshore from the North Otago coast. This basin has had only sporadic drilling (Mogg et al., 2008). This 390 km$^2$ prospect has the potential to contain up to 750 million barrels of recoverable oil or 2.7 trillion cubic feet of recoverable gas and 500 million barrels condensate (Petroleum Review 2010). In 1985, Galleon-1, located 30–50 km from the Carrack-Caravel prospect, produced sub-commercial gas that tested at 10 MMCFGPD and 2300 CFPD of condensate. Another drillhole, Cutter-1, recently completed, is the only offshore exploration well drilled in the Canterbury Basin since the mid-1980s.

Source rocks are typically Cretaceous through to Eocene in age, dominated by terrestrial coal deposits. Reservoir rocks range between fluvial and deep marine, and are as young as Pliocene. Widespread mudstone, marl and limestone of Late Oligocene–Early Miocene age provide regionally extensive seal rocks.

On the western margin of the North Island, progradational successions built out into the rift-related New Caledonia Basin in the Jurassic and Early Cretaceous and are as much as 2,500 m (1.5 sec TWT) thick. This is the so-called "Taranaki Delta" of Uruski (2007), which may include substantial volumes of source rock (Figure 11). Rift basins W of the North Island are common and formed over much of Zealandia in the Early Cretaceous during continental fragmentation of eastern Gondwanaland (e.g., Gippsland, Capel, Faust basins; Uruski et al., 2002; Hashimoto et al., 2010). Mesozoic rocks on mainland New Zealand (e.g. Murihiku Supergroup) have been considered economic basement. However, their source rock potential has long been postulated (Fleming, 1958; Cook et al., 1999); only now are they are being recognised as significant potential source rocks. As

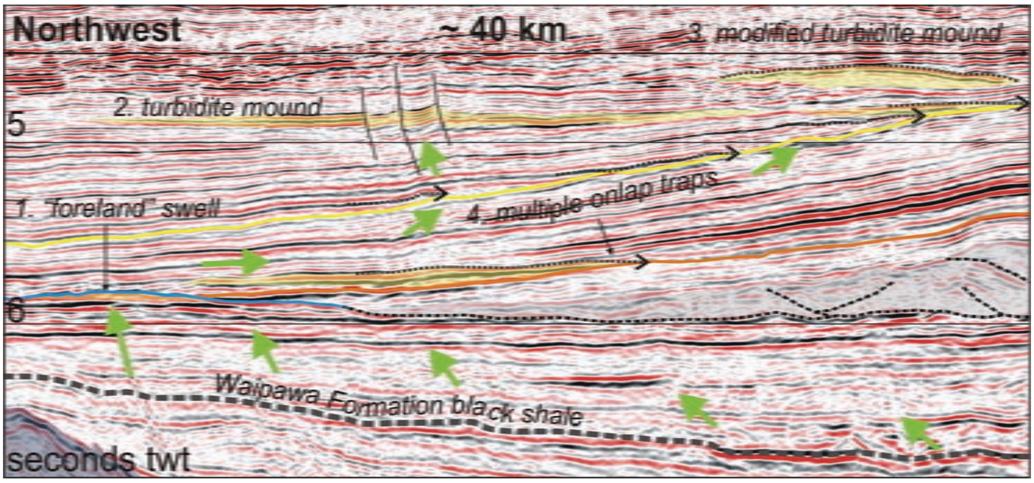


***Figure 10** Seismic line through the Raukumara Basin showing postulated migration pathways (green arrows), from Early Eocene source rocks at depth (Waipawa Formation; a potential source rock), around the toe of the submarine slope failure megasequence (grey shading) and up-dip into possible Neogene reservoir formations. Several potential traps (labelled 1 to 4) are indicated. Depth is in two-way-travel time (TWT) and the image is approximately 40 km across (after Stagpoole et al., 2008).*

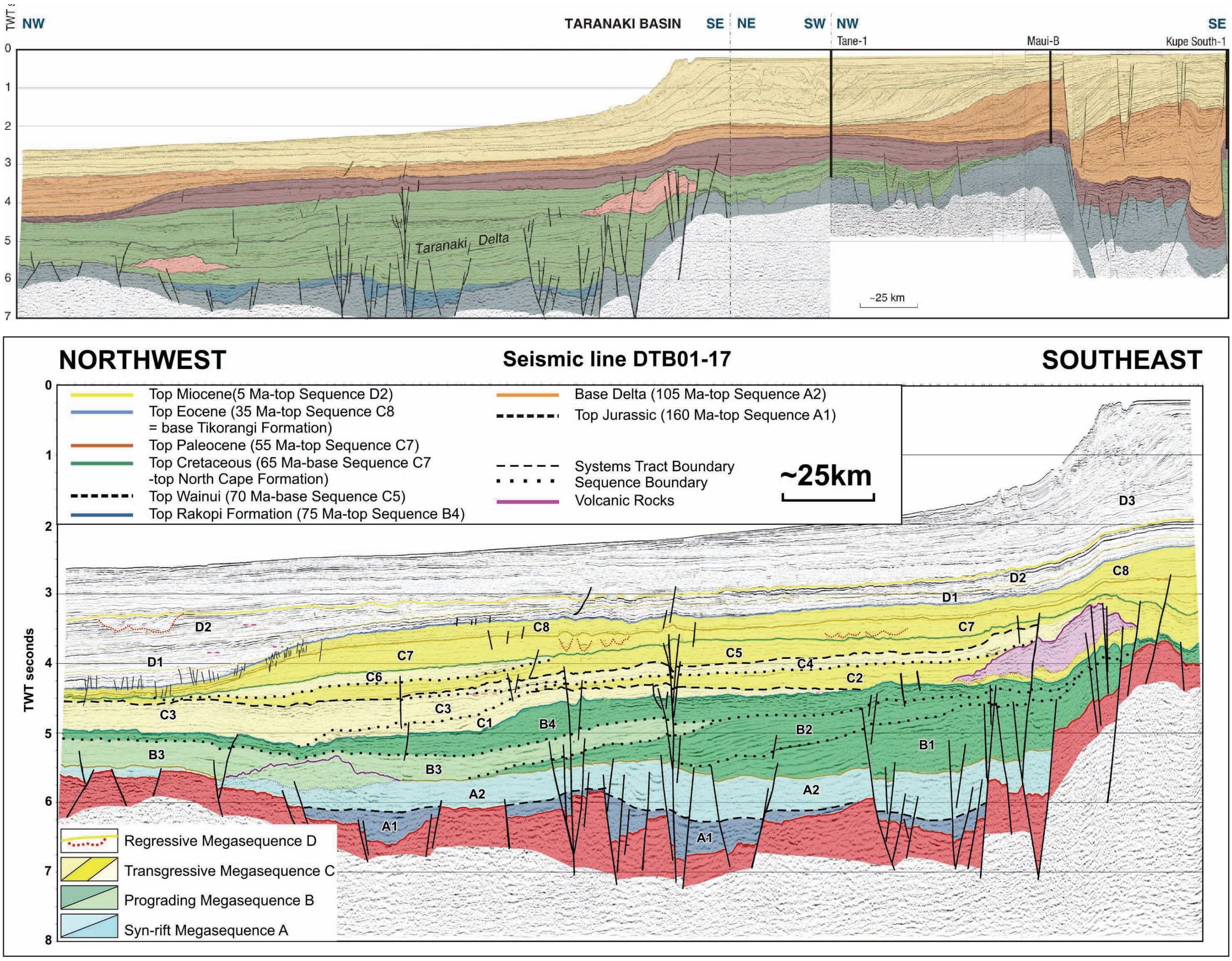


***Figure 11 Representative seismic profile (Line DTB01-17) from Kupe South-1 (SE) to the Deep Water Taranaki Basin (NW), western margin of North Island, New Zealand. Broad-scale stratigraphic interpretation of seismic units is shown (lower left; Uruski, 2007).***

exploration goes into deep water, these older rocks may become an important component of frontier basin petroleum systems.

With the much broader mandate that the 2008 UNCLOS jurisdiction now provides, substantial offshore exploration efforts are planned, especially for frontier areas such as the Deep Water Taranaki, Northland, Reinga, Pegasus and Great South basins. Planned drilling in Deep Water Taranaki Basin will test geological interpretations of Cretaceous non-marine and shallow marine source and reservoir facies, but also the thick deltaic succession postulated to underlie it. Seaward of these deltaic sediments there is potential for pro-delta turbidites that may contain kerogen carried into deeper waters. The largest structure recognised to date is the Romney prospect, with a closure area of 200 km$^2$ in 1600 m water depth, and with the potential to contain up to 1100 to 1650 million barrels of oil-in-place or between 1.7 and 2.7 trillion cubic feet of gas (Uruski et al., 2010).

Some 5,900 line kilometres of 2D seismic (Figure 12) in Northland Basin and the neighbouring Reinga Basin indicate a thick rift succession of likely Jurassic to earliest Cretaceous age and that total sediment thickness in these basins is up to 9 km. Regional mapping indicates several plays ranging from structural to stratigraphic (Uruski et al., 2008; Stagpoole et al., 2009; Uruski et al., 2010). Only one well, Waka Nui-1, has been drilled.

Other basins of note include the Pegasus Basin SE of the North Island, an unusual basin, because it has been largely undeformed by late Cenozoic subduction-related tectonics, despite its location adjacent to the modern day plate boundary and Hikurangi Trench. Also of interest is the Great South Basin SE of the South Island.

The New Zealand region of Zealandia has producing petroleum fields both on land and near to land, but with such a vast and unexplored continental area lying further out to sea, there is the potential to discover continental scale resources.

# Alpine Fault

The Alpine Fault was first recognised and mapped by Harold Wellman and Dick Willett in the 1940s, and Wellman postulated 480 km of displacement (Nathan, 2011). It is the main active structure in the oblique continental collision zone of the South Island. It is continuous at the surface for c. 800 km and accommodates c. 70% of current plate motion (Norris and Cooper, 2001). A 480 km offset of basement rocks implies that the Alpine Fault has accommodated >50% of plate displacement since the Eocene (45 Ma) (Sutherland, 1999).

Observed geophysical, geological, and contemporary kinematic data can be successfully explained by a model involving Pleistocene

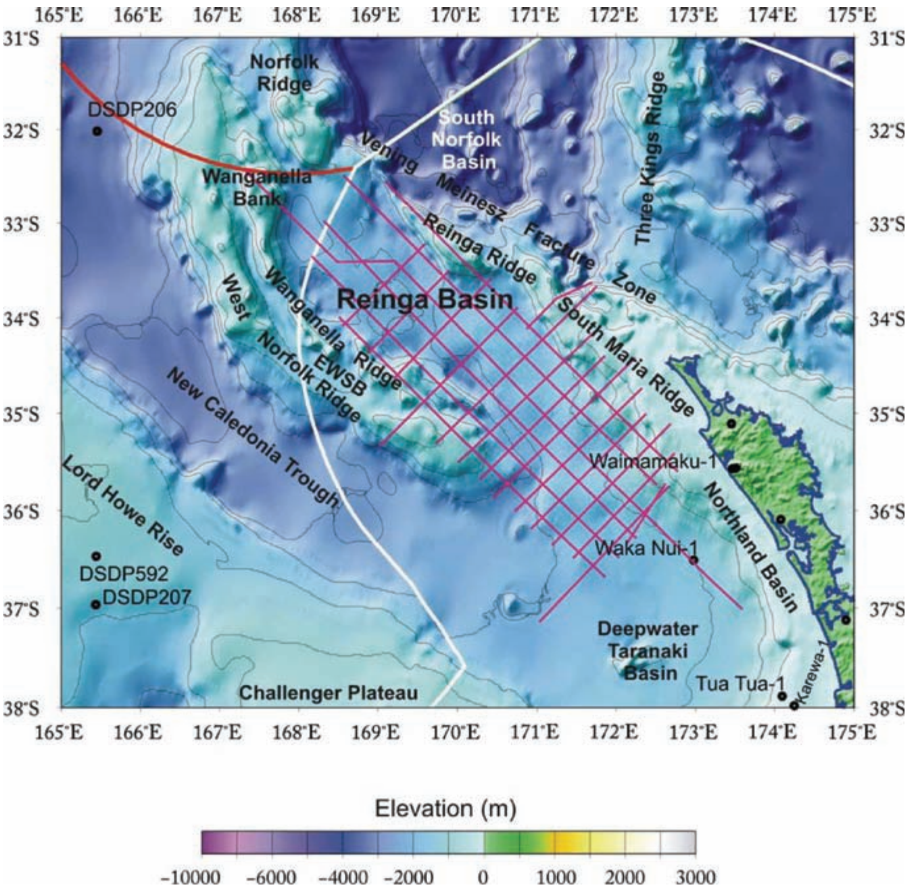


***Figure 12 Physiography of the Reinga Basin and adjacent area showing the Northland Basin and the Deepwater Taranaki Basin. Recent seismic acquisition lines (the REI09 survey) are shown in pink; drillhole locations (mainly DSDP sites) are black dots; the outer limit of the New Zealand EEZ is the bold white line; the outer limit of the New Zealand ECS is the bold red line (after Stagpoole et al., 2009).***

slip on a narrow Alpine Fault zone extending into the lower crust (Beavan et al., 2007).

The Alpine Fault represents a major lithospheric discontinuity within Zealandia that previously separated Paleozoic continental lithosphere of the Challenger Plateau from Eocene–Miocene oceanic lithosphere (Sutherland et al., 2000). Subduction of oceanic lithosphere at the Puysegur Trench and strike-slip motion on the Alpine Fault since c. 25 Ma has translated this passive margin into an active continent–continent collision zone. The orientation of the Alpine Fault is inherited from the Eocene geometry of rifting and is not perfectly oriented to accommodate strike-slip motion. The obliquity of plate motion to this inherited structure has resulted in widespread deformation of the Pacific plate and consequent uplift of the Southern Alps.

Erosion of the Southern Alps by strong westerly weather systems is highly efficient and has deposited vast quantities of sediments offshore (Sutherland, 1999). This erosion has an important role in stabilising localised slip on the Alpine Fault and inhibits growth of a broad mountain range within the Australian Plate (Koons, 1987).

Previous studies of the central Alpine Fault (Figure 13) suggested that the adjacent mid-crust is hot ( Koons, 1987; Toy et al., 2010) and over-pressured (Stern et al., 2001). Hanging-wall seismicity terminates at a depth of c. 10–12 km (Leitner et al., 2001), and focal mechanism analysis suggests that the Alpine Fault sustains shear stresses lower than expected for idealised faulting models (Townend, 2006). Geodetic studies have yielded elastic locking depths as shallow as 5–8 km (Beavan et al., 1999) though partial locking may extend to as deep as 18 km (Wallace et al., 2007).

The near-surface fault plane is segmented on a 1–4 km along-strike-scale into reverse segments with strike 030–050°, and strike-slip segments with strike 070–090° (Norris and Cooper, 1995). It is thought that this segmentation has formed in response to topographic stress perturbations caused by erosional processes (Koons and Kirby, 2007).

The 1 km-thick hanging wall Alpine Fault rock sequence (Figure 14) has been exhumed from depths of <35 km at rates more than 4 mm/year regionally, and locally >10 mm/year during the past 5 Myr ( Little et al., 2005; Beavan et al., 2010b). Large scale plate motions that are driving the deformation have changed only slightly since 20 Ma, therefore the exposed sequence was probably entirely deformed under conditions presently experienced at depth.

The Alpine Fault (Figure 13) has not produced significant earthquakes or measurable creep during New Zealand's relatively short written history, but is thought to fail in large earthquakes (Mw c. 7.9) every 200–400 years, and to have last ruptured in 1717 (Wells et al., 1999).

At least nine key attributes make the Alpine Fault an attractive target for fundamental research into tectonic deformation,

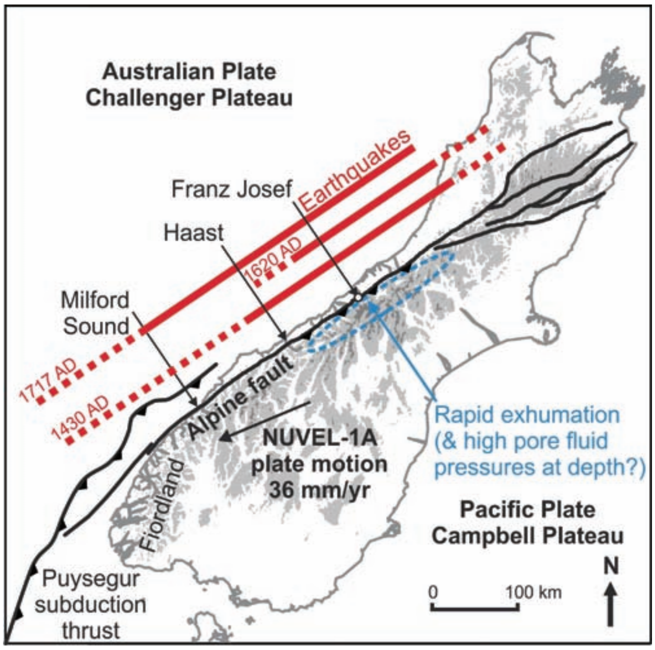


***Figure 13 Map of the Alpine Fault and adjacent structures highlighting the inferred rupture extent of past Alpine Fault earthquakes in 1430, 1620 and 1717 (Sutherland et al., 2009). Light grey shading demarcates topography higher than 800 m.***

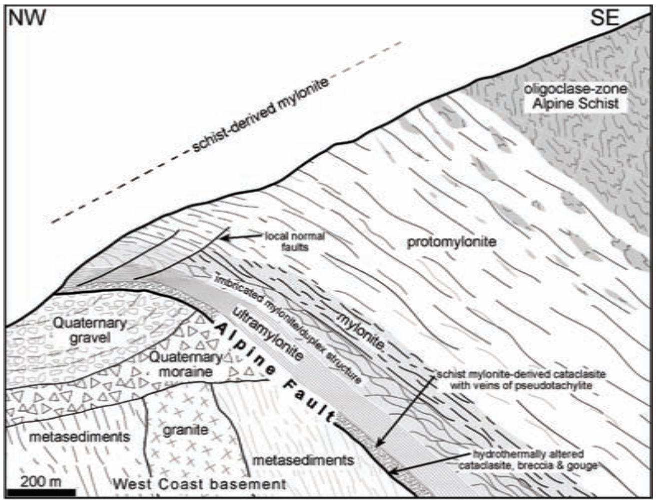


***Figure 14** Sequence of fault rocks, exposed in the hanging wall (Mesozoic basement, Eastern Province; Pacific Plate) of the Alpine Fault that have been exhumed from depths of as much as 35 km and are thrust on top of young (typically <16 kyr) Pleistocene-Recent fluvio-glacial gravels and Paleozoic basement rocks (Western Province; Australian Plate) (Norris and Cooper, 2007).*

seismic hazard, and mineral formation. (1) It has well-determined and rapid (25 mm/year) Pleistocene slip rates; (2) precisely known plate motion history with a single fault that has accommodated most plate boundary displacement for >20 Myr; (3) >300 km along-strike exposure of uniform-composition rocks in the hanging wall and fault

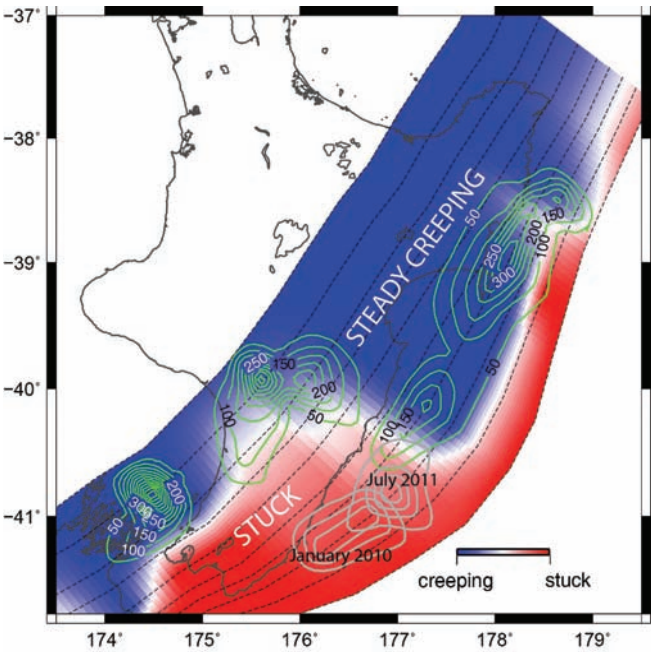


***Figure 15** Inter-seismic coupling on the Hikurangi margin of the westward subducting Pacific Plate interface. Modelling derived from the 1991-2004 GPS campaign (Wallace et al., 2010). The coupling factor is the fraction of inter-block motion being stored as elastic strain energy: high coupling factor (red); low coupling factor (blue). Green contours are previous slow slip events (SSE); grey contours are newly discovered Wairarapa coast SSE. The red area is locked (stuck) subduction interface that is inferred to be building up for a large subduction thrust earthquake.*

rocks derived from them; (4) sequences of fault rocks developed during unidirectional exhumation on well-determined trajectories over relatively short time periods; (5) it has had rapid uplift resulting in advection of crustal isotherms so that brittle–ductile transition processes can be studied at shallower (potentially drillable) depths than normal; (6) it offers the opportunity to monitor a locked fault late in its perceived earthquake cycle with a high (c. 25%) probability of rupture in the lifetime of a fault-zone observatory (c. 30 years); (7) there is an extensive body of existing geological and geophysical knowledge, and a modern nationwide geophysical monitoring network (GeoNet); (8) it has an inclined fault orientation enabling fault penetration with sub-vertical boreholes; and (9) New Zealand has a relatively benign political and physical environment in which to operate, with existing industry and supporting infrastructure.

Understanding what conditions prevail within the interiors of active faults is crucial for elucidating the mechanisms governing long-term fault evolution and, in particular, the earthquake-rupture processes that are of special interest to society (Tullis et al., 2007; Zoback et al., 2007).

This motivation, combined with the factors listed above, led to an ambitious multinational project, the “Deep Fault Drilling Project, Alpine Fault (DFDP)” (Townend et al., 2009). The DFDP project completed Phase 1 by drilling two boreholes (maximum depth 152 m) at Gaunt Creek, c. 20 km NE of Franz Josef Glacier, during January and February 2011.

Phase 2 of the project will determine physical conditions at upper- and mid-crustal depths by drilling to a depth of >1 km, hence seeing through near-surface effects of topographic relief on stress, temperature, and fluid flow and chemistry, and of critical importance for inferring conditions at greater depth (Fulton and Saffer, 2009; Koons and Kirby, 2007; Liu and Zoback, 1992; Norris and Cooper, 1997).

## Current tectonic activity

The Pacific Plate is subducting beneath the Hikurangi Trough at 5–6 cm/year offshore of Gisborne, E coast North Island, making this subduction thrust the fastest slipping fault in New Zealand. Farther S, beneath southern North Island, the subduction rate reduces to 3–4 cm/year and GPS data show the subducting plate is locked by friction on the subduction plate interface to a depth of c. 40 km (Wallace et al., 2010). This is in contrast to the Gisborne region, where GPS data show episodic movement in “slow slip events” to shallow depths (<10 km), and is of concern because locked regions may eventually become unstuck with a resultant large earthquake. Data from around Wellington bear striking resemblance to data from Japan near Sendai before the 11 March 2011 M 9.0 Tohoku Earthquake. However, no conclusive geological evidence has yet been found of such subduction thrust earthquakes in southern North Island.

GPS technology has transformed seismology in the past 14 years and, in particular, understanding of subduction zone kinematics. It has led to the discovery of slow slip events (SSE) also known as slow or silent earthquakes. These are hypothesised to occur around the edges of locked zones and may help to identify those areas on the plate interface where subduction earthquakes may occur in the future.

Active volcanoes and associated geothermal fields in the TVZ are all the products of active subduction of the Pacific Plate along the Hikurangi Trough beneath the Australian Plate and/or back-arc rifting of continental crust of the Australian Plate.

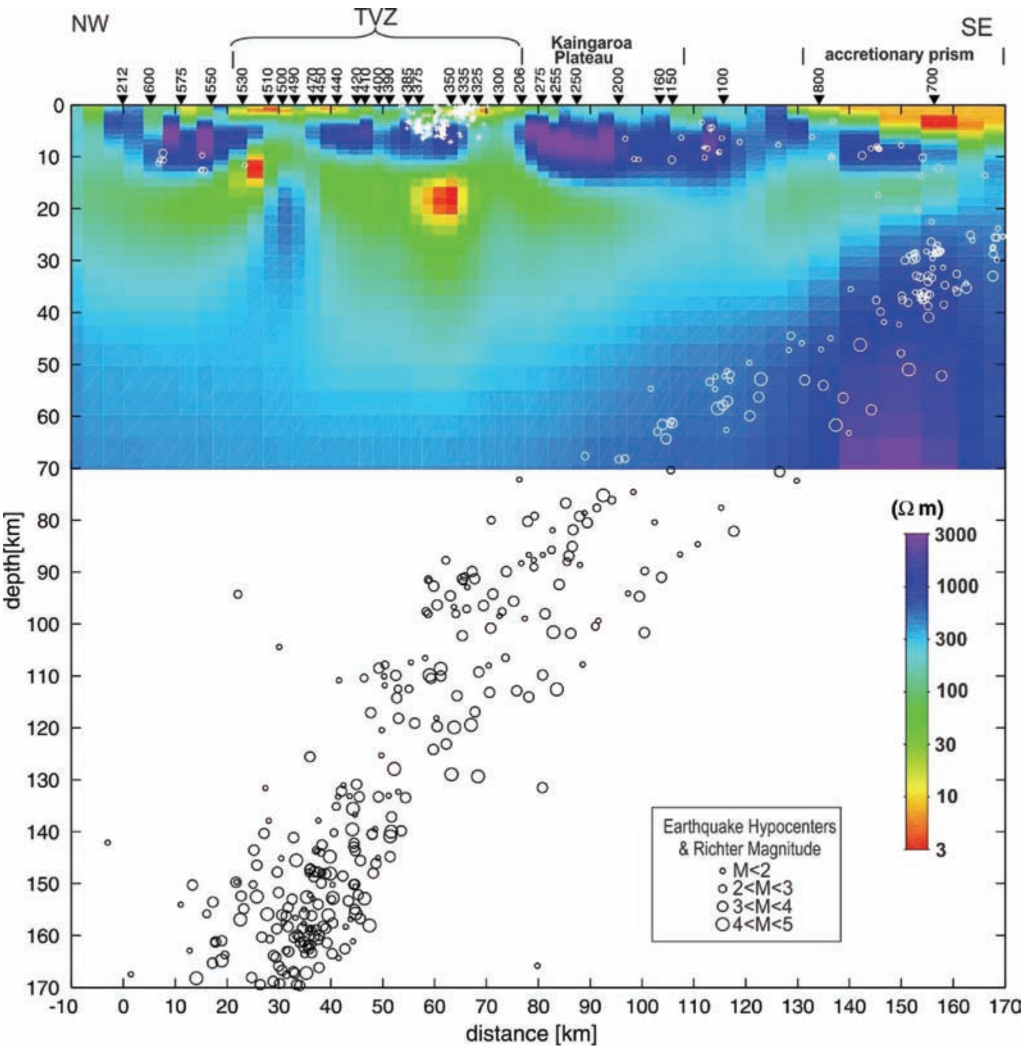


***Figure 16 The Taupo Volcanic Zone as an active rift within continental crust of the Australian plate, which is closely associated with subduction of the Pacific Plate. This NW-SE depth profile shows the conductivity structure within the upper 70 km that has been constructed from 2-D inverse modelling of magnetotelluric (MT) data (Heise et al., 2007). Earthquake hypocentres mark the slab of the subducting Pacific Plate. The colour code is expressed in terms of resistivity. Note the rapid change that occurs beneath the TVZ at 10 km, some 3 km below the base of the seismogenic zone. This is a reasonable and plausible interpretation of the depth to the melt zone (magma) in the thin crust beneath the TVZ.***

Data acquired using magnetotelluric (MT) technology has enabled the TVZ to be mapped in terms of electrical conductivity and has produced a compelling thermal picture of the deep geological structure of the TVZ (Heise et al., 2007). An example of this research is shown along a NW–SE profile across the TVZ (Figure 16). It suggests that the mantle wedge beneath the TVZ is anomalously conductive, compared to the Pacific Plate lithosphere beneath the South Island. This supports the presence of a partially interconnected melt fraction in the mantle wedge above the plate as suggested from seismological data by Reyners et al. (2006). The melt fraction is greater along the SE margin of the TVZ, where the geothermal flux and geothermal activity is greatest.

Subduction along the E coast of New Zealand, recognised on the basis of the pattern of seismicity, effectively ceases S of Wellington and N of Christchurch. Plate motion mechanics are transformed into complicated crustal deformation as the nature of the collision changes from ocean-continent (North Island) to continent-continent (South Island). A series of faults connect the subduction zone to the Alpine Fault. Right lateral motion on the Alpine Fault has involved relative plate displacement of 480 km since the Early Miocene.

The rate of Pacific Plate motion in the South Island is 38 mm/yr W with respect to the N-moving Australian Plate. This continent-continent collision is responsible for the Southern Alps, and the same tectonism also has the potential to activate faults located beneath the alluvium of the Canterbury Plains.

Relief of this westward Pacific Plate stress through fault motion along buried faults has recently generated a number of disastrous earthquakes in the Christchurch area.

The first of these earthquakes occurred at 04:55 on Saturday 4 September 2010 (local time) with a moment magnitude (Mw) of 7.1. The earthquake epicenter was located 10 km SE of Darfield and 40 km W of Christchurch at a depth of 10.8 km. An E-W oriented surface rupture, some 30 km long with an average horizontal displacement of 2.5 m (Quigley et al., 2010) developed 4 km S of the epicenter.

This strike-slip fault was previously unknown and is now named the Greendale Fault. There was no loss of life. The older brick and masonry buildings of Christchurch were badly damaged. Liquefaction, lateral spreading and slumping caused major damage with an estimated $NZ4 billion to repair the damage (Gledhill et al., 2011). The Darfield Earthquake was followed by more than 4,000 aftershocks through January 2011, with 14 aftershocks of local magnitude (ML) 5.0 greater and 155 of ML 4–5 (Figure 17).

A destructive aftershock of local magnitude (ML) 6.3 and shallow depth of 2.8 km (Kaiser et al., 2012) struck c. 6 km SE of downtown Christchurch at 12:51 on Tuesday 22 February 2011 local time causing extensive damage and a loss of 185 lives.

This Christchurch Earthquake was very energetic (Me) 6.7 with a recorded maximum vertical acceleration of 2.2 g near the epicenter (Gledhill et al., 2011), leading to greater destruction and liquefaction than the Darfield Earthquake almost six months earlier. The eastern suburbs of Christchurch were especially hard hit with liquefaction, lateral spreading, flooding and subsidence. More than 900 buildings in the central business district were irreparably damaged as well as 10,000 or more residential homes.

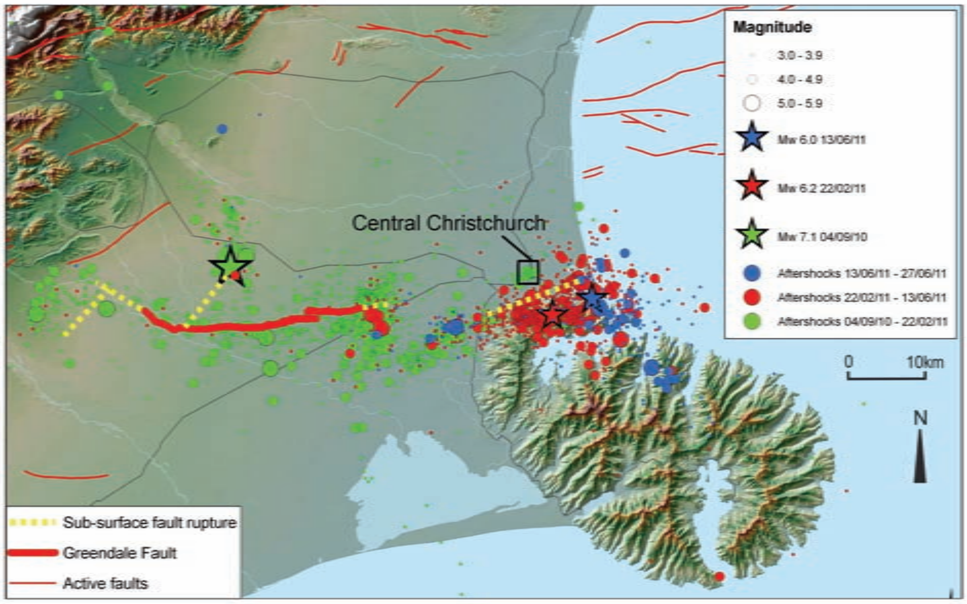


*Figure 17 Central Canterbury showing the aftershocks and faults relating to the Christchurch earthquake sequence. The first major event on 4 September 2010 (green) was associated with rupture of the Greendale Fault was followed by major aftershocks on 22 February 2011 (red), and 6 June 2011 (blue). These latter two events were associated with blind faults proximal to central Christchurch. From Kaiser et al. (2012).*

This earthquake occurred on a previously unrecognised NE-SW oriented blind fault; it did not rupture the surface. However, resultant permanent ground deformation caused relative uplift of the Port Hills and the mouth of the Heathcote-Avon River system, and down-drop of much of eastern Christchurch, which was already low-lying and close to sea-level. Cathedral Square in central Christchurch was 5.50 m asl; it is now 5.15 m asl. With the additional effects of lateral spreading, liquefaction and compaction, parts of eastern Christchurch have subsided by up to one metre.

The 22nd February Christchurch Earthquake caused an additional $NZ15–20 billion of damage making the Christchurch earthquake the most expensive in New Zealand's history.

The E-W orientation of the Greendale Fault and other faults in Canterbury suggest that these recent earthquakes may have occurred on re-activated Cretaceous faults (Reyners, 2011). Numerous E-W normal faults relating to extensional tectonism when Zealandia rifted away from Gondwanaland have been mapped on the Chatham Rise to the immediate E of Banks Peninsula (Wood and Herzer, 1989).

The extinct Late Miocene basalt shield volcano of Banks Peninsula (Figure 17) may have contributed to concentrating the stress field following the Darfield Earthquake (Reyners, 2011). The Christchurch Earthquake involved oblique - reverse faulting. The red dots (Figure 17) show that the aftershocks are consistent with a NE-SW striking fault plane dipping to the SE.

These Christchurch earthquakes are considered to be an unusual sequence. Available geological evidence prior to the Darfield Earthquake indicated no disturbance of this nature for at least 10 kyr. On the other hand, the damage sustained by the particularly violent nature of these earthquakes was not unexpected. Detailed geological and hydrological mapping completed in the 1980s correctly interpreted susceptibility of eastern Christchurch to seismic shaking and liquefaction hazard (Brown and Weeber, 1992; Forsyth et al., 2008).

The tectonism that Christchurch has experienced is due to stress release associated with routine plate boundary deformation. Whereas c. 70% of plate motion is taken up proximal to the plate boundary in the South Island by the Alpine Fault and Southern Alps, the remaining c. 30% is accommodated further afield. For all that the Christchurch experience is topical, it is important to remember that the entire plate boundary zone running through New Zealand is alive.

## Conclusion

The geological history of New Zealand makes sense in terms of the large scale tectonic history of the SW Pacific. Our Gondwanaland heritage lasted from 510–83 Ma and was dominated by active subduction margin tectonism and accretionary growth of the eastern margin of Gondwanaland.

Our Zealandian history 83–23 Ma was utterly different and relates to passive margin tectonism. This was a 60 Myr period of relative tectonic quiescence and sustained subsidence. In fact, rifting of Zealandia commenced at least 40 Myr prior to 83 Ma, so extensional tectonism of eastern Gondwanaland lasted for 100 Myr. The exposed geological record of our Zealandian heritage may be relatively meager, but this belies its much more widespread significance subsurface especially within New Zealand's submarine sedimentary basins.

The New Zealand history, relating to the past 23 Myr is just as interesting, reflecting the return to active subduction-related collision tectonics. However, there is a profound difference from our Gondwanaland heritage. The modern plate boundary involves continent-continent collision as well as ocean-continent collision, with all its attendant hazards. New Zealand is the product of continental Zealandia being split to form two sub-continents: northern Zealandia and southern Zealandia.

The geology of New Zealand can be explained in terms of crustal processes operating on a scale of 10s of kilometres, lithospheric processes occurring on a scale of 100s of kilometres, and mantle processes occurring on a scale of 1000s of kilometres. Just about everything in New Zealand is currently governed by oblique plate convergence at a rate of 3–5 cm/year.

Thanks to New Zealand's unique geological history and active tectonic setting, it is a spectacular natural laboratory for the study of both resource- and hazard-related geological processes, past and present. Advances in geophysics in particular are providing significant new insight into how the crust 'works', especially in the context of rifting and subduction.

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by Dominique Cluzel[1], Pierre Maurizot[1,2], Julien Collot[1,3] and Brice Sevin[1,3]

# An outline of the Geology of New Caledonia; from Permian–Mesozoic Southeast Gondwanaland active margin to Cenozoic obduction and supergene evolution

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*The geological evolution of New Caledonia may be divided into three phases. The Gondwanan phase (Permian–Early Cretaceous), is marked by subduction along the SE Gondwaland margin. At that time, proto-New Caledonia was located in a fore-arc region in which volcanic-arc detritus accumulated; whilst accretion and subduction of oceanic and terrigenous material formed an accretionary complex metamorphosed into the blueschist facies. During the Late Cretaceous–Eocene, marginal rifting isolated New Caledonia, and after a short period of shallow water terrigenous sedimentation associated with minor volcanic activity, only pelagic sediments accumulated. A new NE-dipping subduction appeared to the E of New Caledonia at the Paleocene-Eocene boundary, it generated the eclogite-blueschist complex of northern New Caledonia, consumed the eastern Australian Plate, and eventually ended with Late Eocene obduction when the Norfolk Ridge blocked the subduction zone. Finally, during the post-Eocene phase, New Caledonia definitively emerged; this episode mainly corresponds to prominent regolith development and minor tectonic events that lead to the present morphology and to the development of supergene nickel ores.*

## Introduction

Since Paris' geological synthesis of New Caledonia (Paris, 1981), advances in analytical methodologies and the application of the terrane concept by Aitchison et al. (1995a, b), Meffre (1995) and Meffre et al. (1996) has allowed a reappraisal of the regional geology. During the last thirty years, research groups or individuals from France, Australia, New Zealand and the USA have addressed several points of the geology of New Caledonia and proposed new interpretations; in addition, the recent creation (2006) of the Geological Survey of New Caledonia has provided new impetus and opportunity to gather new field data, harmonise and synthesise already mapped areas, and create geological databases; therefore, knowledge of the geology of New Caledonia has made significant progress.

Since earliest Permian time, the age of New Caledonia's oldest rocks (Aitchison et al., 1998), three phases of development have been recognised: Permian–Early Cretaceous, Late Cretaceous–Eocene, and Oligocene–Holocene. The oldest is related to the evolution of the SE Gondwanaland active margin, Mesozoic marginal basin opening and subsequent closure; the second corresponds to the rifting that isolated slices of the older Gondwanaland margin and extends into the Cenozoic convergence that eventually ends with Late Eocene obduction; the third mainly corresponds to the supergene evolution of New Caledonia and involvement of the Australian Plate in the New Hebrides (Vanuatu) subduction zone. This article presents a simplified review of the geological evolution of New Caledonia mainly based upon recently published data.

*Key for abbreviations:* MORB: mid-oceanic ridge basalt; E-MORB: enriched (or undepleted) MORB; OIB: oceanic island basalt; BABB: back-arc basin basalt; IAT: island-arc tholeiite; HP-LT: high pressure-low temperature (metamorphic rocks).

## New Caledonia in the framework of the Southwest Pacific

New Caledonia is located within a complex set of marginal basins and "continental" or volcanic-arc ridges (Figure 1). It is composed of several islands that are parts of the Norfolk and Loyalty ridges. The main island (or "Grande Terre") belongs to the Norfolk Ridge, which is connected southward to the large continental plateau that also bears New Zealand. The Belep Islands to the N and Isle of Pines to the S of the main island also belong to the Norfolk Ridge. In contrast, the Loyalty Islands represent the emerged part of a sinuous submarine ridge (the Loyalty Ridge) that runs more or less continuously parallel to the Norfolk Ridge over more than 1,500 km, from the d'Entrecasteaux Zone (W of Espiritu Santo,Vanuatu) in the N, to the Cook Fracture Zone in the S (Figure 1). Between the Loyalty and Norfolk ridges, the Loyalty Basin is a narrow, 1.5–3 km deep, oceanic

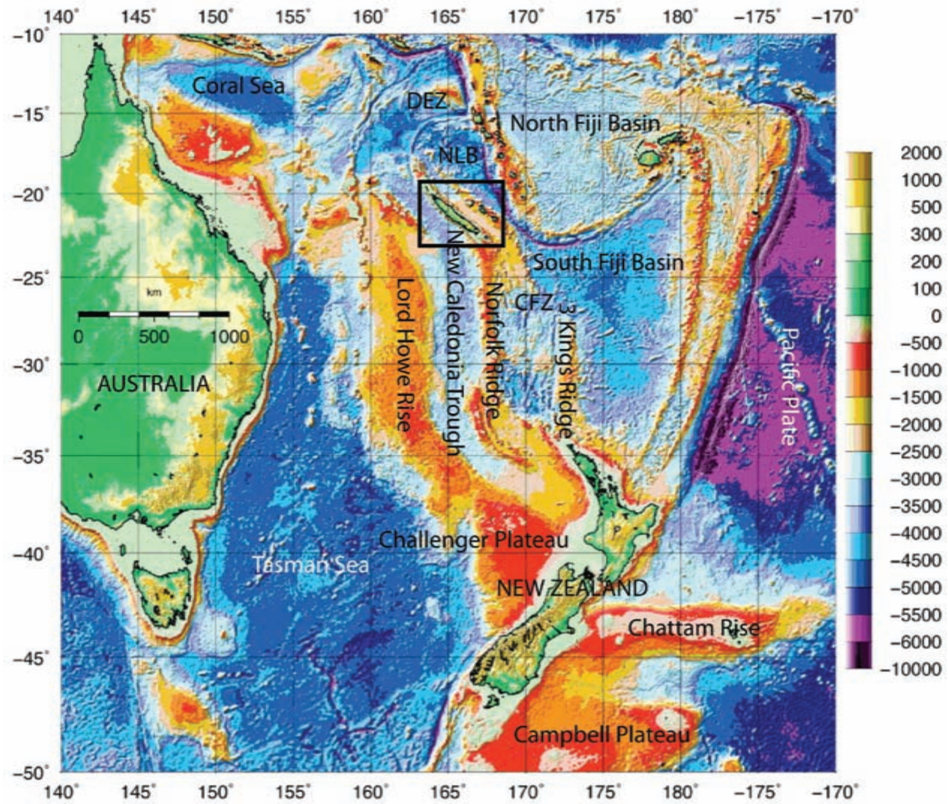


*Figure 1 Bathymetric map of the SW Pacific (Smith and Sandwell, 1997). DEZ: D'Entrecasteaux Zone; NLB: North Loyalty Basin; CFZ: Cook Fracture Zone.*

basin partly filled with 1–3 km of post-Eocene (?) sediments. The NE dip of the Moho, and geometry of seismic reflectors within the sedimentary filling of the basin (Bitoun and Récy, 1982) (Figure 2), are evidence of syn- and post-obduction infill. Over more than 500 km along the E coast of New Caledonia, a positive gravity anomaly (> +100 mGal) (see below) marks the occurrence of a large elongated body of dense rocks, which has been interpreted as the root of the Peridotite Nappe (see below; Figure 3), the latter is thought to be in continuity with the oceanic lithosphere of the Loyalty Basin (Collot et al., 1987).

To the W and SW of the Grande Terre, the New Caledonia Basin (Figure 4) is at present based upon thinned continental crust in its northern part, and oceanic crust in its southern part (Symonds et al., 1999; Auzende et al., 2000; Klingelhoefer et al., 2007). In the northern part of the basin, the pre-Cenozoic sedimentary infill shows evidence for half-graben structures that may be related to Late Cretaceous rifting. Pre-Oligocene reflectors (Figure 4) (Collot et al., 2008) and the Moho as well dip gently to the NE (Klingelhoefer et al., 2007). This may support the hypothesis of an attempted underthrusting (or continental subduction) of the northern Lord Howe Rise below New Caledonia (Cluzel et al., 2005). In contrast, the Oligocene westward onlapping sequence that rests horizontally above a gently NE dipping surface of unconformity (Figure 4), is consistent with the static infill of the basin by post-tilt, and thus post-obduction erosion products (Collot et al., 2008).

Similar to New Zealand, the main island of New Caledonia is a complex mosaic of volcanic, sedimentary and metamorphic terranes (Figures 5 and 6). New Caledonian terranes were assembled during two major tectonic episodes; an Early Cretaceous tectonic collage, which may be time-correlated with the Rangitata Orogeny of New Zealand (Fleming, 1969); and a Paleocene–Late Eocene subduction followed by obduction (Avias, 1967; Paris, 1981; Collot et al., 1987; Aitchison et al., 1995a). Both events included periods of high-pressure metamorphism and are therefore thought to

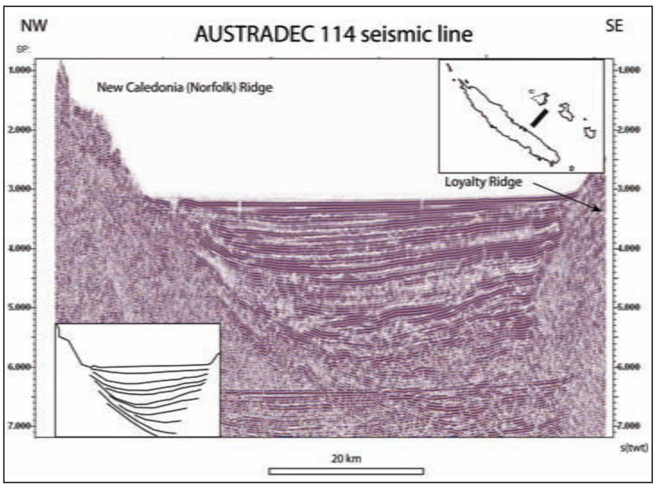


*Figure 2 Seismic profile of the Loyalty Basin (Austradec 114) (after Schor et al., 1971; Pontoise et al., 1982).*

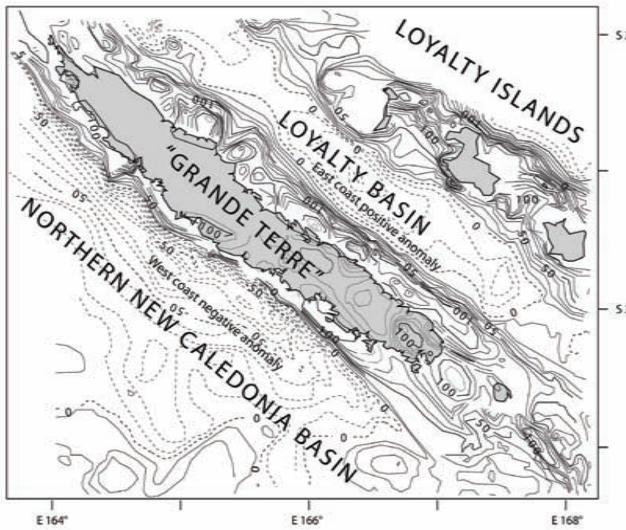


*Figure 3 Free air anomaly gravity map of New Caledonia and adjacent basins (after Collot et al., 1987; Van de Beuque, 1999) to show the prominent asymmetry of the gravity field in New Caledonia. Plain lines: positive anomaly; doted lines: negative anomaly. The E coast positive anomaly is related to the occurrence of peridotites at shallow depth below the eastern lagoon; whilst the negative anomaly of the W coast is related to the > 7 km sediment thickness in the New Caledonia Basin.*

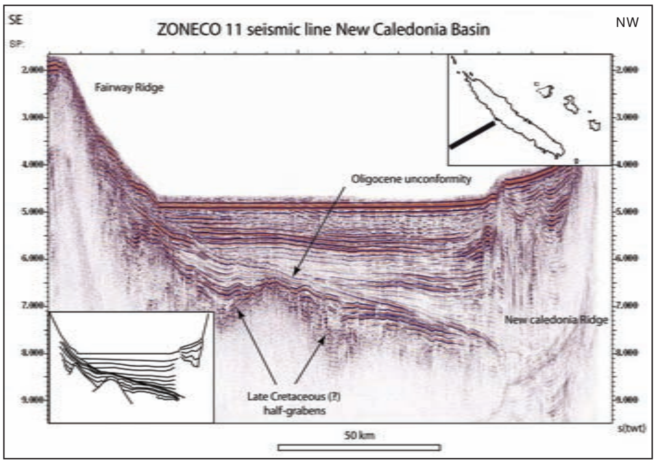


*Figure 4  Seismic profile of New Caledonia Basin (ZONECO 11) (Klingelhoefer et al., 2007).*

have occurred in connection with plate convergence in subduction zones.

## Pre-Late Cretaceous geology of New Caledonia: the Gondwanaland period

Two major terrane groups are distributed along the length of New Caledonia (Figure 5): an older, late Permian–Early Cretaceous group of three sub parallel, elongate terranes on the W coast (Teremba) and in the central mountain range (Koh-Central, and Boghen); and a younger Late Cretaceous–Eocene group, overlying the latter, that formed in response to break-up, drift, convergence and subsequent collision of an island arc (Aitchison et al, 1995a). The pre-Late Cretaceous terranes were formed during a period of subduction/accretion, and show closest biostratigraphic correspondence with Eastern Province terranes of New Zealand.

The three older terranes that form the central mountains of New Caledonia are as follows:

### The Koh-Central Terrane

The Koh-Central Terrane is formed of an "oceanic" basement overlain by a thick volcanosedimentary cover. A disrupted, Early Permian (Aitchison et al., 1998) ophiolite suite (Koh Ophiolite) occurs locally along the centre of the island, comprising gabbro, dolerite, rare plagiogranite, IAT, low-Ca boninite pillow basalts, and undated cherts directly overlying the pillow basalts. The Koh Ophiolite rocks are overlain by a thick succession of volcano-sedimentary rocks: black shale, volcaniclastic turbidite (greywacke), siltstone and chert (Meffre et al., 1996). The black shales are several hundred metres thick, whilst greywackes are generally associated with 20–50% argillite and locally with chert, conferring this terrane a distal and deep-water character. The greywackes are exclusively composed of volcanic lithic (andesite, dacite and basalt) and mineral clasts (feldspar, quartz, amphibole, etc.), whilst plutonic clasts are generally absent, except for one locality (late Early Cretaceous). In general, fossils are extremely rare and poorly preserved; however, scarce Middle Triassic (Anisian), and Late Jurassic faunas have been correlated with those of the New Zealand Murihiku Terrane (Campbell et al., 1985; Meffre, 1995). A fossiliferous succession at Pouembout, formerly correlated with the Late Jurassic, is now considered Early Cretaceous (Adams et al., 2009).

### Teremba Terrane

The Teremba Terrane comprises a succession of very low-grade (zeolite facies), Late Permian–mid-Jurassic, shallow-water, volcaniclastic (calc-alkaline, island arc-derived, andesitic) sedimentary rocks and volcanics (andesites, dacites and rhyolites). The sedimentary rocks are typically medium grained greywackes with only minor (<10%) intercalated argillite, some shallow water volcaniclastic conglomerate and rare black shale, a few tens of metres thick. The mineral, geochemical and isotopic composition of greywackes is closely similar to that of the Koh-Central Terrane (Adams et al., 2009), similarly lack plutonic clasts and most probably come from the same source. In contrast with the Central Terrane, this terrane contains abundant faunas resembling those of the Murihiku Terrane of New Zealand (Grant-Mackie et al., 1977; Paris 1981; Campbell, 1984; Ballance and Campbell, 1993).

### The Boghen Terrane (the "ante-Permien" of Paris, 1981)

The Boghen Terrane is a subduction complex comprising schistose unfossiliferous, volcano-sedimentary rocks (pillow basalts, chert,

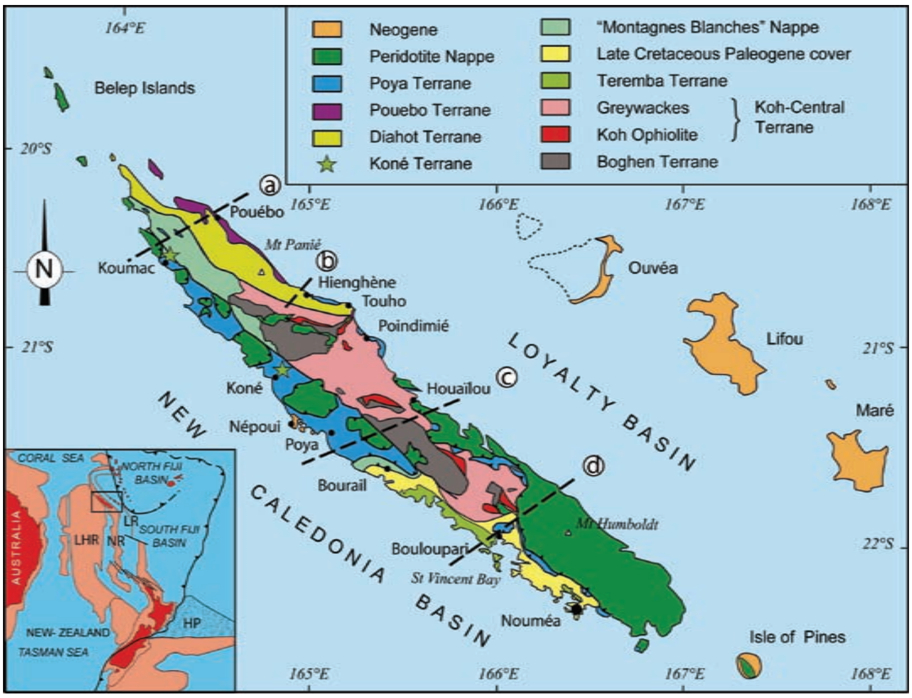


*Figure 5  Tectonostratigraphic terrane map of New Caledonia (after Cluzel et al., 1999; Maurizot and Vendé-Leclerc, 2009). For a more detailed geological map of New Caledonia, see: http://dimenc.gouv.nc/portal/page/portal/dimenc/librairie/documents/telechargement/NC_1000000_A1.pdf).*

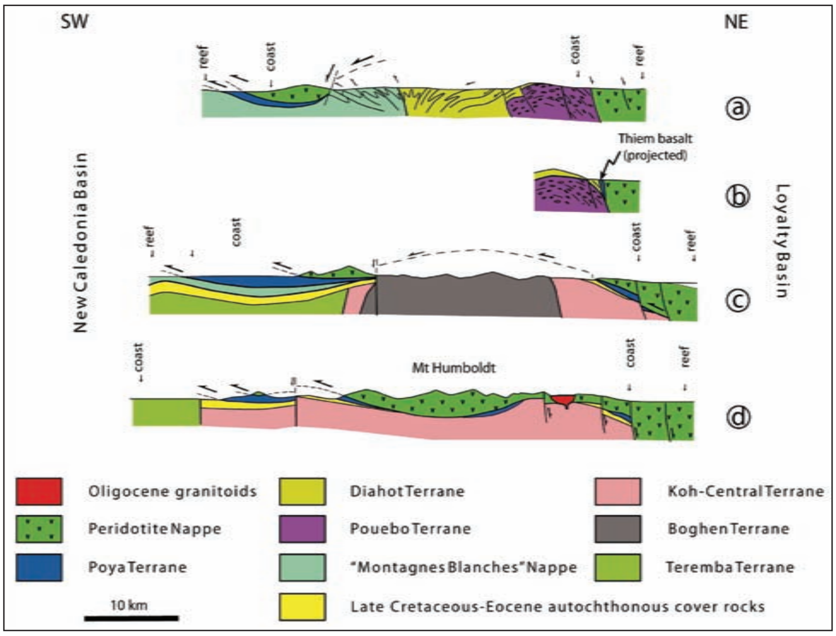


*Figure 6 Simplified cross sections of New Caledonia to show the "basement" terranes overlain by Late Cretaceous to Eocene unconformable sedimentary cover, in turn overthrust by mafic (Poya Terrane) and ultramafic (Peridotite Nappe) allochthons (after Cluzel et al., 2001). For location, see Figure 5.*

black shale, sandstone, tuffs, turbiditic greywackes, mafic melange and peridotite/serpentinite), at a metamorphic grade (lower greenschist to blueschist facies) that is notably higher than the adjacent terranes. Late Jurassic metamorphic ages (c. 150 Ma, whole-rock K-Ar) of the blueschists and metabasalts (Blake et al., 1977) suggest a minimum mid-Jurassic age for this terrane. However, Early Jurassic (Cluzel and Meffre, 2002) and more recently Early Cretaceous (c. 135 Ma) detrital zircon ages (Adams et al., 2009) set a maximum Early Cretaceous depositional age for the original sediments (see also Bryan et al., 2012).

## Reconstruction of the Southeast Gondwanaland active margin in New Caledonia

Because Triassic–Early Cretaceous shallow-water volcaniclastic sediments (Teremba) occur to the W, and deeper-water sediments (with the same origin; Koh-Central) to the E and, volcanic rocks and shallow intrusions are absent in the Koh-Central Terrane, Meffre (1995) and Cluzel and Meffre (2002) have suggested that the Teremba and Central terranes are the onshore and offshore parts of the same fore-arc basin respectively. This view is also supported by a westwards-increasing metamorphic gradient in the HP-LT Boghen Terrane (Guérangé et al., 1975; Paris, 1981), thus suggesting westwards-dipping Mesozoic subduction. At present, the corresponding Permian-Mesozoic volcanic arc is likely buried below younger sediments of the Lord Howe Rise (Figure 1).

In contrast with New Zealand where Mesozoic terranes derive at least partly from the erosion of mainland Australia, the Permian–Mesozoic volcaniclastic sedimentary rocks in New Caledonia mainly come from an intra-oceanic volcanic arc and contain unimodal zircon populations with about the same age as the enclosing sediment (Adams et al., 2009). This view is supported by the endemism of Triassic faunas and floras that is shared with New Zealand easternmost terranes. However, a contribution of continental rocks is present in sandstones associated with black shales that are intercalated at two levels of the sequence in the mid-Triassic and Late Jurassic. To account for zircon provenance, geochemical and isotopic features of volcanic and volcaniclastic rocks, and a possible northern origin of Koh Ophiolite rocks recorded by paleomagnetic data (Ali and Aitchison, 2000) as well, it has been suggested that proto-New Caledonia formed the eastern edge of a volcanic-arc and marginal basin system along the SE Gondwanaland margin. The marginal basin (in which deep-sea fans fed by mainland Australia accumulated) opened during the Permian-Triassic and closed obliquely during the Jurassic–Early Cretaceous to reach its present location marked by the Dun Mountain Suture (Adams et al., 2009).

## Late Cretaceous–Eocene phase: from marginal rifting to obduction

### Late Cretaceous: the marginal rifting

Overlying the three above-mentioned terranes with angular unconformity, there is a prominent Late Cretaceous (Coniacian–Campanian) (Paris, 1981), volcano-sedimentary unit (classically referred to as "Formation à charbon"), which is composed of fining upwards marine shallow water sandstone, coal-bearing siltstone, tuffs and volcanic rocks (Figure 7) that accumulated in a tidal-zone or a near shore deltaic environment. The Late Cretaceous marine siltstones contain endemic faunas (ammonites and inocerams) (see Paris, 1981) that indicate an isolation from Australia, which is confirmed by the local provenance of detrital zircon populations (Cluzel et al., 2011). The pre-Coniacian unconformity post-dates the final amalgamation of the three aforementioned terranes. Exhumation of high-pressure metamorphic rocks of the Boghen Terrane thus occurred between the Barremian (c. 130 Ma; Adams et al., 2009) and the Coniacian (c. 89 Ma). Mafic and felsic volcanic rocks and tuffs occur near the base of the Formation à charbon (pre-Campanian), U-Pb dating of zircons extracted from a rhyolite flow there (88.4 Ma; Nicholson et al., 2011) confirms the Coniacian age of the Late Cretaceous transgression. The geochemical features of Late Cretaceous volcanic rocks contrast with that of pre-Late Cretaceous terranes. Whilst the latter are clearly related to an intra-oceanic arc, the Coniacian-Santonian basalts and felsic volcanics of the former display evidence for subduction affinity and intraplate magmatism as well, with some transitional features. The volcanic rocks have been interpreted as a result of regional W-dipping subduction (Nicholson et al., 2011); however, the occurrence of contemporaneous bimodal arc- and rift-related volcanic rocks, and a mixing trend between them is a common feature of rifted active margins in which mafic and felsic magmas are produced by partial melting of a previously metasomatised mantle wedge and the lower continental crust respectively (Bryan et al., 1997; Cluzel et al., 2011). The end of volcanic activity in New Caledonia which coincides with Maastrichtian–Paleocene thermal subsidence (Aitchison et al., 1995a) is consistent with the latter interpretation.

The magmatic activity ceased before the Campanian and is thus restricted to the 89–83.5 Ma interval, i.e., magmatic activity ceased

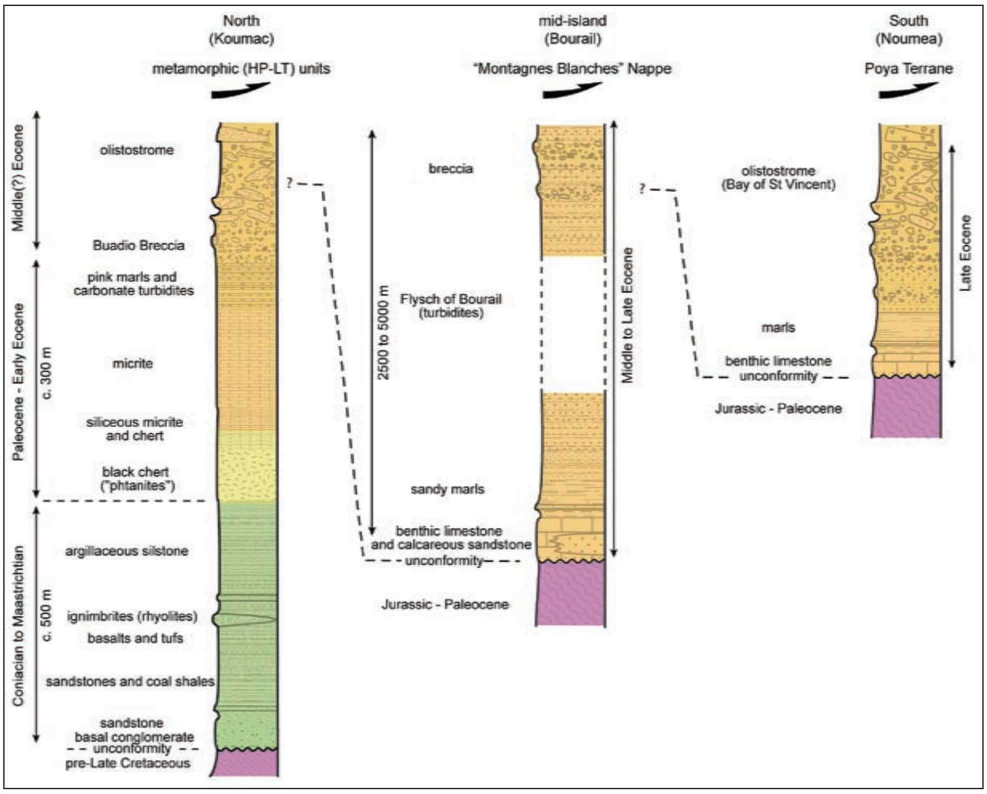


***Figure 7 A comparison of Eocene sedimentary evolution in northern, mid- and southern New Caledonia to show the southward migration of the Eocene unconformity and syntectonic foreland basins (after Cluzel et al., 2001).***

when the Tasman Sea (Hayes and Ringis, 1973) and South Loyalty marginal basins (Cluzel et al, 2001) opened. Meanwhile, sedimentation evolved from shallow-water marine sandstone and siltstone toward pelagic siliceous pelite (Maastrichtian), and micrite (Paleocene–Early or mid-Eocene depending upon diachronous Eocene pre-obduction events). This environmental change may be related to the thermal subsidence that followed the marginal rifting and complete submersion of New Caledonia (Aitchison et al., 1995a; Cluzel et al., 2011), a similar evolution is recorded from the Lord Howe Rise that was above sea level until the Maastrichtian and thereafter was the locus of pelagic siliceous sedimentation (McDougall and Van Der Lingen, 1974).

In northern New Caledonia, development of the Eocene high-pressure metamorphic complex (see below) has dramatically erased most of the primary features of the protoliths. However, the bulk of the Late Cretaceous–mid-Eocene sequence, although metamorphosed to eclogite or blueschist facies, is similar to the unmetamorphosed sequence of the Noumea area; although Late Cretaceous carbonaceous sediments there are considered more distal (Maurizot et al., 1989).

A triple provenance is necessary to account for the composition of Coniacian–Santonian sandstones. Contemporaneous volcanic activity provided most of the felsic component and rare volcanic zircon of the same age; a prominent mid-Cretaceous population (110–95 Ma) is puzzling because rocks of this age are extremely rare in New Caledonia (Cluzel et al., 2010), unless they were completely eroded away before the Late Cretaceous; the older part of the detrital zircon age spectrum very closely resembles those of the sandstones of the Boghen and Koh-Central terranes and has been provided by the erosion of directly underlying basement rocks. Therefore, at variance with previous interpretations, there is no need to advocate an Australian provenance to account for the occurrence of Precambrian detrital zircons (Aronson and Tilton, 1971; Aitchison et al., 1998). New Caledonia was already isolated from Australia by the Late Cretaceous, a feature consistent with the development of faunal and floral endemism at that time (Cluzel et al., 2011).

## Paleocene–Early Eocene: long-term subsidence and pelagic sedimentation

Paleocene black cherts and minor argillite that transitionally overlie the Formation à charbon are classically referred to as "phtanites" in New Caledonia. These pelagic sediments are formed of cryptocrystalline silica, sponge spicules, radiolarians, and rare plagioclase grains. Phtanite deposition marks the end of terrigenous inputs and a complete submersion of New Caledonia; it originally formed at a low latitude and in a cool climate. Progressive global warming (Zachos et al., 2001) and northward drift during the Paleocene probably drove New Caledonia progressively out of the southern circum-polar siliceous ooze belt and allowed siliceous micrite to form; thereafter, the amount of silica progressively decreased and pure micrite accumulated during the Eocene (Figure 7). During that period, the New Caledonia Ridge was a stable isolated plateau in relatively deep waters, a situation that drastically changed during the late Early–Late Eocene, when intra-basinal breccias diachronously appeared and migrated southwards, and New Caledonia emerged again (see below).

## Eocene foreland basins

In the areas where an Eocene sequence (Figure 7) has been preserved, in the N of the island (near Koumac) and in olistoliths of the Baie de St Vincent Olistostrome (south), the pelagic white micrite abruptly changes upwards into late Ypresian (c. 50 Ma) pink marls (Maurizot, 2011). This change records input of illite and minor hematite in the basin, and likely signals local emersion and weathering. Along the Koumac-Ouegoa section in the N of the island, SW verging stacked and isoclinally folded units result from in-sequence duplexing of the Late Cretaceous–Eocene sedimentary cover; therefore, the stacked units record a NE-ward change in basin geometry. The pink marls locally show rhythmical bedding and evidence for turbiditic sedimentation that signals the appearance of a slope in the basin; upwards, i.e., NE-wards in the stacked units, increasing intrabasinal instability is recorded by the occurrence of monogenetic limestone breccia, that rework the pink marl and white micrite as well. This breccia is quickly followed by intraformational polygenic breccia that mainly reworks the immediately underlying limestone. Finally, the Buadio breccia, which is composed of limestone and phtanite clasts changes upwards into an olistostrome formed of 100 m- to 1000 m-scale olistoliths enclosed in a matrix of coarse conglomerate. The evolution of clasts provenance and size records the progressive involvement of deeper (i.e., older) parts of the sedimentary cover in a series of short-lived syntectonic basins located in front of SW- or SSW-prograding thrusts (Maurizot, 2011).

In contrast, in the S of the island (from Bourail to Noumea), unconformable shallow water limestone forms the base of a mid- to upper Eocene turbidite sequence that rests upon eroded older rocks (Figure 7). Pre-mid–Late Eocene erosion has been interpreted as a consequence of emersion due to fore-arc bulge (Cluzel et al., 1998) and foreland basin inception. The basal limestone youngs southward from Early to Late Eocene (Figure 7), a feature consistent with the progressive involvement and northward subduction of the northern tip of Norfolk Ridge (Cluzel et al., 2001). The 3–5 km-thick turbidite sequence that overlies the basal limestone is referred to as "Bourail Flysch" (Paris, 1981). The arenites and breccias are mainly composed of upwards coarsening clasts of various lithologies that include easily recognisable angular fragments of Paleocene limestone and black chert (phtanite). In contrast with breccias of the northern units, which only contain elements derived from the autochthonous sedimentary cover, the Bourail Flysch contains an upwards increasing amount of mineral and lithic clasts derived from a mafic source. In the upper third of the turbidite sequence, arenites may contain up to 50% clinopyroxene clasts. The geochemical features of mineral (clinopyroxene) and lithic clasts (basalt, dolerite and red chert) are closely similar to those of Poya Terrane rocks, which are therefore the most likely source (Cluzel et al., 2001). The top of the Bourail Flysch is an olistostrome, which in turn is overthrust by the allochthonous Poya Terrane (see below).

Thus, southward younging, syntectonic Eocene sequences record the in-sequence overthrusting of the Late Cretaceous–Early Eocene sedimentary cover of the northern Norfolk Ridge, and in southern units, the final out-of-sequence overthrusting of the Poya Terrane (Figure 8). It is worth noting that no ultramafic clasts have been recorded in the foreland basin; it clearly remained out of reach of ultramafic detritus issued from the fore-arc mantle lithosphere (i.e., the future Peridotite Nappe) that was obducted later.

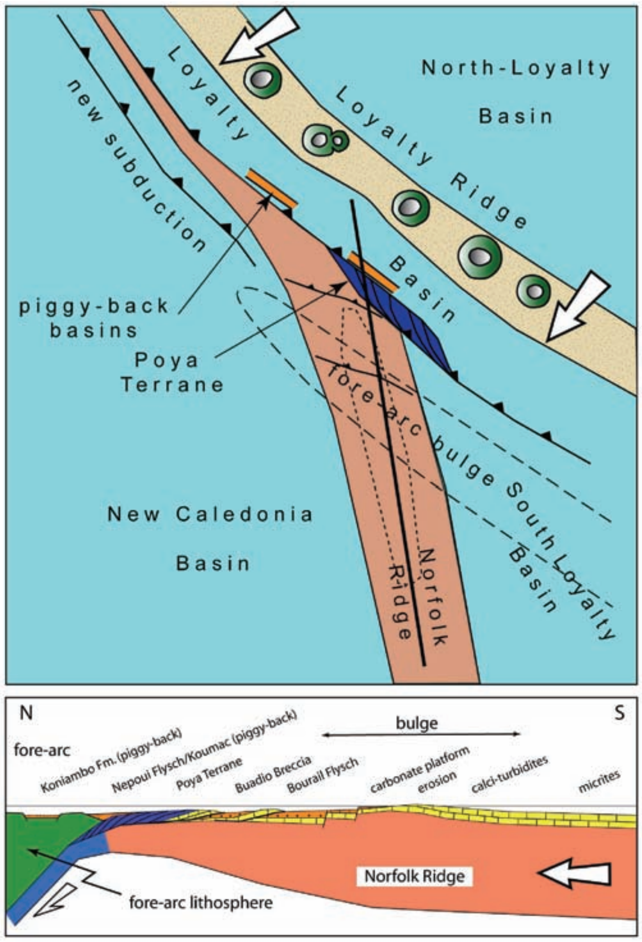


***Figure 8** Tentative paleo-geographic reconstruction of New Caledonia and the Loyalty Ridge during Eocene time, to show the development of foreland and piggy-back basins, and southward migration of sedimentary facies during the pre-obduction period.*

## The subduction-obduction complex

The occurrence and tectonic features of ultramafic and mafic allochthons, age migration of foreland basins, polarity and exhumation kinematics of the HP-LT metamorphic complex, all suggest that Late Eocene obduction was preceded by N- or NE-dipping subduction.

### Peridotite Nappe

From a geological and an economic point of view as well, the Peridotite Nappe (Avias, 1967) is certainly the most prominent terrane of New Caledonia as it represents more than 25% of the surface of the Grande Terre and is of major economic importance for the country. This dominantly ultramafic terrane consists of a southern unit, the "Massif du Sud" and a number of tectonic klippes spread along the W coast of the island (Figures 5 and 6). This unit is dominantly formed of upper mantle rocks (harzburgite and rare lherzolite) with minor ultramafic (pyroxenite, wehrlite and dunite) and mafic (layered gabbro) cumulates (Prinzhoffer, 1981). A high temperature foliation associated with dunite/harzburgite layering exists throughout the terrane, and it generally dips gently (i.e., <20°) forming broad gentle folds. This foliation locally bears a high-temperature mineral/stretching lineation oriented N-S; it is marked by the preferred orientation of orthopyroxene in harzburgite, and chromite aggregates in dunite (Prinzhoffer et al., 1980). This lineation associated with a gently dipping foliation probably results from the detachment-like tectonics that occurs at the base of the newly formed lithosphere (the 1,200°C isotherm) close to the oceanic ridge, as a consequence of differential motion between lithosphere and asthenosphere. Therefore, owing to the average N-S trend of the stretching lineation, the oceanic ridge that generated the ophiolite is thought to have had an E-W trend.

The upper mantle and lower oceanic crust rocks reflect a complex history that includes several stages of melting, rock-melt interaction, and re-melting (Marchesi et al., 2009; Ulrich et al., 2010) that finally lead to an extreme overall depletion. Therefore, age constraints on ophiolite formation are scarce, and except a few poorly reliable K-Ar ages (120–50 Ma; Prinzhoffer, 1981), the only direct dating has been attempted by Prinzhoffer (1987), who provided a 131±5 Ma age based on a Sm-Nd rock-mineral isochron from a gabbro of the Montagne des Sources (Massif du Sud). An Early Cretaceous age is problematical considering the geodynamic setting of the SW Pacific at that time and requires validation by other methods.

The Peridotite Nappe is crosscut at all levels by Early Eocene basalt (dolerite), micro-diorite and diverse felsic dykes emplaced within a narrow span of time (55–50 Ma; U-Pb dating of zircon) that suggest a minimum Late Paleocene age for the oceanic lithosphere (Cluzel et al., 2006). Dolerite dykes are dominantly IAT-like and probably represent the youngest product of magmatic-arc activity (c. 50 Ma, whole rock K-Ar; Prinzhoffer, 1981). Microdiorite results from the hydrous melting of a similar supra-subduction source whereas most felsic dykes display the geochemical features of slab melts (Cluzel et al., 2006). Thus, a transient thermal pulse in the fore-arc region during earliest Eocene time probably generated this short-lived magmatic event with arc and fore-arc magmatic affinities.

Locally, amphibolite lenses, c. 200 x 10–50 m, appear at the base of the serpentinite sole, above Poya basalt. Back-arc basin basalts (BABB) geochemical composition and $^{40}Ar$-$^{39}Ar$ hornblende cooling ages at 56 Ma suggest that these amphibolites originated from the Poya Terrane basalts at, or near a spreading ridge at subduction inception, a feature consistent with synchronous formation of the slab melts mentioned above (Cluzel et al., 2012).

Partial serpentinisation (20–60%) appears throughout the Peridotite Nappe, it is generally interpreted as a consequence of cooling and low temperature hydration of the oceanic mantle lithosphere; in addition, a more extensive serpentinisation occurs in the tectonic sole with the development of porphyroclastic mylonite 20–200 m thick, which likely formed during obduction, and locally indicates SW-directed shearing. In general, the Peridotite Nappe overlies the Poya Terrane and the autochthonous rocks above a sub-horizontal fault (Figure 6), with the very simple structure sharply contrasting with the complexity of underlying terranes.

## Poya Terrane

The Poya Terrane is an allochthonous set of m- to km-scale upright tectonic slices of pillow and massive basalt (99% of the terrane) associated with thin inliers of bathyal sediment (red, black and green chert and argillite); it is always located below the Peridotite Nappe (Figures 1 and 2). The main body of this unit is located along the W coast where it is 10 km wide, and extends over a length of 250 km from Bourail to Koumac; smaller and less continuous units are located along the E coast (Figures 5 and 6).

Based upon paleomagnetic evidence, these rocks are thought to have erupted to form an oceanic floor (South Loyalty Basin; Cluzel et al., 2001) to the NE of New Caledonia, at a latitude c. 300 km to the N of its present location (Ali and Aitchison, 2000). Inliers of bathyal sedimentary rocks, which are closely associated with pillow basalts contain Campanian–latest Paleocene or earliest Eocene radiolarians (Aitchison et al., 1995b; Cluzel et al., 2001). Poya Terrane basalts are dominantly undepleted MORB (E-MORB), which represent remnants of an oceanic or marginal basin crust (Cluzel et al., 2001). Subordinate BABB and intra-oceanic alkaline basalts (OIB) display carbonate interpillow material with Late Paleocene–Early Eocene microfaunas. These basalts are highly vesicular, and often reddened; therefore, they probably represent seamount lavas erupted upon Late Cretaceous-Paleocene oceanic crust. K-Ar whole rock apparent ages that range from c. 61–38 Ma (Guillon and Gonord, 1972; Eissen et al., 1998), are young relative to fossil ages and are thus probably meaningless. These apparent ages probably reflect a complex evolution due to both excess argon input due to fluid-rock interaction during lithosphere cooling, and diverse thermal resetting phases.

Except in its northernmost occurrence along the E coast (Thiem basalt, SE of Hienghene) where blueschist facies minerals scarcely appear (Meffre, 1995), the Poya Terrane is not regionally metamorphosed and does not display any ductile deformation. The local occurrence of lower greenschist or zeolite facies mineral associations is likely due to intra-oceanic water-rock interaction referred to as ocean floor metamorphism (Nicholson et al., 2000). Slices of Poya Terrane rocks have been first scraped off the down-going plate during E-dipping subduction of the South Loyalty Basin, then accreted in a fore-arc region (see below) and thereafter thrust onto New Caledonia during the Late Eocene, prior to the obduction of ultramafic rocks (Cluzel et al., 2001). Accretion of km-scale slices of coherent crustal material may be related to the Cordilleran-type ophiolite formation.

According to analog experiments for material transfer in accretionary wedges (Gutscher et al., 1998) the peeling of the upper part of the down-going slab remains a possibility, which may result either from an extreme roughness of the down-going plate, or more probably from a very shallow dip of the subduction zone. Off-scrapping of the Poya Terrane in the Loyalty fore-arc probably took place because subduction started at, or near the oceanic ridge (Ulrich et al., 2010; Cluzel et al., 2012) and was buoyant at least at the beginning.

Small-scale piggy-back basins, represented by the Nepoui Flysch and Koumac Olistostrome are closely associated with the Poya basalt. Both basins are pinched between the Poya Terrane and the Peridotite Nappe. The Nepoui Flysch starts with 2–5 m-thick biocalcarenite of uncertain age (Bartonian–Priabonian; Meffre 1995; Cluzel, 1998) that rests directly upon serpentinite. Significantly, the basal limestone contains large benthic foraminifers, and some detrital serpentine and chromite grains. It is overlain by pale brown argillite, and alternating coarse arenite and dolomicrite. On top of arenite beds, thin layers or lenses of reworked red argillite very closely resemble those of the Poya Terrane. Arenite clasts are derived from three distinct sources: Poya Terrane basalt (clinopyroxene, ilmenite and magnetite), an adjacent shallow water platform (carbonate bioclasts), and serpentinite. It is worth noting that the arenites contain no fresh peridotite rock or mineral clasts such as chromite. The dolomicrite beds, 2–10 cm thick, have provided a Late Eocene (Priabonian) pelagic microfauna (Maurizot, unpublished data, 2009).

The Koumac Olistostrome, c. 300 m thick, is composed of several mass flow breccia units, 5–10 metres thick, composed of basalt boulders embedded in a matrix of sandy breccia made of pillow and chert fragments with some lenses of red blocky argillite. Basalt fragments display the same geochemical features as Poya basalt (E-MORB; Cluzel et al., 2001). Occasionally, decimetre-sized clasts of felsic and amphibole rich magmatic rocks and scarce serpentinite appear in the breccia. These rocks closely resemble those of the Early Eocene dyke system of the Peridotite Nappe. On top of the mass flow unit, 50 m of whitish sands occur but contain a few basal basalt boulders. One single dolomicrite horizon, c. 3 metres thick, appears within the mass flow units which therefore resemble the Nepoui Flysch in terms of clast provenance and basinal conditions.

Both the Nepoui Flysch and Koumac Olistostrome may have accumulated in piggyback basins located near the boundary between the Poya Terrane and exhumed serpentinites of the fore-arc region (Figure 8).

## Kone Terrane

Discrete slices of a parautochthonous Late Cretaceous hemi-pelagic unit composed of siltstone, argillite and chert, referred to as

"Formation de Koné" (Carroué, 1972; Paris, 1981) occur near Koné and Koumac (Figure 5). The unit has no known basement, contains no carbonaceous material, and has more distal facies than the typical autochthonous rocks of the same age. Currently, it is situated above the Montagnes Blanches Nappe (see below), Paleocene autochthonous sedimentary rocks and below the allochthonous Poya Terrane and everywhere has faulted boundaries. It possibly accumulated originally on a continental slope, or directly upon older oceanic crust, to the NE of New Caledonia and was picked up and thrust upon New Caledonia by the Poya Terrane, and finally overthrust by the latter.

## The "Montagnes Blanches" Nappe

The "Montagnes Blanches" Nappe (Maurizot, 2011) is systematically intercalated between the top of the Paleogene flysch and the overlying obducted ophiolite units throughout the Grande Terre. It consists of Late Cretaceous black argilites, cherts, overlain by Paleocene micrites passing upwards into Early Eocene calci-turbidite. It represents an unmetamorphosed lateral equivalent of the Diahot Terrane. Named after the range marking the northern flank of the Bourail anticline (Figure 5), where it is particularly well exposed, this unit is parautochthonous and rooted in the Koumac area and allochthonous farther south. In the Bourail and Nouméa areas, the uppermost levels of the flysch contain elements of this allochthonous sedimentary unit broken up into olistoliths, as well as elements of the Poya Terrane.

## Eocene metamorphic complex

The HP-LT metamorphic complex in the N of the island comprises the metamorphic Diahot and Pouebo terranes (Figure 1). In spite of a much higher metamorphic grade (blueschist and eclogite), the Diahot Terrane rocks do not greatly differ from the Cretaceous–Eocene sedimentary cover, although they have a more distal character during the Late Cretaceous. It shows the same Cenozoic sedimentary sequence except that the Eocene turbidite-olistostrome is dominantly composed of breccias, which started earlier than in the S (late Ypresian; c. 50 Ma; Maurizot, 2011), and characteristically do not contain any mafic rock. The Pouebo Terrane is a subduction melange composed of 0.1–100 m-scale boulders of mafic rocks derived from the Poya Terrane (Cluzel et al., 2001; Spandler et al., 2005), embedded in a meta-serpentine (talc-schist) or metasedimentary matrix. The Diahot Terrane rocks have been subducted at depth c. 50 km (1.7 GPa - 550°C) (Fitzherbert et al., 2003; 2005) whereas eclogites in the Pouebo Terrane have metamorphic mineral associations which suggest a much higher grade and maximum burial depth of c. 80 km (2.4 GPa - 650°C) (Clarke et al., 1997; Carson et al., 1999, 2000). At present, the Pouebo Terrane, which was subducted at greater depth, appears in the core of a regional-scale foliation antiform wrapped by the Diahot Terrane (Figure 6) as shown by the occurrence of E-dipping Diahot carbonaceous schists, siliceous schists (metamorphosed "phtanites") and recrystallised limestones along the eastern shoreline between Hienghene and Touho (Maurizot et al., 1989; Cluzel et al., 1995a, b, 2001; Rawlings and Lister, 1999, 2002). In turn, the eastern flank of the antiformal structure is overlain by slices of the Poya Terrane (Thiem unit; Maurizot et al., 1985; Cluzel et al., 2001; Figures 5 and 6) and finally by the Peridotite Nappe, which is rooted in the Loyalty Basin (Collot et al., 1987). The regional-scale antiformal structure of the metamorphic complex may be due either to a late folding of the tectonic pile (Rawlings and Lister, 2002); or alternatively, to the exhumation of the high grade eclogitised melange (Pouebo Terrane) that dragged the overlying rocks (Diahot Terrane) towards the surface (Figure 9; Cluzel et al., 1995a, b; Baldwin et al., 2007). Along the Koumac-Pouebo section, i.e. from SW to NE across the island, the metamorphic grade changes abruptly in connection with SW-verging ductile thrusting; in contrast, it evolves much more gradually along strike, and blueschist facies rocks crop out to the SE of Poindimié (E coast; Figure 5). Therefore, exhumation of HP-LT rocks was likely to occur in a scissors- or fan-like fashion.

Thermochronological and radiochronological data allow the exhumation of the HP-LT complex to be better constrained. Eclogites of the Pouebo Terrane started to exhume at c. 44 Ma (U-Pb age of zircon overgrowths; Spandler et al., 2005); whilst zircons of Diahot Terrane rocks display younger overgrowths at c. 38 Ma (Cluzel et al., 2010). The closing temperature of phengite (450°C) in both terranes is recorded by $^{40}Ar$-$^{39}Ar$ thermochronology at c. 36 Ma (Ghent et al., 1994; Baldwin et al., 2007). Exhumation at c. 7 km depth occurred at c. 34±4 Ma as shown by apatite fission track data (c. 85°C; Baldwin et al., 2007; Figure 9).

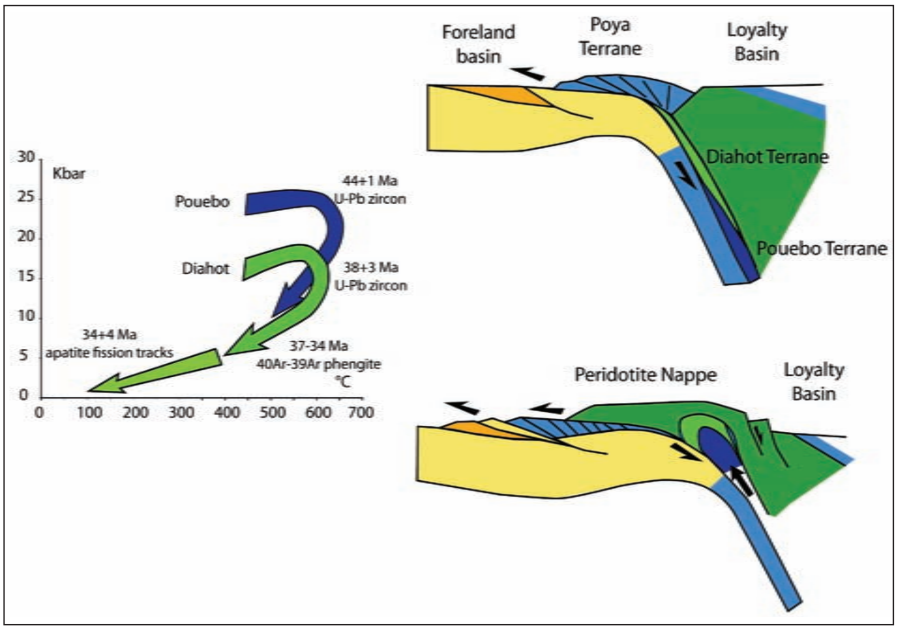


***Figure 9** A conceptual model for the subduction and exhumation of the HP-LT terranes of New Caledonia to account for the P-T-t paths of Pouebo and Diahot terranes.*

## Loyalty Ridge

The geology of the Loyalty Ridge is still poorly known due to the lack of basement outcrops and a thick carbonate cover. It may be extended northwards into the Southern D'Entrecasteaux Zone and is connected southwards to the Three Kings Ridge beyond the Cook Fracture Zone (Kroenke and Eade, 1982). Geophysical and swath bathymetry data indicate that this ridge is formed of seamounts, the size and spacing of which are similar to those of most island arcs, (Bitoun and Récy, 1982; Lafoy et al., 1996). Seamounts are overlain by Miocene limestone plateaus, which are in turn capped and fringed by Pliocene–Recent coral reefs. The Bougainville Seamount located at the eastern end of the South d'Entrecasteaux Ridge, has been drilled (ODP 831), and Eocene andesite was found underneath 700 m thick Oligocene to Holocene limestone (Dubois et al., 1988; Greene et al., 1994). In the North Loyalty Basin, at the DSDP 286 site, c. 500 m-thick Middle–Late Eocene andesitic volcaniclastic turbidites have been drilled (Andrews et al., 1975). Because of these features, and in spite of lack of outcrop evidence, the Loyalty Ridge should be considered as an Eocene island arc and the North Loyalty Basin the associated back-arc basin (Maillet et al., 1983; Cluzel et al., 2001; Schellart et al., 2006; Paquette and Cluzel, 2007; Whattam et al., 2008; Whattam, 2009). Intra-oceanic alkaline (OIB; Baubron et al., 1976; Maurizot and Lafoy, 2003) basalt and dolerite that crop out on Mare Island were thought to represent the volcanic basement of the island (Rigolot, 1988). However, intraplate volcanism is younger that some of the carbonate cover and not older than Middle Miocene (Maurizot and Lafoy, 2003). In addition, the Late Miocene (11 Ma; Baubron et al., 1976) dolerite and basalt dykes that appear at the surface crosscut the limestone and develop contact metamorphism (Chevallier, 1968). Therefore, these basalts are unlikely to represent the basement of Loyalty Ridge; instead they may be part of a N-S trending hot spot trail and younger than the Loyalty Ridge itself (Meffre, 1995). The Loyalty Ridge is at present involved in the fore-arc bulge of the Vanuatu Arc which results in diachronous emersion and uplift of the Loyalty Islands (Dubois et al., 1974). Incipient oblique collision with the arc/trench system (Lafoy et al., 1996) results in the development of a complex fracture set (Bogdanov et al., 2011).

## Oligocene–Present: the post-obduction phase

### Late Oligocene post-obduction plutonism

In southern New Caledonia, post-obduction km-size plutons are intruded into the ultramafic allochthon and its autochthonous basement as well. U-Pb dating of magmatic zircons provided 27.5 and 24 Ma (Late Oligocene) for St Louis granodiorite (near Noumea) and Koum-Borindi adamellite (E coast) respectively (Paquette and Cluzel, 2007) (Figure 6c). The high-K to medium-K calc-alkaline granitoids display the geochemical and isotopic features of volcanic-arc magmas uncontaminated by crust-derived melts. These magmas were possibly generated during Oligocene subduction of the oceanic part of the northern New Caledonia Basin (Cluzel et al., 2005) that may also account for the north-eastward dip of the Moho (Klingelhoefer et al., 2007) (Figure 10). Although no alternative interpretation has been proposed for these granitoids, this model does not meet a general agreement (e.g., Sutherland et al., 2010). Strontium, Nd and Pb isotopic ratios indicate derivation from an isotopically homogeneous mantle wedge; but some variation in trace-element ratios uncorrelated to differentiation, are diagnostic of source heterogeneity. Prominent HREE depletion of some of the younger granitoids may be due to an equilibrium with garnet-bearing sub-crustal material (granulite) found as xenoliths (Paquette and Cluzel, 2007); whilst a relative Nb, Ta and Hf enrichment, irrespective of crystal fractionation, may be related to either a modest contamination

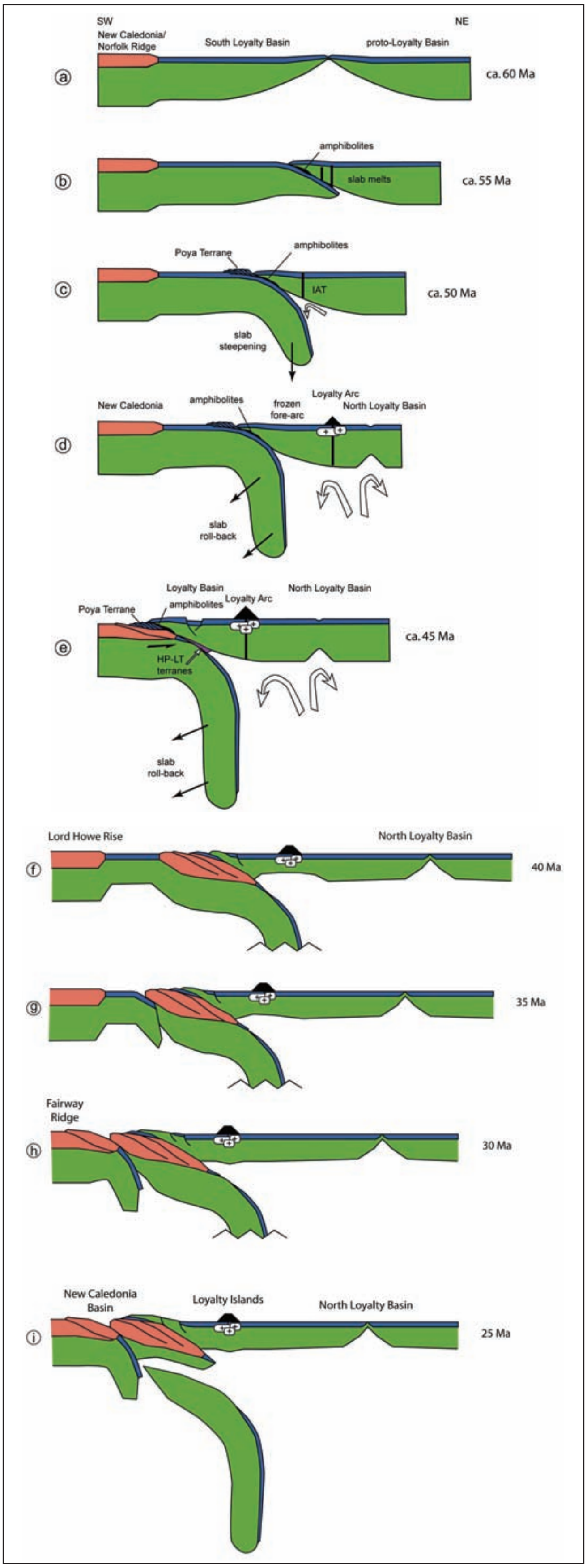


***Figure 10 A model of lithospheric evolution of New Caledonia from Late Paleocene subduction inception to Late Oligocene slab break off.***

by previously underplated mafic material, heterogeneous hydration of the mantle wedge, or mixing with uplifted Nb-rich mantle. Post-obduction slab break-off is advocated to account for sub-lithospheric mantle mixing and subsequent heterogeneity (Cluzel et al., 2005).

## Supergene evolution: a polyphase weathering history

The emersion of New Caledonia during, or soon after obduction prevented marine sediments to accumulate except in coastal areas; instead, thick regolith developed and probably covered the whole island. Weathering profiles that develop upon peridotites are important not only because they record tectonic stability and hence are good proxies for regional dynamics, but also because they are important Ni resources. The regolith is comprised of several stepped planation surfaces (Wirthmann, 1966, 1970); the oldest surface being the most elevated (Trescases, 1975; Latham, 1986). Each step of surface development was related to repeated uplift phases and/or sea level changes (Chardon and Chevillotte, 2006; Chevillotte et al., 2006).

Paleomagnetic dating methods have been successfully applied to date ferricrust development over obducted peridotites. This method involves a comparison of remanent magnetisation preserved in Fe oxides to the apparent polar wandering curve of Australia, assuming that Australia and New Caledonia have been rigidly fixed since the Early Oligocene. Preliminary results reveal that a phase of ferric crust development probably ended at c. 25±5 Ma (Late Oligocene); the associated weathering surface is located on the Tiebaghi Massif (near Koumac; Figure 5), one of the northern klippes of the W coast. This surface has been deeply eroded and dissected; whilst in the rest of the island and especially in the southern lowlands and endorheic basins, weathering continued until Recent time (Sevin et al., 2011).

## Early Miocene coastal deposits and vigorous uplift

Miocene sediments crop out in cliffs of the Mueo and Pindaï peninsulas and islets of the Nepoui Bay (Coudray, 1976; Figure 5); these are the only known Miocene marine sediments of this age in New Caledonia. The Nepoui series consists of two subunits, the lower unit, the base of which does not crop out, consists of lagoonal limestone and calcareous sands that contain Aquitanian (Early Miocene) foraminifers (Maurizot, unpublished data, 2011). The upper unit, c. 100 m thick, has an erosive lower boundary and starts with slightly unconformable torrential cobble conglomerate, c. 100m thick, overlain by reddish to yellow calcareous sands with conglomerate lenses, topped by pre-Late Miocene yellow calcareous sands (Coudray, 1976). In the Mueo Peninsula, gently S-dipping conglomerate and coarse sands overlie severely sheared serpentinite, which represents the tectonic sole of the Ophiolitic Nappe, Poya Terrane basalts and the folded Late Eocene Népoui Flysch, with angular unconformity. Well-rounded pebbles are mainly composed of peridotite, minor dolerite and rare amphibole-bearing felsic rocks, which belong to the Early Eocene dyke system of the Peridotite Nappe (Cluzel et al., 2006). The most significant feature of Early Miocene conglomerate is the occurrence of, pebbles and sand-sized particles of ferricrete and silcrete that reflect the erosion of older regolith. The Goa N'doro Formation which crops out on the E coast near Houaïlou (Figure 5) is formed of two sub-units (Orloff and Gonord, 1968), the younger of which is composed of a conglomerate similar to that of Nepoui. This conglomerate has been tentatively correlated to the Late Oligocene by Chardon and Chevillotte (2006), but could be a lateral equivalent of the Early Miocene Nepoui conglomerate as well.

The occurrence of Nepoui conglomerate and its possible E coast correlative, and the new time constraints on regolith formation infer that (1) peridotite weathering started soon after obduction (a fact signalled by Coudray, 1976), (2) this stage of regolith development stopped locally at c. 25±5 Ma and (3) this event may be roughly correlated with the occurrence of Early Miocene dramatic erosion that may be due to either climatic (sea level drop) and/or tectonic (uplift) events.

## Neogene–Recent tectonics and morphology of New Caledonia

Late Oligocene granitoids (Cluzel et al., 2005), Early Miocene sediments (Chardon et al., 2008), and regolith, including re-sedimented laterite deposits (Cluzel and Vigier, 2008; Chardon and Chevillotte, 2006) are crosscut by fault sets. These tectonic events may be due to Miocene uplift, but time constraints are not always available and additional work is necessary to better constrain this fault set that partly controls the weathering of peridotites (Leguéré, 1976; Elias, 2002), and is therefore of some importance for the nickel resource.

The asymmetrical morphology of the "Grande Terre", which is tilted SW-wards, may be related to distinct events: (1) a Late Oligocene–Early Miocene post-obduction extension (Lagabrielle et al., 2005); (2) a syn- to post-Early Miocene transtension associated with subduction reversal and opening of the North Fiji Basin (Chardon and Chevillotte, 2006); and, (3) the recent involvement of the Australian Plate in the fore-arc bulge of the Vanuatu (New Hebrides) volcanic arc, as revealed by uplift of the Loyalty Islands and by the uplifted fringing reefs located in the SE of the Grande Terre (Dubois et al., 1974).

It is worth noting that the oldest known reefal construction on the margins of Grande Terre is not older than 1.4 Ma (Cabioch et al., 2008); whilst the Present barrier reef and lagoon are not older than 0.4 Ma (Frank et al., 2006) and underwent several periods of emersion during Holocene sea level low stands (Coudray, 1976; Chardon et al., 2008; Cabioch et al., 1999, 2008; Le Roy et al., 2008). During the latter periods, the lagoon was drained and rivers formed canyons across it. The paths of these rivers form submerged meanders that are now clearly visible.

Undated terrestrial deposits referred to as "Fluvio-lacustrine System" fill in paleo-valleys and endorheic weathering cells of the Massif du Sud (Trescases, 1975). Those sediments, 50–100 m thick, with coarse torrential conglomerate at the base, mainly consist of reworked laterite (actually Fe oxide), detrital chromite, and supergene silica clasts. They have been tentatively correlated to the Late Oligocene by Chardon and Chevillote (2006). At present, they are actively and unequally eroded.

## Mineral resources of New Caledonia

In contrast to many islands of the Pacific Rim, New Caledonia has no large volcanogenic ore deposits and only minor polymetallic orebodies of the sedimentary exhalative type are located in Late Cretaceous metasediments (Diahot Terrane). Gold was discovered in 1863 near Pouebo and, later, the Fern Hill mine (near Ouegoa) produced 214 kg of Au between 1873–1900. Many other Au occurrences have been discovered, but none of them have economic importance. Copper, Pb and Ag were discovered and mined in the Diahot Region during the late 19th century but all mines closed in the 1930's. Chromite lenses associated with the Peridotite Nappe have been mined in many places of the Massif du Sud, with that in the Tiebaghi Massif (North) producing 3.3 Mt chromite between 1902–1976.

However, these resources are now worked out and New Caledonia's mining industry depends upon the Ni contained in weathered peridotites. New Caledonia ranks fifth among the world's nickel ore producers (Elias, 2002); nickel concentration results from the hydrolysis of olivine (0.4% Ni) and pyroxene (0.025% Ni) by weathering solutions and subsequent concentration as silicate minerals or adsorbed mineral phases in the lower part of the weathering profile. The lateritic profile developed on obducted peridotite in New Caledonia belongs to the hydrous Mg silicate type (Brand et al., 1998; Freyssinet et al., 2005). It is characterized by an absolute enrichment or concentration of Ni in the saprolite zone, which consists of secondary serpentine, neoformed goethite, smectitic clays and garnierite. Much of the nickel is re-precipitated within the saprolite by substituting Ni for Mg in secondary serpentines (which can contain up to 5% Ni) and in garnierite, which can grade over 20% Ni (Pelletier, 1996). This silicate (or garnieritic) ore has been actively mined since the late 19th century but reserves are rapidly being depleted. Lesser Ni concentrations are found in oxidised (lateritic) ore (1.0–1.5% Ni); however, reserves are enormous and lower grade lateritic ore will represent the bulk of Ni reserves of New Caledonia in the future.

## Summary and conclusions

During Permian–Early Cretaceous time, New Caledonia had almost continuous volcaniclastic sedimentation. Its location at the eastern edge of a marginal basin system is consistent with discontinuous connections with continental sources of sediments; in

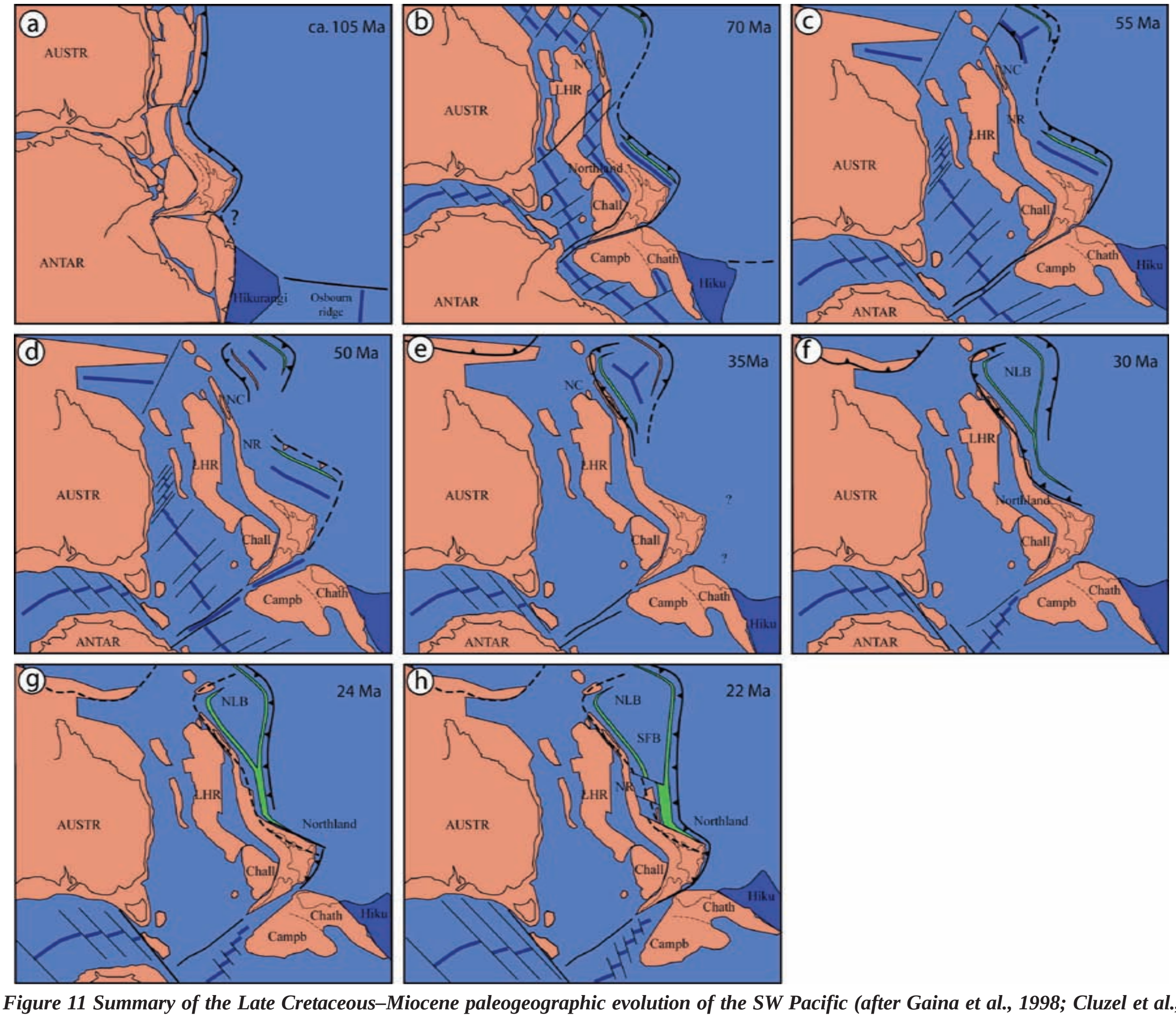


***Figure 11** Summary of the Late Cretaceous–Miocene paleogeographic evolution of the SW Pacific (after Gaina et al., 1998; Cluzel et al., 1999; Hall, 2002; Sdrolias et al., 2003; Crawford et al., 2003). Campb: Campbell Plateau; Chall: Challenger Plateau; Chatham Rise; Hiku: Hikurangi Plateau; LHR: Lord Howe Rise; NC: New Caledonia; NLB: North Loyalty Basin; NR: Norfolk Ridge; SFB: South Fiji Basin.*

contrast, long periods of isolation are responsible for prominent endemism.

During the Late Cretaceous, the Australian margin was the site of marginal break-off and marginal basin opening; it is worth noting that all marginal basins opened almost synchronously; and that no Late Cretaceous volcanic-arc is hitherto known, therefore, it may be suggested that the eastward flow of the upper asthenosphere triggered by rapidly stretching lithosphere prevented a "normal" mantle wedge to form (Cluzel et al., 2011). At the end of the Paleocene, a new NE-dipping subduction zone appeared at or close to the spreading ridge of the South Loyalty Basin (Figures 10 and 11c). During the Paleocene and Eocene, back-arc spreading of the North Loyalty Basin induced the roll-back and SE-ward migration of the Loyalty arc-trench system (Figure 11d). The trench reached the northern tip of the Norfolk Ridge in the Early Eocene and continental subduction/obduction started (Figure 11e); new subduction along the W coast relayed the Loyalty arc-trench system subduction blocked by the Norfolk Ridge (Figure 11f), this short-lived subduction was finally blocked by the Lord Howe Rise. The Eocene jamming of subduction and actively spreading North Loyalty and South Fiji basins provoked the activation (or reactivation) of the Tonga subduction; meanwhile, the system propagated southwards and reached the eastern Lord Howe Rise and northern New Zealand in Oligocene (Figure 11f) and Miocene times (Figure 11g) respectively. During the Miocene, the Tonga subduction was active and complex back-arc basins opened until it reached its present location and was eventually blocked by the collision of the Australian Plate with South Island and Hikurangi Plateau.

The pre-Oligocene geology of New Caledonia within the framework of SW Pacific evolution provides a good example of preserved pre-collision stages because no crustal thickening happened after the development of foreland basins, HP-LT complexes, melanges, obduction etc. It may be prominently useful to elucidate the more complex evolution of many accretionary orogens such as those of Paleozoic Central Asia.

During Oligocene–Recent time, New Caledonia emerged and underwent several phases of weathering that formed thick regolith, some of which is of major economic importance. The predominance of peridotite during the early stages of resettling by vegetation probably resulted in the unique endemic flora, which is at present located in ultramafic massifs.

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*by Hugh L. Davies*

# The geology of New Guinea - the cordilleran margin of the Australian continent

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*The island of New Guinea is the mountainous margin of the Australian continent. Paleozoic and Proterozoic Australian craton extends northward beneath the shallow waters of the Arafura Sea to underlie the southern plains of New Guinea and, with overlying sediments, to form the dramatically sculpted southern slopes of the central range in a great fold and thrust belt. The fold and thrust belt marks the outer limit of the autochthon. Beyond, to the N, E and W, is an aggregation of terranes that have accreted since the Late Cretaceous, driven by oblique convergence between the Pacific and Indo-Australian plates. The terranes comprise continental fragments and blocks of oceanic volcanic arc and of oceanic crust and mantle origin, and include two great ophiolites. The plate boundary itself is a complex system of microplates, each with separate motion, and marked by every kind of plate boundary. In the E the opening of the Manus Basin is associated with rapid clockwise rotation of New Britain, and the opening of the Woodlark Basin causes extension of continental crust in the Papuan peninsula and islands. This has resulted in the development of low-angle extensional faults and domal structures in metamorphic rocks and the exhumation of Pliocene eclogite. Remarkably similar extensional structures and the exhumation of Pliocene eclogite are seen in the Bird's Head area of western New Guinea (Wandamen Peninsula). Flat and shallow oblique subduction at the New Guinea Trench has caused the deformation of Plio-Quaternary sediments in the Mamberamo Basin, deformation and Pliocene igneous activity in the central range, and the southwestward motion of the Bird's Head. The island has significant resources of economic minerals and hydrocarbons.*

## Introduction

The island of New Guinea is the alpine and, in part, Andean margin of the Australian continent. In plan view, the island resembles a great bird flying westward (Figure 1). It is the second largest island in the world (2,200 km long and up to 750 km wide) and one of the most mountainous, with peaks to c. 4.9 km above sea level (Figure 2).

Politically, the island is divided between the Independent State of Papua New Guinea (PNG) in the E and the Republic of Indonesia in the W, with a boundary that coincides, for the most part, with the 141°E meridian. The western half was known as Irian Jaya and is now known as Papua and Western Irian Jaya; Western Irian Jaya is the Bird's Head and Neck.

The geology of New Guinea can be considered in three parts:

1. a western part that includes the Bird's Head and Neck and adjacent islands;
2. a broad central part that adjoins the Australian continent (Figure 1); and
3. an eastern part that includes the Papuan peninsula and islands.

All three parts have a similar geometry with sedimentary basins on continental basement in the S and a hinterland of metamorphic and oceanic rocks including ophiolite and volcanic arc assemblages in the N.

Islands that lie to the NE and E, the Bismarck Archipelago and Solomon Islands, do not fit this pattern. They are thought to have formed solely by intra-oceanic volcanic arc activity and, in the case of the Solomon Islands, accretion of a mostly-submarine volcanic plateau (see below).

Geological maps of the New Guinea and adjacent islands at 1:1 million scale (Bain et al., 1972; Dow et al., 1986) and of PNG at 1:2.5 million scale (D'Addario et al., 1976) are available, as are map series maps at larger scales. A useful bibliographic data base is Van Gorsel (2011).

## Geology of the Western Part: the Bird's Head peninsula and islands

Pieters et al. (1983) discussed the geology of the Bird's Head in terms of an oceanic province in the N, a continental province in the S, and a transitional zone between the two. The continental province occupies the greater part of the Bird's Head and includes Misool Island (Figure 3). It is bounded on the N by the E-W Sorong Fault (SF in Figure 3) and on the E by a N-S fault that parallels the W coast of Cendrawasih Bay and connects to the Weyland Overthrust (WT in Figure 3).

### The Continental Province

The Continental Province comprises sedimentary rocks over a Paleozoic basement. The basement is exposed in the mountains S of

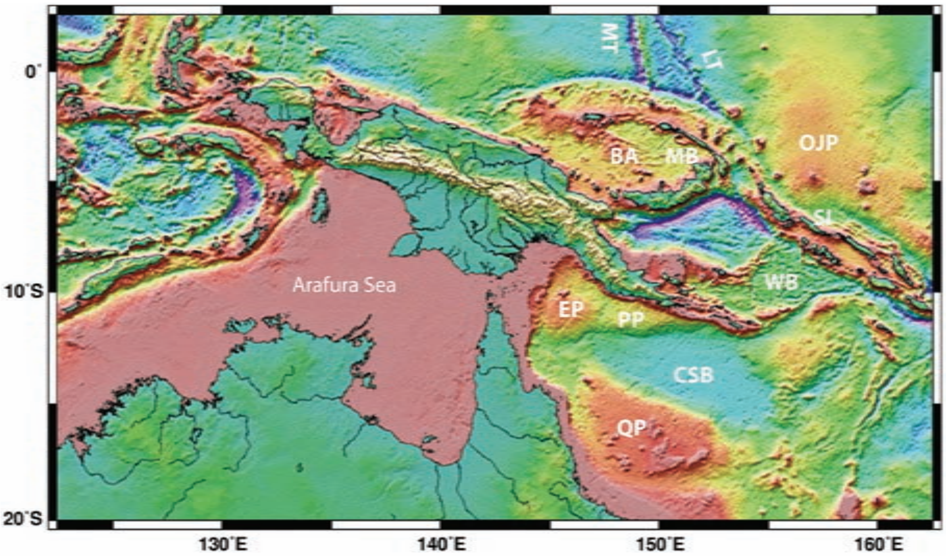


*Figure 1 Physiographic map of New Guinea. BA Bismarck Archipelago, CSB Coral Sea Basin, EP Eastern Plateau, LT Lyra Trough, MB Manus Basin, MT Mussau Trench, OJP Ontong Java Plateau, PP Papuan Platform, QP Queensland Plateau, SI Solomon Islands, WB Woodlark Basin. Seafloor topography from Smith and Sandwell (1997).*

*Figure 2 The morning sun lights part of the ice field in the summit area of Puncak Jaya, 4.08°S latitude, 4,884 m elevation, the highest peak in New Guinea. The ice field rests on S-dipping Middle Miocene limestone. This view to NW was taken in 1995. (Photograph H. Davies).*

the Sorong Fault and on Misool Island and comprises folded low-grade regionally metamorphosed turbidites that have been dated by Silurian graptolites and Devonian ostracods (Pieters et al., 1983). On the Bird's Head, the sedimentary sequence is of Permian and Mesozoic platform sediments, Eocene to Mid-Miocene limestone, and Late Miocene to Pliocene and Quaternary siliciclastics that are part turbiditic and part non-marine (Pieters et al., 1983; Bailly et al., 2009). The Cenozoic sediments form the Salawati and Bintuni basins (SB and BB in Figure 3) that are separated by a N-S basement ridge.

The sedimentary sequence on Misool Island differs from the mainland. Here "a unique and almost continuous sequence of deep-water and shallow marine sediments extends from Triassic times to the present day" (Pieters et al., 1983). The older sediments were folded in the Late Triassic and Early Jurassic (Visser and Hermes, 1962).

In the SE part of the Bird's Head the entire sedimentary section has been deformed and locally metamorphosed by contractional tectonics in the W-facing Lengguru fold belt (LFB in Figure 3). Bailly et al. (2009) interpreted the deformation to be the result of Late Miocene E-dipping subduction on the line of the present shoreline of Cendrawasih Bay. Miocene contraction was followed by Pliocene extension, the development of normal faults, and the unroofing of the Wandamen metamorphic core complex at 4–2 Ma (Bailly et al., 2009).

In the SW part of the Bird's Head the Cenozoic carbonates have been arched upward to form the karstified limestone antiforms of the Onin and Kumawa peninsulas (Ratman, 1998). The antiforms trend NW towards Misool Island.

## Oceanic Province

The Oceanic Province, N of the Sorong Fault includes Paleogene volcanic arc rocks and younger sediments, Triassic granitoids and, in the adjacent islands, ophiolite (Pieters et al., 1983).

## The Transition Zone

The Transition Zone rocks E of the N-S fault include fault slices of Paleozoic(?) metamorphosed sediments intruded by Early Jurassic (197 Ma) granite and, on the Wandamen Peninsula, the Plio-Pleistocene metamorphic core complex with grades as high as eclogite (Bailly et al., 2009). Further to the SE, on the E-W part of the Bird's Neck, Transition Zone rocks above the S-facing Weyland Overthrust include highgrade metamorphic rocks, ophiolite slices and Miocene diorite. The metamorphic rocks are pelitic and include staurolite-garnet-mica schist. Beneath and S of the thrust fault are footwall Paleozoic to Cenozoic sediments on continental basement; this is a westward arm of the Papuan Basin (Pigram and Panggabean, 1989).

# Geology of the Central Part of New Guinea: 136–145°E

The central part of the island is made up of the Papuan Basin in the S and a hinterland of mostly crystalline rocks in the N; the hinterland rocks are extensively overlain by Neogene sediments.

## Papuan Basin

The Papuan Basin occupies all of autochthonous New Guinea – the southern part of the bird's body (Figure 3). Sediments of the Papuan Basin underlie the southern plains and are exposed in the adjacent fold belt.

The basin is underlain by Australian craton of Precambrian age in the W and of Paleozoic age in the E; the boundary between Precambrian and Paleozoic basement is at around 141°E, with the exception that there is an exposure of Paleozoic basement at 140.3°E (Eilanden Metamorphics; Parris, 1996b). In the W the sedimentary section is 16 km thick and has late Proterozoic strata at base (Table 1). In the E the sedimentary section is 4 km thick and has Triassic and Jurassic sediments at base (Table 2). Hill et al. (2004) described the basin and its hydrocarbon potential.

### *Western Papuan Basin*

The sedimentary sequence is known from exploration wells on the foreland platform (Kendrick and Hill, 2001) and from mapping of the fold and thrust belt, where the rocks are exposed in the eroded core of a frontal anticline (Mapenduma Anticline; Parris, 1994a). An almost complete Paleozoic and Mesozoic section is exposed along the Freeport Grasberg mine access road (Martodjojo et al., 1975; Parris, 1994b; Cloos et al., 2005) and the Cenozoic section is exposed near the mine (Quarles van Ufford and Cloos, 2005).

The rock units of Cambrian and older age, notably the Kariem, Nerewip and Awitagoh formations, are known only from isolated exposures and there is doubt about their inter-relationships (Table 1). However the younger rock units, beginning with Otomona Formation, appear to be part of a paraconformable sequence that extends from Late Proterozoic or Cambrian to the Mid or Late Cenozoic.

The older sediments were deposited in a shelf environment. Break-up began in the Permian and continued in the Triassic and Early and Middle Jurassic (Pigram and Panggabean, 1989), and is recorded in the sediments of the Tipuma Formation (Parris, 1994a). Break-up was followed in Middle Jurassic and Cretaceous by the deposition of sag phase sediments of the Kembelangen Group (Table 1).

Carbonate sedimentation began in the Maastrichtian and Paleocene and persisted until mid-Miocene. An interval of clastic sedimentation (Sirga Formation) in the early Oligocene is correlated with a fall in sea level at the time of the first of the Cenozoic glacial maxima (Cloos et al., 2005). The later transition from carbonate to mixed pelitic and carbonate sedimentation at the beginning of the Late Miocene can be correlated with the fall in sea level at the time of the Late Miocene glacial maximum, though uplift associated with the first stages of mountain-building probably was a contributing factor. The emergence of the mountain mass in Late Miocene, Pliocene and Quaternary led to the rapid deposition of mostly molasse-type clastic sediments to S and N, notably in the Mamberamo Basin where total thickness may exceed 10 km (Visser and Hermes, 1962).

### *Eastern Papuan Basin*

The Mesozoic–Cenozoic basin evolution and sedimentary sequence in the eastern Papuan Basin (Home et al., 1990) is similar to that in the W, though not identical (Table 2). Break up in Late Triassic and Early–Mid-Jurassic was followed by sag phase siliciclastic sedimentation through Late Jurassic and Cretaceous, carbonate sedimentation from Eocene–mid-Miocene, and development of the fold belt accompanied by molasse-type sedimentation and some volcanism in late Miocene, Pliocene and Quaternary.

A feature that is seen only in the eastern Papuan basin is the rift-related uplift at the end of the Cretaceous and resultant erosion of Cretaceous section. The rifting and uplift were precursors of the Paleocene opening of the Coral Sea basin. Another feature seen only in the E is the develop-ment of Pleistocene strato- and shield volcanoes.

The strike of the fold belt changes at the international border perhaps coincident with the transition from Paleozoic basement in the E to Precambrian basement in the W. Structural style changes at 142°E from a broad asymmetric S-facing basement-thrust-bounded anticline upon which are superimposed lesser structures in the W to a thin-skinned thrust belt of parallel thrust-bounded anticlines and valley-and-ridge topography in the E (Figure 4).

Oil and gas in the eastern basin are sourced from Jurassic Imburu Formation and have accumulated in uppermost Jurassic and lowermost Cretaceous (Neocomian) sands that developed in the mudstone environment during sea-level lowstands; the fluids migrated into structural traps in the Pliocene (Hill et al., 2004).

### *Jimi-Kubor and Bena Bena blocks*

The Triassic to Cretaceous sedimentary rocks that are exposed in the Kubor Range and in the Jimi Valley, N of the Kubor Range, share some features in common with the Papuan Basin but there is much that is distinctive (Table 3). The distinctive character suggests that the Jimi-Kubor block is a terrane – probably a para-autochthonous terrane that broke from the Paleozoic Australian craton and was re-joined by collision in the late Paleocene or early Eocene (Davies et al. 1996; 1997). The basement that is exposed in the Kubor Range comprises metamorphosed Permian sediments intruded by Middle Triassic granitoids (Van Wyck and Williams, 2002).

Granitoids that intrude the Jimi-Kubor rocks are known as Kubor Granodiorite but form two populations, as indicated by K-Ar age, one c. 240 Ma and the other c. 220 Ma. Crowhurst et al. (2004) determined that the older suite is volcanic-arc-related and the younger is rift-related, as shown by Sr isotope and Sm-Nd data.

The Bena Bena terrane is a mountainous area of greenschist facies metamorphic rocks (Goroka and Bena Bena Metamorphics; Tingey and Grainger, 1976) in the eastern part of the area mapped as Jimi-Kubor in Figure 3. The protolith of the Bena Bena Metamorphics is part Late Triassic (221 Ma; Van Wyck and Williams, 2002) and the metamorphics are intruded by Jurassic gneissic granite (172 Ma; Page, 1976).

## The Hinterland of the Central Part of the Island

The hinterland of the Papuan Basin extends from Cendrawasih Bay in the W (CB in Figure 3) to the Finisterre Range in the E (FR in Figure 3). The hinterland is entirely allochthonous, or may be para-autochthonous in part, and is made up of terranes that have accreted to the Australian craton in a succession of collisions beginning in the Late Cretaceous (Pigram and Davies, 1987; Davies et al., 1996).

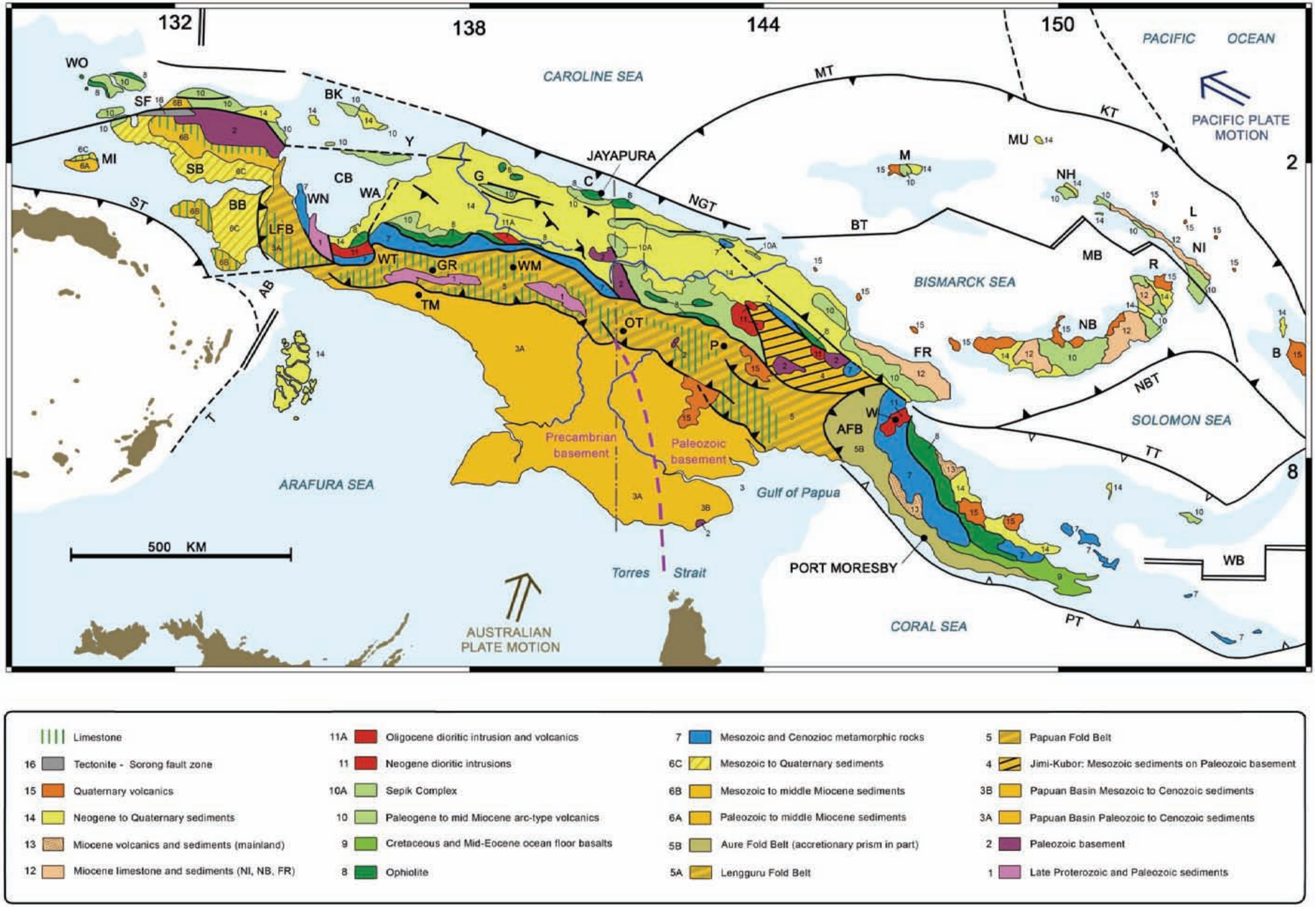


*Figure 3 Geological map of New Guinea. AB Aru Basin; AFB Aure fold belt; B Bougainville; BB Bintuni Basin; BK Biak; BT Bismarck Sea Transform; C Cyclops Mountains; CB Cendrawasih Bay; FR Finisterre-Saruwaged Range; G Gauttier or Foja mountains; GR Grasberg Mine; KT Kilinailau Trench; L Lihir Island (mine); LFB Lengguru Fold Belt; M Manus; MB Manus Basin; MI Misool; MT Manus Trench; MU Mussau; NB New Britain; OT Ok Tedi; P Porgera; PT Pocklington Trough; R Rabaul; SB Salawati Basin; SF Sorong Fault; ST Seram Trench; T Timor Trough; TT Trobriand Trough; W Wau; WA Waipona Basin; WB Woodlark Basin; WM Wamena; WN Wandamen Peninsula; WO Waigeo; WT Weyland Thrust; Y Yapen. (Map drawn by Randall Betuela).*

***Table 1 Stratigraphy of the Indonesian part of the Papuan Basin.***

<table>
<tr><th>Age</th><th colspan="3">Description and thickness</th></tr>
<tr><td>Late Miocene–Quaternary</td><td colspan="3">Molasse-type sediments derived from erosion of the emerging mountain mass. Mamberamo Group on N side of Central Range includes mid–late Miocene Makats Formation (1.8 km) and late Miocene to Pleistocene Mamberamo Formation (10 km; Visser and Hermes, 1962; as reported by Parris and Warren, 1996).</td></tr>
<tr><td>Late Miocene–Pliocene</td><td colspan="3">Buru Formation: basal calcareous mudstone and sandy shale grades upward to well-bedded lithic sandstone and mudstone with a 200 m interbed of limestone; parts mapped as Iwoer (Iwur) Formation, Kau Limestone, and Birim Formation; conformable on New Guinea Limestone Group. Thickness up to 3 km.</td></tr>
<tr><td rowspan="6">Paleocene–Middle Miocene, locally lower limit is Maastrichtian</td><td colspan="3">New Guinea Limestone Group. Gradational contacts with Ekmai Sandstone below and Buru Fm above; 1.6 km. Imskin Formation is deepwater equivalent (Pieters et al., 1983)</td></tr>
<tr><td></td><td>Cloos et al., 2005; Parris, 1994b</td><td>Pieters et al., 1983</td></tr>
<tr><td>Oligocene–Mid Mio</td><td>Kais or Ainod Fm, massive limestone, locally abund. fusulinids, forams, 1.1 km</td><td>Yawee Limestone (lateral equivalent of Kais and Ainod) massive limestone, reef complex.</td></tr>
<tr><td>Early Oligocene (Tc stage)</td><td>Sirga Sandstone, quartz-rich; 100 m. Erosional disconformity at base.</td><td>Adi Member sandstone and mudstone, 150 m.</td></tr>
<tr><td>Eocene (Ta–Tb stage)</td><td>Faumai Formation, well bedded arenaceous limestone, commonly muddy, 250 m (Pieters et al., 1983).</td><td>Lower part of Yawee Limestone is equivalent of Faumai Formation.</td></tr>
<tr><td>Paleocene–Eocene; part Maastrichtian (Parris, 1996b)</td><td></td><td>Waripi Fm, well-bedded sandy oolitic calcarenite, 700 m, most has no fossils, Late Cretaceous age from forams (Parris 1996b).</td></tr>
<tr><td rowspan="5">Middle Jurassic–<br>Late Cretaceous</td><td colspan="3">Kembelangen Group: grey variably argillaceous, glauconitic, calcareous, micaceous and pyritic sandstone and siltstone, black calcareous mudstone to limestone, quartz sandstone and orthoquartzite; 4.5 km. Kembelangen component rock formations are (Cloos et al., 2005; Parris, 1994b):</td></tr>
<tr><td rowspan="2">Late Cretaceous</td><td colspan="2">Ekmai Sandstone: pyritic and glauconitic quartz sandstone; 650 m.</td></tr>
<tr><td colspan="2">Piniya Mudstone: micaceous and glauconitic mudstone rare foraminifera; 700 m.</td></tr>
<tr><td>Early–Late Cretaceous</td><td colspan="2">Woniwogi Sandstone: orthoquartzite with belemnites; 200–400 m.</td></tr>
<tr><td>Middle–Late Jurassic</td><td colspan="2">Kopai Formation: quartz sandstone, siltstone, mudstone, belemnites, gastropods, pelecypods, ammonites, limestone with star crinoids; shallow marine; 300 m.</td></tr>
<tr><td>Late Permian and Middle Jurassic; (alternatively Triassic)</td><td colspan="3">Tipuma Formation. Maroon and green mudstone, lithic sandstone and pebble conglomerate, part non-marine. Comprises two rock units separated by disconformity, one late Permian and the other Middle Jurassic; ages by palynology (Parris, 1994a). Fossils in lower part include Glossopteris; thickness 2 km (Cloos et al., 2005). The disconformity may represent a rifting event (Parris, 1994a). Mapenduma Fm may be equivalent to Tipuma Fm; it is the field name given to thick sequence of grey turbidites in Wamena area, part Triassic and part probable Middle Jurassic from palynology (Parris, 1994a).</td></tr>
<tr><td>Permian</td><td colspan="3">Aiduna Formation. Lithic sandstone part feldspathic, part micaceous, interbedded with black shale, biocalcarenite, polymict conglomerate and coal; overlies Modio Fm and underlies Tipuma Fm; Permian age from brachiopods and plant fossils (Parris, 1994a); paraconformable on Modio Fm and grades upwards into Tipuma Fm (Pieters et al., 1983); up to 2.2 km (Parris, 1994b; Cloos et al., 2005). Part of Aifam Group in Bird's Head basins (Pieters et al., 1983)</td></tr>
<tr><td>Devonian and possibly Silurian</td><td colspan="3">Modio Formation: fossiliferous dolostone and crinoidal grainstone with siliciclastic sediments towards the top; Devonian age (part Frasnian, part pre-Frasnian and possibly Silurian), age from corals (Oliver et al., 1995) and poorly preserved conodonts; 2 km approx. (Pieters et al., 1983; Parris, 1994b; Cloos et al., 2005).</td></tr>
<tr><td>Ordovician</td><td colspan="3">Tuaba Formation: (136.6 –138°E) coarse quartz sandstone with some conglomerate, interlaminated sandstone and siltstone, and at top red laminated siltstone and mudstone; overlies Otomona and underlies Modio Fm; 1 km thick (Parris, 1994b; Pieters et al., 1983). Kora Fm (138.4–139.3°E) black slatey mudrocks, may be equivalent of Tuaba Fm, Ordovician age from nautiloids and graptolites (Parris, 1994a, 1996a).</td></tr>
<tr><td>Late Proterozoic–<br>Cambrian</td><td colspan="3">Otomona Formation upward-coarsening turbidites exposed at 136.6–137.4°E, basal part weakly metamorphosed, one fission track zircon age is 675 Ma; thickness >3 km (Parris, 1994b). This suggests that the Otomona Fm may be older than Nerewip Fm and therefore could be equivalent to Kariem Fm (Parris, 1994a).</td></tr>
<tr><td>Late Proterozoic–<br>Cambrian</td><td colspan="3">Awitagoh (140°E) and Nerewip (137°E) formations: basaltic lava, pillow lava and argillites 600 m thick. Nerewip Formation partly schistose and metamorphosed to greenschist facies. Awitagoh Fm overlies Kariem Fm; altered basalt has minimum K-Ar age of 486 ± 17 Ma (Parris, 1994a); may underlie Kariem Fm (Cloos et al., 2005).</td></tr>
<tr><td>Late Proterozoic<br>> 820 Ma</td><td colspan="3">Kariem Formation: pyritic mudstone and dolomitic mudstone with dolerite sills, known only in the eastern cenrtral Range at 139.5–140.4°E. K-Ar ages 820 ± 21 Ma and 847 ± 5 Ma; thickness >2.5 km (Parris, 1996b).</td></tr>
<tr><td>Proterozoic</td><td colspan="3">Basement (not exposed).</td></tr>
</table>

*Table 2 Stratigraphy of the eastern Papuan Basin.*

| Age | Description and thickness |
|---|---|
| Quaternary | Scattered major central volcanoes and volcanic fields, none now active, lava compositions are K-rich basalt to andesite. |
| Late Miocene– Quaternary | Siliciclastic sequence up to 1.5 km thick, marine at base grading upwards to non-marine molasse type sediment derived from erosion of emerging mountains and from subaerial volcanic activity. |
| Eocene–Middle Miocene | Darai Limestone: 1–1.5 km. Most is Late Oligocene to Mid Miocene; overlain by 600 m siltstone and 200 m limestone, cf. Buru Fm of western part of basin. E of longitude 144°E, Darai Limestone gives way to a thick sequence of rapidly deposited polymict, part volcanogenic clastic sediments, informally referred to as Aure Group. |
| Paleocene | Mudstones and lithic sandstones in NE, slope environment. |
| Cretaceous | Toro Sandstone and Ieru Formation: mudstone and quartz sandstone, transgressive-regressive sequence, part volcanogenic, 1.5 km. |
| Late Jurassic | Koi-Iange Sandstone and Imburu Formation: Syn-rift and post-rift carbonaceous siltstone and sandstone, 400 m on shelf thickens to >5 km in N and NE (Om Formation). |
| Early and Middle Jurassic | Syn-rift mudstone and sandstone intersected in wells in SW (on platform), part non-marine. |
| Mid and Late Triassic | Syn-rift Kana Volcanics: bi-modal composition, 3.5 km in Jimi-Kubor Block; related rift-related intrusives (220 Ma); Mid–Late Triassic sediments on Jimi-Kubor Block. |
| Permian | Basement of low-grade metasediments, Permian in part, intruded by Early Triassic (240 Ma) granodiorite. The Early Triassic granodiorite is arc-related (Crowhurst et al., 2004). |

The terranes include one or more fragments of Paleozoic craton, ophiolites, a variety of metamorphic rocks, dioritic and granodioritic intrusives, island arc volcanic rocks and associated sediments, and oceanic crustal rocks. They are partly covered, unconformably, by as much as 10 km of sediment of Early and Middle Miocene and younger age.

## Derewo Metamorphics

The Derewo or Ruffaer Metamorphics (Warren and Cloos, 2007; Weiland, 1999) extend for 550 km along the N side of the Central Range in West Papua. They include two distinct and probably unrelated metamorphic rock units: one is the great mass of contorted quartz-veined phyllitic and micaceous graphitic schists that comprise the bulk of the rock unit, and the other is a series of higher-grade metamorphic rocks including high-temperature amphibolite, blueschist and eclogite, that are found close to the contact with the ophiolite (Weiland, 1999; Parris and Warren, 1996).

The phyllitic and graphitic schists are entirely greenschist facies but metamorphic grade increases northward (Warren and Cloos, 2007). They derive from and are transitional southward into unmetamorphosed Jurassic-Cretaceous Kembelangen Formation. In some places the contact is faulted (Warren and Cloos, 2007).

The amphibolites probably are part of the metamorphic aureole of the ophiolite. They appear to be similar to the high-temperature amphibolites and hornblende granulites that are found at and near the base of the Papuan Ultramafic Belt (PUB) ophiolite in PNG (Davies, 1971; Lus et al., 2004).

On the geological map (Figure 3) the Derewo Metamorphics appear to be truncated at the international border but this is not the case. Rather, the quartz-veined graphitic schists continue across the border where they were mapped as metamorphosed Jurassic Om Formation (Davies, 1982) and on Figure 3 are included within the area mapped as Papuan basin fold belt (rock unit 5). Metamorphic grade in metamorphosed Om Formation increases northwards and the schists have both gradational and faulted contacts with the unmetamorphosed protolith.

The blueschists and eclogites of the Derewo Metamorphics are not continuous across the border but reappear 100 km E of the border at 142°E (Tau Blueschist; Ryburn, 1980; Davies, 1982).

## The West Papua Ophiolite

Ultramafic and gabbroic rocks of the West Papua ophiolite are exposed for a length of 440 km on the N side of the central range, and

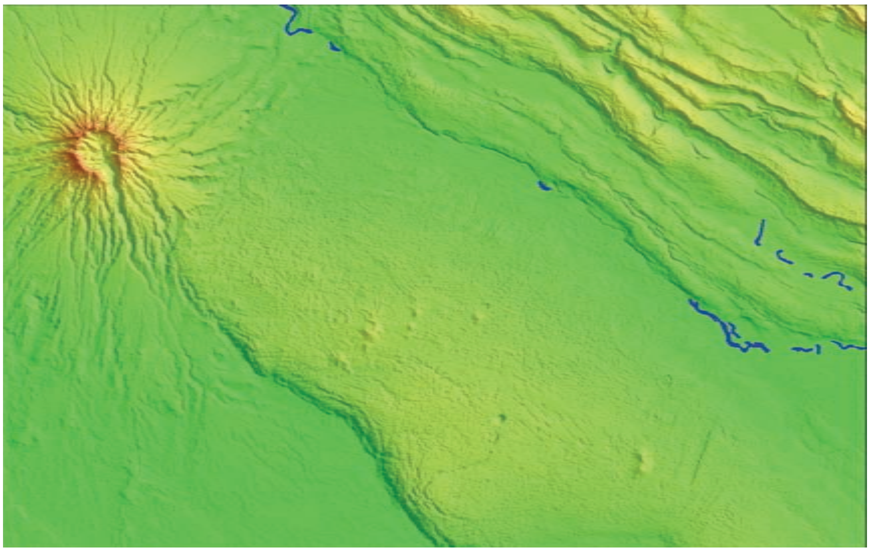

***Figure 4 The Pleistocene stratovolcano Bosavi, 2,397 m, overlooks the Darai uplift, a broad basement-involved anticline (inverted graben) expressed in Oligo-Miocene limestone. To the NE, beyond the Kikori River, thin-skinned fold belt is expressed as valley-and-ridge topography formed on successive S-facing thrust-bounded anticlines. Relief data from Shuttle Radar Topography Mission; image prepared by P. L. Shearman.***

*Table 3 Stratigraphy of the Jimi-Kubor and Bena Bena blocks.*

<table>
<tr><th>Age</th><th colspan="4">Description</th></tr>
<tr><td>Late Cretaceous</td><td colspan="4">Chim Formation and Asai Shale: massive finely laminated calcareous grey shale; some volcanics; Cenomanian–Maastrichtian and Early Paleocene; 3 km; (Davies, 1983).</td></tr>
<tr><td>Early Cretaceous</td><td colspan="4">Kondaku Formation: tuff and volcanically-derived sandstone; cuesta-forming; 2.5 km; Aptian–Albian (Bain et al., 1975)</td></tr>
<tr><td>Late Jurassic</td><td colspan="4">Maril Shale: dark calcareous siltstone and shale; 2 km; Kimmeridgian bivalves (Davies, 1983).</td></tr>
<tr><td>Mid Jurassic</td><td colspan="2"></td><td colspan="2">BB Block: Karmantina gneissic granite, 172 Ma (Page, 1976), intrudes Bena Bena Metamorphics.</td></tr>
<tr><td>Early Jurassic</td><td colspan="2">Balimbu Fm: dark grey volcanolithic sandstone and siltstone; 300 m; Sinemurian–Pliensbachian age (Pigram et al., 1984).</td><td colspan="2">BB Block: Metamorphic event, Bena Bena Metamorphics</td></tr>
<tr><td>Late Triassic</td><td colspan="2">Kuta Fm: limestone, conglomerate, arkose; 250 m; Late Norian or Rhaetian (Skwarko et al., 1976); unconformable on Kubor Gd and Omung Mtmcs.</td><td colspan="2">“Kubor Granodiorite(2)” rift-related granitic intrusive rocks (Crowhurst et al., 2004); c. 220 Ma; includes younger parts of Kimil Complex.</td></tr>
<tr><td>Middle–Late Triassic</td><td>Kimil Complex intrusives: K-Ar ages 242, 231 and 213 Ma (Pigram et al., 1984)</td><td colspan="2">Kana Volcanics: 3.5 km, rift-related, part contemporaneous with Kimil Complex. Protolith of Bena Bena Metamorphics 221 Ma; (Van Wyck and Williams, 2002)</td><td>Jimi Fm: grey sandstone, mudstone, 800 m, lateral equivalent of Kana Volcanics, Ladinian-Norian (Pigram et al., 1984)</td></tr>
<tr><td>Middle Triassic</td><td colspan="2">Yuat Fm: black to dark grey mudstone; 100 m; Late Anisian ammonites; grades upward into Jimi Fm (Pigram et al., 1984).</td><td colspan="2">“Kubor Granodiorite(1)” volcanic-arc-related mafic granitic rocks (Crowhurst et al., 2004); 240 Ma</td></tr>
<tr><td>Late Permian</td><td colspan="4">Omung Metamorphics (Van Wyck and Williams, 2002)</td></tr>
</table>

as outliers within the Weyland Overthrust (Dow et al., 1986; Monnier et al., 2000). The ophiolite is in fault contact with Derewo Metamorphics to the S. Metamorphosed gabbro that is faulted against the metamorphics immediately W of the international border may be an outlier of the same ophiolite (Parris, 1996b; rock unit not shown in Figure 3).

The western sector of the ophiolite, W of 138.5°E, is mostly ultramafic rocks, and the eastern sector, E of 138.5°E, mostly mafic rocks (Dow et al., 1986; based on interpretation of aerial photographs). Weiland’s (1999) foot traverses and spot landings generally confirmed these boundaries but picked out an area of sheared serpentinite with blocks of blueschist (137.8°E) and eclogite (138.05°E), just W of the Gauttier Offset. (The Gauttier Offset is an ENE-trending left-lateral fault that may extend from the central ranges beneath Mamberamo Basin sediments to the eastern end of the Gauttier (Foja) mountains; Dow et al., 1988.)

Weiland (1999) described the ultramafic rocks as variably serpentinised peridotites. He searched for but did not find pillow lavas. Monnier et al. (2000) found harzburgite and dunite with minor wehrlite at two sites near 136.5°E, and gabbro with cumulus texture and some basalt near 138.7°E.

The ophiolite is thought to represent Late Cretaceous oceanic crust and mantle and to have been emplaced in the Late Cretaceous as indicated by ages of the high-temperature metabasites (68 Ma; Weiland, 1999; Bladon, 1988) or in the Eocene, 44 Ma, as indicated by the age of blueschist facies metamorphic rocks (Weiland, 1999).

Late Miocene arc-type volcanic rocks and associated dioritic intrusive rocks adjoin the ophiolite in the NW (136.6°E; Parris and Warren, 1996), and late Eocene to Oligocene arc-type volcanic rocks adjoin the ophiolite further E (138.7°E, Dabera complex; Parris 1996d).

## *Fragments of Paleozoic craton*

The Border Mountains are an isolated block of Paleozoic metasediments and Triassic intrusive rocks that extend W from the international border at 3.75°S (the Idenburg Inlier of Dow et al., 1988). The metamorphic rocks include amphibolite gneiss, garnet-muscovite leucogneiss and ‘greenschist’. The intrusive igneous rocks include a layered mafic complex with layered ultramafic rocks, troctolite, gabbro, diorite, hornblende-quartz diorite and granodiorite, cut by andesitic dykes; most K-Ar ages are 250-240 Ma, and one is near 230 Ma (Davies, 1990; Parris, 1996c).

The fragment of Paleozoic craton extends E across the border to Amanab where metamorphic rocks are overlain unconformably by Late Cretaceous (late Campanian or Maastrichtian) limestone (Wilson et al., 1993) and may extend SSE to the Landslip Range, as is shown in Figure 3. There is no age evidence to confirm this but the Landslip Range has served as a barrier against which the generally westward tectonic trends have been sharply deflected to the N (Davies, 1982; Davies et al., 1997) and thus may represent older stronger lithosphere. Crowhurst et al. (2004) concluded that Sr and Nd isotopic values for the Amanab and Landslip metamorphic rocks indicate a mixed partly cratonic provenance.

## *Mamberamo Basin*

The low-lying area N of the central range is occupied by the Mamberamo Basin, a sedimentary basin with as much a 10 km thickness of sediments. Middle Miocene turbidites are unconformably overlain by rapidly deposited late Miocene–Quaternary clastic sediments; these show extensive diapirism (Williams and Amiruddin, 1984). Mamberamo Basin sediments have been subjected to Pliocene-Quaternary N-S contractional and transpressional tectonics to produce folds and N-facing thrust faults. Isolated blocks of basement rocks protrude from the basin and partially deflect the deformation. The Gauttier or Foja mountain block is one such. It comprises ultramafic rocks with some Paleogene andesitic–basaltic volcanics, volcaniclastic sediments, and minor limestone and is unconformably overlain by E-W-trending Neogene sediments (P.E. Pieters, pers.comm., 2008).

The Efar and Sidoas mountain blocks, adjacent to the N, are of ultramafic rocks.

The Cyclops Mountains, near Jayapura, comprise an ophiolite assemblage and moderate to high grade metamorphic rocks: amphibolite, gneiss and schist (Baker, 1956; Monnier et al., 1999). Boninite associated with the ophiolite has a K-Ar age of 43 Ma (Monnier et al., 1999). Immediately adjacent are Miocene sediments and Cenozoic volcanic arc rocks. Ultramafic rocks of the ophiolite may extend across the border as indicated by an exposure of sheared serpentinite in an erosional window just E of the border (Norvick and Hutchison, 1980).

### *Sepik Complex 141–146 °E*

The basement rocks of the Sepik Valley are grouped together as Sepik Complex (Davies, 1990; Rogerson et al., 1987), an assemblage of arc-volcanic, ultramafic, metamorphic, dioritic intrusive and sedimentary rocks that is interpreted to have formed in the Cenozoic in two arc-continent collisions.

In the southern part of the Sepik Complex the volcanic rocks and associated sediments are of Mid-Eocene to basal Late Eocene age and have been mapped as 'Salumei Formation". Associated metamorphic rocks include blueschist and eclogite (Ryburn, 1980) and have K-Ar ages of 44-38 Ma and 27-23 Ma (Davies, 1990). The Salumei rock assemblage is thought to be a product of a late Eocene arc-continent collision. This suite of rocks underlies the Sepik Valley sedimentary basin and extends northward as far as the southern slopes of the northern ranges (Bewani-Torricelli ranges).

The northern slopes of the Bewani-Torricelli ranges expose arc-volcanic rocks and intrusives that are of generally late Eocene and Oligocene age. They have been mapped as Bliri Volcanics and are thought to have been emplaced by arc-continent collision at the end of the Oligocene. Oligocene Bliri and Eocene Salumei arc volcanic rocks are not readily distinguished from one another in the field and the distribution of each may be revised by future mapping. The Bliri Volcanics are overlain unconformably by Early Miocene and younger sediments. The prominent calc alkaline intrusive rocks of the Bewani Torricelli ranges are mostly Late Eocene–Mid-Oligocene in age: 13 samples have K-Ar ages in the range 40–30 Ma; two others are Late Cretaceous and three are Early–Mid-Miocene (Hutchison, 1975).

### *The youngest accretion events*

Pliocene MORB-type basalts that are exposed on Kairiru Island at 143.5°E (John, 2006); and Plio-Pleistocene siltstone on the adjacent mainland at 3.41°S, 143.42°E (Klootwijk et al., 2003) are the youngest of the accreted terranes. The siltstone had moved S from an initial position near the equator (Klootwijk et al., 2003).

### *Cretaceous high-grade metamorphic rocks and associated ultramafics*

Cretaceous moderate to high-grade metamorphic rocks are known in the NE part of the Sepik Complex in the Prince Alexander Range at 3.4–3.5 °S, 143.0–143.5 °E (rock unit 7 in Figure 3). The rock types include amphibolite gneiss, orthogneiss and subordinate mica schist (Hutchison, 1975). Age from a number of K-Ar determinations is Aptian–Albian (c. 110 Ma). Other lower grade metamorphic rocks along strike to the E are exposed in fault contact with ultramafic rocks (Mt Turu Complex), and across the Sepik Valley to the S (Hunstein Range). These generally lower grade rocks were mapped as Ambunti Metamorphics. Their age is not known but Sr and Nd isotopes indicate mixed continental and oceanic provenance (Crowhurst et al., 2004) and the possibility remains that these too are Mesozoic or even late Paleozoic. The Mount Turu ultramafic rocks appear to be of igneous cumulate origin (Hutchison, 1975).

### *Miocene granodiorite: the Maramuni arc*

Large bodies of Miocene granodiorite intrude the SE part of the Sepik Complex and extend to ESE in the Jimi-Kubor terrane. These intrusives are very likely related to Miocene arc volcanic rocks that form a major part of the cover sequence and the two together have been given the name Maramuni arc.

### Marum Ophiolite

The Marum Ophiolite is a layered sequence of ultramafic and gabbroic rocks that dips NE beneath the plains of the Ramu River and on the SW side is faulted against low-grade metamorphics (5.5°S, 145°E; Jaques, 1981). Lateritic soils on the ultramafic rocks are currently being developed as a source of Ni and Co.

### *Adelbert-Finisterre-Saruwaged ranges*

The Adelbert, Finisterre and Saruwaged ranges comprise Oligocene–early Miocene arc volcanic rocks overlain by Miocene and younger limestone (Jaques and Robinson, 1977; Abbott, 1995). Fault deformation of the volcanic rocks preceded deposition of limestone. The volcanics and limestone form a great S-facing thrust-based antiform marked by limestone dip-slopes on the N side and by faulted and rapidly eroding volcanic rocks on the S side.

## Geology of the Eastern Part of New Guinea: the Papuan Peninsula

The tail of the bird comprises a SE-trending peninsula and chain of islands. The Owen Stanley Range extends along the peninsula as a mountainous spine though broken into two parts by an area of lower ground at 148.25–148.8°E; peaks in the NW part of the range rise to 4 km, and in the SE part to 3 km.

The main rock units of the peninsula are a great mass of generally low-grade metamorphics (Owen Stanley metamorphic complex), a major ophiolite (Papuan Ultramafic Belt (or PUB) ophiolite), a great mass of Late Cretaceous and Mid-Eocene submarine basalts with minor pelagic limestone, Oligo-Miocene sedimentary cover, Miocene and Pliocene dioritic and granodioritic and shoshonitic intrusive stocks, and two Quaternary stratovolcanic complexes (Lamington-Hydrographers and Victory-Trafalgar).

### Owen Stanley Metamorphic Complex

The greater part of the metamorphic complex is made up of greenschist to low greenschist facies felsic rocks, some clearly derived from felsic volcanic detritus (Kagi Metamorphics; Pieters, 1978) with minor limestone and conglomerate. Along the NE margin of the Owen Stanley Range the felsic metamorphics are structurally overlain by a

1–2 km-thick carapace of blueschist-greenschist metabasites (Emo Metamorphics) that dips NE beneath the ophiolite. Metabasites also occur at scattered localities within Kagi Metamorphics.

The age of the protolith of the Owen Stanly metamorphic rocks is mid-Cretaceous in part. The U-Pb age of zircons of probably volcanic origin in the NW Owen Stanley Range is Aptian–Albian (120–107 Ma; Kopi et al., 2000) and the age of macrofossils preserved in the metamorphics in the same general area is Aptian–Cenomanian (Dow et al., 1974). The timing of metamorphism is probably Late Cretaceous or Paleocene and may extend into the Late Eocene, or has been followed by a discrete Eocene event. Metamorphic minerals give radiogenic cooling ages of 68–22 Ma, with some clustering at 44–32 Ma (Davies and Williamson, 2001).

### *Kemp Welch Formation*

In the SE, at 147.2–148.2°E, Eocene low-grade metapelites mapped as Kemp Welch Formation (Pieters, 1978) may have the form of a thrust-bounded duplex that underlies the greenschist-facies Kagi Metamorphics, or may grade into Kagi Metamorphics. The age is from foraminifera.

## Aure Fold Belt and Poreporena accretionary complex

The Aure Fold Belt (AFB in Figure 3) is an arcuate W-facing thrust and fold belt that is made up of a thick sequence of rapidly deposited late Oligocene–Miocene and Pliocene, mostly-clastic sediments (Dow et al., 1974). The sediments were folded and faulted in response to westward movement of the Papuan peninsula. The Aure Fold Belt extends offshore and SE as far as 146.8°E.

East of 146.8°E the Aure Fold Belt gives way to thrust-bounded strike ridges of Paleocene and Eocene fine, mostly siliciclastic, sediments, minor serpentinite, and minor Oligocene coarser sediments, intruded by Oligocene gabbro. The sediments and gabbro are partly metamorphosed to low greenschist facies. The sequence is interpreted to have formed as an accretionary prism above a late Eocene–Oligocene NE-dipping subduction system. The complex is best exposed in the Poreporena Highway road-cuts in Port Moresby.

## Papuan Ultramafic Belt ophiolite (PUB ophiolite)

The ophiolite is a layered sequence of ultramafic, gabbroic and basaltic rocks that dips away from the Owen Stanley Range to E and SE at a shallow angle (Davies, 1971). It comprises 4–8 km of tectonite ultramafics overlain by about 4 km of gabbro which is in turn overlain by about 4 km of basalt. The tectonite ultramafics are mostly harzburgite with some orthopyroxenite veining. A discontinuous, 500 m thick layer of cumulus-textured ultramafic rocks intervenes between the tectonites and the gabbro in places. Gabbro is commonly layered and mostly shows some effect of deformation during crystallization, as in the smearing out of crystal-cumulus textures. The sheeted dyke complex that might be expected between gabbro and basalt layers is seen at only a few localities.

The crystallization age of the igneous part of the ophiolite is thought to be Maastrichtian, c. 71–65 Ma, as indicated by foraminifers in fine sediments associated with the basalt; one K-Ar age is 56 Ma (Davies and Williamson, 2001). Paleocene tonalite stocks and plutons intrude the gabbroic part of the ophiolite and mostly do not penetrate the basalt.

Overlying the ophiolite are Middle Eocene arc-type volcanic rocks, middle Miocene and younger volcanics, rapidly deposited Miocene and younger clastic sedimentary rocks and some limestone, and Pliocene–Quaternary volcanic rocks, including those of the intermittently active major volcanoes, Lamington and Victory.

## Metamorphic aureole of the ophiolite

At the base of the ophiolite there is intermittently exposed a 300 m thick thermal metamorphic aureole that comprises granulite and hornblende granulite grading to amphibolite away from the contact. The protolith is thought to be Emo Metamorphics. The cooling age of the contact metamorphic event is tightly constrained at 58.3 ± 0.4 Ma (Lus et al., 2004).

## Owen Stanley Fault

The Owen Stanley Fault is an E and NE-dipping fault that separates the metamorphic rocks from the ophiolite. It is interpreted to be a former subduction system thrust fault and is now, for most of its length, an extensional fault that has allowed the emergence of the metamorphic rocks in Late Miocene–Quaternary time.

## Suckling-Dayman Massif

East of 148.2 °E, the Owen Stanley Range falls away to lower ranges (Musa Valley area) but resumes at 148.8°E as an E-W elongated antiformal range – the Suckling-Dayman Massif. The remarkable smooth arched surface of the range, on older maps mistaken for a stratovolcano, is interpreted to be an exhumed and anticlinally folded subduction thrust fault (Davies, 1980; Daczko et al., 2009). Greenschist facies metabasite at the frontal fault gives way to variably metamorphosed basalt and limestone with some high pressure indicator minerals away from the fault (Davies, 1980). The limestone is moderately metamorphosed but foraminifera of Maastrichtian age are preserved in places.

## Submarine basalts

Much of the remainder of the Papuan peninsula, SE of the Suckling-Dayman Massif, is made up of a great volume of little-deformed submarine basalts with a thickness greater than 1 km. Much of the basalt is Mid-Eocene and some associated limestone shows low-angle extensional faulting. Mid-Miocene shoshonitic stocks intrude the basalts (Smith and Davies, 1976).

## D'Entrecasteaux Islands, Louisiade Islands, Trobriand Islands and Muyua

The association of exhumed metamorphic rocks and ophiolite extends from the mainland NE to the D'Entrecasteaux Islands and ESE to Misima, Sudest and Rossel islands (Davies and Warren, 1988; Hill et al., 1992). Eclogite is preserved in the generally amphibolite facies metamorphic rocks of the northern D'Entrecasteaux Islands (Davies and Warren, 1992); some is coesite-bearing and some has been shown to have formed in the late Miocene or Pliocene (Baldwin et al., 2004, 2008; Little et al., 2011).

The Trobriand Islands are raised coral probably underpinned by Plio-Quaternary volcanic rocks. Woodlark (Muyua) Island comprises Quaternary limestone cover on Cenozoic volcanic arc rocks.

## Small Ocean Basins

The Coral Sea Basin opened in the Paleocene (Weissel and Watts, 1979) leaving rifted fragments of Australian craton as submarine plateaus on the NW and SW margins of the basin (EP, PP, QP in Figure 1; Drummond et al., 1979). The Solomon Sea basin, N of the Trobriand Trough, opened in the early Oligocene, 34–28 Ma (Joshima et al., 1986) and has been subducted at both the New Britain Trench and the Trobriand Trough. The antiformal Solomon Sea plate can be traced WNW beneath mainland New Guinea for at least 400 km to 142 °E (Cooper and Taylor, 1987). The Manus Basin (eastern Bismarck Sea) opened by asymmetric spreading in the last 3.5 Myr and Woodlark Basin in the last 6 Myr (MB and WB in Figure 3; Taylor, 1979; Taylor et al., 1995, 1999).

## Bismarck Archipelago, Bougainville-Buka and the Ontong Java Plateau

The islands of the Bismarck Archipelago, together with the volcanic arc rocks of the Bewani-Torricelli and Adelbert-Finisterre-Saruwaged ranges, are thought to have a common origin and to possibly have formed as a single linear volcanic arc, referred to as the Bewani-Torricelli-Baining arc by Klootwijk et al. (2003). Each has a basement of volcanic arc rocks of Eocene or Oligocene age, variably deformed (faulted, tilted), and intruded by dioritic or granodioritic stocks. Commonly there is a partial cover of Middle Miocene limestone unconformably on the basement volcanics and above the limestone are Pliocene and Quaternary clastic sediments, volcanic rocks, active volcanoes, and fringing Quaternary reef limestone. Literature sources for New Britain include the 1:250,000 geological map series and Page and Ryburn (1977); for New Ireland – Hohnen (1978) and Stewart and Sandy (1988); for Manus Island – Jaques (1980); and for Bougainville and Buka islands – Blake and Miezitis (1967) and Rogerson et al. (1989).

The Ontong Java Plateau (Figure 5), the world's largest volcanic oceanic plateau, is a continent-scale submerged mass that comprises a sequence of basaltic lava flows overlain by 1 km of pelagic sediments. The lavas were emplaced in one remarkable magmatic event over a period of less than 7 Ma at c.122 Ma (Early Cretaceous, Aptian) with a lesser pulse at 90 Ma (Mahoney et al., 2001). The plateau is entirely submerged except for isolated atolls. WNW motion of the Pacific plate brought the plateau into contact with the Solomon Islands and Bougainville-Buka in the Late Miocene (Mann and Asahiko, 2004). Faulted parts of the Plateau are exposed in the adjacent Solomon Islands.

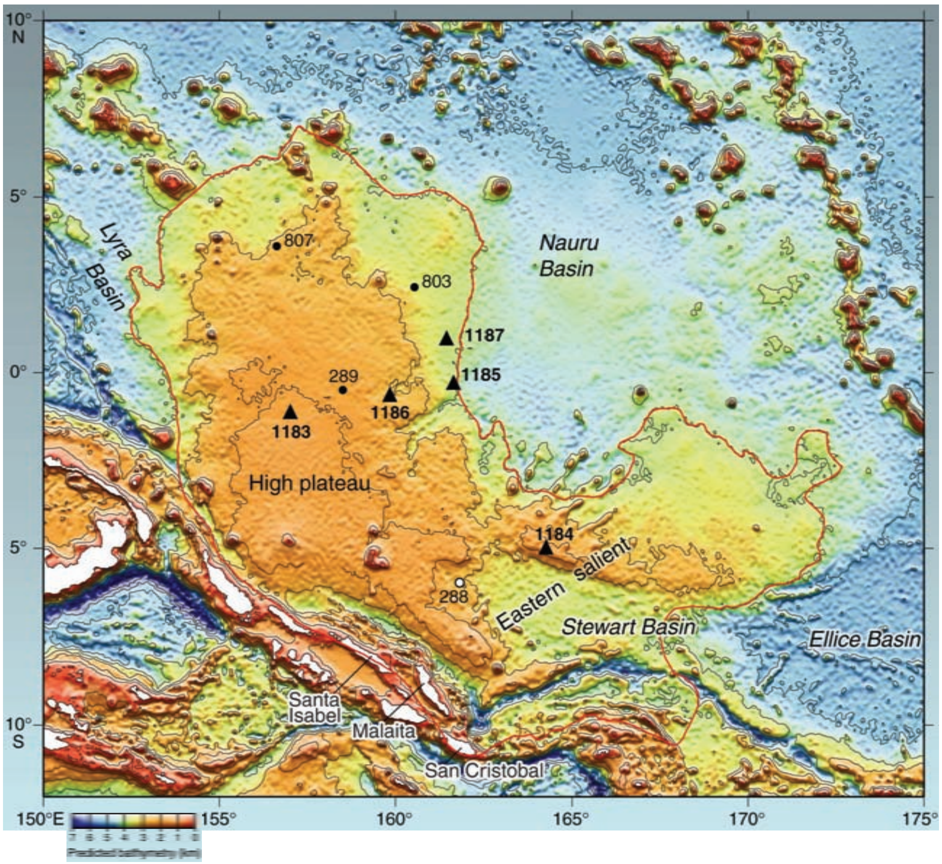


*Figure 5 Ontong Java Plateau showing Ocean Drilling Program drill sites. Those drilled on ODP Leg 192 are in bold type. The plateau is outlined in red. From Mahoney et al. (2001).*

## Plate tectonic setting, evolution, neotectonics and volcanic activity

New Guinea is at the interface of the Pacific, Australian, Philippines Sea, SE Asian and Banda Sea plates (Hamilton, 1979; Hill and Hall, 2003; Bird, 2003). Convergence between the Pacific and Australian plates is at a rate near 110 mm/yr on azimuth near 070 degrees. The present plate boundary comprises a series of micro-plates (Figure 3), each demarcated by earthquake activity (Figure 6). Figure 6a shows clearly the Solomon Sea plate lithosphere dipping N beneath New Britain and extending W beneath the mainland as far as 142°E. Figure 6b, with fewer earthquakes, shows more clearly the NW-trending Mamberamo Fault (centred at 3°S, 140°E) and the SW-trending fault that runs through Nabire (3°S, 136°E) in south-eastern Cendrawasih Bay. Shallow earthquakes coincide with all of the trenches and with the front of the fold belt.

Convergence has led to a succession of collisions of the Australian craton with volcanic arcs, ocean crust and microcontinents including fragments of Australian craton that had rifted from the craton and then docked again (Figure 7; Pigram and Davies, 1987; Davies et al., 1997). Silver and Smith (1983) pointed to the remarkable parallelism between the accretion of terranes to the New Guinea margin in the Cenozoic and the accretion of terranes to the North American margin in the Mesozoic. In both cases the accreted

(a)

(b)

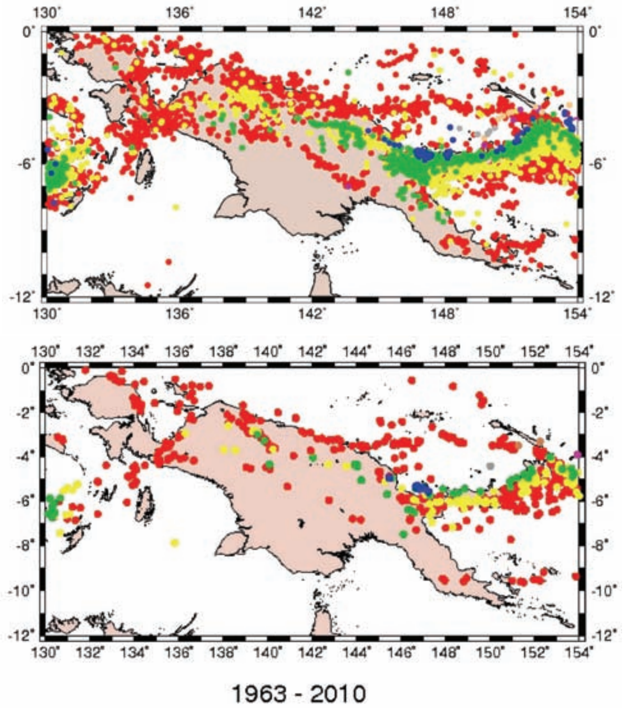


***Figure 6 Earthquake epicentres. (a). Earthquakes M5 and greater 1963–2004. (b). Earthquakes M6 and greater 1963–2010 showing detail in western part of island. Focal depths: Red <50 km, yellow <100 km, green <200 km, blue <300 km, purple <400 km, brown <500 km, grey >500 km. Shallow earthquakes (<50 km) mark active plate boundaries. (Maps by Emile Okal).***

blocks have been rotated, anticlockwise in the case of New Guinea (Klootwijk et al., 2003) and clockwise in the case of North America (Jones et al., 1982).

## Neotectonics

In NW New Guinea the oblique convergence between the Australian and Pacific plates is accommodated partly by left-lateral strike-slip motion on an E-W fault system that connects the Bismarck Sea Transform Fault in the E with the Sorong Fault in the W; partly by left-lateral strike-slip motion on another E-W fault system on the S coast – the Tarera-Aiduna fault system that extends W from the Weyland Overthrust (Pubellier and Ego, 2002); partly by subduction at the New Guinea Trench; partly by transpressional and strike-slip faulting in the fold belt (Abers and McCaffrey, 1988) and by folding and thrust faulting in the Mamberamo Basin.

The lithosphere of the Caroline Sea or Pacific plate is subducted southwestward at the New Guinea Trench (see also Milsom et al., 1992). Seismic tomography shows the subducted slab to dip at a shallow angle and to extend beneath the island of New Guinea to near the line of the S coast (Tregoning and Gorbatov, 2004). The tomography model is partly supported by mapping of earthquake foci (e.g., Pegler at al., 1995). If this interpretation is correct then the igneous activity in the Papuan Basin fold belt, including the Grasberg and Ok Tedi mineralised intrusive rocks, can be seen as slab-related (Davies, 2009 a, 2010), rather than related to slab break-off (Cloos et al., 2005). A shallow-dipping slab that is partly coupled to the upper plate also would explain the deformation of Plio-Quaternary sediments within the Mamberamo Basin, the transfer of convergent motion southward for a distance of 400 km from the line of the New Guinea Trench to the southern front of the fold belt and the observed WSW movement of the Bird's Head.

The Bird's Head is moving WSW at a rate of 86 ± 9 mm/yr relative to Australia (Stevens et al., 2002). This is slower than the motion of the Pacific Plate but has the same azimuth and is most likely explained by some degree of coupling between the Bird's Head and the underlying subducted slab. This motion in turn requires subduction of Bird's Head lithosphere at the Seram Trough (ST in Figure 3; Stevens et al., 2002). The WSW motion also has caused the opening of Cendrawasih Bay, the development of the Waipona sedimentary basin, and the change from Miocene contractional to Pliocene extensional tectonism in the Wandamen Peninsula and Lengguru Fold Belt (Bailly et al., 2009).

In NE New Guinea, collision between the Bismarck volcanic arc and the leading edge of the island of New Guinea is causing uplift of the north coast of the Huon Peninsula at (averaged) rates of 1–3 mm/yr. (Study of raised coral terraces on the peninsula (Figure 8; Chappell, 1974) yielded a high-quality record of fluctuations in sea level during the Late Quaternary.) The same convergence causes the Finisterre-Saruwaged mountain mass to ride southward (Abbott et al., 1997) and results in loading and downwarping of the northern end of the Papuan peninsula, the coast of which is subsiding at a rate of 5 mm/yr (Webster et al., 2004).

Sea-floor spreading in the Manus Basin in the eastern Bismarck Sea (MB in Figure 3; Taylor, 1979) and retreat of the New Britain Trench allow relatively rapid clockwise rotation of the island of New Britain (Tregoning et al., 1998; Wallace et al., 2004). Sea-floor mineralisation in the Manus Basin was investigated by the Ocean Drilling Program Leg 193 (Barriga et al., 2007).

The westward advance of sea-floor spreading in the Woodlark Basin (WB in Figure 3) has caused rifting and extension of the Papuan peninsula and adjacent islands and the development of domes and antiforms of layered metamorphic rocks ('metamorphic core complexes'). The easternmost continental extensional structures were investigated by Ocean Drilling Program Leg 180 (Taylor and Huchon, 2002) and with a seismic array experiment that showed crustal thinning and extension within the upper mantle (Abers et al., 2002).

## Volcanic Hazards

Fourteen active and 22 dormant volcanoes are a danger to an estimated 250,000 people (Saunders and Itikarai, 2006). Of the active volcanoes, 6 have been categorized as high-risk volcanoes. Five of these are in the Bismarck volcanic arc which extends along the N coast of New Britain and WNW to the islands offshore from mainland New Guinea. The five are the island volcanoes, Manam and Karkar, and the New Britain volcanoes Pago/Witori, Ulawun and Rabaul (Figure 9). The sixth high risk volcano is Lamington volcano on the NE coastal plains of the Papuan peninsula, NE of Port Moresby.

The caldera collapse volcanoes Long Island, Dakataua, Witori and Rabaul have been the source of devastating eruptions in the past, including a remarkable sequence of major eruptions from the three New Britain caldera volcanoes in the 7th century (McKee et al., 2011). Ash emission from an eruption of Long Island in c.1665 was sufficiently voluminous to block out the light of the sun for several days, and is recalled in legend in the PNG highlands as a time of darkness (Blong, 1982).

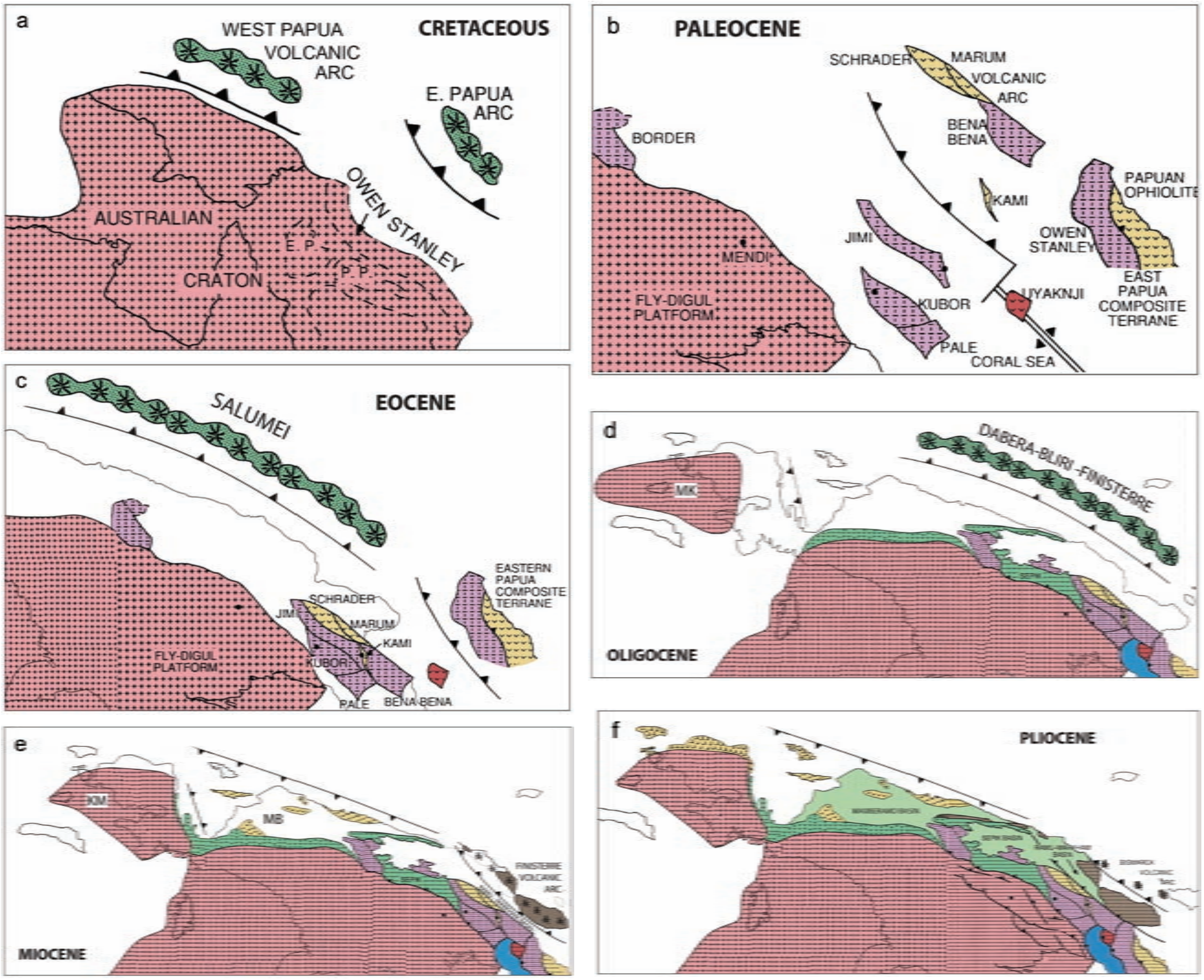


***Figure 7 Six stages in the evolution of allochthonous New Guinea. The evolution of other parts of New Guinea notably the epi-cratonic Papuan Basin is discussed in the text. (a) In Late Cretaceous, the West Papua ophiolite was emplaced by arc-continent collision. The East Papua arc lay offshore and was to be confronted in the Paleocene by a rifted fragment of Australian craton (Owen Stanley). EP and PP broke away from the continental margin before drifting north to form the Eastern Plateau and Papuan Platform. (b) In the Paleocene arc-continent collisions formed the Eastern Papua Composite Terrane (EPCT) and joined the Marum volcanic arc to the Bena Bena block and Schrader terrane. The Coral Sea-Uyaknji-Kami ocean basin formed, and more fragments separated from the Australian craton (Jimi, Kubor and Pale). (c) In the Eocene the rifted continental fragments joined with Kami and Jimi-Kubor to form a composite terrane which, in turn, docked with the craton. The Uyaknji small ocean basin persisted. The Eocene Salumei volcanic arc was accreted in the late Eocene or early Oligocene. (d) In the Oligocene the Kemum and Misool terranes lay west of their present location and joined to form a composite terrane (KM). The Dabera-Bliri-Finisterre (DBF) volcanic arc developed and the Solomon Sea opened. Toward the end of the Oligocene the EPCT moved west to close the Uyaknji basin and dock with the craton. (e) In the Miocene, the KM composite terrane moved eastward in response to subduction causing development of the Lengguru fold belt. Accretion of the DBF volcanic arc was followed by development of the Mamberamo, Sepik and Ramu-Markham successor basins, and was followed by arc-reversal and development of the New Guinea Trench and the Maramuni volcanic arc. Late Miocene uplift of the central range resulted in rapid sedimentation to north and south. (f) In the Pliocene, the New Britain Trench and Bismarck volcanic arc developed and were active. Continuing convergence between the New Guinea mainland and the Bismarck volcanic arc caused uplift and thrust-faulting within the northern ranges. Terranes, mostly of oceanic origin, were accreted in the western half of the island. Eclogites were exhumed by extension of crust in the SE Bird's Head and in SE PNG. (Maps drawn by Lian Brown).***

## Natural Resources

The island is richly endowed with natural resources. The Grasberg mine in Indonesian Papua contains the largest recoverable reserves of Cu and the largest Au reserve in the world and Papua New Guinea ranks eleventh in the world in the annual production of Au. In addition both PNG and the Indonesian provinces are about to become major exporters of LNG. In the Indonesian provinces oil is produced from Miocene sediments in the Salawati and Bintuni basins and gas for LNG from Jurassic sediments in the giant Tangguh field (Robertson,

*Figure 8 One of the scenic wonders of the world, the coral terraces have formed as the NE Huon Peninsula emerged in response to arc-continent collision. The terraces rise to a maximum elevation of 600 m and preserve a record of sea-level fluctuations for the last 300 kyr. Photograph: J. Chappell.*

*Figure 9 The violent eruption of the twin volcanoes that flank Rabaul Harbour on 19 September 1994 caused the destruction and abandonment of much of Rabaul town and the temporary relocation of 90,000 people; four lives were lost. The ash column rose to 18 km. This view N to Vulcan cone shows the base of the ash column and, on the right, the pyroclastic flows that moved eastward across the harbour. Photograph: M. Phillip and B. Alexander.*

2006). In PNG hydrocarbons are produced as oil from the Kutubu field and will be produced as gas for LNG from the Hides anticline (142.8 °E) and other structures in the fold belt, and from Miocene reefs in the eastern fold belt. Copper and Au are produced from mines at Grasberg and Ok Tedi (Figure 10), Au from Porgera, Hidden Valley, and Lihir Island (locations shown on Figure 3) and, commencing in 2012, Ni and Cu will be produced from the Ramu lateritic deposit, SW of Madang. Mineral deposits likely to be developed in the near future are Cu-Au at Wafi-Golpu (SW of Lae), at Frieda River (NE of Ok Tedi), and at Yandera (SW of Madang; Cu and Mo), and Cu-Au-bearing sea-floor massive sulfides in the Manus Basin.

*Figure 10 A view to the SE across the Ok Tedi mine open pit. Waste rock from the pit is discharged into the Ok Mani River which joins the Ok Tedi River at top left. The mine mill is in the small valley at left centre. Photograph: Ok Tedi Mining Limited.*


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*by Benjamin E. Cohen*

# The scenic rim of southeastern Queensland, Australia: A history of mid Cenozoic intraplate volcanism

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*Intraplate volcanism was widespread in southeastern Queensland during the mid Cenozoic, leaving a legacy of variably eroded volcanoes and rugged topography known locally as "The Scenic Rim". These plume-derived volcanoes provide a detailed record of northward Australian plate velocity, and indicate a major slowdown commencing at 26 Ma and persisting until 23 Ma, correlated with initial collision of the massive Ontong Java plateau with the northern subduction margin of the Australian plate. Despite traversing over 36 km of continental crust, trace element and isotopic signatures indicate little or only minor contamination for most units, with the exception of rhyolites formed during the period of slow plate velocity. Nevertheless, the thick continental crust allowed magmas to stall and fractionate during ascent, often producing highly evolved rocks (e.g., comendites) containing extreme concentrations of incompatible elements, including >2000 ppm Zr. Meanwhile, isotopic and trace element results from mafic units are consistent with melting and mixing of depleted upper mantle and an EM1-like source. Alkaline mafic eruptions also often contain abundant upper mantle and lower crustal xenoliths, providing excellent samples of these otherwise inaccessible regions. Denudation has produced good exposures of the subsurface magmatic architecture, a variety of landscapes, and diverse wildlife habitats; as a result many of the volcanoes are contained in National Parks, including the World Heritage listed Gondwana Rainforests of Australia.*

## Introduction

Australia is located in the centre of its tectonic plate, and has a widespread public image of being "a wide and sunburnt country" that is geologically all but dead (e.g., Bryson, 2000). Nevertheless, the Cenozoic history of the continent is not as quiet as is often assumed. Instead, eastern Australia is the site of one of the most extensive intraplate volcanic regions on the globe, with the products of volcanism extending over 4,000 km, covering a present-day area of 1.6 million km$^2$ (Figure 1a) (Johnson, 1989). Eruptions occurred throughout the Cenozoic (Wellman and McDougall, 1974; Vasconcelos et al., 2008), with the most recent activity as little as 4.3 kya (Blackburn et al., 1982; Robertson et al., 1996).

The products of intraplate volcanism were particularly abundant in SE Queensland during the mid Cenozoic, extending over the state border into New South Wales (Figures 1 and 2). The largest Cenozoic volcanoes on the continent are located in this region; the Main Range volcano is over 80 km N-S, and Tweed has a diameter of c. 100 km (Figure 1b). This region also hosts some of the greatest diversity of magma types found in eastern Australia, and an unusually large volume of highly fractionated rhyolites (Ewart et al., 1985; Ewart and Grenfell, 1985; Ewart et al., 1988). Erosion of the volcanoes has produced a wide range of spectacular landforms, exposed the magmatic plumbing systems, and resulted in a variety of habitats and ecosystems (Figures 3, 4) (Hunter, 2004). These factors make SE Queensland and northern New South Wales some of the most important intraplate volcanic zones on the continent. This paper reviews the chronology, petrology, and geomorphology of volcanism in this region.

## Time-space distribution of volcanism, and the plate tectonic record

Eastern Australia is an ideal natural laboratory to investigate the origins of intraplate volcanism, as three types of volcano are recognised (Figure 1): central volcanoes, lava fields, and leucitites, with each type displaying differences in morphology, chronology, and chemistry (Wellman and McDougall, 1974; Johnson, 1989). Central volcanoes comprise a central zone of silicic intrusions and flows, surrounded by a variably eroded shield of mafic lavas (Wellman and McDougall, 1974; Cas, 1989). Lava field provinces, in contrast, are comprised almost exclusively of mafic rocks (Wellman and McDougall, 1974; Ewart, 1989), and are generally aerially extensive and thin, although piles up to 1000 m thick were formed in some early Cenozoic cases (Cas, 1989). Leucitites are volumetrically minor, comprising a cumulative total of <3 km$^3$ scattered over small outcrops in inland New South Wales and Victoria (Figure 1) (Wellman and McDougall, 1974; Cohen et al., 2008).

There are no well-defined time-space groupings or progressions for lava field eruptions, with ages spanning from Late Mesozoic to Holocene (Wellman and McDougall, 1974; Johnson, 1989; Vasconcelos et al., 2008). A variety of formation mechanisms have been proposed (e.g., Lister and Etheridge, 1989; Sutherland, 1991,

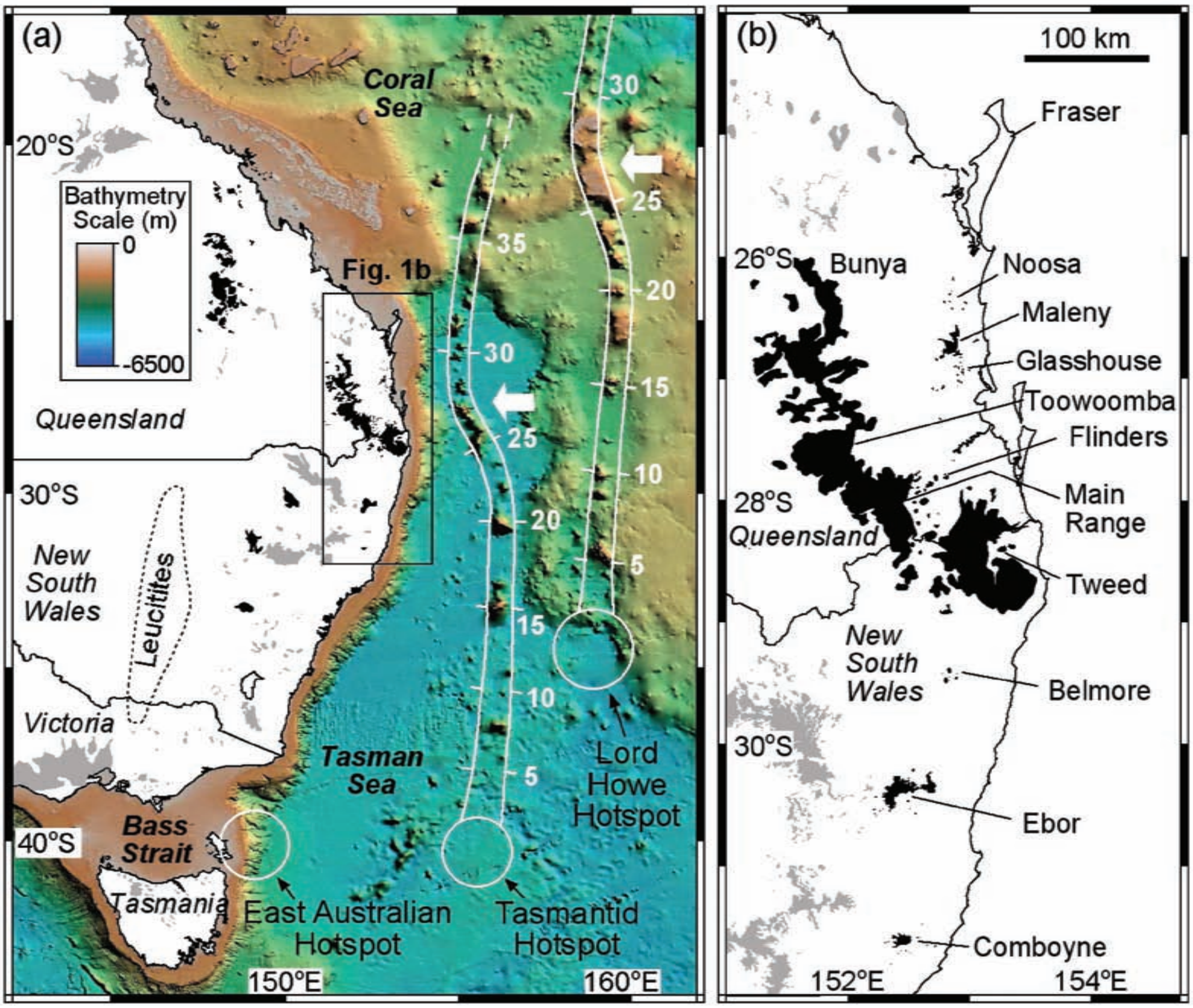


***Figure 1 Widespread intraplate volcanism in a) the Australian region and b) SE Queensland. Central volcanoes are black; lava fields are grey (after Knesel et al., 2008).***

1992, 1993; Demidjuk et al., 2007; Conrad et al., 2011) although none of these models explain the complete spectrum of spatial, geochemical, and temporal features of the lava fields (see Johnson, (1989) and Vasconcelos et al., (2008) for reviews).

Central volcanoes, on the other hand, become progressively younger towards the S, and record the northward velocity of the Australian plate over a mantle plume that is presently located beneath Bass Strait (Figure 1a) (Wellman and McDougall, 1974; Knesel et al., 2008). Plume-derived volcanism also occurs offshore, with two age-progressive seamount trails in the seas E of the continent (Figure 1a) (McDougall and Duncan, 1988; Quilty, 1993). Leucitite $^{40}$Ar/$^{39}$Ar ages are similar to the adjacent central volcanoes, and also become younger to the S, consistent with a hotspot-related origin (Cohen et al., 2008).

In SE Queensland central volcanoes dominate, with only minor occurrences of lava field eruptions (Figure 1b). $^{40}$Ar/$^{39}$Ar geochronology of these central volcanoes indicates that shield-forming mafic eruptions occurred over a protracted period of 3–5 Myr, whereas the silicic units were emplaced within ≤1 Myr, towards the end of the main period of mafic activity (Cohen et al., 2007; Knesel et al., 2008). As such, the silicic rocks provide consistent and distinctive markers in the history of central-volcano activity, allowing detailed age comparisons to be made between volcanoes.

$^{40}$Ar/$^{39}$Ar ages for these silicic units become younger towards the S, but this migration was not constant, with a >50% decrease in migration rate occurring from 26 until 23 Ma (Figure 2). At the same time, the Tasmantid and Lord Howe seamount chains have abrupt and brief changes in direction, with a distinct NW trend from 26–23 Ma (indicated by the large white arrows in Figure 1) that is in sharp contrast to the NNE trends before c. 26 and after c. 23 Ma (Knesel et al., 2008). Plate tectonic modelling shows that the Tasmantid hotspot was fixed relative to the Pacific hotspot reference frame for at least the last 40 Ma (Gaina and Müller, 2000), indicating that the bends in the seamount chains (Figure 1) reflect changes in Australian plate motion. In addition, paleomagnetic analyses of the Main Range and Tweed volcanoes also demonstrate a westward plate excursion at this time (McElhinny et al., 1974; Wellman, 1975). Taken together, these pieces of evidence indicate that during the mid Cenozoic, the Australian plate underwent abrupt and brief changes in plate velocity, from fast NNE velocity prior to 26 Ma, to slow NW motion between 26–23 Ma, and recovering to fast NNE velocity after 23 Ma (Knesel et al., 2008).

This transient ‘crash’ in Australian plate velocity is correlated to oblique collision of the Australian plate with the massive (Greenland-sized) Ontong Java plateau, which sits on the Pacific plate (Knesel et al., 2008). This collision acted to cease N-dipping subduction of the

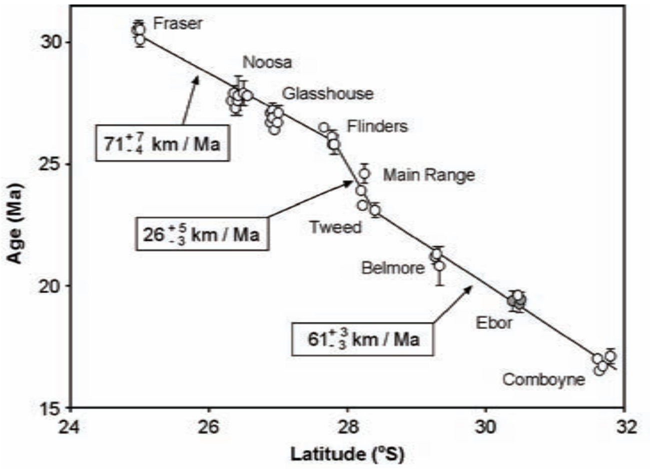


***Figure 2 Ages for the volcanoes become steadily younger towards the S, tracking the Australian plate velocity over a mantle plume (after Knesel et al., 2008).***

Australian plate (Knesel et al., 2008) and also changed the motion of the Pacific plate, as indicated by bends of the same age in Pacific seamount trails, including the Hawaiian and Louisville chains (Kroenke et al., 2004). The high-resolution $^{40}Ar/^{39}Ar$ study of Australian central volcanoes by Knesel et al. (2008) demonstrates the power of such analyses in investigating plate-tectonic and geodynamic processes, as also exemplified by geochronologic analyses of other intraplate volcanic chains (e.g., Koppers et al., 2004; Sharp and Clague, 2006).

## Magma petrogenesis: influence of changing plate velocity, and interaction of plume-derived magmas with continental lithosphere

In addition to these tectonic observations, the Australian central volcanoes also provide an excellent case study of plume-derived magmas interacting with the continental lithosphere. Lavas erupted in SE Queensland and northern New South Wales include tholeiitic basalts, mildly alkaline basalts, highly alkaline mafic units (e.g., nepheline hawaiites, leucite basanites), and highly evolved silicic units (metaluminous trachytes, peralkaline rhyolites, and peraluminous rhyolites) (Ewart et al., 1980; Ewart et al., 1985; Ewart and Grenfell, 1985). Intrusive equivalents are also present. This magmatic diversity reflects the influence of different mantle reservoirs, degrees of partial melting, fractional crystallisation, and crustal contamination, as reviewed below. The changes in plate velocity (Figure 2) also affected the characteristics of volcanism in the region.

The Tweed and Main Range volcanoes (i.e., those erupted during the 26–23 Ma period of slow plate velocity; Figure 2) have been recognised as being both anomalously large and containing an unusually high abundance of rhyolitic rocks (e.g., Ewart et al., 1987; Duggan et al., 1993). These two edifices cover a substantially larger area than other central volcanoes (Figure 1), and have a correspondingly greater volume – between 3,000–4,000 $km^3$ in the case of Tweed, compared to 10–800 $km^3$ for other central volcanoes (Duncan and McDougall, 1989). Rhyolitic rocks at these volcanoes also have trace element and isotopic signatures that are indicative of

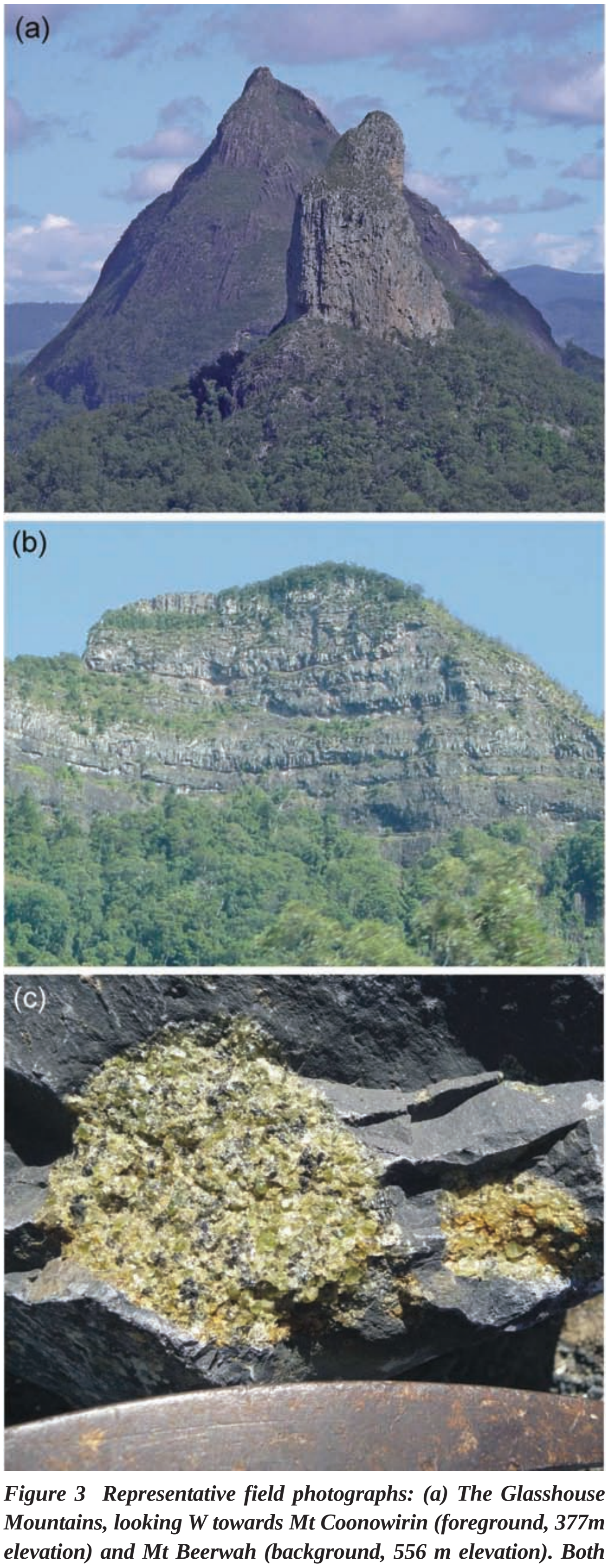


***Figure 3 Representative field photographs: (a) The Glasshouse Mountains, looking W towards Mt Coonowirin (foreground, 377m elevation) and Mt Beerwah (background, 556 m elevation). Both peaks feature very large cooling columns (Photo courtesy of Warwick Willmott). (b) Excellent stratigraphic exposure of mafic lavas at Mt Mitchell in the Main Range volcano. (c) Ultra-mafic xenoliths at Euston Road Quarry, Toowoomba (hammer for scale).***

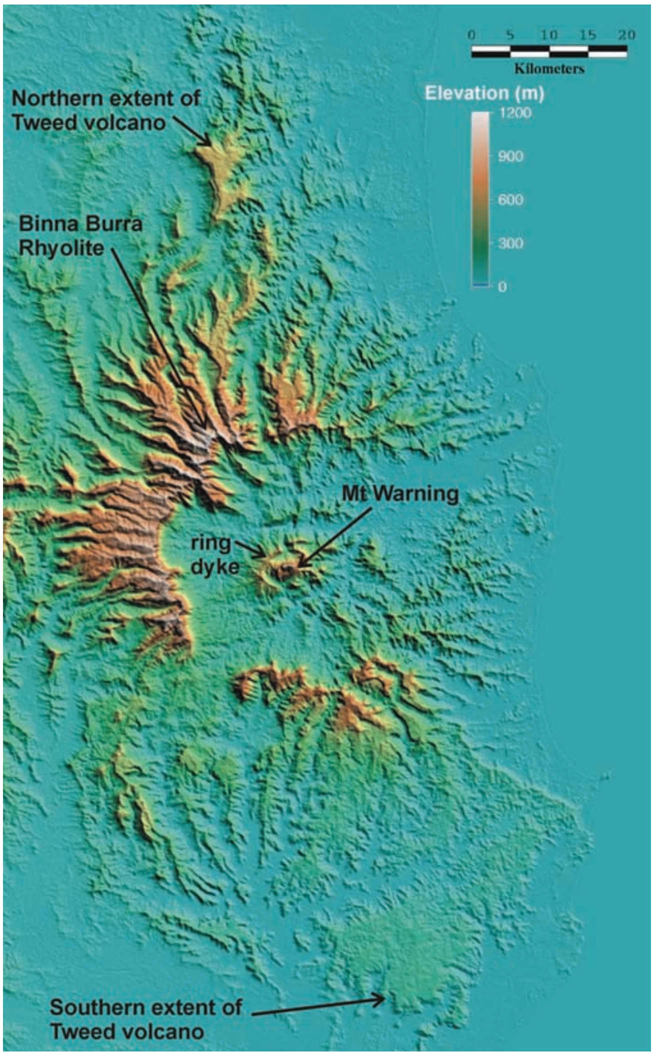


*Figure 4 Digital elevation model of the Tweed volcano showing the original shield shape, with erosion exposing the subvolcanic intrusive complex that is centered on Mt Warning, and surrounded by ring dykes.*

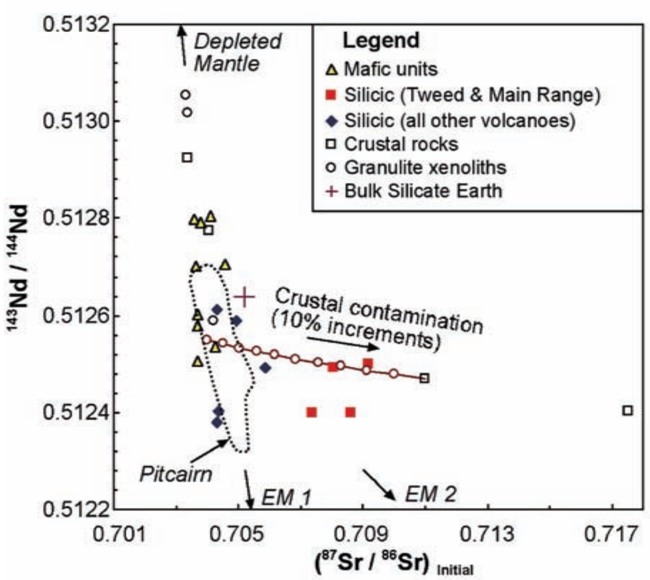


*Figure 5 Sr vs. Nd isotopes for lavas, granulite xenoliths, and crustal materials in SE Queensland (data from Ewar, 1982 and Ewart et al. 1988). Isotopic data from the EM1 hotspot at Pitcairn (Eisele et al., 2002) shown for comparison.*

substantial crustal melting or contamination, including high $(^{87}Sr/^{86}Sr)_{initial}$ values (Figure 5). Such high amounts of crustal influence are not commonly recognised in volcanoes erupted outside the 26–23 Ma period; the implication is that the slower plate velocity, and therefore longer time over the mantle plume, allowed the construction of larger volcanoes, as well as promoting increased melting of the crust (Knesel et al., 2008).

Other than the abovementioned rhyolites from Tweed and Main Range, trace element and isotopic (Pb, Sr, Nd, O) values indicate minimal or no crustal contamination for silicic and mafic magmas (Ewart, 1982; Ewart et al., 1988). This is illustrated by the Sr and Nd isotopic signatures of trachytic and rhyolitic rocks (excluding those from Tweed and Main Range), which largely overlap with the values for the mafic lavas and are isotopically similar to the oceanic Pitcairn hotspot (Figure 5).

Continental crust in SE Queensland is c. 36 km thick, with a further 6–10 km of strong coherent reflecting horizons representing the transitional zone to the mantle (van der Hilst et al., 1998). Although evidently only providing minimal material contribution to most magmas (Figure 5), this thick buoyant continental lithosphere still influenced the magma petrogenesis by acting as a substantial barrier to the upward transport of magmas. Most major and trace element trends point to variable amounts of fractional crystallisation (Ewart et al., 1985). In some cases, particularly for the peralkaline rhyolites, extreme degrees of fractional crystallisation are indicated, with greater than 2000 ppm Zr and 400 ppm Rb recorded – but <1 ppm Sr – indicating extreme enrichments and depletions of incompatible and compatible elements, respectively (Ewart et al., 1985; Ewart and Grenfell, 1985). Rare earth element concentrations are also enriched in these rocks; one peralkaline rhyolite from the Glasshouse Mountains has a La concentration over 700 times that of chondritic meteorites (Figure 6). The rhyolitic rocks also have considerable Europium anomalies (Eu/Eu* as low as 0.0023 for the Binna Burra rhyolite from the Tweed volcano; Figure 6), which can be reproduced by a feldspar-dominated fractionation assemblage (Ewart et al., 1985). These extreme degrees of fractional crystallisation have resulted in distinctive mineralogical assemblages, including low Ca-anorthoclase, Zr-aegirine, fluor-ardvedsonite, ± aenigmatite, ± fayalite, ± hedenbergite, and high-Zn ilmenite (Ewart, 1985).

As a result of extensive fractional crystallisation processes, as well as crustal melting, the various silicic rocks comprise on average c. 30 volume % of the central volcanoes (Wellman, 1971) – a much higher percentage than typically observed at oceanic intraplate islands (e.g., Clague, 1987). Nevertheless, mafic lavas still comprise the bulk of the volcanic products. Variations in trace element and isotopic data among central-volcano basalts in SE Queensland, and elsewhere in eastern Australia, indicate mixing between two distinct source components: an EM1-like source and a depleted upper-mantle source (Figure 5) (McDonough et al., 1985; Ewart et al., 1988; Sun et al., 1989). In some models, the enriched-mantle signature has been interpreted to arise from interaction of plume-derived melts with the subcontinental lithosphere (e.g., McDonough et al., 1985; Ewart et al., 1988; Sun et al., 1989). However, EM1 signatures have also been recorded from the Tasmantid seamounts (Eggins et al., 1991), which sit on oceanic crust, suggesting the EM1 signature may derive directly

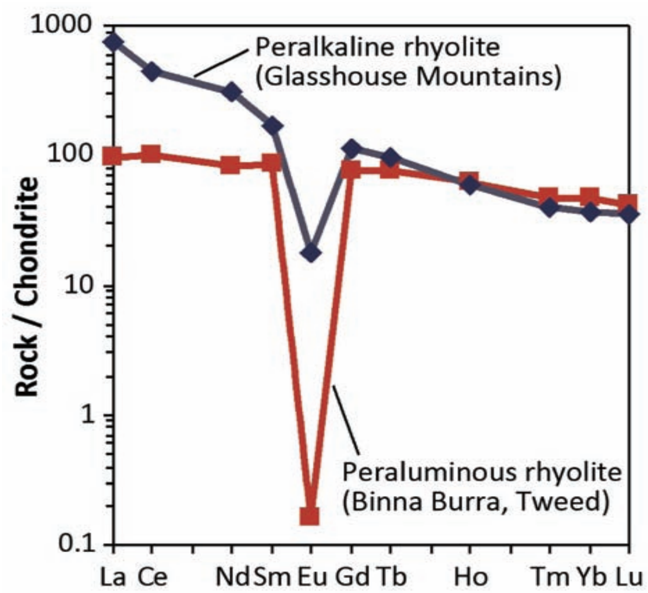


***Figure 6 Chondrite-normalised rare earth element patterns for a comendite from the Glasshouse Mountains, and the Binna Burra high-silica rhyolite from the Tweed volcano (data from Ewart et al., 1985).***

from the plume. The exact source of the EM1 signature therefore remains unresolved. An important – but as yet unused – mechanism to investigate the isotopic systematics of the lithospheric mantle beneath SE Queensland is provided by the abundant xenoliths and megacrysts in this region, which are described below.

## Xenoliths, megacrysts, and alkaline mafic eruptions

Cenozoic volcanoes in eastern Australia have yielded a rich record of xenoliths and megacrysts (O'Reilly and Griffin, 1987) and SE Queensland is no exception, with the Toowoomba and Main Range areas being especially prolific (Figure 3c) (Green et al., 1974; Ewart and Grenfell, 1985; O'Reilly and Griffin, 1987; Sutherland et al., 1990). These inclusions are hosted in silica-undersaturated mafic eruptions of hawaiite, olivine nephelenite, nepheline hawaiite, nepheline benmoreite, leucite basanite, and camptonite (Ewart and Grenfell, 1985). These alkaline eruptions typically commenced with an initial explosive phase producing tuff and breccia, with subsequent more effusive lavas filling the earlier-formed vent (Willey, 2003). These eruptions clearly crosscut or overlie earlier phases of activity, and $^{40}Ar/^{39}Ar$ geochronology of one such eruption from Main Range indicates eruption 3 Ma, after the bulk of the volcano was constructed (Knesel et al., 2008); a timeframe analogous to the postshield- and rejuvenated-stage eruptions on the Hawaiian Islands (Clague, 1987).

Xenolith phases recorded in SE Queensland are dominantly from the upper mantle or lower crust, comprising: Cr-spinel lherzolite, clinopyroxenite, wehrlite, websterite, kaersutite pyroxenite, metasomatised clinopyroxenite, hornblendite, meladiorite, two-pyroxene granulite, and gabbro (Ewart and Grenfell, 1985; Sutherland et al., 1990). Megacrysts have also been recovered, comprising: Ti-Al augite, orthopyroxene, feldspar, garnet, kaersutite, mica, and Fe-Ti oxides, as have a small number of accidental upper-crustal xenoliths (e.g., quartzite, granite) (Ewart and Grenfell, 1985; Sutherland et al., 1990).

Given the volumetric abundance of each xenolith type, these xenolith suites indicate the upper mantle in SE Queensland is dominated by spinel lherzolite, with a granulitic lower crust (Ewart and Grenfell, 1985; Sutherland et al., 1990). Only spinel lherzolites have been recorded in SE Queensland, although garnet lherzolites are found elsewhere in Australia (O'Reilly and Griffin, 1985; Griffin et al., 1987). The east Australian xenolith suites define a paleogeothermal gradient that is substantially hotter than is typical for continental regions (O'Reilly and Griffin, 1985; Griffin et al., 1987). However, no trace elemental or isotopic studies have been undertaken on the ultramafic xenoliths in SE Queensland, which could yield information on the composition of the sublithospheric mantle in this region. As such, this is a prime research target awaiting further investigation.

## Volcanic geomorphology

Despite erosion since the mid Cenozoic, the shield structure is still apparent in some volcanoes, with the original surfaces remaining as elevated plateaux that are cut by a series of radial streams. The Tweed volcano is an excellent example, with the shield flanks remaining on the northern, western, and southern sides (Figure 4). Bunya also retains a shield shape, while Toowoomba and Main Range have only the western flanks preserved (Ewart and Grenfell, 1985). Coastal erosion has progressed further at Noosa and Glasshouse-Maleny, with only small expanses of mafic lava remaining (Figure 1) (Cohen et al., 2007). Meanwhile, in the case of Fraser, the only remaining volcanic outcrops are three isolated headlands comprised of trachytic lavas cut by mafic dykes, which are completely surrounded by Quaternary sand (Cohen et al., 2007). At this volcano, all other volcanic units have either been eroded, or exist below sea level (Grimes, 1982).

Although removing the original mafic shield, this erosion has enabled excellent exposure of the stratigraphy and subsurface architecture of the volcanoes, often also producing spectacular scenery (Figure 3a, b). At the Tweed volcano this erosion has revealed the magmatic plumbing system beneath the shield (Figure 4), which comprises an intrusive complex of gabbros, syenite, granite, and comendite (Duggan et al., 1993). Circular ring dykes surround the complex, and form significant relief (Figure 4). The intrusive complex extends from almost sea level to a maximum elevation of 1159 m at the summit of Mt Warning, revealing a wide range of intrusive environments. A large subvolcanic intrusive complex also exists at Mt Barney in the southern Main Range volcano (sometimes called the Focal Peak volcano, e.g., Ewart et al., 1987).

Other volcanoes in SE Queensland apparently lack a single main subvolcanic intrusion, instead containing multiple plugs, sills, dykes, and laccoliths. The Glasshouse Mountains are an excellent example, where over 15 eroded hypabyssal plugs and laccoliths rise abruptly from the coastal plain (e.g., Figure 3a). Numerous subvolcanic plugs are also present in the Fassifern Valley E of Main Range, and in the Noosa region (Ewart and Grenfell, 1985; Cohen et al., 2007).

The rugged eroded volcanic landforms are the most prominent geographic landmarks in SE Queensland. As a result, the volcanoes feature in Aboriginal legends, and were also observed by early European explorers. Mt Warning, the prominent core of the Tweed volcano (Figure 4), was sighted on the 16th of May 1770 by the British explorer, James Cook, and named as a warning to subsequent mariners of dangerous shoals that exist off the coast in this region

(Hawkesworth, 1773). This peak, also known as Wollumbin, is sacred for the Bundjalung Aboriginal people. Further N, the Glasshouse Mountains were named by Cook on the 17th of May 1770, who described the peaks as “very remarkable on account of there [sic] singlar [sic] form of elevation which very much resemble glass houses which occasioned my giving them that name” (Hawkesworth, 1773). The peaks themselves are known after characters in a Dreamtime story of the Gubbi Gubbi (or Kabi) people, with distinctive names such as Miketeebumulgrai, Coochin, Beerburrum, Ngungun, Tibrogargan, Coonowrin, and Beerwah.

## Volcanic influence on flora, fauna, and human land-use

The rock types and geomorphology of the volcanoes also significantly influence the vegetation and land use patterns in this region. Volcanic plateaus underlain by mafic lavas have fertile soils, supporting dense subtropical rainforests, especially in regions with greater rainfall (Willmott, 2003; Hunter, 2004). In contrast, regions underlain by rhyolite have poorer soils, and wet or dry sclerophyll eucalypt forests have developed, adding to the biodiversity of the region (Willmott, 2003; Hunter, 2004). Steep and inaccessible areas frequently remain uncleared, and, as a result of rugged scenery and diverse biology, many of the volcanic regions have been protected as National Parks, which are important tourist attractions in the region. Of particular significance are those parks on either side of the New South Wales/Queensland state border, which are part of the World Heritage listed “Gondwana Rainforests of Australia” (formerly the “Central Eastern Rainforest Reserves of Australia”, Hunter, 2004).

Meanwhile, some fertile regions developed on mafic lavas were cleared for farming, especially on the Maleny plateau, around Toowoomba, and the flanks of the Tweed volcano. Where cleared, steep slopes are subject to landslides, a hazard that has increased in recent years as the volcanic regions become increasingly attractive for residential and rural housing (Willmott, 2010). Volcanic rocks are quarried for road and concrete aggregate, boulder stone, perlite (hydrated volcanic glass), as well as diatomite from sediments deposited between flows. Quarries in mafic plugs are particularly common around Toowoomba (Willey, 2003). Some rhyolite or trachyte plugs around Noosa, the Glasshouse Mountains, and Main Range have also been exploited, but most of these quarries are now inactive. In all cases, quarrying is limited by the often-steep volcanic terrain, preservation in National Parks, and by the increasing urbanisation of SE Queensland.

# Conclusions

The mid-Cenozoic volcanoes in SE Queensland provide a valuable record of intraplate volcanic activity on continental Australia. The time-space distribution of the volcanoes indicate a major reduction of northward Australian plate velocity from 26–23 Ma, correlated to initial collision of the massive Ontong Java plateau with the northern subduction margin of the Australian plate. The rocks also record the interaction of the plume-derived magmas with continental lithosphere, with varying degrees of melting, fractionation, and crustal contamination producing a wide variety of magmas. Erosion of the volcanoes has created diverse rugged landscapes with excellent exposures of subvolcanic magmatic systems. These eroded volcanoes also host considerable biodiversity, and the volcanic landforms feature prominently in both Aboriginal legends and the early European exploration of the region. As a consequence of this varied geologic, biologic, and human history, these volcanoes represent important sights of research and tourism, both in the past and into the future.

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*by John S. Jell and Gregory E. Webb*

# Geology of Heron Island and Adjacent Reefs, Great Barrier Reef, Australia

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*Heron Island has been the focal point for research on the southern Great Barrier Reef Province for the last 80 years. Heron Reef is an excellent example of a lagoonal platform reef with a sand cay developed on its leeward end, displaying typical reef morphological, sedimentological and ecological zonations allowing comparison of their windward and leeward development. Limited subsurface data indicate that the total reef section is only 150 m thick, consisting of stacked limestone packages, with a gently eastward sloping solution disconformity delineating the base of c. 15 m of Holocene reef growth. Holocene reef growth does not appear to fit the "classical" model, with evidence of much progradation on the windward margin relative to the associated leeward margin. Large dredged blocks of reefal material provide new data on the abundance of in situ framework in much of the reef and the importance of microbialite in the unification process.*

## Introduction

Although Heron Reef is one of the most thoroughly studied modern coral reefs, sparse subsurface data suggest that it has much to teach us about late Quaternary reef development. The geomorphology of modern coral reefs reflects their evolution through Holocene sea-level rise. Various reef growth models have been proposed but variable settings (tectonics, tidal range, wave energy, storm activity, etc.) suggest potentially very different histories in different regions (e.g., Montaggioni, 2005; Montaggioni and Braithwaite, 2009). Although reefs are known to contain abundant sand and rubble, generally more or less enclosed within a 'framework' composed of in situ coral facies (e.g., Davies and Hopley, 1983), the degree to which in situ reef framework or reef rubble dominates reefs is controversial (e.g., Hubbard et al., 1990; Blanchon et al., 1997; Riegl, 2001). Unfortunately, most of what is known about Holocene reef evolution results from analysis and dating of relatively few drill cores (e.g., Hopley et al., 2007; Montaggioni and Braithwaite, 2009). Also, cores commonly target a particular facies within a reef, thus not allowing the lateral evolution of reef facies through sea-level rise to be determined (Hopley et al., 2007). Additionally, differentiating in situ coral framework and larger coral rubble in cores combined with general poor core recovery complicates the interpretation of reef cores (Hubbard et al., 2001).

Where multiple cores have been recovered from a single reef, growth models can be constructed. The reef growth model of Marshall and Davies (1982) suggests that following Pleistocene substrate inundation c. 8 ka, high-energy coral head facies and lower energy branching coral facies aggraded on the windward and leeward sides, respectively, with both initially lagging behind sea-level rise and the windward facies attaining modern sea-level sooner than the leeward after sea-level began to stabilise. Once sea-level was attained on the windward margin, progradation was directed primarily towards the backreef/lee. However, Hopley et al. (2007) noted from sparse data, that many GBR reefs evolved differently. Heron Reef affords observations of reef framework in large (m+) blocks dredged from the reef flat on the western end during enlargement of the boat harbor in 1988. These blocks provide good examples of the in situ coral framework. This paper proposes to place the Heron Reef framework and limited coring data in the context of regional geology of the southern Great Barrier Reef Province (GBRP).

## Regional Setting

Heron Reef, in the southern GBRP had its geology outlined by Jell and Flood (1978) and the origin and evolution of the GBR have been reviewed by Davies (2011). The GBRP includes eastern Queensland's continental shelf that has been occupied or influenced by reefs and reef-derived sediments at any time since the initiation of reef growth. It represents the western margin of the Northeast Australian Carbonate Platform System, which also includes the reef provinces of the Gulf of Papua and the Eastern, Queensland, and Marion plateaus (Davies et al., 1988; Figure 1). The offshore plateaus and intervening troughs appear to be modified and subsided continental crust that originated from late Cretaceous–Paleogene rifting of the northeastern Australian Plate (Symonds et al., 1983).

At 2,300 km long, from Bramble and Anchor cays in the Gulf of Papua (latitude 9$^o$15'S) to S of Lady Elliot Island (latitude 24$^o$10'S) and covering 265,000 km$^2$, the GBRP is the largest reefal and carbonate-siliclastic shelf system in the present oceans, and is comparable in size to some of the large carbonate platforms in the geological record.

The continental shelf is mostly a shallow (0–90 m, average 40 m), partially rimmed, high-energy platform, sloping gently (<1–4 m/km) from the coast to the shelf edge and deepening southward. Apart from the latter and the N being wetter than the S, there is little latitudinal variation. It is narrowest off Cooktown broadening to the N across Torres Strait into the Gulf of Papua and to the S off Broad Sound where the reefs occupy only mid to outer shelf areas. The shelf narrows considerably at its extreme southern margin where it is embayed by the Capricorn Channel. Throughout the length of the

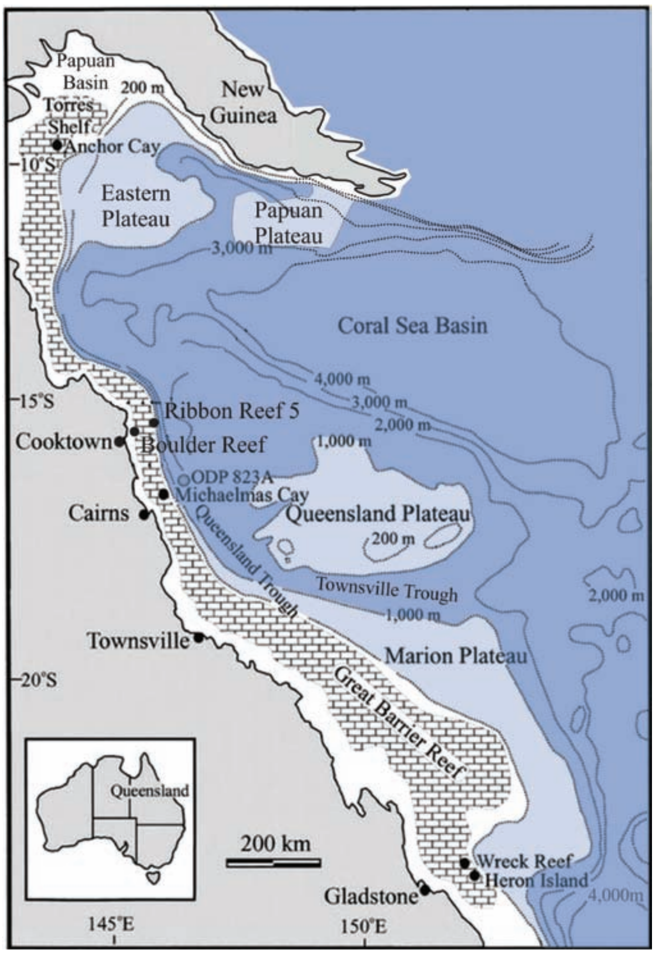


***Figure 1 Location Great Barrier Reef within the Northeast Australian Carbonate Platform System showing the boreholes referred to in the text (after Braithwaite et al., 2004).***

GBR there is a noticeable cross-shelf regional variation influenced primarily by wave energy and water quality.

Approximately 3,500 individual reefs occupy c. 25,000 km$^2$ within the GBRP, less than 10% of the Provence. More than 86% of the reefs occur in a 25–65 km-wide band along the eastern margin of the shelf. Maxwell (1969) referred to this zone of maximum reef development as the *Reef Zone*, which is separated from the *Near-shore* Zone by what is commonly referred to as the *Shelf Lagoon*. Hopley et al. (2007) provided detailed spatial data of the reefs for the Great Barrier Reef Marine Park based on available bathymetric surveys and satellite imagery.

Heron Reef at 23° 27'S, 151°55'E is one of the Capricorn Group, which with the Bunker Group, Lady Elliott Reef (24°07'S) and four northern shoals comprise the southernmost GBRP (Figure 2). Fourteen cays occur on 12 of the reefs with leeward sand cays more common than windward rubble cays. The reefs occur on the mid to outer shelf 80 km offshore. Water depths are 35–60 m and the reefs are 10–20 km W of the incised continental shelf edge (100 m isobath) in a zone of pure carbonate sediment (Maiklem, 1970). The continental slope is gentle and terraced with algal buildups at depths of 80 to 120 m (Davies et al., 2004). No drowned reefs have been reported from the continental edge in this region, although apparently in situ reef material was reported from the continental slope at 175 m (Veeh and Veevers, 1970; Yokoyama, et al., 2006) and drowned coral reefs occur farther N (Beaman et al., 2008; Abbey et al., 2011). The reef zone is separated from the terrigenous sediment-dominated inner shelf, by the Curtis Channel. Reefs are aligned on intersecting lineaments. The Bunker High extends 125 km from Lady Elliot Reef to North Reef. The other highs trend SW perpendicular to the shelf edge and parallel to prominent bathymetric trends in deeper water, with one line extending from Sykes Reef to Polmaise Reef (50 km) and another, shorter line, extending from Broomfield Reef to Northwest Reef.

The sea floor between reefs is flat or nearly so, except where reefs are close together with the bottom rough and commonly steep. Reef slopes in the channel between Heron and Wistari reefs are up to 45º. In general, leeward reef slopes are gentler than windward. Shoal areas connect many of the reefs, the most prominent being between Sykes and Heron reefs and the Isbell Shoal, connecting Northwest, Wilson and Broomfield reefs. In both cases the shoals rise from the shelf floor to –15 m. Prominent surfaces and terraces occur around the slopes of Heron, Erskine, Masthead, Polmaise, Wistari, Sykes, Wreck and Tryon reefs at –5 to –7 m, with an additional prominent –15 m surface around Heron, Sykes, Lamont and Fitzroy reefs.

## Geological framework

Basement of the GBRP is part of the Tasman Fold Belt, which forms the tectonic framework of eastern Queensland. Its history dates from the Ordovician and is represented by a series of NNW trending fault-bounded basins and highs. In the Mesozoic, intra-cratonic downwarps developed across the present coastline, such as the Maryborough Basin containing late Triassic-early Cretaceous clastic and some volcanic rocks deposited in continental to paralic and occasional shallow marine environments.

The present structure of the southern GBRP was shaped by continental break-up and rifting in the Late Cretaceous-Paleocene with the development of the Capricorn Basin between the western Bunker High above the Maryborough Basin and the Swains Reef High to the E (Maxwell, 1968). Rifting was initially associated with the spreading of the Tasman Sea to the SE, and then Palaeocene spreading of the Coral Sea Basin. Spreading terminated by the end of the Palaeocene–early Eocene and most subsequent movement has been vertical. Accompanying this spreading was the uplift of the adjacent mainland together with widespread Cenozoic volcanic activity.

The Cenozoic of the GBRP is represented along its central part by five depositional packages of sediment: (1) A thin syn-rift sequence of ?late Cretaceous–Paleocene, mainly continental to marginal marine, alluvial deposits; (2) Paleocene–Late Eocene marine onlap sequence, transgressive at its base but becoming open marine in its upper part, formed during continued cooling and subsidence; (3) A sequence of progradational-recessive phases of Late Oligocene, late Miocene, and late Pliocene–early Pleistocene age with progradation of fluviatile and wave-dominated shelf-margin deltaic sedimentation, and regressive phases of marine onlap facies; (4) A shelf aggradation/progradation phase of Pleistocene age representing sea-level fluctuation, concurrent subsidence and deposition, becoming more marine in its upper section; (5) Late Pleistocene–Holocene reef development with pro-delta sediments on the inner shelf (Symonds, et al. 1983).

Until a few decades ago the GBR was widely considered to be millions of years old, even Miocene. However, Marshall (1983) and Symonds et al. (1983) suggested that it may be no older than

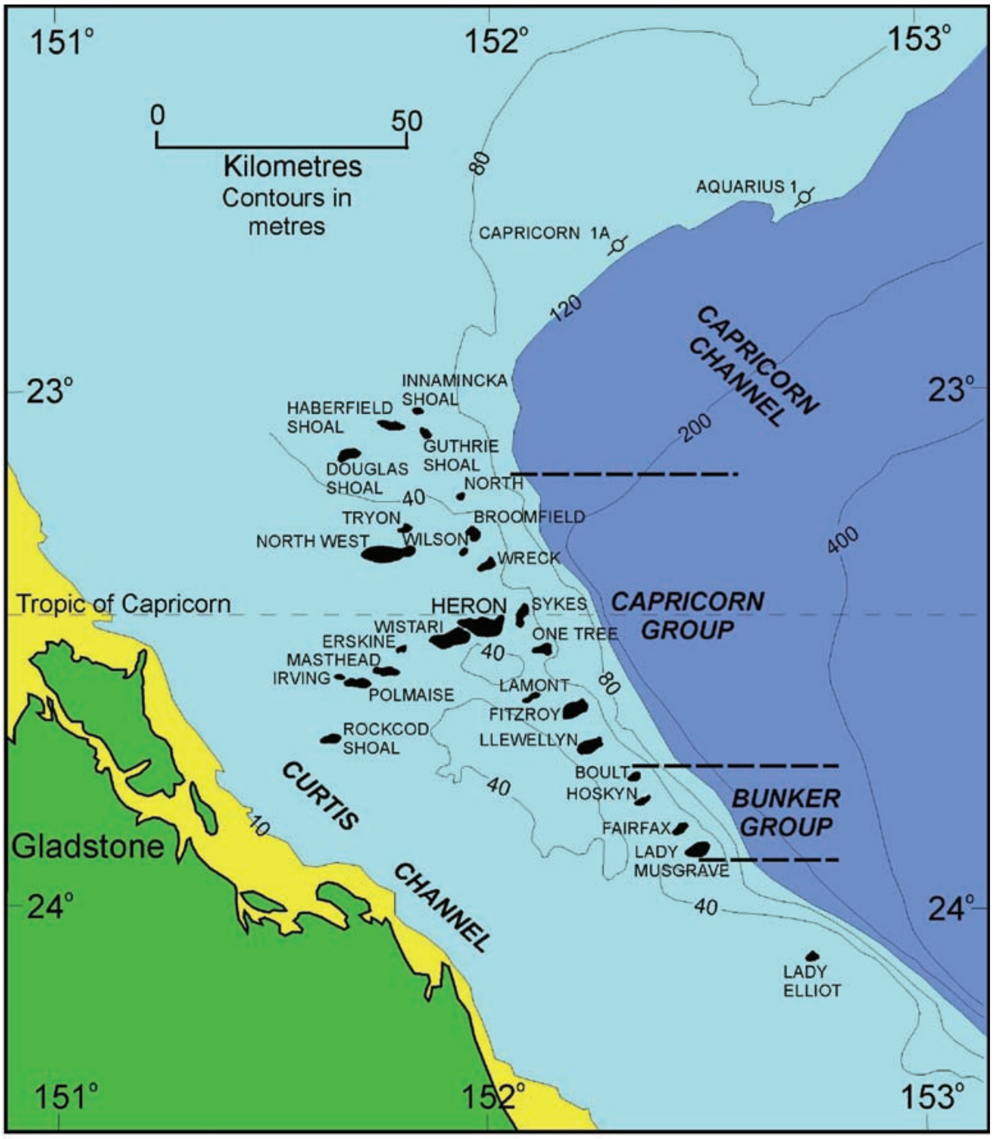


*Figure 2 Map of the reefs of the southernmost Great Barrier Reef showing position of Heron Reef.*

Pleistocene. Seismic data and sediment cores from the continental slope suggested that the reef may have initiated as late as c. 500 ka (Davies and Peerdeman, 1998). However, 'deep' cores from the GBR itself (Ribbon Reef 5 and Boulder Reef) constrained initiation of the central GBR to 452–365 ka (see Braithwaite et al., 2004). Those dates were questioned by Dubois et al. (2008) who suggested an age of initiation for southern central GBR at 670–560 ka on the basis of sediment cores from the Marion Plateau.

A core from Heron Reef in 1937 reached a total depth of c. 220 m including reef limestone to 156 m (Richards and Hill, 1942). Unfortunately, recovery was poor and the remains of the core are unlabelled rubble; no reliable dates were recovered. Hence, the age of initiation of the southernmost part of the GBRP remains unconstrained. Davies (1974) recognized the Pleistocene-Holocene boundary at c.15-20 m depth. The broad shoal that supports both Heron and Sykes reefs rises to c. 15 m depth (Jell and Flood, 1978), dipping slightly eastward, suggesting that both reefs represent Holocene growth on the same older Pleistocene reef basement. Ribbon Reef 5 core, in the northern GBR, suggests four to six packages of reef growth in the upper c. 96 m with the upper four successions separated by unconformities associated with glacial lowstands (see Braithwaite and Montaggioni, 2009). Flood (1993) recognised five solution unconformities within the Heron bore reefal sequence and it is assumed that latest Pleistocene aggradational reef facies of the southern GBR reefs consist of stacked highstand reef packages separated by unconformities associated with intervening lowstands.

## Prevailing physical conditions

Maxwell (1968) and Hopley (1982) summarised the hydrology, climate and weather of the GBRP and Wolanski (1994) reviewed its oceanography and hydrodynamics. Maiklem (1968), Brandon (1973) and Marshall (1977) provided general details for the Capricorn Group. The Commonwealth Bureau of Meteorology, has recorded data since 1963 from weather stations on Heron Island, North Reef and Lady Elliot Island.

The climate is subtropical, with distinct summer and winter with monthly average maximums of 21.5–30°C and minimums of 16.5–24.2°C. Sea surface temperature averages 20°C in winter and 27–28°C in summer. Rainfall averages 1,028 mm/year spread throughout the year with February to June the wettest months. The prevailing wind is the South East Trade, which blows for approximately 70% of the year with a mean velocity of 12–18 km/hr but with velocities greater than 36 km/hr not uncommon. The summer months experience calms or NE–NNW winds and occasional cyclones. During winter, strong westerlies develop for short periods. Cyclones with winds up to 180 km/hr superimpose their effects over the trade winds and significantly influence reef morphology (Flood and Jell, 1977).

The dominant South East Trades induce westward reef surface drift, but their velocities are generally low and their effect strongly affected by tidal current oscillation. Around Heron Island, the tidal range for spring tides is 2–3.3 m and for neaps 0.8–1.6 m. On the outer shelf the flood tide sets W and the ebb E producing strong currents through the Capricorn reefs; currents in the channel between Heron and Wistari reefs can reach 7.5 km/hr. As the tide falls on the reef top, drainage is crudely radial until the rim becomes exposed when water flows through any gaps in the rim which are mainly leeward. The lagoons experience slack water for several hours during each tidal cycle. Thus shelf waters are replaced twice daily with oceanic waters. Two surface oceanic currents influence the region: the S-flowing East Australian Current, which flows parallel to, and only slightly impinges on, the shelf; and a northerly current during spring and into summer, which circulates water around the reefs.

Ocean swells of 1–3 m predominate from the ESE. Waves breaking on the reef may exceed 2 m and they refract around the reef producing lateral transport of sediment from windward to leeward. Sediment is deposited and may accumulate where wave sets converge (e.g., sand cays). Even at low tide with the reef top waters isolated from the ocean swell and waves, the wind shear over the shallow waters is sufficient to agitate sand-sized particles and to keep silt-sized particles in suspension.

## Reef types

Fairbridge (1950), Maxwell (1968), Hopley, (1982), and Hopley et al. (2007) described three major reef types: (1) Linear or Ribbon reefs that grow along the shelf edge and only grow landward because water depths increase too rapidly on the continental slope; (2) Platform reefs that rise from the shelf floor and grow in any direction depending on the hydrological regime; and (3) Fringing reefs that grow seaward away from land or continental islands.

The reefs of the Capricorn-Bunker groups are platform reefs. Hopley (1982) recognised six stages of platform reef development, and apart from the submerged reefal shoals, all the southern reefs are mature or senile (i.e., lagoonal or planar). Their size and shape depend on that of their underlying foundation or 'antecedent platform'. They vary in maximum dimension 1–11 km with corresponding surface areas of c. 1–40 $km^2$. Reefs with diameters <3 km are all planar, as reef growth from their perimeter coalesces. Larger reefs trend from: a) deep lagoons with widely spaced, large isolated patch reefs; through b) shallow lagoons with numerous isolated patch reefs; through c) very shallow lagoons with reticulate patch reefs; to d) planar reefs without lagoons (e.g., westward Fitzroy–Heron–Wistari–Masthead). This progression primarily represents stages of lagoon infilling by reef detritus, and may depend on topography and depth of the antecedent platform, subsidence of the foundation, rate of reef growth or production rate of detritus. Subsurface data are too limited to assess this. The other factor influencing reef form is the hydrological regime, differentiating reefs on the relative development of windward and leeward features.

Heron Reef (Figure 3) is a moderate sized (9.5 x 4.5 km), lagoonal, platform reef with a vegetated sand cay (800 x 300 m) on its leeward end. The Pleistocene surface on which the reef grew was 10–20 m (bmlw) as identified by refraction profiles adjacent to the island (Harvey, 1986), reflection seismic profiling (Smith et al., 1998) and the solution unconformity in the Heron Island bore. The unconformity dips 0.03° to the E at c. 14 m (bmlw) and can be traced across the intervening shoal into Sykes Reef where it is slightly higher. The Heron lagoon is 3–3.5 m deep and has numerous patch reefs. Heron reef has a well developed windward margin broken only in the SE by a narrow channel, and a less well developed leeward margin indented and breached in several places such as at Blue Pools NE of the cay.

## Reef morphology

Maxwell (1968) and Hopley (1982) outlined variation in GBRP reef morphology, showing the inter-relationship between morphological, ecological and sedimentological zones that parallel the line of intersection of the reefal mass and the water surface at low tide. Maiklem's (1968) zonation of Capricorn Group reefs was followed by Jell and Flood (1978) who provided a detailed summary for Heron Reef, that may be even further subdivided (Figure 4).

### *Windward Reef Slope*

The reef slope is steeply inclined (10–40°) with two distinct terraces 4–6 m and 15–20 m (bmlw). The *Reef Front Subzone* is commonly covered with a profusion of staghorn coral, patches of platy forms on ledges and domal massive corals on the upper terrace and at the top of the slope. In places spur-and-groove structures are well developed. Exposed during low spring tides, the upper surfaces of spurs, support luxuriant low-profile growths of *Acropora* spp. and massive corals, whereas the growing edges and the terrace support branching and platy forms. Large, massive, domal corals are found on the fronts of spurs and on the terrace onto which the grooves open. Sandy gravel groove floors are devoid of coral growth. Corals decrease markedly below –10 m, and the *Lower Reef Slope Subzone* is a coral veneered cemented limestone mass with fans of coral debris, sloping gently down to the shelf floor. On the eastern margin, the reef slope flattens out at –15 m onto the shallow platform between Heron and Sykes reefs.

### *Reef Rim*

The reef rim is the highest part of the reef top being up to 10 cm above reef flat coral growth, and 60 cm above the lowest spring tide level. It is continuous around the perimeter of the reef except for gaps along the western tip and northeastern section. It slopes gently seaward and is irregularly terraced. Three subzones are typically recognizable and variously developed around the reef:

1. The outer *Coralgal Subzone* faces the oncoming breaking waves


***Figure 3 Composite aerial photograph of Heron Island and surrounding reef, showing zonation and position of boreholes and source of reef blocks.***

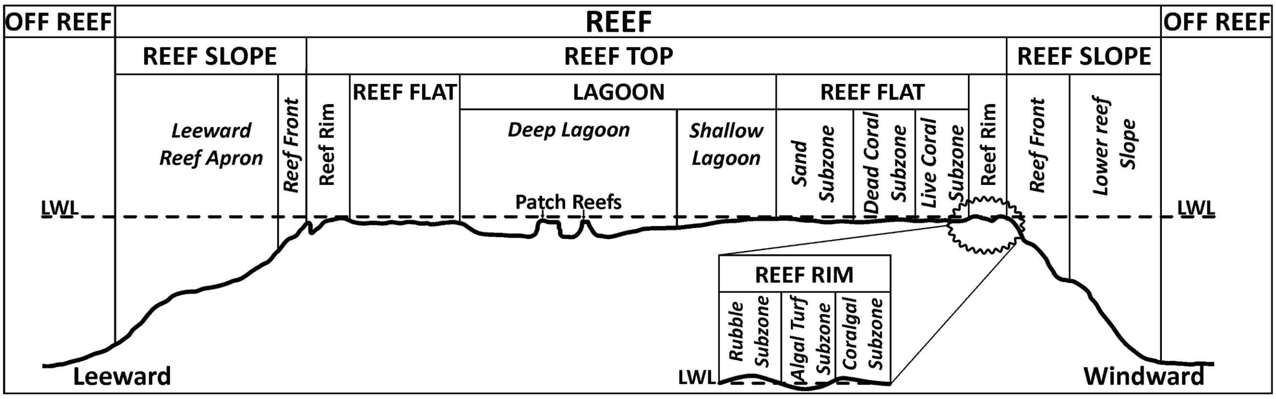


*Figure 4 Generalised cross-section of a lagoonal platform reef such as Heron Reef showing the recognisable zones and subzones.*

and is a smooth terraced pavement consisting of low profile corals heavily encrusted by coralline algae. The pavement is crossed by narrow runoff channels, and broken by sand floored pools rimmed with luxuriant coral growth. Rhodoliths occur in small pools.

2. The *Algal Turf Subzone* is a slightly lower area of the pavement encrusted by a mat of algae and some domal corals growing in its deeper parts. It has been referred to as the outer moat.
3. The *Rubble Subzone* is the highest varying from only a few to tens of metres wide and from thin sheets 10 cm thick to 1.5 m rubble banks. Extensive tongue-like shingle banks, at right angles to the eastern and southeastern margins, are common. Other banks occur W of the harbour entrance and W of Blue Pools NE of the cay. Rubble varies from *Acropora* shingle (5 x 1 x 1 cm) to coral plates (20 x 10 x 2) cm to reef blocks >2 m in diameter. This subzone's outer boundary parallels the reef edge representing the inner extent of transport of the largest blocks whereas the inner boundary is typically finger-like reflecting transport of finer material into slightly deeper water. Rubble in the tongues is usually encrusted with coralline algae and coral growth is restricted to small pools within the rubble.

The Reef Rim isolates reef top waters from open shelf waters for several hours each low tide with water dammed up to 50 cm above outside waters creating strong runoffs through any gaps in the rim and influencing sedimentation patterns. In several places both windward and leeward double rims are developed.

## Reef Flat

The reef flat is exposed at low tide within the reef rim. Corals grow to the level of the dammed water and are commonly heavily encrusted by coralline algae. The encrusted surface provides a very level surface. Four subzones are:

1. The *Outer Living Coral Subzone* has >50% coral cover with more living than dead displaying a radial pattern perpendicular to the reef edge with coral pools and channels up to 80 cm deep. Its outer boundary is difficult to distinguish because of areas of coral growth on reefal substrate and growth of corals and coralline algae on tongues of rubble extending from the rubble subzone. This subzone has the highest density and diversity of corals.
2. The *Dead Coral Subzone* has similar coral coverage, but dead predominate over live corals and fleshy increase over coralline algae. It also becomes shallower and coral diversity decreases dramatically.
3. The *Sand Subzone* has <50% coral cover, rare microatolls and bioturbation is extensive.
4. In places an *Inner Coral Subzone* is marginal to the lagoon, e.g., off the NE end of the cay. It has slightly deeper water and c. 50% coral cover with *Halimeda* and fleshy algae common.

## Lagoon

The central lagoonal zone is a complex system of varying water depths and patch reefs. Two subzones (Smith et al., 1998) are:

1. The *Deep Lagoon* is 3–3.5 m (bmlw) with numerous 6–25 m patch reefs predominantly in the N and E where they occupy up to 50% of the surface. Patch reefs are mainly reefal limestone encrusted by coralline algae with scattered corals, *Halimeda* and fleshy algae. They supported much more coral growth before late 1960's cyclones. The northeastern lagoon is isolated by an E-W ridge that was thought to be a relic coral-algae ridge, but recent inspection in two areas indicates a thin rib of rubble. In the NW the deep lagoon is divided by an offshoot of the shallow lagoon. The fine lagoonal sediment contains a rich infauna and extreme bioturbation.
2. The *Shallow Lagoon* is typically 1 m deep and separated abruptly from the Deep Lagoon. It can be further subdivided into areas with and without sporadic coral growth. The former may represent coral growth on recently buried patch reefs. After calm weather the sediment surface is commonly blanketed with a yellow-brown cyanobacterial mat.

## Leeward Reef Slope

The leeward reef slope is typically gentler (10–20°) than the windward, except where strong currents sweep around the ends of the reef. Seismic profiles perpendicular to the reef edge show that the slope is formed by an accretionary wedge of reefal sediment abutting the reef as the *Reef Apron Subzone*. The *Reef Front Subzone* extends as tongues from the reef rim on top of the *Apron* and may have spur-and-groove appearance. Massive *Porites* heads several metres in diameter and other patches of coral growth are common in water to –10 m. The apron may extend 0.5 km from the reef edge. Ryan et al. (2001) detailed the sediment apron to the lee of Wistari Reef.

## *Reefal Shoal*

The platform between Heron and Sykes is in waters 10–14 m deep with large sand dunes migrating over its undulating surface. Large stands of corals including *Acropora* form small reefal patches along the platform edge (see Jell and Flood, 1978).

## *Sand Cay*

The sand cay rises abruptly from the southern beach to a height of 4.5 m (mhwl) and slopes gently to the N. The beach is 10–25 m wide and bordered by beachrock in a 9–20 m wide belt from E of the harbour to past the eastern end of the island. Its strike parallels that of the beach and dips seaward at 2–12°, decreasing seaward. The beachrock is bordered by a 50 cm deep moat. The moat is caused by wave scour. Strong tidal currents channel along the moat to and from the harbour adding to the scour. The moat is populated sparsely by molluscs and a few corals including *Porites lutea* that rolls around the moat with the coral living on the part of the colony that is uppermost at the time.

Shark Bay, at the eastern end, is a wide beach accreting to the NE. The flat area above the beach is favourable for nesting turtles and they significantly disturb the sediment. A wide tear-drop shaped sand bank off the NE beach displays tidal sedimentary structures and is strongly bioturbated. Several generations of beachrock crop out around the NW end of the Island which is being eroded since the harbour was installed to the extent that cement walls are necessary to protect the Resort.

# Sediments

Maxwell (1968) provided a comprehensive regional study of GBRP sedimentology. Mathews et al. (2007) reviewed the next 40 years of studies providing an up-to-date synthesis. Maxwell and Maiklem (1964), and Maiklem (1970) investigated the inter-reef sediments of the Capricorn/Bunker groups. Maxwell et al. (1961, 1964) described the Heron Reef reef-top sediments. Sediments of the reef slope have not been studied.

## Shelf sediments

The continental shelf sediments are influenced by their mainland provenance, a submarine dune source, a reefal source, and in situ benthic foraminifers, bryozoans and molluscs. Three facies (Jell and Flood, 1978, text-fig. 13) are:

1. Terrigenous: <60% carbonate, inner shelf adjacent to the mainland, detrital quartz sand a major constituent derived from coastal dunes and ancient dune systems of the shelf and older fluviatile deposits.
2. Carbonate: > 80% carbonate, on the outer shelf around the reefs, reef-derived bioclastic carbonate sand and gravel, in situ benthic foraminifers, bryozoans, molluscs, *Halimeda*, etc. A low-mud subfacies is restricted to the immediate proximity of the reefs, and consists of moderately sorted carbonate sand and shingle (i.e. coral sticks of gravel size or coarser). A variable mud sub-facies consists of poorly sorted carbonate sand, and it represents the area free of coarse-grained reef derived skeletal debris.
3. Mixed (or transitional): 60–80% carbonate, transitional between the other two facies.

## Reef sediments

Maxwell et al. (1964) showed that corals, coralline algae, foraminifers, molluscs and *Halimeda* account for c. 90% of reef top sediment. A range of factors influence particle size and degree of sorting. Coralline algae (calcitic), which are relatively resistant to mechanical break-down, contribute to coarse sands and gravels on the reef rim. Corals (aragonitic) break down rapidly into distinct size modes (shingle sticks and very coarse, fine to very fine sand). Winnowing by breaking waves and translatory waves leaves boulder and gravel particles as a lag deposit (Rubble Subzone) on the reef rim and outer reef flat, and transports coarse sand into the Sand Subzone and Shallow Lagoon. Very fine sand and silt are carried in suspension to the Deep Lagoon where it settles during periods of slack water, or off the reef onto the continental shelf. Consequently, there is a generally fining size gradient from windward reef rim to central lagoon. Factors differentiating particle size also promote segregation of calcitic detritus (coralline algae and foraminifers) as coarser sediments, remaining near source from aragonitic detritus (corals and *Halimeda*) as finer material transported towards the lagoon. This concentric pattern is modified by tidal currents. Sediment thickness varies from very thin on the Reef Flat, <10 cm, to 4 m below the Shallow Lagoon and >5 m in the Deep Lagoon (Smith et al., 1998).

# Internal architecture

The 1937 Heron Island bore provided no reliable dates and little data on the internal structure of the reef.

## Shallow cores

Cores from within 20 m of the western reef margin to the NE of the cay (Nothdurft et al., 2010) call into question the ‘classical’ reef model of Marshall and Davies (1982). U-series dates of the cores are 7.3–4.2 ka with the reef flat having reached its current elevation 5.4 ka at the innermost position and 4.2 ka 20 m towards the margin. Hence, the leeward reef reached sea level relatively early and prograded c. 20 m at current sea level in c. 1 ka. A single core, 25 m inward from the windward reef near the cay dated at 1.9 ka at 0.1 m and 3.0 ka at 0.7 m depths, suggesting much more significant progradation on the windward margin at the channel between Heron and Wistari reefs than on the opposing leeward margin. Greater windward progradation may account for the wider windward reef flat relative to the lee.

## Reefrock blocks

Blocks of shallow (upper 2–3 m) reefrock were dredged to enlarge the boat harbour in 1988. Three large blocks (dimensions 0.5–1.2 m) and c. 30 smaller blocks were collected and many more remain in retaining walls on Heron Island. The three large blocks contain corals in growth position and their upper surfaces had been the top floor of the reef flat. They were sawn into 20 cm-thick slabs and cut surfaces were logged for constituents and macro-porosity using a 5 cm grid. Selected microbialite crusts and skeletal substrates were dated by $^{14}C$ accelerator mass spectrometry (AMS) (Webb and Jell, 2006). Although the samples may not be a statistically valid subset of the overall reef framework or volume, they are a representative set,

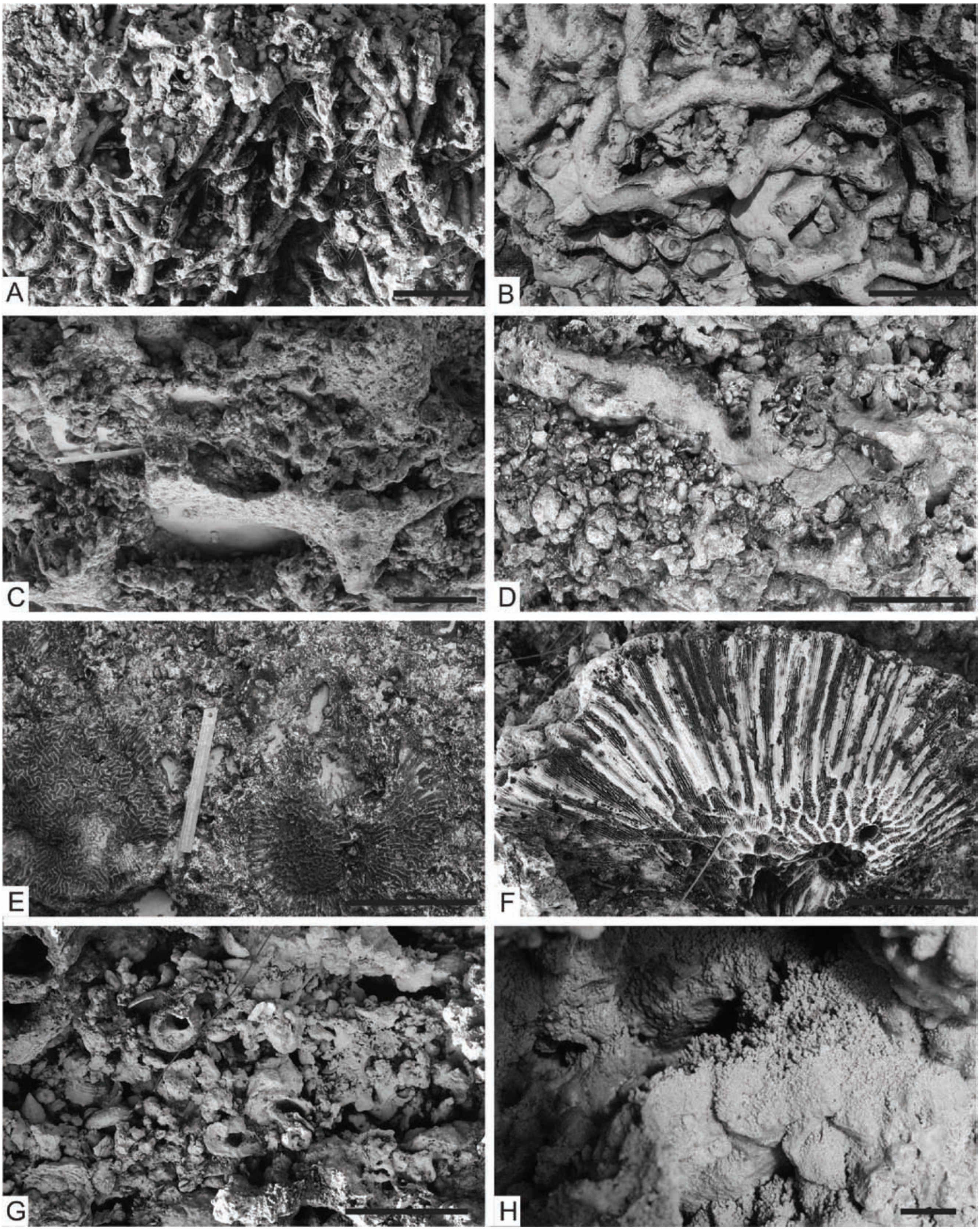


*Figure 5 Reef blocks, Heron Island boat channel. Scale bars = 15 cm (except as noted). A). Branching Acropora framework with thick coralline algal crusts and microbialite, but little entrapped rubble. B). Underside of branching Acropora framework with some entrapped rubble. C).* A. hyacinthus *complex framework with large laterally contiguous cavity system. D).* A. hyacinthus *complex framework with abundant coarse rubble bound into extensive cavity system. E). Massive coral facies. F). Massive coral facies –* Goniastraea sp. *Scale bar= 5 cm. G). Stratified rubble bound by microbialites. Scale bar = 10 cm. H). Thrombolitic microbialites in framework cavity. Scale bar = 2 cm.*

indicative of the relative percentages of coral, algae, rubble, microbialite and cavity in average well-lithified reefrock in the vicinity of the boat harbor.

The reefrock blocks are generally well lithified with abundant open porosity and consist of varying mixtures of in situ (i.e., growth position) corals and encrusting organisms (e.g., coralline algae, foraminifers), coarse debris, and microbialite. Any unconsolidated sediment originally within cavities had been removed, presumably during dredging. Individual coral frameworks in the blocks include (1) branching acroporids (Figure 5A, B), 2) overlapping vase- to table-shaped acroporids (e.g., *A. hyacinthus* group) (Figure 5C, D), and (3) massive corals (Figure 5E, F). Coarse rubble fills the bases of some growth cavities, but dominates rare blocks in laterally continuous strata (Figure 5G). Although coralline algae are important binders in these framework blocks, coarse debris was bound primarily by microbialite, which is a conspicuous component within most observed cavity systems (Figure 5H).

The three quantified blocks represent two major types of reef framework and composition data are shown in Table 1. Two of the blocks consist primarily of in situ corals (21–26% volume) dominated by *A. hyacinthus* group corals with subordinate branching acroporids, rare massive and encrusting corals, scattered coarse debris (5–9% volume), and a large laterally connected cavity system (30–36% volume). Cavities >10 cm in diameter are common and are generally better connected laterally than vertically owing to the overlapping plate-like coral habit. Similar *A. hyacinthus* group and staghorn acroporids dominate coral taxa in shallow reef slope environments near the harbor. The third block contains more lithified coarse debris (18%), mainly in pockets within a growth framework constructed by more scattered branching and plate-like corals (7% volume). The cavity system in the debris-rich framework is more irregular, with smaller dimensions, but accounts for 39% by volume. Coralline algae are volumetrically important in all blocks (15–17%), and microbialites that preferentially line the lower surfaces and walls of cavities (Figure 5H) make up 6–8% by volume.

Frameworks with in situ corals dominate the assemblage. Blocks with branching coral frameworks generally contain the least bound coarse debris and contain very open cavity systems between horizontal ledges made up of wedged debris and or encrusting corals. *Acropora hyacinthus* frameworks contain rare or abundant coarse debris and massive corals, which are less abundant within the blocks, are commonly associated with more abundant coarse debris (Figure 5D). Blocks dominated by coarse debris are rare, but potential bias exists wherein in situ growth frameworks may be more easily preserved as large blocks than debris accumulations. The more open cavity systems of growth framework may have been more conducive to unification by microbialites, but coarse debris also hosts significant cavity systems and microbialites played a very significant role in unifying the well lithified debris-rich blocks. Contrary to some other reef environments, in situ coral framework is an abundant part of the reefrock on the western end of Heron Reef.

***Table 1 Volumetric percentages of components based on point counts of reefrock fabric in three excavated blocks from reef flat, harbour channel, Heron Reef.***

| Heron Reefrock Block component | Percentage on slab Block 1 | Block 2 | Block 3 | Total |
|---|---|---|---|---|
| Coral in growth position | 21.2 | 7.2 | 26.3 | 17.2 |
| Red algae crust | 17.5 | 14.9 | 15.1 | 15.9 |
| Coarse debris | 9.4 | 17.6 | 4.6 | 11.3 |
| Vermetid gastropod | 0.5 | 0.0 | 0.0 | 0.2 |
| Encrusting foraminifer | 0.9 | 0.9 | 2.0 | 1.2 |
| Microbialite | 8.5 | 6.3 | 6.6 | 7.2 |
| Framework cavity | 30.2 | 38.7 | 35.5 | 34.8 |
| Boring in skeleton/debris | 11.8 | 14.4 | 9.9 | 12.3 |

n=586

Skeletal substrates date at 6610–6050 yrs BP (uncorrected ages) (Webb and Jell, 2006). Microbialite growth began in enclosed framework cavities within c. 200–300 years of the substrates (corals) having been alive and then accreted at 28–34 µm/yr (mean = 29 µm/yr or 2.9 mm/100 yr). Microbialites grew for 228–319 years (Webb and Jell, 2006). Hence, microbialites were emplaced relatively soon (i.e., <300 years) after death of the skeletal substrates as the framework accreted, and microbialite growth continued for only a few hundred years once established. A similar pattern of growth was found in Tahitian microbialites (Seard et al., 2011). As microbialites were not significantly affected by subsequent cementation after they stopped growing (Webb et al., 1998), their role in framework binding is best considered the final, but integral, phase of framework growth rather than a subsequent diagenetic lithification event.

## Conclusions

Heron Reef is an excellent example of a lagoonal platform reef developed on a partially rimmed platform in a high energy environment with continued water turnover due to high tidal flows and clear access to open ocean. All geomorphological units are easily recognizable and the sedimentology is well known. The 1937 Heron Island bore penetrated the complete though thin (c. 150 m) reef and showed that Holocene growth was c. 15 m, but additional coring is required to gain further information on its age of initiation, Pleistocene structure and complete developmental history. The Holocene/Pleistocene unconformity has been geophysically identified under the full length of Heron Reef across the shoal area to beneath Sykes Reef. However, more sensitive equipment is needed to obtain details of the topography of the antecedent platform to assess its possible influence on the morphology of the reef. Rare cores through the Holocene suggest that its development does not fit the 'classical' model. Large reef blocks reveal the fabric of the upper part of the Holocene reef and suggest that abundant in situ coral framework occurs in the shallow reef rock, that microbialites were important in the final unification stage of framework development and that lithified rubble occurs but is relatively rare.

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*by Laurie J. Hutton[1], Terence J. Denaro[1], Courteney Dhnaram[1] and Geoff. M. Derrick[2]*

# Mineral Systems in the Mount Isa Inlier

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*Northwest Queensland contains several world class mineral deposits, being one of the worlds leading producers of Zn, Pb, Cu and Ag. Rather than focus on mineral deposit models, as has been done in the past, we are using the mineral system approach (Barnicoat, 2008), where the whole system is studied at a variety of scales and a variety of processes, which culminate in the deposition of mineralisation. Seven Mineral Systems are identified, namely:*

1. *Shale/siltstone/dolomite hosted Zn-Pb-Ag systems – Western Fold Belt*
2. *Ag-Pb-Zn in high-grade metamorphic terrains – Eastern Fold Belt Province*
3. *Structurally-controlled epigenetic iron oxide-Cu-Au – Eastern Fold Belt and Kalkadoon-Ewen Provinces*
4. *Structurally-controlled epigenetic Cu±Au mineralising system – Western Fold Belt Province*
5. *Phosphate Mineralisation in the basal Georgina Basin sequence*
6. *U and Rare Earth element (REE) mineralisation.*
7. *Fe Ore – South Nicholson Group*

## Introduction

The NW Queensland Mineral and Energy Province (NWQMEP) can be regarded as the premier Zn-Pb-Cu region in the world (Geological Survey of Queensland, 2011).

The NWQMEP evolved as a Paleoproterozoic–Mesoproterozoic province of the North Australian Craton (NAC), from c. 1900–1500 Ma, in a largely far-field extensional back-arc to intracontinental setting over-riding the NE-dipping convergent margin of the Gawler Craton to the far S (Betts et al., 2003). Three major stacked superbasins developed on c.1900–1860 Ma crystalline basement – the Leichhardt Superbasin (1800–1750 Ma), the Calvert Superbasin (1740–1670 Ma), and the Isa Superbasin (1670–1595 Ma), containing extensional and sag-phase sedimentary packages with some volcanics and magmatic rocks, separated by unconformities. From c. 1680 Ma, an E-facing rifted continental margin may also have developed along the eastern margins of the NAC.

Basin development was largely terminated by compressional tectonism of NNW-SSE and E-W orientations from 1600–1500 Ma, accompanied by major felsic magmatism in the E. These events – the Isan Orogeny - produced the current geological setting of the Mt Isa Inlier, comprising a central Kalkadoon-Leichhardt basement block flanked by eastern and western fold belts (Figure 1). More comments on the geodynamic setting of the Mount Isa Inlier are given in Gibson et al. (2012).

Thick pre-mineralisation clastic successions became important metal reservoirs and fluid aquifers (Polito et al., 2006) during formation of the world-class Pb-Zn deposits of the Isa Superbasin. Extensional environments also provided a complex network of syndepositional faults that were later reactivated to provide fluid conduits feeding metalliferous oxidised brines into reduced third-order sag basins of the Isa Superbasin (Figure 2), or the deeper water rifted basins of the eastern craton margins.

Most Cu deposits formed in the late tectono-magmatic cycle in shear, breccia and dilational structural environments, in new and reactivated fault systems. Oxidised warm to hot saline fluids (± Fe) of magmatic and metamorphic origins were highly effective transporters of Cu and Au into structurally favourable (i.e., brittle) and reactive (e.g., magnetite-hematite) host rocks.

As well as being a major producer of Cu-Pb-Zn, the NWQMEP is also a significant producer of Ag and Au, the former as byproduct of the Pb-Zn deposits, and the latter as part of the Cu deposits. Younger Fe ore and phosphate deposits formed within and adjacent to the Mt Isa Inlier in Mesoproterozoic and Cambrian times respectively. More detailed aspects of mineralisation and mineral systems appear below.

Singer (1995) defined various criteria for the ranking of mineral deposits. World class Zn deposits for example contain a minimum of 1.7 Mt of Zn metal, and at least 6 such deposits occur in the NWQMEP – George Fisher, Century (Figure 3), Mt Isa (Figure 4), Dugald River, Cannington, and Lady Loretta, all of which make the NWQMEP the largest known repository of economically mineable Zn in the world. The George Fisher, Mt Isa, Cannington and Century deposits are also world class Pb deposits (>1Mt Pb), with the first 3 also containing large resources of Ag (632, 643 and 870 M ozs respectively). The province is also host to significant Cu deposits, including the world class (>2 Mt Cu) Mt Isa deposit (255 Mt @ 3.3% Cu), and Ernest Henry (127 Mt @ 1.1% Cu, 0.55 g/t Au).

Several Mineral Systems are postulated to occur within the NWQMEP. The following is a description of these systems.

## Shale/siltstone/dolomite hosted Zn-Pb-Ag systems: Western and Eastern Fold Belt

These deposits are characterised by stratiform to stratabound massive sulfide lenses in carbonaceous shales and dolomitic siltstones at varying stratigraphic levels within the Isa Superbasin. They include the Mount Isa Pb-Zn (Isamine, George Fisher), Century, Duguld River, Kamarga, and Lady Loretta deposits.

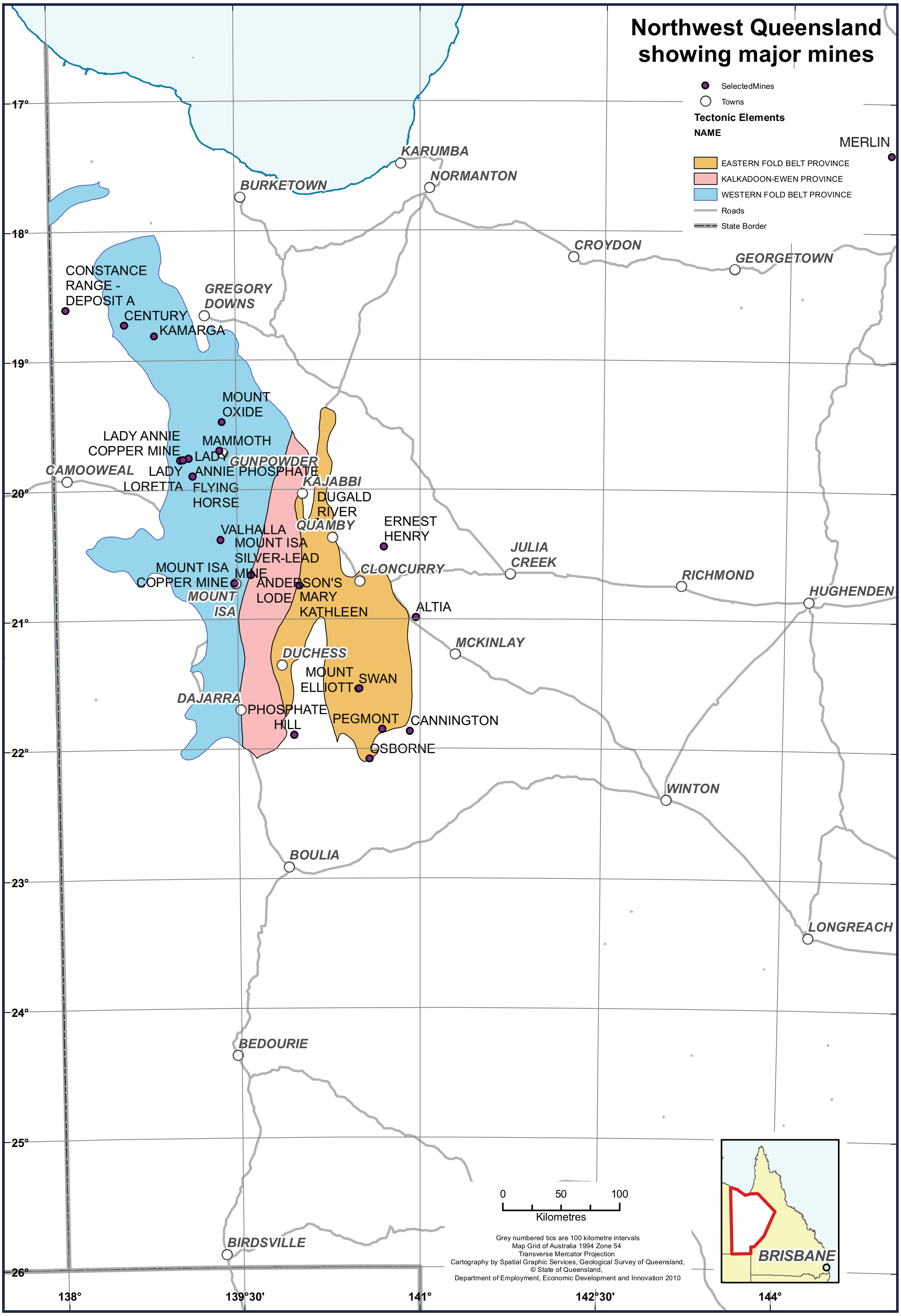


*Figure 1 Map showing selected mines and deposits in NW Queensland.*

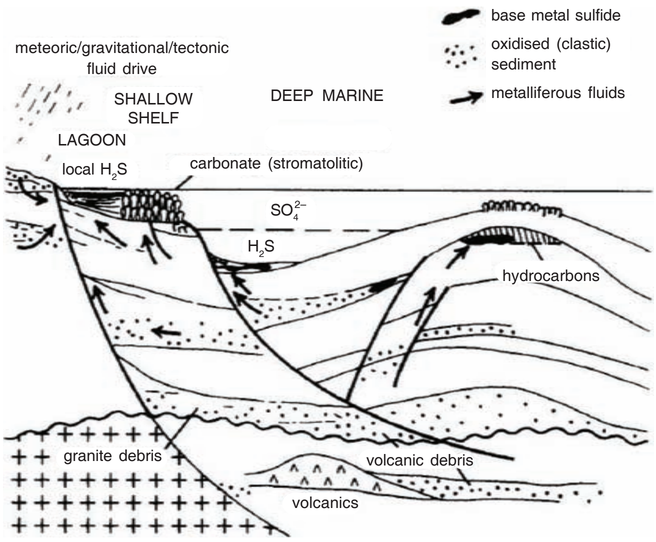


***Figure 2 Diagram showing aspects of genesis of the sediment hosted Pb-Zn deposits in the Mount Isa Inlier.***

## Tectonic/Geological Environment

The deposits typically occur in an intracontinental rift to passive margin environment. There is strong basement control on basin architecture and the orientation of faults active at the time of basin formation. The rift environment provides a source for the fluids and fluid pathways; deposition of the orebodies typically occurs late in the extensional cycle and may be related to either sedimentation or inversion of the basins.

All significant occurrences are hosted by 2–8 km thick successions of the Isa Superbasin, ranging in age from c. 1660 Ma (Dugald River, Lady Loretta) to 1590 Ma (Century) (Queensland Department of Mines and Energy et al., 2000). These successions are interpreted to represent the products of sedimentation related to thermal subsidence following extension and rifting (Queensland Department of Mines and Energy et al., 2000). Within the sag successions, Zn-Pb-Ag deposits are typically localised in districts or 'sub-basins' of 100–200 km$^2$ area that are characterised by:

1. Underlying clastic and silty units in the Calvert Superbasin commonly show a network of growth faults and subtle half-graben structures upon which the sag-phase successions were deposited, with local unconformity.
2. Sag-phase basins form large accumulations of basal clastics overlain by thick and extensive packages of siltstone, dolomite and dolomitic siltstone.
3. Within the broader sag-phase basins smaller (third order) sub basins develop which are characterised by abundant pyritic and carbonaceous shale and dolomitic siltstone which are immediate host rocks to Pb-Zn mineralisation (Huston et al., 2006).

## Source

The source of the metals in most sediment-hosted Zn-Pb-Ag deposits has usually been attributed to clastic rocks in upper crustal sequences which underlie the deposits (for example, Zn from shale, basalt; Pb from arkose, grit, felsic volcanics, granites; Derrick, 1996).

***Figure 3 Aerial View of the pit at Century Mine, NW Queensland.***

*Figure 4 View of the Black Star pit – part of the Isa Mine complex.*

## Fluid Pathways

The pre-existing Leichhardt and Calvert Superbasins provided permeable aquifers and fluid reservoirs for many of the metals that are hosted by the Isan Superbasin or deposited during the Isan Orogeny (Polito et al., 2006).

## Depositional Mechanisms

Depositional mechanisms for Zn-Pb-Ag mineralisation are varied:

- Extension and thermal input during early superbasin development did not result in the formation of mineral deposits but rather were the storage compartments for fluids drawn down into the system (Murphy et al., 2011; Polito et al., 2006).
- The main processes contributing to Zn-Pb-Ag mineral deposition were fluid cooling, dissolution of host rock carbonate (with consequent pH increases) and thermochemical sulfate reduction due to the interaction of oxidised Zn-Pb-Ag-transporting saline fluids with organic matter and also mingling with migrated but locally sourced hydrocarbons; inorganically precipitated carbon was also produced (Hinman et al., 1994; Dixon and Davidson, 1996; Hinman, 1998; Broadbent et al., 1998). These processes emphasise the significance of organic-rich and calcareous successions as potential hosts and reductants.
- Deposits such as Century and Mount Isa exhibit paragenetic stages from early, layer-parallel sphalerite, sphalerite breccias with minor galena and pyrrhotite to vein and breccia-hosted galena with sphalerite, pyrrhotite and euhedral pyrite. A paragenetic evolution from early sphalerite to late galena with euhedral pyrite is consistent with a thermally prograding event, increasing extent of thermochemical sulfate reduction and saturation of hydro-thermal pyrite (Murphy et al., 2011).
- While layer-parallel mineralisation is widespread in the deposits, coarser-grained layers and veins of galena and lesser sphalerite are also developed as a consequence of proximity to possible feeder zones along bounding faults, and/or to recrystallisation and replacement of earlier sulfides by later generations of sulfides formed during metamorphism, especially at Mt Isa and George Fisher.
- Carbonate-hosted replacement deposits such as Kamarga may have formed by neutralisation of hot acid fluids (Jones et al., 1999).
- Although synsedimentary to early diagenetic mineralising processes have generally been favoured for formation of these deposits (Waltho and Andrews, 1993; Hinman et al., 1994; Dixon and Davidson, 1996; McGoldrick and Large, 1998; Betts and Lister, 2002), there is a growing body of evidence for a late diagenetic (reminiscent of Mississippi Valley-type deposits) to syntectonic replacement mechanism as an alternative explanation for the formation of the stratabound Zn-Pb-Ag orebodies (Xu, 1996; Broadbent et al., 1998; Rohrlach et al., 1998; Jones et al., 1999). The timing of the mineralising Pb-Zn systems has been fully discussed by Large et al. (2005) and Huston et al. (2006). Alternatively, syngenetic–early diagenetic mineralisation may have been remobilised and enriched, in a manner similar to models generally invoked for Broken Hill-style deposits. At Duguld River, the deposit occurs in metamorphosed carbonaceous shale with a substantial part of the resource resulting from significant structural upgrading, perhaps during the Isan orogeny.

# Ag-Pb-Zn in high-grade metamorphic terrains: Eastern Fold Belt Province

Deposits comprise massive to semi-massive galena, sphalerite, pyrrhotite and pyrite and/or magnetite layers or stacked lenses hosted by thin-bedded calcareous paragneiss and migmatitic quartzo-feldspathic gneiss, considered to be metamorphosed immature siliciclastic sediments. Amphibolite, porphyry and pegmatite lenses occur within the gneissic terrain. The complex gangue mineralogy includes calc-silicate mineral assemblages containing garnet( Mn-enriched)-fluorite-hedenbergite-pyroxmangite-quartz-magnetite-fayalite-pyrrhotite-gahnite (Walters et al., 2002). These stratabound deposits are typically thin, but laterally extensive and were deformed and metamorphosed together with their host rocks (Hoy, 1996). Deposits in the Mount Isa Inlier include Cannington, Pegmont, and Altia (Figure 1). These deposits have similarities to world class Broken Hill mineralistation >1,000 km south (Gibson et al., 2012 ) and are commonly referred to as 'Broken Hill Type' (or BHT) deposits. They are important sources of Pb, Zn and Ag with Cannington being the world's largest and lowest cost single mine producer of Ag and Pb and a significant producer of Zn.

## Tectonic/geological environment

Broken Hill-type deposits appear to be restricted to the eastern margin of Proterozoic Australia (Fraser et al., 2007), where mineralisation formed in feldspathic clastic rocks that were deposited in a deep water turbiditic basin. The region is modelled in extension as a thin, brittle upper crust above a thermally weakened lithosphere, where connectivity between the two vertically stacked domains appears to be largely along steep crustal scale faults (Murphy et al., 2011). While siliciclastic sedimentary packages are dominant, they also contain rift-related basic layered sills and exhalative Fe formations enriched in Mn and P (Hatton and Davidson, 2004).

## Depositional Mechanisms

Two models for the generation of BHT deposits are:

- The modified synsedimentary/syndiagentic model (Boden, 1996; Bailey, 1998) with initial introduction and zoning of base metal and Ag mineralisation with Zn dominant and Pb-Ag dominant horizons. This pre-metamorphic zoning could have been developed by processes associated with a volcanogenic sulfide system or a basin dewatering diagenetic system with mineralisation controlled by primary porosity or matrix replacement, associated with the emplacement into the sequence of a series of tholeiitic basic sills (amphibolite). Introduction of the mineralisation is followed by regional deformation and metamorphism. During a post-metamorphic metasomatic event, initial metasomatism of mineralised rocks resulted in anhydrous alteration characterised by hedenbergite-garnet-quartz and the deposition of very minor pyrrhotite and rare sphalerite. This was followed by high and low temperature hydrous stages.
- The skarn model (Williams et al., 1996) comprises an original metasedimentary package (consisting of an Fe-Mn-(Ca)-rich fraction) and an outer Fe- and Mn-rich peraluminous metasediment derived from quartz-pelite mixtures with local feldspathic fractions; regional deformation with peak metamorphism reaching upper amphibolite facies; peraluminous anhydrous Fe-rich alteration (quartz-sillimanite-potassium feldspar-biotite-garnet-graphite); anhydrous Ca-rich alteration (quartz-apatite-pyroxmangite-hedenbergite-fayalite-hornblende-garnet); hydrous Fe-Ca-K alteration (hornblende-biotite-pyrosmalite-dannemorite); and a mineralising phase with sphalerite, galena, pyrrhotite, chalcopyrite.

A large coherent halo of stratabound almandine (pink garnet)-quartz-apatite-biotite-graphite alteration occurs as an envelope around the mineralised package at Cannington. This alteration has penetrative tectonic fabrics and is overprinted by later alteration. Quartz-garnet-pyroxene-pyroxenoid alteration affects partial melt segregations that occurred during peak metamorphism, suggesting that these skarn-like alteration assemblages developed under fairly deep-seated (ductile) conditions at a late stage of the Isan Orogeny. Peraluminous Fe-Mn-rich metasediments form compositionally banded K-feldspar-sillimanite-quartz-biotite-garnet assemblages. Goerthite-quartz assemblages are also diagnostic of this mineralisation. Recent studies and age dating continue to favour a synsedimentary/syndiagenetic origin for BHT deposits (Huston et al., 2006), formed at or just below the sea floor. Despite the extensive polyphase folding and high grade metamorphism evident in BHT deposits, feeder zones and a replacement/exhalative footwall system have been recognised at Broken Hill (Groves et al, 2008).

# Structurally-controlled epigenetic iron oxide-Cu-Au systems (IOCG): Eastern Fold Belt and Kalkadoon-Ewen Provinces

Deposit styles within these systems comprise epigenetic mineralisation as hydrothermal replacements, veins and breccias. Two major but contrasting groupings are identified:

1. An economically significant grouping of larger deposits commonly referred to as IOCG deposits (iron oxide-Cu-Au), and which include the world-class Ernest Henry deposit (Figure 5), Osborne, Mt Elliott, Roseby, Eloise, Rocklands, Mt Dore and Starra deposits. Recent discovery of the Merlin Mo-Rh deposit adds to the economic significance of this deposit grouping.
2. Smaller deposits (i.e., unlikely to ever achieve production status in the foreseeable future) are voluminous and widespread throughout the Kalkadoon-Ewen basement province, and in the Eastern Fold Belt. They form as narrow 1–5 m quartz vein-type deposits in N-, NW- and NE-trending shear zones, and are closely associated with dilatant structures along the margins of dolerite

***Figure 5 View of part of the open cut pit at Ernest Henry.***

and amphibolite bodies which occupy the same structures. Host rocks include older granite and volcanics, and metasedimentary cover rocks.

Many of these deposits formed the basis of a historic small-scale ("gouger") Cu mining industry during the early to mid-20$^{th}$ Century.

## Tectonic/geological environment

The grouping of smaller deposits throughout the basement and eastern fold belts are located in largely intracontinental and continental margin environments within the North Australian Craton.

Most of the larger IOCG deposits formed in an E- or SE- facing passive continental margin dominated by shallow shelf to deeper water turbiditic sequences. Within the shelf and slope, local rifting promoted the stacking of and juxtaposition of chemically reactive lithologies, including ironstones, carbonaceous siltstones, volcaniclastics, carbonates and mafic sills and dykes (Davidson and Large, 1998). These sequences are 1760–1650 Ma, equivalent to the Calvert and Isa Superbasins.

Basin inversion from 1600–1500 Ma resulted in deformation, crustal thickening and the intrusion of voluminous, mainly felsic, magmas to crustal depths of 5–10km (Mark et al., 2006) – the Williams and Naraku Batholiths. Duncan et al. (2011) suggest that metal-rich reservoirs formed in the SE end of the subduction zone, along the southern margin of the NAC, to be tapped by metamorphic and magmatic events during the Isan Orogeny.

Most IOCG deposits are related spatially to the Williams/Naraku batholiths (1545–1490 Ma) but mineralising fluids could have been metamorphic (e.g., Osborne, 1600 Ma) or magmatic in origin (e.g., Ernest Henry main stage Cu-Au, 1525 Ma; Duncan et al., 2011). Most IOCG deposits formed in the time range 1550–1500 Ma, as part of the Isan Orogeny.

## Depositional Mechanisms

The major mineralising event in this system occurred within the Isan Orogeny (1600–1500 Ma, from peak metamorphism 1600–1550 Ma through to major granite intrusion (Williams and Naraku Batholiths) from 1545–1490 Ma. Most Cu-Au mineralisation is preceded by region-wide Na-Ca alteration manifested as albite-diopside-calcite-actinolite assemblages which are overprinted by Cu-Au±Fe.

The region-wide small-scale Cu deposits are generally unrelated to granite plutons, but form as narrow (1–5 m) quartz-calcite-chlorite filled shears in a diverse range of host rock age and composition. They contain little or no magnetite, and show a spatial and possible genetic relationship to basic dykes and sills; in addition, mineralising fluids were likely to be saline because of metamorphism of extensive scapolitic (?evaporative) metasedimentary cover sequences, and likely contributions from regional deep-crustal sources.

The IOCG deposits by contrast contain abundant magnetite (and hematite), and formed in larger structures associated with dilatancy, rheology contrasts, brecciation and replacement of brittle host rocks (e.g., 1740 Ma intermediate volcanics at Ernest Henry). Ore fluids were high temperature (300–500$^{o}$C), highly saline (26–70 wt% NaCl) and oxidised (Mark et al., 2006). Magnetite formed in some deposits from mineralising fluids (e.g., Ernest Henry), while in others fluid replaced existing host-rock ironstones. Metal formed by pH changes due to wallrock interaction, redox changes and reduction of some fluids by carbonaceous rocks (e.g., Mt Dore).

Two IOCG events are recognised:-

1. Ironstone-hosted deposits such as Osborne and Starra formed c. 1600–1565 Ma (Perkins and Wyborn, 1998; Gauthier et al., 2001; Duncan et al., 2009; Baker et al., 2010), and possibly as early as 1680 Ma (Oliver and Rubenach, 2009).
2. Breccia and shear-hosted deposits such as Mt Elliott, Ernest Henry, Lady Ella and Mt Dore formed post-peak metamorphism and synchronous with the period of granite emplacement (1555–1485 Ma) (Perkins and Wyborn, 1998; Wang and Williams, 2001; Duncan et al., 2009; Baker et al., 2010).

Older mineralising events in this system are sparse. The Tick Hill Au-only deposit (511,000 ozs mined at 22 g/t Au) formed in a high strain domain possibly related to the Wonga event extension from 1750–1730 Ma, and could be related to roof zones of Wonga-age granites (Forrestal et al., 1998).

# Structurally-controlled epigenetic Cu±Au mineralising system: Western Fold Belt Province

At Mount Isa, the ore forming system involves similar chemical processes to other sediment-hosted Cu systems, but represents the relatively high temperature end of the spectrum of syn-diagenetic to low-grade metamorphic ore-forming environments (Queensland Department of Mines and Energy et al., 2000). Historically, theories on the genesis of the Mount Isa Cu orebodies have ranged from igneous telemagmatic replacement to syngenetic deposition followed by remobilisation. A variation on this model is the progressive build-up of the Cu ores as a feeder system to syngenetic Pb-Zn. Today, the deposit is almost universally regarded as replacement late in the deformation history of the Isan Orogeny (Perkins, 1990).

## Tectonic/geological environment

There is strong fault control on deposit location at a range of scales. The regional faults are numerically modelled as fluid pathways which, in extension, draw down fluids; in compression, the convective cells break down and fluids are expelled upwards, typically ponding in permeable hanging wall positions. Discrete element modelling at the district to deposit scales indicates that stress anomalies associated with a particular compression direction during $D_4$ deformation played a critical role in the localisation of Cu deposits (Murphy et al., 2011). Fault bends, jogs and intersections are regarded as key localisation features.

Derrick (2008) has shown that the Isa and Mammoth Cu deposits are controlled by an array of earlier growth faults developed in basin extension from 1770–1700 Ma, reactivation and inversion of these normal faults in the Isan Orogen at 1500 Ma produced favourable sites of folded faults and accompanying dilatancy and jogs e.g., along the folded Paroo Fault which forms the immediate footwall to Isa Cu mineralisation.

The Lady Annie, Mount Kelly group of deposits occur in the Paradise Creek Formation, a time equivalent of the Lower Mount Isa Group. The Mammoth and Esperanza deposits occur in fractured quartzite in the Myally Formation, a setting different to the Mount Isa and Lady Annie deposits. The Mammoth orebody is the only significant deposit not hosted by Mt Isa Group equivalents. It occurs in the c. 1760 Ma Whitworth Quartzite of the Myally Subgroup, in which dilatant vein sets form a crackle-fractured breccia in competent and brittle feldspathic quartzite, within a complex fault zone.

## Source

The Mount Isa Cu mineralisation occurs within the Urquhart Shale within the Mount Isa Group. The deposit comprises crosscutting chalcopyrite within a zoned siliceous to dolomitic alteration halo ("silica-dolomite"). Within the Isa mine, the mineralisation lies above a shallow basement fault separating the Mount Isa Group from the Eastern Creek Volcanics (Perkins, 1990). A common interpretation is that the Cu has been sourced by leaching from the Eastern Creek Volcanics (e.g., Smith and Walker, 1971) and therefore the proximity to this unit is a prerequisite for Cu ore formation. Traces of chalcopyrite and either bornite or pyrite locally occur in veinlets, mainly in intensely hematitised metasediments within the Eastern Creek Volcanics (Heinrich et al., 1995).

## Fluid Pathways

For Mount Isa-style deposits, a protracted development of an alteration system beginning with an early K-feldspar and mica alteration, then formation of fractures and dolomite veins and ending with late massive proximal dolomitisation and silicification occurred during the Isan Orogeny. The phase of dolomitic alteration in the host rocks was associated with epidote-sphene and chlorite-albite alteration in the Eastern Creek Volcanics (Heinrich et al., 1995). As the ore fluids moved away from their source they were focussed along brittle/ductile shear zones, interacting to varying degrees with a range of rock types, partly modifying their character. Limited silica-dolomite alteration is also evident in some of the smaller deposits (e.g., Mt Kelly, Lady Annie), in crack-seal breccia and fibrous extensional veins (van Dijk, 1991). Many of the smaller deposits are hosted within a 1670–1655 Ma stratigraphic triplet, comprising basal carbonaceous siltstone, massive ?algal chert and dolomite. Late mineralising Cu-rich fluids intersecting this zone in fault and shear settings may deposit Cu in dilatant sites along competency boundaries of the chert, by pH change from dolomites, and reduction of the fluid by carbonaceous matter.

## Depositional Mechanisms

It is postulated (Heinrich et al., 1993; Matthai et al., 2004; Wilde et al., 2006) that ore deposition was due to mixing of an oxidised brine that circulated within metabasalts with a sulfur-rich fluid from overlying Mount Isa Group metasedimentary rocks or a younger Mesoproterozoic basin at the site of deposition.

Copper deposition was primarily a function of wall-rock reaction; mixing is a necessary consequence of the evolving permeability and porosity regime rather than an essential element of ore deposition. Cu ore precipitation involved one or a combination of depositional mechanisms including:

- cooling;
- a number of wall-rock reactions (reduction by carbonaceous matter, replacement of quartz and dolomite). Dissolution of carbonate minerals, feldspar, and micas buffered pH at somewhat neutral values, optimising Cu extraction (Wilde et al., 2006);
- fluid mixing between magmatic and one or more fluids of a different origin (mantle/metamorphic/basinal evaporate/meteoric) (Kendrick et al., 2006).

# Phosphate Mineralisation in the basal Georgina Basin sequence

Phosphorite deposits in the Georgina Basin have been described by de Keyser and Cook (1972), Southgate (1988), Southgate and Shergold (1991) and Draper (1996). The deposits in the Mount Isa region are in the Cambrian age Beetle Creek Formation, Border Waterhole Formation and Thorntonia Limestone. They consist of beds of consolidated pelletal phosphorites interbedded with chert, carbonate, shale, siltstone and volcanic materials. The phosphorite beds average 11m (but range up to 36 m) thick and consist of dense pellets of apatite in a cherty and carbonate matrix. The phosphorites range from dense pelletal rocks consisting almost exclusively of francolite (one of the collophane group minerals) to siliceous and calcareous phosphorite, phosphatic chert and phosphatic siltstone, and grade into fossiliferous limestone. Chert (silica) and clay are the main dilutants and the deposits have comparatively low levels of heavy metals (for example, <5 ppm Cd). The phosphorites comprise apatite + fluorapatite + francolite + dolomite + calcite + quartz + clays (montmorillonite or illite) ± halite ± gypsum ± Fe oxides ± siderite ± pyrite ± carnotite (Queensland Department of Mines and Energy et al., 2000).

## Tectonic/geological environment

Phosphate deposits occur in an intracontinental or shallow continental margin setting and require predominantly carbonate sedimentation (Draper, 1996; Southgate and Shergold, 1991).

General criteria for phosphate deposition are as follows:

- A low paleolatitude
- A broad shallow downwarp adjacent to a seaway
- High productivity in the vicinity
- Minimal terrigenous sedimentation in a shallow marine environment
- A major transgression
- A trap such as a bay or carbonate bank.

Early Cambrian NW-SE rifting initiated widespread sedimentation in the Georgina Basin and the phosphatic sediments developed in shallow water basins and shelves adjacent to the Proterozoic land mass. The Duchess-Phosphate Hill deposit formed in the S, in the Burke River embayment, while other deposits (e.g., Lady Annie, Lady Jane, Thorntonia, Phantom Hills) formed along the W and NW margins of the Proterozoic land mass.

## Depositional Mechanisms

Phosphate deposits and occurrences are present in two predominantly carbonate sequences. In each of these sequences, the 'retrogradational parasequence sets of the transgressive sytems tract (Southgate and Shergold, 1991)' comprise a repeating suite of phosphorite, phosphatic limestone and organic rich shales. There is a subaerial exposure surface between the two sequences. The phosphate bearing facies were controlled by relative sea level, paleogeography and paleotectonics and there is evidence of structural compartmentalisation of phosphatic facies.

Recently, a blanket of Y+ REE-rich material has been found overlying phosphate mineralisation in the Georgina Basin in western Queensland. As well as Y, the deposit also contains Neodymium (Nd) and Dysprosium (Dy) (Alston, 2011). The origin of the REE enriched blanket is not known.

# Uranium and Rare Earth element (REE) mineralisation

Uranium mineralisation is known from several different settings in the Mount Isa Inlier. These are:

## Tectonic/geological environment

### *Unconformity-related mineralisation*

The unconformity-related mineralisation at Westmoreland (Hills and Thakur, 1975; Rheinberger et al., 1998; Wall, 2006; Polito et al., 2005) is spatially related to either:

- NE-trending structures with proven or suspected tholeiitic dyke filling;
- NE- and NW-trending structures;
- volcanic sills;
- E-trending structures with volcanic dyke filling;
- quartz breccias of NW-trending regional faults; and/or
- proximity of the contact between the uppermost unit of the Westmoreland Conglomerate and the overlying Seigal Volcanics.

Faults at the deposit scale may be related to larger strike-slip fracture zones extending for tens of kilometres. Mineralised zones do not show any signs of pervasive deformation but are displaced by later faulting.

Mineralisation in the principal deposits is present as horizontal, vertical or hybrid styles. Horizontal-style mineralisation is relatively extensive and sheet like, up to 20 m thick, within the uppermost portion of the Westmoreland Conglomerate and close to the Seigal Volcanics contact. This style of mineralisation flanks the NE-trending Redtree Dyke and is best developed immediately adjacent to and on one side of the dyke only. Vertical-style mineralisation forms subvertical, relatively irregular lenses to 30 m thick that are hosted by sandstone of the Westmoreland Conglomerate, although some mineralisation extends into the dolerite dykes. These lenses are adjacent to the Redtree Dyke and their geometry closely mimics that of the dyke-joint system. Hybrid mineralisation is developed in the overlap zone between the horizontal and vertical styles of mineralisation and is, in detail, a combination of both styles. The overlap zone can be up to 50 m thick (Queensland Department of Mines and Energy et al., 2000).

### *Shear-hosted mineralisation*

Lenticular to tabular, stratabound uraniferous beds and zones are hosted by metamorphosed basic volcanics and pelitic and psammitic sediments of the Eastern Creek Volcanics in the Leichhardt River Fault Trough in the Calton Hills-Paroo Creek and Spear Creek-Mica Creek areas. Secondary U mineralisation is generally not readily discernible at the surface of the known deposits, which were located with radioactivity detectors. Most deposits are uneconomic to subeconomic, but some such as Valhalla, Skal, Anderson's Lode (Counter) and Warwei-Watta represent significant U resources.

### *Skarn-hosted mineralisation*

The Mary Kathleen U deposit lies S of the $D_3$, NE-trending Cameron Fault, and is sited in the axial surface of a tight, slightly asymmetrical syncline (the Mary Kathleen Syncline) that can be traced southward for >5 km. The western limb of this structure is cut off by the Mary Kathleen Shear, and the eastern limb by the 1737±15 Ma Burstall Granite. Slightly younger rhyolite dykes W of the granite have similar compositions and an identical radiometric age (Solomon et al., 1994). The Burstall Granite and associated rhyolite dykes also have elevated U contents (7 and 12 ppm U, respectively).

The orebody is hosted by a reduced (magnetite-poor) calcic exoskarn formed by replacement of calcareous rocks of the Corella Formation. The ore comprises fine-grained uraninite disseminated through allanite-apatite enriched rocks that cross-cut the garnet-diopside skarn (Queensland Department of Mines and Energy et al., 2000).

Similar skarn-hosted REE-Cu-Au mineralisation occurs to the S, at the Elaine Dorothy prospect.

## Mineralisation

### *Unconformity-related Mineralisation*

Pitchblende is the main ore mineral and occurs in both the Westmoreland Conglomerate and altered basic dyke rocks. In the sandstones, it occurs interstitial to detrital grains, along fractures, and in veins up to 10 mm thick. It is present as massive, structureless, or rarely euhedral grains, as colloform masses and as thin films of sooty pitchblende. Pitchblende in the dyke rocks occurs as fine aggregates, as thin films and as veins. Secondary U minerals occur as fine disseminations and filling pore spaces. The most abundant secondary U minerals are torbernite, metatorbernite and carnotite. The upper, weathered parts of mineralised systems contain uraninite, torbernite and carnotite, with traces of autunite, bassetite, ningyoite and coffinite. The deeper and unweathered portions of the deposits contain uraninite, autunite, ningyoite, bassetite and coffinite, and minor brannerite. Other ore minerals include pyrite, marcasite, chalcopyrite, galena, sphalerite, Co-Ni sulfarsenides, bismuth, bismuthinite, bornite, chalcocite, digenite, covellite and Au. Thorium is present in alteration products of detrital Th-bearing minerals as thorogummite and florencite. Hematite is abundantly present as the specular type or as a finely disseminated earthy variety and is intimately associated with the primary mineralisation (Queensland Department of Mines and Energy et al., 2000).

### *Shear-hosted mineralisation*

Shear-related deposits are hosted in metabasalts and interbedded metasediments within N-trending to E-W structures in the Eastern Creek Volcanics, and in steep N-S trending mylonite zones in metabasalt and metasediments at Valhalla.

### *Skarn-hosted mineralisation*

The orebody at Mary Kathleen consists of elongate lensoidal ore shoots that are up to 50 m thick and roughly parallel the margins of a broader garnet mineralised zone. The relationship of the ore shoots to stratigraphy is obscured by garnetisation in the upper part of the orebody but the ore lenses are broadly stratiform at depth. The ore is largely a replacive breccia with clasts of early skarn breccia in an allanite-garnet ore matrix.

The spatial relationships between ore, the Mary Kathleen Shear and the axial trace of the Mary Kathleen Syncline indicate that ore formation postdated major folding and was synchronous with shearing under amphibolite facies conditions, consistent with a syn-regional metamorphic age for ore genesis.

The primary structural control on ore formation was the development of ore in and around tensile veins and/or secondary shears in a competent skarn host, along a major boundary between skarn-dominated rocks to the E and regionally metamorphosed, 'un-skarned' metasediments and Wonga Granite to the W (Oliver et al., 1986). Uraninite-bearing ore at Mary Kathleen has a U-Pb age of 1550–1500 Ma (late $D_2$-$D_3$), compared with 1737± 15 Ma for the Burstall Granite, 1700± 60 Ma for banded skarn and 1620–1500 Ma for the main regional metamorphism and deformation.

### Depositional Mechanisms

U-REE enrichment is related to reaction of highly saline and oxidised fluids (Isan Orogeny) with earlier, slightly reduced (magnetite-poor) skarn (Oliver et al., 1986).

## Iron Ore: South Nicholson Group

Oolitic Fe formations occur in the Mesoproterozoic South Nicholson Group of the South Nicholson Basin in the Constance Range area. Up to 10 (generally <4) lenticular, Fe-rich beds occur in the 45–180 m thick Train Range Ironstone Member, some 275–520 m above the base of the Mullera Formation. The Train Range Ironstone Member also contains thinly bedded, alternating dark grey shales, siltstones and sandstones. One to four ironstone beds are present at any one place and the potentially economic ore occurs in the "Main Ironstone Member" – the lowest Fe-bearing unit of significant thickness (Harms, 1965).

### Tectonic/geological environment

Limited observations of the Train Range Ironstone Member suggest that much of the ironstone represents deposition in the upper parts of shallowing-up cycles, i.e., in prograding parasequences during sea level highstands. The presence of both chamositic and sideritic ooidal ironstones indicates growth of Fe minerals on siliceous nuclei in shelfal environments, perhaps on offshore or nearshore bars. The existence of sandstones with rip-up clasts of ironstone as an intraformational conglomerate suggests that erosion and redeposition of pre-existing layers occurs, indicating either a renewed transgressive phase, or local development of channels within an overall prograding succession. Sediment starvation at times of maximum flooding also generates Fe-rich deposits (Burkhalter 1995), and Carter and Zimmerman (1960) state (p 13) that "some of the smaller lenses appear to be concretionary". Although they speculate that these could result from later weathering, it is also possible that they represent sediment-starved horizons, i.e., maximum flooding surfaces within the basinal sediments (Sweet, 2012).

### Mineralisation

Outcropping ironstones are a variable mixture of ochrous red hematite, finely crystalline blue-black hematite, limonite, quartz grains, quartz cement, shale and clay minerals, and rare relict siderite. The ironstones vary in appearance from oolitic forms to a sandstone with a hematite matrix, and have been derived from primary ironstone by surface weathering. Grades range from 20–62% Fe, depending on the silica content of the parent rock (Harms, 1965). Oxidised ironstone extends to 12–30 m vertically. The transition zone appears to have some Fe enrichment, and the near-surface zone has probably been enriched in silica.

Below the water table, the ironstones contain oolites of ochrous or finely crystalline hematite, siderite and/or chamosite, and silica grains in a matrix of siderite, hematite, minor microcrystalline quartz and carbon. Oolites range from 0.2–3 mm in diameter and successive shells may consist of different Fe minerals. Veins of quartz-pyrite, siderite-pyrite and calcite cut the ironstones. Disseminated syngenetic pyrite occurs along bedding planes, especially in carbonaceous shales associated with the ironstone beds, and in siderite-rich bands. Siderite partially or completely replaces some or all of the other Fe minerals. It also replaces quartz grains and appears to have formed late in the deposition or during diagenesis.

The highest grade beds are oolitic and contain 50–55% Fe at the surface. Lower grade beds contain <20–25% Fe and are siliceous. Fifteen individual deposits have been investigated and resources were calculated for three deposits, which contain a total resource of 368 Mt @ 45.4% Fe and 9.1% $SiO_2$, including 40 Mt of oxidised ore @ 57.0% Fe and 10.0% $SiO_2$ (Queensland Department of Mines and Energy et al., 2000).

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# Paleoproterozoic–earliest Mesoproterozoic basin evolution in the Mount Isa region, northern Australia and implications for reconstructions of the Nuna and Rodinia supercontinents

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*Paleoproterozoic–earliest Mesoproterozoic sequences in the Mount Isa region of northern Australia preserve a 200 Myr record (1800–1600 Ma) of intracontinental rifting, culminating in crustal thinning, elevated heat flow and establishment of a North American Basin and Range-style crustal architecture in which basin evolution was linked at depth to bimodal magmatism, high temperature-low pressure metamorphism and the formation of extensional shear zones. Rifting initiated in crystalline basement ≥ 1840 Ma old and produced three stacked sedimentary basins (1800–1750 Ma Leichhardt, 1730–1670 Ma Calvert and 1670–1575 Ma Isa superbasins) separated by major unconformities and in which depositional conditions progressively changed from fluviatile-lacustrine to fully marine. By 1685 Ma, a deep marine, turbidite-dominated basin existed in the E and basaltic magmas had evolved in composition from continental to oceanic tholeiites as the crust became increasingly thinned and attenuated. Except for an episode of minor deformation and basin inversion at c. 1640 Ma, sedimentation continued across the region until onset of the Isan Orogeny at 1600 Ma. A near-identical record of crustal thinning and basaltic magmatism accompanied basin formation (lower Willyama Supergroup) in the formerly contiguous Broken Hill region from 1730–1670 Ma. This was followed by further extension and a second phase of basin development that lasted until at least 1640 Ma. Modern-day rifted continental margins preserve a comparable record of crustal thinning and near-continuous basin formation over 100–200 Myr timescales, supporting suggestions that the late Paleoproterozoic–early Mesoproterozoic rift basins of Mount Isa and Broken Hill similarly evolved to continental breakup and formed part of a continental margin sequence no later than 1640 Ma and possibly as early as 1670 Ma. This rifted margin predates assembly and breakup of the Neoproterozoic Rodinia supercontinent to which Australia once belonged and best accords with a pre-Rodinia, SWEAT-like supercontinent (Nuna) that matches the E-facing late Paleoproterozoic–early Mesoproterozoic rift sequences of eastern Australia against rocks of comparable age in western Canada. Reconstructions of Rodinia (AUSWUS) based on the distribution of Grenville-age orogenic belts that coincidently position the continental rift sequences of Broken Hill along strike from more juvenile 1700–1650 Ma accreted terranes in the SW United States (Yavapai and Mazatzal provinces) are only possible if the proposed alignment of terranes is not original but an artefact of Neoproterozoic supercontinent assembly. The SWEAT hypothesis avoids this complication but, like AUSWUS, presupposes that eastern Australia and western Laurentia remained juxtaposed throughout the Mesoproterozoic until onset of Rodinia breakup after 830 Ma.*

## Introduction

Reconstructions of the Neoproterozoic supercontinent Rodinia, and its Paleoproterozoic predecessor Nuna (or Columbia by which it is sometimes known) (Ernst et al., 2008; Rogers and Santosh, 2009; Zhao et al., 2002), are typically based on matching conjugate continental rift margins with similar geological histories and polar wander paths (e.g., Dalziel, 1991; Wingate et al., 2002). For the pre-Phanerozoic east Australian rift margin, potential matches have been sought in western Laurentia (Figure 1), South China and Mexico where basaltic dyke swarms, syn-rift sedimentary sequences and older basement rocks of the same age and isotopic composition are known to occur (e.g., Burrett and Berry, 2000; Karlstrom et al., 2001; Moores, 1991; Park et al., 1995; Thorkelson et al., 2001; Wang et al., 2011; Wingate et al., 2002). Australian basement rocks for which Laurentian equivalents have already been proposed (e.g., Burrett and Berry, 2000; Karlstrom et al., 2001; Betts et al., 2008) include the late Paleoproterozoic–early Mesoproterozoic basin sequences of Broken Hill and Mount Isa (Figure 1a), both of which lie just inboard of the Tasman Line (Figure 1b) and E of which Proterozoic continental crust is thought not to occur. Although geographically distant from each other (Figure 1), these two regions exhibit strikingly similar geological histories and potential field signatures (Figure 2a) (Baker et al., 2010; Gibson et al., 2008; Giles et al., 2002; Henson et al., 2011; Laing, 1996). This has led many researchers to propose that their constituent rock sequences were formerly contiguous (Figure 2b) and originated through extensional processes in the same back-arc basin (Betts and Giles, 2006; Betts et al., 2008; Gibson et al., 2008; Giles et al., 2002).

Some researchers have further argued that this basin was located above a N-dipping subduction zone along the southern margin of the Australian craton (Betts et al., 2008, 2011; Giles et al., 2002). A near-identical tectonic environment has been proposed for the late Paleoproterozoic–early Mesoproterozoic basins of interior North America (e.g., Thelon, Athabasca) N of the Cheyenne suture (Duebendorfer and Houston, 1987; Karlstrom et al., 2001; Karlstrom and Bowring, 1988; Rainbird et al., 2007; Thorkelson et al., 2005), reinforcing earlier observations that the Paleo–Mesoproterozoic rocks of eastern Australia and western Laurentia share too many geological similarities to have developed in isolation of each other (e.g., Bell and Jefferson, 1987; Betts et al., 2008, 2011; Dalziel, 1991). Rather, there is a strong possibility that the older 1800–1600 Ma rift basins of eastern Australia and western Laurentia are genetically related and share a common evolutionary history linked to breakup of a pre-Rodinia supercontinent (Zhao et al., 2002). Other researchers favour links between Laurentia and Serbia (Sears et al., 2004). Here, we give a brief account of basin evolution in the late Paleoproterozoic–earliest Mesoproterozoic sequences of the Mount Isa region along with a more tightly constrained kinematic and tectonic framework whereby the Paleoroterozoic rocks of eastern Australia might be further compared with their North American counterparts and used as a test of competing supercontinent reconstructions. Such comparisons have been carried out before but more usually in the context of Rodinia reconstructions, including AUSWUS (Australia-western United States)(Burrett and Berry, 2000; Karlstrom et al., 2001) and SWEAT (SW United States-East Antarctica) (Dalziel, 1991;

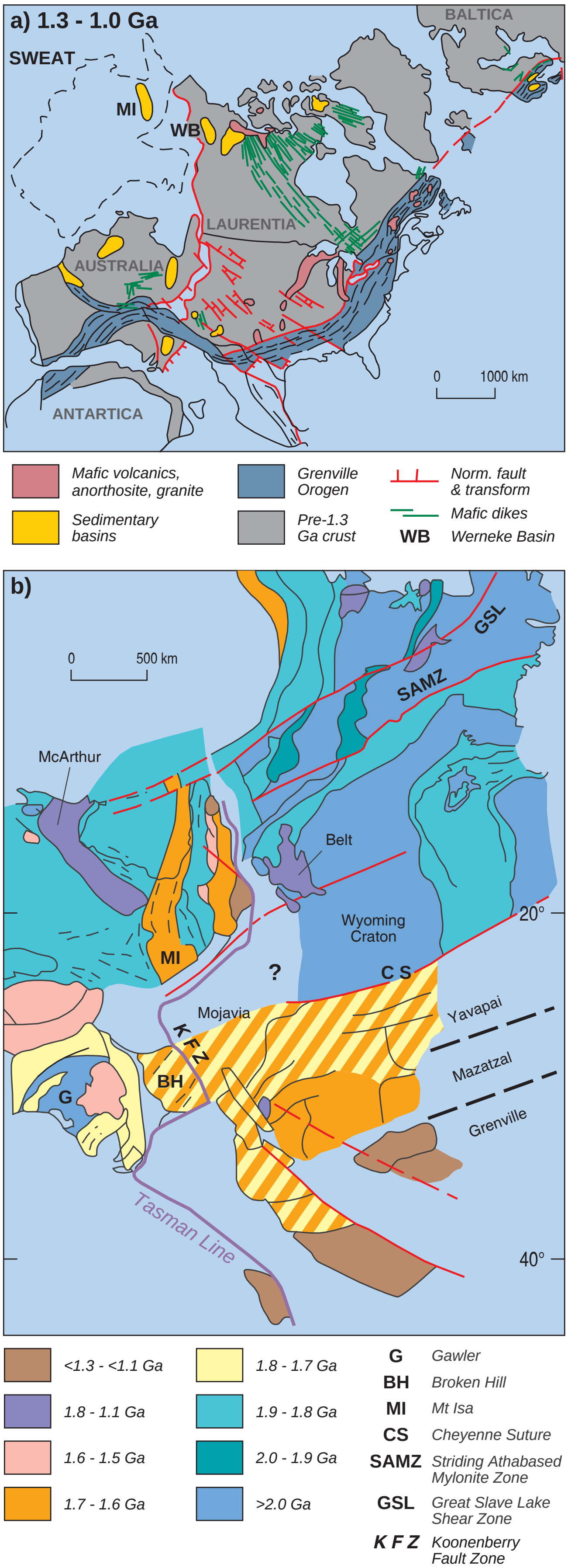


***Figure 1 (a) AUSWUS versus SWEAT reconstruction of Australian and Laurentian rifted margins for Neoproterozoic time. SWEAT restores Australia (dotted outline) and Mount Isa to much the same position opposite NW Canada as existed at the time of the Paleoproterozoic Nuna supercontinent. (b) More detailed AUSWUS reconstruction (Burrett and Berry, 2000) with Broken Hill Block juxtaposed against terranes of equivalent age and isotopic composition in southern Laurentia. WB=Wernecke Basin.***

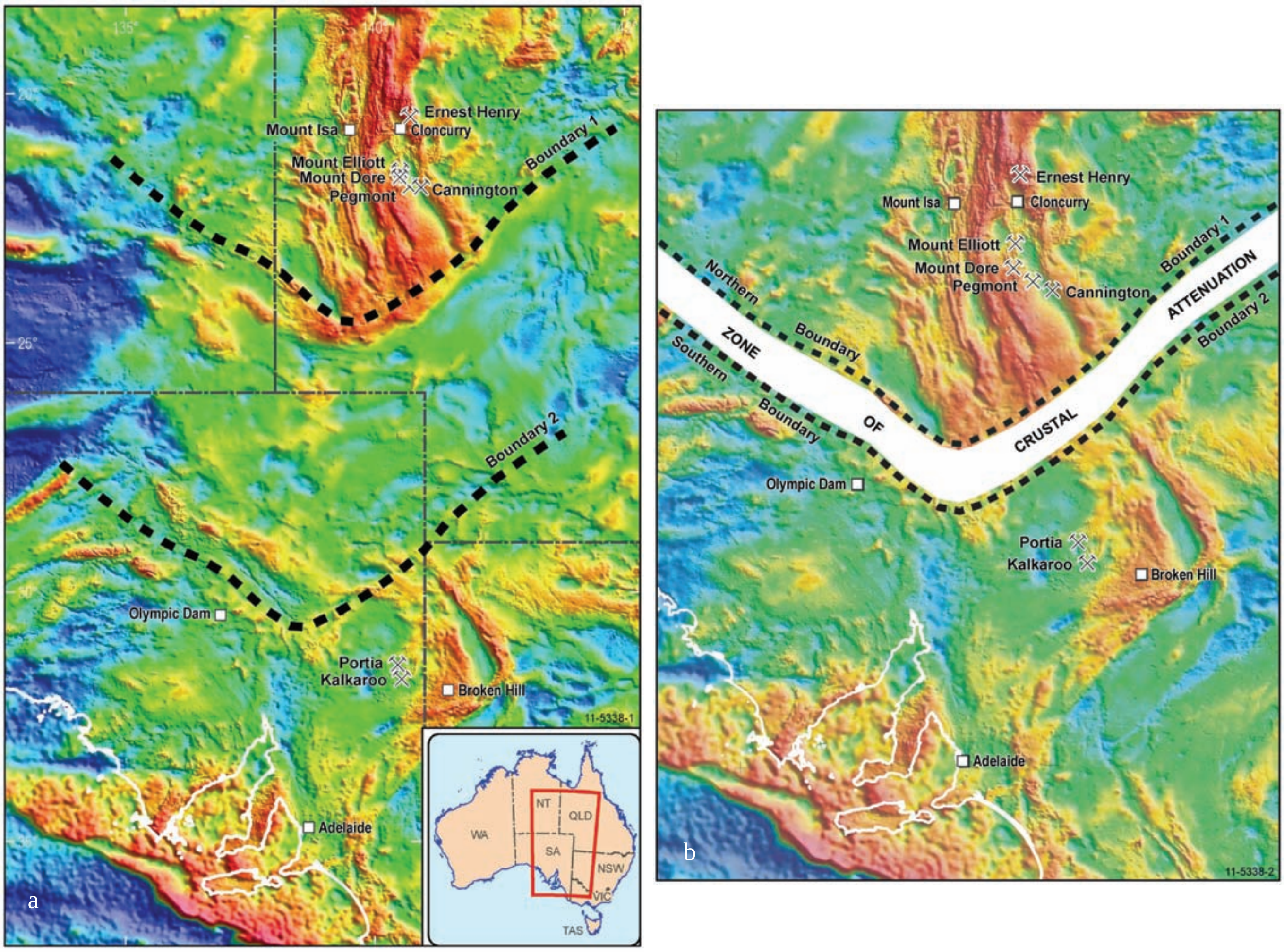


***Figure 2 (a) Uninterpreted gravity anomaly image with Mount Isa and Broken Hill in present-day configuration. (b) Gravity image with Broken Hill basement terrane restored to its pre-Rodinia breakup position opposite the Mount Isa region (after Henson et al., 2011). Note coincidence of Cloncurry-Cannington and Broken Hill gravity trends.***

Moores, 1991) where the primary constraint on reconstruction was not the distribution of the Paleoproterozoic–Mesoproterozoic rift basins but the orogenic belts of Grenville-age along which super-continent assembly is interpreted to have taken place (Figure 1).

## Regional geology and tectonic setting of the Mount Isa region

The Mount Isa region (Figure 3) combines an older crystalline basement (Kalkadoon-Leichhardt Block) affected by the ≥ 1840 Ma Paleoproterozoic Barramundi Orogeny with three variably deformed and vertically stacked superbasins ranging in age from 1790–1575 Ma (Leichhardt, Calvert and Isa superbasins). Equivalents of the Kalkadoon-Leichhardt Block and Leichhardt Superbasin have yet to be identified in Broken Hill even though there are good geological and geophysical grounds for concluding that the two terranes were once continuous with each other (Figure 2b) and evolved in a common tectonic environment (Betts et al., 2011; Gibson et al., 2008; Giles et al., 2002; Henson et al., 2011). Orogenesis in both regions peaked around 1600–1585 Ma (Isa and Olary orogenies) and obscures an earlier history of syn-extensional magmatism, deformation and low pressure-high temperature metamorphism linked to basin formation and normal faulting at higher crustal levels (Conor and Preiss, 2008; Forbes et al., 2008; Gibson and Nutman, 2004; Gibson et al., 2008; Neumann et al., 2009; Page et al., 2005).

Some researchers (Gibson et al., 2008; Holcombe et al., 1991; Passchier, 1986; Passchier and Williams, 1989) have also argued that basin evolution in the Mount Isa terrane was linked at depth to the formation of extensional shear zones best exposed E of the Kalkadoon-Leichhardt Block (Wonga Extensional Belt), inviting comparison with the North American Basin and Range Province where such linkages have been more comprehensively documented (Wernicke, 1985). Nevertheless, it is evident that such comparisons cannot be carried too far because the youngest extensional shear zones at Mount Isa formed no later than 1670 Ma (Gibson et al., 2008; Neumann et al., 2006) and thus relatively early in basin history. A further 70–80 Myr of predominantly deep water marine sedimentation followed, possibly linked to post-extensional thermal subsidence. This and other aspects of basin formation indicate that the Mount Isa terrane is not simply a deformed intra-continental rift or continental back-arc basin (Giles et al., 2002) but shares many similarities with present-day rifted continental margins and evolved almost to the point of sea-floor spreading. In the absence of any direct evidence for seafloor spreading

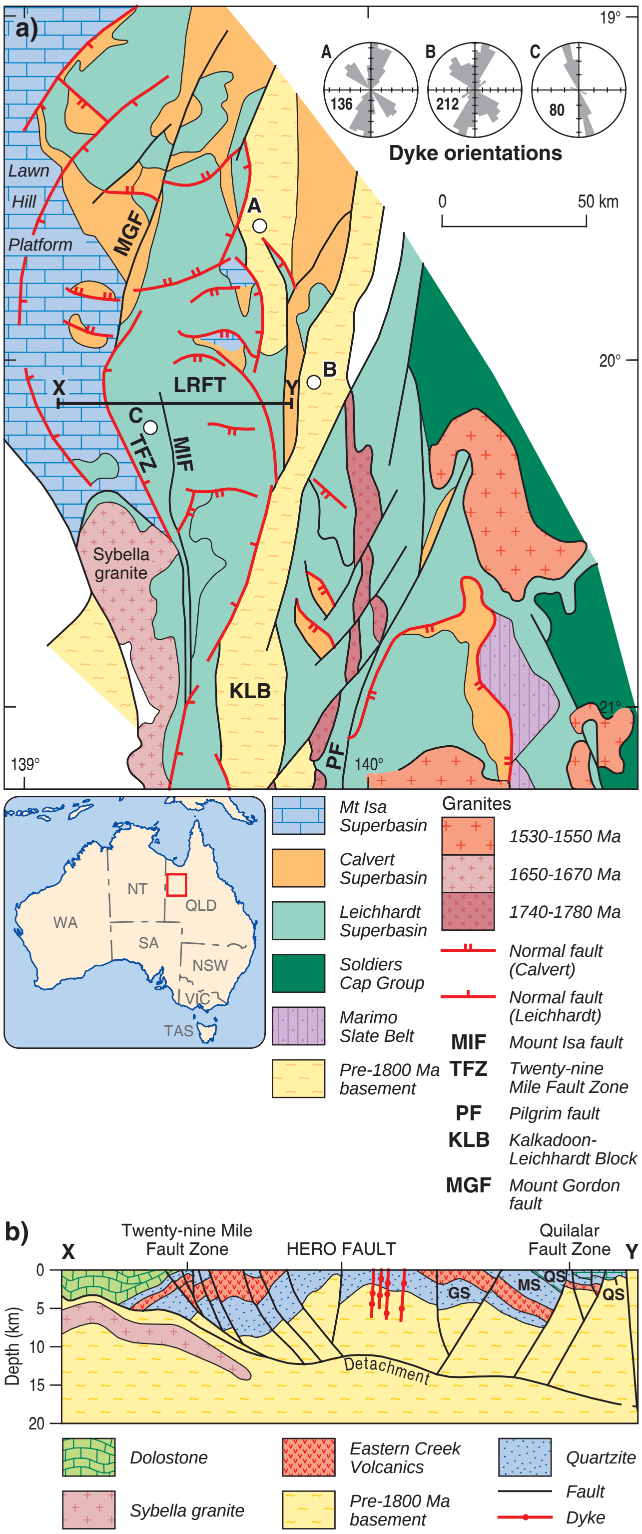


***Figure 3** Simplified geologic map and section for northern part of Leichhardt River Fault Trough (LRFT). Note switch in direction of stratal thickening into Quilalar and Twenty-nine Mile Fault zones at c. 1775 Ma, and parallelism between measured dike orientations (insets A, B and C) and normal fault trends in the LRFT and adjacent Kalkadoon-Leichhardt basement block.QS = Quilalar Supersequence; MS = Myally Supersequence; GS = Guide Supersequence*

in the Mount Isa region itself, rifting may have stepped further outboard and been re-established E of Georgetown (Baker et al., 2010) where continental breakup actually took place, possibly as late as 1650 Ma and no earlier than 1670 Ma (cf. Betts and Giles, 2006).

## Basin history of Mount Isa region

The Leichhardt, Calvert and Isa superbasins each comprise several unconformity-bounded sedimentary packages or supersequences (Jackson et al., 2000; Southgate et al., 2000). These include both syn- and post-rift packages (Betts and Giles, 2006; Blake, 1987; Eriksson et al., 1993; O'Dea et al., 1997) although opinion remains divided about basin architecture and the kinematic framework in which successive packages were deposited. Many of these same packages are recognised here (Figure 4) but with different conclusions drawn regarding basin evolution and tectonic history. Unlike some earlier studies, no definitive evidence was obtained for widespread basin inversion between deposition of the Leichhardt and Calvert superbasins (Betts, 1999, 2001). Instead, a major unconformity between successive extensional regimes is recognised across which there was a switch in the principal extensional direction from ENE-WSW to NE-SW (Figure 4). This switch brought about a major change in the pattern of sedimentation from c. 1730 Ma onward (Figure 4), and superimposed a differently oriented set of extensional structures on a pre-existing rift template (Leichhardt Superbasin). Deep seismic reflection profiles across Mount Isa support the case for a change in extensional direction between deposition of the Leichhardt and Calvert superbasins and show little evidence for significant basin inversion before c. 1640 Ma (Gibson et al., 2010; Queensland Geological Survey, 2011). Basin inversion at this time is further supported by a prominent 1640 Ma hairpin bend in the N Australia polar wander path (Idnurm, 2000) and corresponding change in sedimentation patterns (Southgate et al., 2000).

Further deformation and inversion of basin architecture occurred during crustal shortening and strike-slip faulting accompanying the polyphase 1600–1550 Ma Isa Orogeny but to different degrees on either side of the Kalkadoon-Leichhardt Block which trends N-S and subdivides the Mount Isa region into western and eastern successions (Figure 3). Eastern succession rocks preserve much less of the original basin architecture and have generally undergone deeper burial and more intense deformation than rocks of the same age farther W: peak metamorphism in these rocks occurred at c. 1585 Ma and ranges up to the amphibolite facies whereas greenschist to sub-greenschist facies conditions predominate in the western succession (Foster and Austin, 2008; Rubenach et al., 2008). Together, the eastern and western successions represent an oblique section through the crust whereby structures formed at mid-crustal depths (eastern succession) can be compared to structures formed at higher structural levels in the W.

## Leichhardt Superbasin (1800–1750 Ma)

The Leichhardt Superbasin (Figure 4) developed between 1800–1750 Ma (Neumann et al., 2006) and is best known from the Leichhardt River Fault Trough (LRFT) and southern Lawn Hill Platform (Figure 3) where some 5–7 km of continental flood basalts (Eastern Creek Volcanics) and syn-rift sediments accumulated in an elongate, fault-bounded basin 50–80 km wide (Blake, 1987; Derrick, 1982; Eriksson et al., 1993; Jackson et al., 2000; Scott et al., 2000).

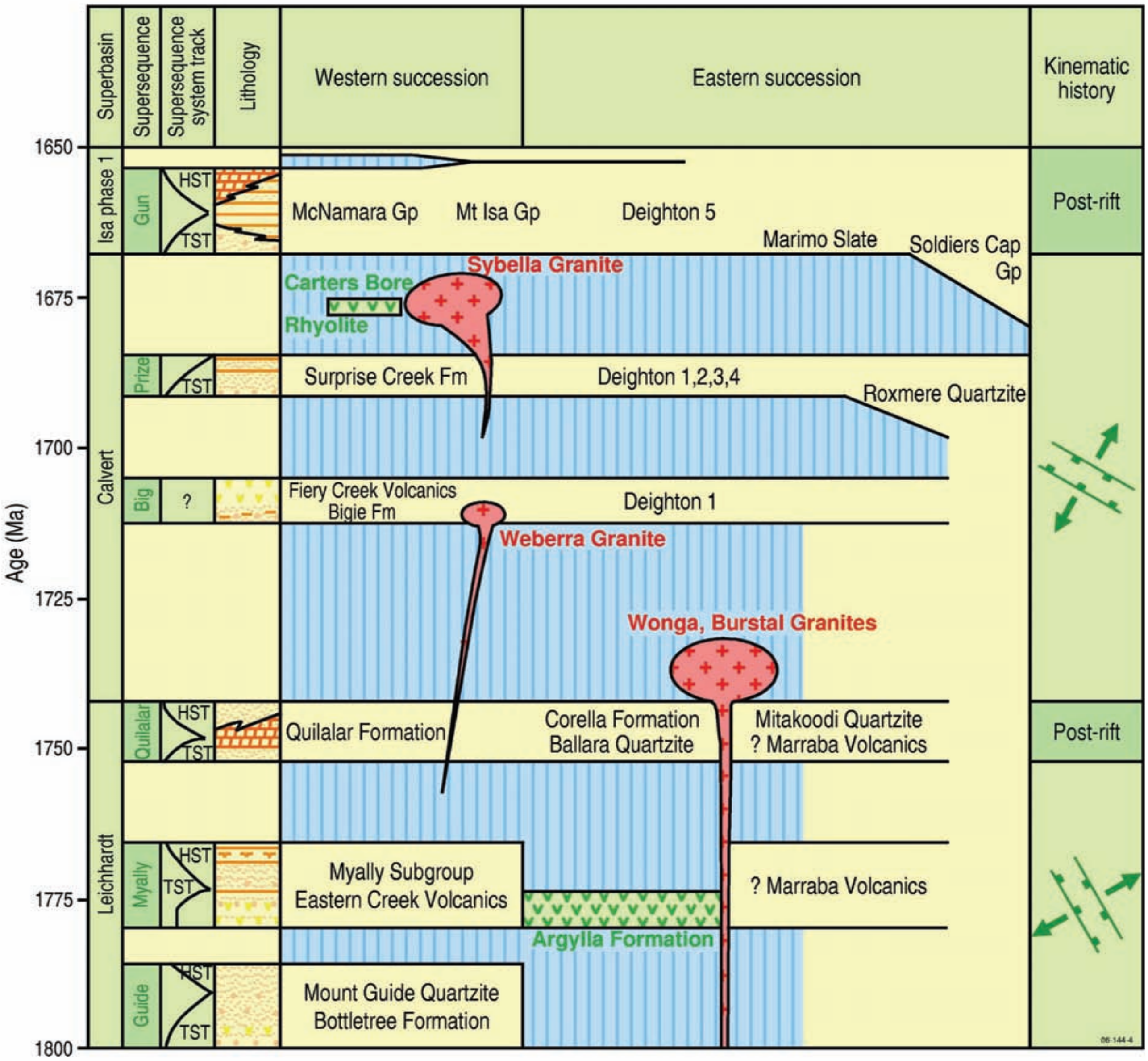


***Figure 4 Chronostratigraphy and interpreted kinematic history for Leichhardt, Calvert and lowermost part of Isa superbasins in Mount Isa region.***

Basin-bounding faults trend NNW and belong to a family of steep, mainly inward-dipping growth faults across which there has been appreciable vertical displacement resulting in half-graben formation (Gibson et al., 2008) and abrupt changes in sedimentary and volcanic thicknesses from the hanging to footwall (Figures 3, 5 and 6). Hangingwall displacements typically range from 100s of metres to a few kilometres (Figure 3) but despite such displacements, topographic relief appears to have been subdued with no evidence for the existence of deep water sedimentary basins. Rather, environmental conditions favoured the deposition of fluviatile to lacustrine sedimentary packages (Guide and Myally supersequences; Figure 4) in which cross- and trough-bedded quartzite and feldspathic sandstone are the dominant lithologies. Red beds with minor amounts of intercalated stromatolitic dolostone (Lochness Formation) commonly occur towards the top of the Myally Supersequence (Figure 4) and most likely represent local excursions into evaporitic or shallow marine conditions (Derrick, 1982; Jackson et al., 2000). Individual half-graben within the Leichhardt Superbasin have dimensions comparable to modern rift basins (Bosworth, 1992) and are up to 70 km long and 30–50 km wide (Figure 3).

Basaltic rocks and interbedded siliciclastic sediments of the 1780–1775 Ma Eastern Creek Volcanics (Figure 4) thicken westwards in the LRFT (Bain et al., 1992; Gibson et al., 2008) and are not known to occur any farther W than the Twenty-nine Mile fault zone (Figure 3). The basalts were extruded under subaerial or shallow water conditions. Their inferred correlatives in the eastern succession are basaltic lava flows of the Marraba Volcanics which are similarly overlain by shallow water quartzite and sandstones (e.g., lower Mitakoodi Quartzite) and contain some layers of clastic sediment (Figure 4). A few interbeds of shallow marine stromatolitic dolostone have been observed within the Marraba Volcanics near its base although in most other respects the depositional environment in the eastern succession at this time does not appear to have been appreciably different to that in the western succession. Felsic volcanic rocks (Argylla Formation) with ages of 1760 and 1780 Ma (Neumann et al., 2009; Page, 1983) at the base of the eastern succession have been widely interpreted as former ignimbrites (Blake, 1987) but have no obvious compositional equivalent in the Eastern Creek Volcanics farther W. Notwithstanding this important difference, both the Argylla Formation and older parts of the Leichhardt Superbasin (Guide Supersequence) have been extensively intruded by dolerite dikes (now metamorphosed) with orientations that match the inferred direction of extension in the Leichhardt Superbasin (Figure 3).

Deposition of the Myally Supersequence was followed (Figure 4) in the western succession by an episode of thermally-induced regional subsidence, leading to marine transgression and burial of the syn-rift sequences beneath a sheet-like cover of fluviatile-shallow marine sediments dominated by clean, well-sorted quartzite and well-bedded, stromatalitic limestones and redeposited calcareous sandstones (Quilalar Supersequence). The eastern equivalents of these rocks (Figure 4) are the Ballara Quartzite and platform carbonate sequences of the Corella Formation (Blake, 1987; Derrick et al., 1980). Detrital zircon ages and intrusion of this platform sequence by the 1740 Ma Burstall Granite (Page, 1983) constrain the age of this marine package to between c. 1755–1740 Ma (Figure 4).

## Calvert Superbasin (1740–1670 Ma)

Onset of rifting and localised uplift in the Calvert Superbasin is marked in the western succession by a major regional unconformity, deposition of fanglomerates and coarse sandstones in fault-angle depressions and fluviatile environments (Bigie Formation), and a rejuvenation of bimodal magmatism (Figure 4) including extrusion of the 1710 Ma Fiery Creek Volcanics (Hutton and Sweet, 1982; Jackson et al., 2000) and intrusion of the 1710 Ma Weberra Granite (Neumann et al., 2006). This was followed by several cycles of upward-fining, mainly siliciclastic sedimentation (Prize Supersequence; Figure 4), during the course of which the depositional environment changed from near-shore to deltaic or shallow marine (Hutton and Sweet, 1982; Southgate et al., 2000). With further deepening of the sedimentary basin(s), increasingly greater amounts of thinly laminated carbonaceous shale or rhythmite were deposited. Stratal thickening of these sequences into E- or NE-trending growth faults points to a syn-rift origin for much of the Calvert Superbasin (Betts et al., 1998; Derrick, 1982; Gibson et al., 2008; O’Dea et al., 1997). Magmatic rocks emplaced during the later stages of rifting include the 1678 Ma Carters Bore Rhyolite and < 50 cm syn-sedimentary peperitic intrusions dated at c. 1690 Ma (Page et al., 2000). These dates provide the best available age constraint on sedimentation in the Calvert Superbasin and are only marginally older than the 1670 Ma age obtained from the Sybella Granite (Neumann et al., 2006) which intrudes basement and/or the Eastern Creek Volcanics near the base of the underlying Leichhardt Superbasin (Figure 3).

Accompanying and/or immediately following the cessation of deposition in the Prize Supersequence (Figure 4), the main sedimentary depocentre shifted eastward into the region now occupied by the Soldiers Cap Group in the eastern succession. This group consists predominantly of metamorphosed deep water siliciclastic turbidites and intercalated carbonaceous sediments which have no direct lateral or temporal equivalent among the shallower water sedimentary facies preserved farther W in the LRFT (Figure 4). Rather, this sedimentary facies is restricted to the eastern succession where it has been intruded by basaltic dikes and sills, including variably metamorphosed 1685 Ma dolerite with highly evolved, Fe-enriched compositions (Baker et al., 2010). Compositionally, these mafic rocks resemble modern-day oceanic tholeiites or basalts extruded through thin sialic crust preceding continental breakup (Barberi et al., 1975; Sinton et al., 1983). Their magmatic age is identical to 1685 Ma detrital zircon ages obtained from their host rocks (Neumann et al., 2009), indicating that sedimentation, crustal thinning and basaltic intrusion were all coeval in at least part of the Soldiers Cap Group (Figure 4). Turbidite deposition in Soldiers Cap Group is consequently viewed here as a response to the same syn-rift extensional processes that gave rise to accommodation space now preserved as Prize Supersequence deposits in the LRFT, despite the slightly younger age (Figure 4) and consequent stratigraphic position above preserved Prize Supersequence in the LRFT. The absence of a preserved temporal equivalent may indicate that parts of the Prize Supersequence have been removed through erosion at the break-up unconformity and now reside farther E in the Soldiers Cap Group (Figure 5).

## Isa Superbasin (1670–1590 Ma)

The Isa Superbasin is best represented on the Lawn Hill Platform (Figure 3) where it comprises 8 km of rhythmically-bedded turbidites, carbonaceous shales and stromatolitic dolostone deposited in a shallow to deep water marine environment (Hutton and Sweet, 1982; Krassay et al., 2000). Farther S, the basal component of this supersequence comprises a transgressive package of fluviatile to shallow marine sandstones, siltstones and dolostones with subordinate amounts of black shale represented by the lower parts of the Gun Supersequence (Southgate et al., 2000). This transgressive package rests unconformably on rocks of the Calvert Superbasin and is widely considered to be of post-rift origin (e.g., Jackson et al., 2000). However, unlike the older Quilalar Supersequence with which it shares many similarities, this transgressive package also shows clear evidence of stratal thickening into the same E-W-trending structures that controlled deposition of the underlying Calvert Superbasin (Southgate et al., 2000). Either these structures continued to be active during the marine transgression or they were simply buried along with any remaining accommodation space during thermal subsidence and deposition of the fluviatile-shallow marine sequence. Possible correlatives of the Isa Superbasin in the eastern succession include thinly laminated carbonaceous slates and siltstones of the Marimo Slate Belt which, like its inferred western correlatives, lacks coeval igneous intrusions.

## Basin evolution and detachment faulting

Concomitant with initial basin evolution in the LRFT, mid-crustal extensional shear zones in the eastern succession (Figure 5) were intruded by bimodal magmatic rocks with ages ranging between 1740–1780 Ma (Gibson et al., 2008; Neumann et al., 2009; Pearson et al., 1991). These shear zones separate a locally exposed lower plate containing mylonitised 1780 Ma Argylla Formation from a brittle upper plate cut by normal faults that penetrate no higher than lowermost Corella Formation (Holcombe et al., 1991; Passchier, 1986). Footwall mylonites give a top-to-the-S or SW sense of shear (Pearson et al., 1991; Gibson et al., 2008) and are constrained by their magmatic host rocks to have formed no later than 1740 Ma and possibly as early as 1780 Ma (Neumann et al., 2009). More importantly, these data indicate that the mylonites and their associated

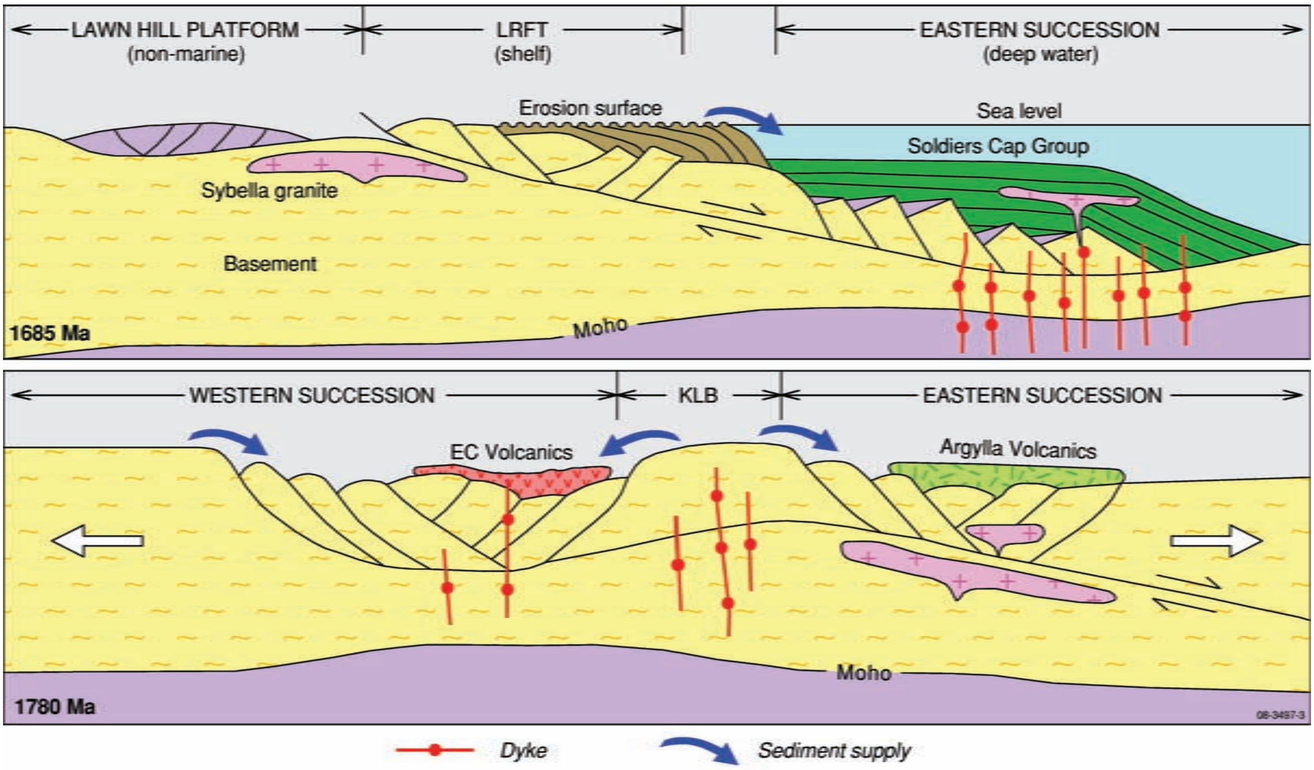


***Figure 5** Generalised W-E crustal sections illustrating links between basin formation, magmatism and syn-extensional detachment faulting in Leichhardt (bottom) and Calvert superbasins (top). Note erosion and reworking of previously deposited platform and near-shore sequences (Prize Supersequence) to form deep water turbidites farther outboard (Soldiers Cap Group). Dolerite dikes are concentrated in region of greatest crustal thinning and deepest water sedimentation.*

shear zones overlap in age with basaltic magmatism and basin formation at higher crustal levels in the Leichhardt Superbasin (Figure 5). A genetic as well as temporal relationship between half-graben formation, bimodal magmatism and development of these mid-crustal extensional detachments is indicated.

Half-graben formation, bimodal magmatism, and the formation of mid-crustal detachments beneath a brittle, extended upper plate are all features shared by other extensional terranes such as the North American Basin and Range Province (Wernicke, 1985). Particularly apt are comparisons with the Rio Grande Rift which shares a similar record of basaltic volcanism followed by fluviatile to lacustrine sedimentation in a narrow intracontinental rift (May and Russell, 1994). It serves as an excellent modern analogue for the Leichhardt Superbasin and, like the latter, comprises a series of half-graben bounded by normal faults that extend downward into a major detachment of extensional origin (Figure 6). Notwithstanding such striking similarities, there is no evidence that formation of the Leichhardt Superbasin was ever accompanied by uplift and the exhumation of metamorphic core complexes as was the case in the Basin and Range Province. Rather, the detachment and associated lower plate mylonites of the Leichhardt Superbasin (e.g., Double Crossing Metamorphics) remained buried until exhumed during a later phase of extension and/or by deformation accompanying the 1600 Ma Isa Orogeny.

By 1685 Ma, basin geometry in the Calvert Superbasin was well established, driven by NE-SW extension and accompanied in the eastern succession by intrusion of basaltic magmas (Figures 4 and 5) into deep marine basins filled by turbiditic sediments (Soldiers Cap Group) (Gibson et al., 2008). Farther W in the LRFT, near-shore, shallow water conditions persisted until arrested by a thermal perturbation at c. 1670 Ma accompanying intrusion and extensional unroofing of the Sybella Granite and its country rocks from mid-crustal depths (Gibson et al., 2008). Unroofing took place on an ENE-dipping detachment surface (Figure 5) that brought about erosion and reworking of rocks belonging to the Leichhardt Superbasin and

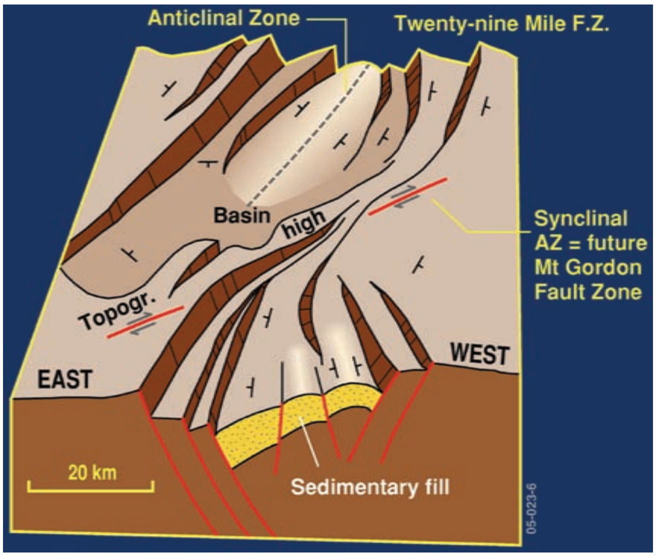


***Figure 6** Schematic representation of basin architecture and fault geometry in the northern part of the Leichhardt River Fault Trough looking south. The future Mount Gordon Fault Zone formed along the accommodation zone (AZ) located between the two sub-basins.*

older parts of the Calvert Superbasin, and their subsequent redeposition in half-graben elsewhere in the basin. Shear fabrics in the Sybella Granite and rotated tilt blocks in Calvert age rocks above the detachment on which unroofing took place further indicate that extension during this stage of basin evolution involved displacement of the upper plate towards the ENE (Gibson et al., 2008) and thus on a detachment that dipped oceanward in the same direction as overall deepening of the sedimentary basin (Figure 5). Less obvious is whether this detachment is the same (reactivated) structure that accommodated extension and normal faulting during formation the older Leichhardt Superbasin. Oceanward-dipping detachments such as these are thought to underlie all sedimentary basins formed in continental margin settings and are a predictable consequence of asymmetric crustal extension and basin development (Lister et al., 1991). The Calvert Superbasin would appear to be a case in point, leading us to conclude that by 1670 Ma basin geometry in the Mount Isa region had evolved beyond a simple intracontinental rift or Basin and Range-type setting into a fully-fledged back-arc basin in which the crust had become appreciably thinned and attenuated, possibly almost to the point of seafloor spreading. In keeping with this interpretation, basaltic rocks in the Leichhardt through to Calvert Superbasin (Eastern Creek Volcanics through Soldiers Cap Group) exhibit compositional changes consistent with extrusion through progressively thinner continental crust (Baker et al., 2010). Equally importantly, the Gun unconformity defining the base of the Isa Superbasin is markedly transgressive and bears a striking similarity to the continental breakup unconformities illustrated by Lister et al. (1991). This unconformity marked the onset of shallow marine conditions across much of the Mount Isa region and was followed by rapid deepening of the depositional environment with sedimentation thereafter dominated by open marine conditions and increasing deposition of turbidite sequences.

## Implications for reconstructions of the Nuna and Rodinia supercontinents

Notwithstanding their obvious great difference in age, the Nuna and Rodinia supercontinents both assume that eastern Australia and western Laurentia represent conjugate rift margins (Betts et al., 2008, 2011; Dalziel, 1991; Karlstrom et al., 2001; Rogers and Santosh, 2009; Zhao et al., 2002). It follows that their constituent terranes were once contiguous and share a common geologic history. In this context, the 1800–1600 Ma history of intracontinental rifting and consequent rift margin formation outlined in this paper for Mount Isa and Broken Hill becomes important because the attendant events pre-date assembly of Rodinia and pertain only to the older Nuna supercontinent. This implies that the assumed longterm connectivity of 1800–1600 Ma orogenic belts between Australia and Laurentia inherent to the AUSWUS and SWEAT reconstructions of Rodinia is incorrect because the rocks in question are unlikely to have remained in their original pre-Rodinia configuration. Rather, following the breakup of Nuna, these rocks and their continental hosts would have dispersed before being reassembled in a different configuration during formation of Rodinia in the Neoproterozoic. A well constrained reconstruction of Rodinia based on matching events and orogenic belts of Grenville age need not work for older rocks such as those reported on here.

This is especially evident in the case of AUSWUS which provides a reasonable match between the Neoproterozoic rift basins of S-central Australia and the southern United States (Figure 1b) but positions the 1800–1600 Ma continental rift basins of eastern Australia, and Broken Hill in particular, along strike from more juvenile accreted terranes of near-identical age in southern Laurentia (Burrett and Berry, 2000; Karlstrom et al., 2001). As pointed out by Betts et al. (2008), the 1800–1600 Ma basins of eastern Australia are more analogous to the basins of interior North America and thus better accommodated in a pre-Rodinia SWEAT-like configuration (Figure 1a) that matches the rocks of Mount Isa against NW Canada (cf. Thorkelson et al., 2001). In this interpretation, the Mojave, Yavapai and Mazatzal terranes have no connection with Broken Hill (Figure 1) and lie much farther S with respect to Australia, occupying a position off the East Antarctic Shield where 1700 Ma eclogites and other collisional rocks have been reported (Goodge et al., 2002). These collisional rocks were interpreted to have formed through the same tectonic processes that accompanied terrane accretion in southern Laurentia and were originally continuous with terranes of the same age developed along the southern margin of the North Australian Craton. In effect, these terranes and their North American counterparts formed a broad continuous belt of 1700–1650 Ma accreted terranes passing from southern Laurentia through Antarctica (Mawsonland) into southern and central Australia. As with the basins of Paleo–Mesoproterozoic age N of the Cheyenne Suture in Laurentia (Figure 1b), the temporally equivalent rift basins in the Mount Isa and Broken Hill regions formed through extension in a back-arc position located above a N-dipping subduction zone (Betts et al., 2008, 2011; Giles et al., 2002). Subduction roll-back, followed by the accretion of continental ribbons and successive juvenile terranes and/or magmatic arcs, were identified as the major drivers of orogenesis along the respective southern continental margins (Barth et al., 2000; Betts et al., 2008, 2011; Karlstrom and Bowring, 1988).

Whether similar far-field stresses could have produced the basins of similar age in NW Canada is not entirely clear although it is interesting to note that the late Paleoproterozoic–early Mesoproterozoic Wernecke Supergroup and Hornby Bay Group both developed on extended crystalline basement of similar age (≥ 1840 Ma) and magmatic character (MacLean and Cook, 2004; Thorkelson et al., 2005) to the Kalkadoon-Leichhardt Block of Mount Isa. As with Mount Isa, the Wernecke Basin (Figure 1a) also experienced basaltic magmatism at 1710 Ma (Fiery Creek vs Bonnet Plume Intrusives) and orogenesis at 1600 Ma (Isa vs Racklan orogeny). Further supporting a connection between the Paleoproterozoic sedimentary rocks of Mount Isa and NW Canada are strikingly similar detrital zircon (U-Pb) ages (Rainbird et al., 2007). Thorkelson et al. (2001) also suggested that the Quilalar Formation and Mary Kathleen Group at Mount Isa are sedimentary equivalents of basin fill in the Wernecke Basin although this makes no distinction between syn-and post-rift components and offers no explanation as to why the much thicker syn-rift sequences of the Leichhardt and Calvert superbasins are not more widely developed in western Canada. This may simply reflect the fact that much of the 1800 Ma western edge of the North American craton lies buried beneath younger rocks of Late Neoproterozoic–early Paleozoic age (Cook et al., 2005) and is only occasionally sufficiently deeply exhumed to expose the underlying late Proterozoic–early Mesoproterozoic basins as in eastern Australia.

Recently published gravity images (Henson et al., 2011) indicate that the late Paleoproterozoic–early Mesoproterozoic sequences of Broken Hill and Mount Isa were not only originally continuous along strike (Figure 2) but formed part of a much more regionally extensive belt of similarly aged rocks that extended southward along the eastern margin of the Gawler Craton (upper Hutchison and Walleroo Groups; Spzunar et al., 2011) into formerly adjacent parts of Antarctica (Peucat et al., 1999). This would imply that the rifted continental margins formed through the breakup of Nuna were oriented grossly N-S (present-day co-ordinates) and had a strike length of several thousand kilometres. Moreover, their orientation was almost orthogonal to the accreted terranes developed along the southern margins of Australia and Laurentia, ruling out any possibility that major Laurentian structures such as the Cheyenne Suture (Figure 1) originally extended into central Australia or has any correlative along the southern margin of the North Australian Craton (Gibson et al., 2008). Rather, the Laurentian terranes would appear to cut across the N-S trend of the Australian late Paleoproterozoic–Mesoproterozoic basins and possibly truncate them. The strike-length of these basins is such that any corresponding conjugate rift margin formed in western Canada at the time of Nuna breakup must be similarly well endowed in an impressive and regionally extensive set of late Paleoproterozoic–earliest Mesoproterozoic rift sequences over and above those currently known and exposed as inliers within the North American Cordillera.

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# Early-mid Cretaceous tectonic evolution of eastern Gondwana: From silicic LIP magmatism to continental rupture

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*The Early–mid Cretaceous marks the confluence of three major continental-scale events in eastern Gondwana: (1) the emplacement of a Silicic Large Igneous Province (LIP) near the continental margin; (2) the volcaniclastic fill, transgression and regression of a major epicontinental seaway developed over at least a quarter of the Australian continent; and (3) epeirogenic uplift, exhumation and continental rupturing culminating in the opening of the Tasman Basin c. 84 Ma. The Whitsunday Silicic LIP event had widespread impact, producing both substantial extrusive volumes of dominantly silicic pyroclastic material and coeval first-cycle volcanogenic sediment that accumulated within many eastern Australian sedimentary basins, and principally in the Great Australian Basin system (>2 Mkm$^3$ combined volume). The final pulse of volcanism and volcanogenic sedimentation at c. 105–95 Ma coincided with epicontinental seaway regression, which shows a lack of correspondence with the global sea-level curve, and alternatively records a wider, continental-scale effect of volcanism and rift tectonism. Widespread igneous underplating related to this LIP event is evident from high paleogeothermal gradients and regional hydrothermal fluid flow detectable in the shallow crust and over a broad region. Enhanced $CO_2$ fluxing through sedimentary basins also records indirectly, large-scale, LIP-related mafic underplating. A discrete episode of rapid crustal cooling and exhumation began c. 100–90 Ma along the length of the eastern Australian margin, related to an enhanced phase of continental rifting that was largely amagmatic, and probably a switch from wide–more narrow rift modes. Along-margin variations in detachment fault architecture produced narrow (SE Australia) and wide continental margins with marginal, submerged continental plateaux (NE Australia). Long-lived NE-trending cross-orogen lineaments controlled the switch from narrow to wide continental margin geometries.*

## Introduction

Rifting, magmatism and sedimentation have fundamentally shaped the modern eastern Australian margin and many, often disparate studies have pin-pointed the Early Cretaceous as when these tectonic-driven processes began affecting and shaping the modern-day eastern margin. As a result of the spatial-temporal overlap of volcanism and extension (e.g., Ewart et al., 1992; Bryan et al., 2000; Och et al., 2009), the eastern margin of Australia has been reinterpreted as a volcanic rifted margin (Bryan et al., 1997), but which shows some important differences to better known volcanic rifted margins such as those surrounding the Atlantic and Red Sea (e.g., Menzies et al., 2002). In general, all these magmatically-enhanced rifted margins extend laterally over distances of 2,000 km (Armitage et al., 2010) and potentially have high significance to Earth history, being linked to mass extinctions, changes in global climate, and natural resource generation (Svensen et al., 2004; Storey et al., 2007).

The purpose of this contribution is to emphasise some important aspects of the Early Cretaceous history of eastern Gondwana: (1) the close spatial-temporal relationship of Silicic Large Igneous Province (LIP) magmatism to continental rifting and passive margin formation, and how complexities in the rifting process have hindered an appreciation of the immensity of magmatism during the Early Cretaceous in eastern Australia; (2) the eruptive output of this igneous province and to provide an upward revision of extrusive volume estimates following recent studies along the eastern Australian margin;

(3) that a large proportion of the erupted products are preserved as huge volumes of coeval volcanogenic sediment in adjacent sedimentary basin systems; and (4) important long-term volcanic and compositional trends of magmatism.

## The Eastern Australian Volcanic Rifted Margin

The Cretaceous–Cenozoic eastern Australian volcanic rifted margin is c. 4,500 km long and mirrors the Jurassic–Cretaceous basalt-dominated volcanic rifted margin of western Australia some 2,500–4,000 km to the W. Over most of its length, the eastern margin is flanked by a major continental-scale physiographic feature (the Great Dividing Range; Figure 1; e.g., Ollier, 1982). The origins of both these rifted margins were initially located in intraplate continental positions, but subsequent rifting to form new ocean basins has also resulted in large portions of extended continental crust and syn-rift successions now residing offshore (Figure 1). The Whitsunday Silicic LIP was the major magmatic precursor event to rifting along the eastern Australian margin, and gives the rifted margin its volcanic character. This section summarises a number of aspects of the development of this rifted margin beginning in the Early Cretaceous.

### Nature of pre-rift margins to the Whitsunday Silicic LIP

Two major continental sedimentary basin systems, the Great Australian and Otway-Gippsland-Bass basins in NE and SE Australia, respectively (Figure 1) flanked the Whitsunday Silicic LIP along its western margin. These pre-existing basin systems were major repositories for huge volumes of volcaniclastic material erupted from the Silicic LIP (Bryan et al., 1997). In contrast, the eastern margin of the Silicic LIP was an elevated margin or likely bordered by rift basin complexes developed coincident with Silicic LIP magmatism. Compiled geophysical data indicate that several offshore basins (Central and Western Rift provinces of the Lord Howe Rise, New Caledonia, Queensland and Townsville) began opening in the Early Cretaceous (Wellman et al., 1997; Willcox and Sayers, 2001; Willcox et al., 2001; Lafoy et al. 2005). However, recent ODP drilling (leg 194) on the Marion Plateau to the E of the Whitsunday Silicic LIP (Figure 1) intersected Jurassic (162 ± 0.9 Ma; Karner and John, unpublished data, 2002) intraplate alkali basalts beneath Cenozoic Oligocene–Miocene sedimentary rocks (Isern et al., 2002), defining an eastern limit to the Silicic LIP. These Jurassic intraplate alkali basalts reinforce an intraplate passive margin setting for NE Australia during the Jurassic and Cretaceous and preclude the Whitsunday Silicic LIP as representing some back-arc extensional magmatic event.

### Duration and relative timing of rifting and magmatism

Rifts in the interior of continents that evolve to form large ocean basins typically last for 30 to 80 Myr and longer before complete rupture of the continent and onset of sea-floor spreading (Umhoefer, 2011). Silicic LIP magmatism immediately preceded large-scale continental rifting that began in the middle Cretaceous and led to: (1) the opening of a series of basins, some floored by oceanic crust

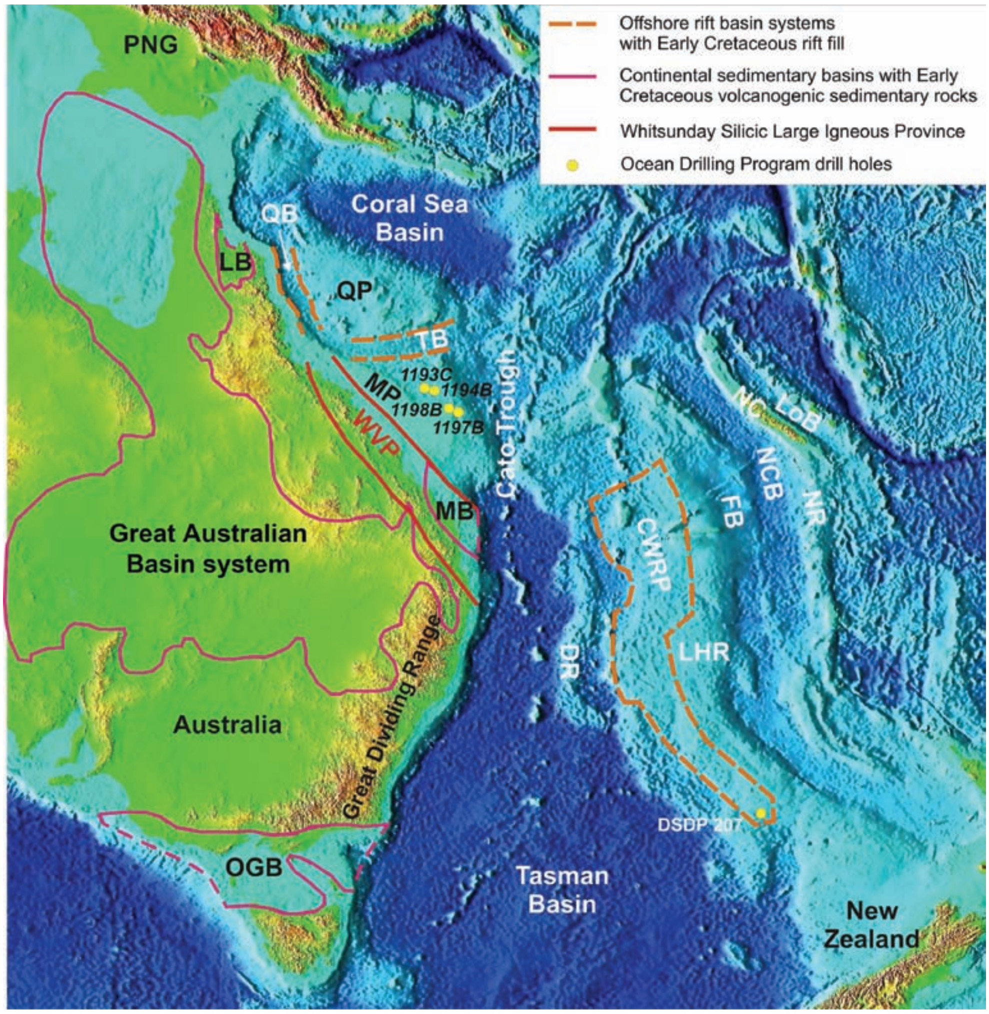


*Figure 1 Topographic-bathymetric map of the SW Pacific with the tectonic elements referred to in text labelled. Drill holes, 1193C, 1194B, 1197B and 1198B are from ODP Leg 194 on the Marion Plateau (MP). Abbreviations: CWRP, Central and Western Rift Provinces of the Lord Howe Rise (LHR); DR, Dampier Ridge; FB, Fairway Basin; LB, Laura Basin; LoB, Loyalty Basin; MB, Maryborough Basin; NC, New Caledonia; NCB, New Caledonia Basin; NR, Norfolk Ridge; OGB, Otway-Gippsland-Bass basins; PNG, Papua New Guinea; QB, Queensland Basin; QP, Queensland Plateau; TB, Townsville Basin; WVP, Whitsunday Volcanic Province.*

(Tasman Basin-Cato Trough-Coral Sea Basin system, South Loyalty Basin) with others by extended continental crust (New Caledonia, Lord Howe, Middleton, Queensland, Townsville and Capricorn Basins), and (2) the dispersion of rifted microcontinents (e.g., Lord Howe Rise, Dampier Ridge, New Caledonia-Norfolk Ridge, Queensland Plateau) away from the eastern margin of Australia (Figure 1). The Tasman-Cato Trough-Coral Sea Basin system is the largest of the basin systems, with sea floor-spreading occurring between the Cretaceous and Paleocene (chron 33–24, 84–52 Ma; Lafoy et al., 2005). The Tasman Basin is triangular in outline, the result of a sea floor-spreading system that propagated northward in a zipper-like fashion over time, resulting in the separation of the Lord Howe Rise microcontinent (Figure 1) from eastern Australia (Willcox et al., 2001).

The end-result of Cretaceous–Cenozoic rifting of eastern Gondwana is that the Silicic LIP is now largely dismembered and only part of the province remains intact (Figures 1 and 2). This is primarily because break-up and sea floor-spreading processes were complex, involving the movement of several microplates, the failure of several rifts and consequent ridge jumps by the sea floor-spreading system (Gaina et al., 1998, 2003; Willcox et al., 2001). Of interest is that initial half spreading rates of the Tasman Basin from 83–79 Ma are indicated to have been unusually slow (4 mm/year) causing the basin to only open by c. 30 km during this time (Gaina et al., 1998). The initially slow rates of rifting may be an important factor in explaining the extended period between Silicic LIP magmatism (c. 130–95 Ma) and the first appearance of new oceanic crust in the earliest Campanian (c. 84 Ma). Understanding the complexities of the rifting processes and dismembering of the Silicic LIP are important because they have been major factors in hindering the recognition of: (1) the scale and magnitude of Early Cretaceous silicic magmatism; and (2) the tectonic setting and relationship of separated but coeval igneous and volcanosedimentary terranes in eastern Gondwana. Past interpretations of individual terranes in isolation have led to contradictory tectonic models for the Early Cretaceous of eastern Gondwana.

## Regional thermal and hydrothermal effects

In addition to the main magmatic expression, this Silicic LIP and rifting event also had regional thermal and hydrothermal effects associated with mantle $CO_2$ degassing. Numerous apatite fission track thermochronology studies in eastern Australia have recorded a major Early Cretaceous heating event (coincident with Silicic LIP magmatism), with increases in maximum Cretaceous paleotemperatures toward the margin, followed by mid-Cretaceous cooling beginning c. 100–95 Ma (e.g., O'Sullivan et al., 1995, 1996, 1999; Marshallsea et al., 2000; Kohn et al., 2002, 2003). These studies have generally concluded that: (1) Cretaceous heating was due to a greater depth of burial and increased paleogeothermal gradient (crustal heat flow), and (2) mid-Cretaceous cooling occurred in response to kilometre-scale denudation associated with rifting along the eastern Australian margin, leading to the formation of a passive margin mountain range (the Great Dividing Range of eastern Australia). A widely recognised paleomagnetic overprint affecting the crust of southern and eastern Australia is also attributed to this Early to mid-Cretaceous thermal event (e.g., Thomas et al., 2000). Significant regional mantle $CO_2$ degassing and mineral trapping occurred during the Cretaceous, which is manifested by widespread authigenic carbonate and clay mineral precipitation throughout the late Permian

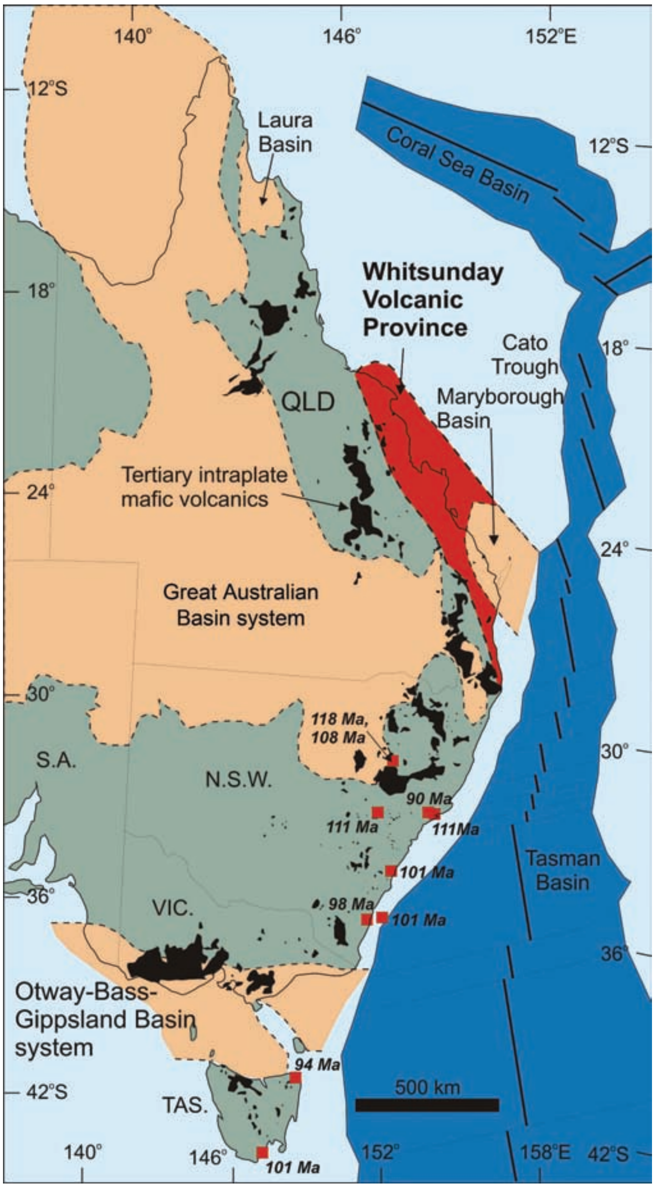


***Figure 2** Location of the Whitsunday Volcanic Province and Early Cretaceous sedimentary basins of eastern Australia that contain >1.4 Mkm$^3$ of coeval LIP-derived volcanogenic sediment (Bryan et al., 1997). Red squares are locations of dated igneous rocks (ages in italics) along the SE margin of Australia. Intraplate alkaline volcanism (80–0 Ma) is shown in black.*

in the Bowen-Gunnedah-Sydney basin system and well inboard of the Whitsunday Silicic LIP along the eastern Australian coast (Baker et al., 1995; Uysal et al., 2000, 2011; Golab et al., 2006). These new data raise the possibility that mantle and magmatic gas emissions may not be restricted to eruptive events in the LIP and much greater greenhouse gas emissions may occur over much larger areas during LIP and rifted margin events.

## Post-rift intraplate alkaline volcanism

The Whitsunday Silicic LIP is like many other LIPs in being followed by asthenospheric-derived or "hotspot"-style mafic volcanism. This intraplate alkali basaltic volcanism shows a clear spatial association with the passive margin mountain range, forming a broken belt 4,400 km long along the 'highlands' of eastern Australia (Figures 1 and 2). Intraplate alkaline volcanism has continued to the Holocene, occurring within 500 km of the coastline, and has an extrusive volume of >20,000 km$^3$ (e.g., Johnson, 1989). However, this volcanism is part of a much broader, mostly basaltic and alkaline igneous province emplaced across continental and oceanic lithosphere in the SW Pacific Ocean region during the Cenozoic (Finn et al., 2005). For eastern Australia, several features of note are that: (1) an c. 15 Myr hiatus occurred between the terminal phases of Whitsunday Silicic LIP magmatism and the first expressions of intraplate mafic volcanism; this hiatus correlates with a period of uplift, erosion and abrupt crustal cooling of the eastern margin, based on the apatite fission track data; (2) the widespread eruption of intraplate alkali basalts overlapped in time with sea floor-spreading in the Tasman Basin, and in particular, that the onset of intraplate mafic volcanism at c. 80 Ma (Sutherland et al., 1996) coincided with a more than five-fold increase in the spreading rate to 22 mm/year in the Tasman Basin at 79 Ma (Gaina et al., 1998; Willcox et al., 2001); (3) some of the youngest intraplate mafic volcanism has occurred in northern and southern Australia and at the extremities of the intraplate volcanic belt; and (4) the most primitive basalt geochemical signatures from the Whitsunday Silicic LIP overlap those of the younger within-plate alkaline basalts of eastern Australia (Ewart et al., 1992), indicating a 'geochemical connection' between pre-break-up Silicic LIP magmatism and post-break-up intraplate volcanism (Bryan et al., 2000). In summary, the time-space relationships between magmatism, highlands uplift and sea floor-spreading are most readily explained by detachment models where eastern Australia is interpreted as an upper-plate passive margin (e.g., Lister and Etheridge, 1989).

## The Whitsunday Silicic Large Igneous Province

Early–mid Cretaceous (c. 135–95 Ma) magmatic products along the eastern margin of Australia define the Whitsunday Silicic LIP, which is the largest of the world's Silicic LIPs where the eruptive output (>2.5 Mkm$^3$) and preserved areal extent of volcanism and its products surpass that of many other LIPs (Bryan et al., 1997, 2000; Bryan, 2007). Silicic LIP magmatism occurred as a within-plate, silicic-dominated pyroclastic volcanic belt, >2,500 x 300 km, roughly coincident with the present eastern Australian plate margin (Figure 1). The Whitsunday Silicic LIP is defined by the following igneous, volcanosedimentary and tectonic elements of Early Cretaceous age: (1) the Whitsunday Volcanic Province of NE Australia; (2) scattered igneous intrusions and volcanic rocks along the SE Australian margin; (3) the Great Australian and Otway-Gippsland-Bass basin systems of NE and SE Australia, respectively; and (4) submerged volcanic and rift-fill sequences on marginal continental plateaux and troughs.

### Whitsunday Volcanic Province

The Early Cretaceous Whitsunday Volcanic Province is defined here to encompass all igneous rocks preserved in the relatively intact extension of the Silicic LIP along the NE Australian margin (Figures 1 and 2). It was preserved on the Australian continental margin following an eastward ridge jump of the Tasman spreading ridge into the Cato Trough at c. 65 Ma (Falvey and Mutter, 1981).

The uplift and exhumation history, related to rifting of the margin, has played an important role in the preservation and general character of Early Cretaceous igneous rocks in Queensland. Recent dating efforts, principally as part of regional mapping programs, have increasingly recognised a much greater extent and number of Cretaceous igneous rocks through eastern Queensland (Figure 3). Many of these units for example, had previously been mapped as Carboniferous or Permian in age. Distinct differences in preservation occur along the Queensland margin (Figure 3), and a boundary between dominantly volcanic successions and granitic intrusions can be delineated. This newly compiled distribution of Early Cretaceous igneous rocks (Figure 3) demonstrates very different and abrupt changes in exhumation history from W–E across the northern (Whitsunday) region, with volcanic rocks limited to coastal sections and the Whitsunday Islands. Further S through central Queensland, volcanic rocks are preserved much further inboard of the margin, whereas in SE Queensland, volcanic rocks are principally restricted in occurrence to near the margin (Maryborough Basin) and intrusive rocks predominate. These along margin variations relate to major cross-orogen lineaments (Figure 3), which have partitioned extensional deformation and exhumation during the Cretaceous–Cenozoic; these lineaments also partitioned contractional deformation during the Permo–Triassic Hunter-Bowen Orogeny (Holcombe et al.,

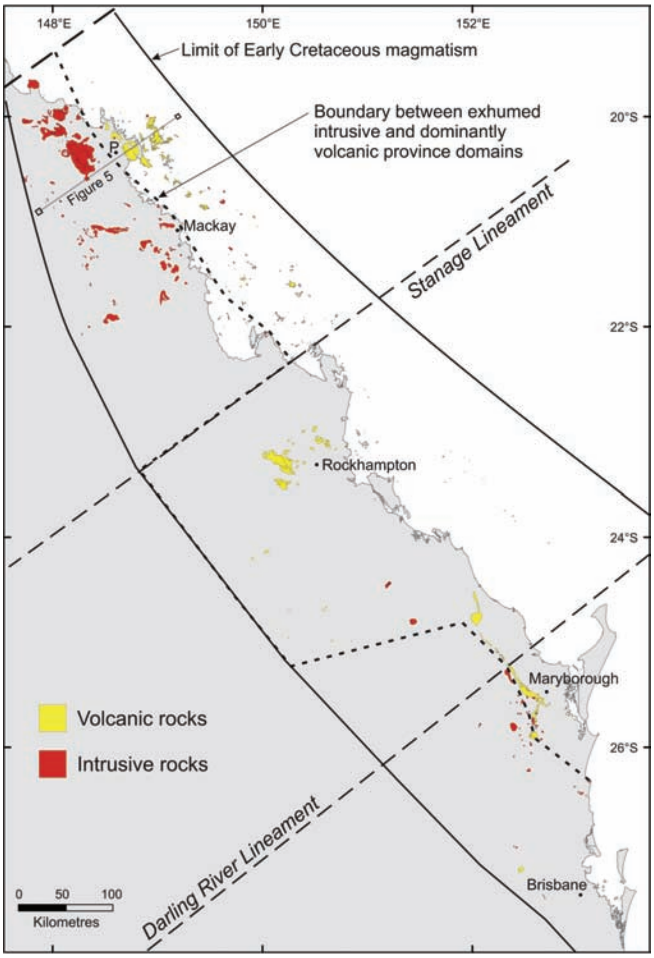


***Figure 3 New distribution map of Early Cretaceous igneous units in Queensland, based on new mapping and dating. Different erosional levels occur along the Queensland margin and dominantly volcanic versus dominantly intrusive domains can be recognised, which are related to major cross-orogen lineaments. P, Proserpine.***

1997). The distribution of Cenozoic intraplate volcanics (Figure 2) and the watershed of the Great Dividing Range (passive margin mountain range) in part mirror this along margin partitioning.

## Volcanic Geology

Lithologically, the Whitsunday Volcanic Province is volumetrically dominated by welded dacitic–rhyolitic and relatively lithic-rich ignimbrite (Bryan et al., 2000), and most exposures often present monotonous sequences of stacked, welded ignimbrite units forming km-thick sections in the northern part of the province. The eruptive powerhouses of the province were several, relatively large (10–20 km diameter) calderas that form a NW-trending belt through the northern part of the province. Some intracaldera ignimbrite units are up to 1 km thick (Clarke et al., 1971; Ewart et al., 1992; Bryan et al., 2000), whereas extracaldera units are 10s to less commonly, 100s metres thick. Coarse lithic lag breccias containing clasts up to 6 m diameter (Ewart et al., 1992) cap the ignimbrites in proximal sections and record caldera collapse episodes. Exhumed caldera sections (e.g., Whitsunday Island) reveal multiple caldera-forming ignimbrite eruptions producing caldera fill successions of at least 3–4 km thick. Several calderas were flooded, being occupied by shallow lakes and rhyolite domes during eruptive hiatuses. Phreatomagmatic eruptive phases characterised some caldera-forming eruptions as a result of explosive magma interaction with the caldera lakes. Intercalated with the silicic pyroclastic rocks are basaltic and silicic lavas, with andesite–dacite lavas rare. Basalt lavas are uncommon in island (eastern) exposures but are volumetrically more abundant in mainland (western) exposures. Associated with the volcanics are locally significant thicknesses of coarse volcanogenic conglomerate and sandstone (Clarke et al., 1971; Bryan et al., 2000). The sedimentary rocks are texturally and compositionally immature, reflecting the local volcanic provenance, and sedimentation appears to have been in poorly confined, high-energy alluvial environments or caldera lakes. Dyke swarms and sills are an integral feature of the volcanic sequences, with individual dykes ranging up to 50 m in thickness, and have a strong N-S orientation (Ewart et al., 1992; Bryan et al., 2000, 2003). Silicic dykes predominate in the exhumed caldera sections, whereas mafic dykes and sills are more abundant in central and western exposures where basaltic lavas are also more common. Early Cretaceous granites also intrude the volcanic sequences (Figure 3). Overall, the volcanic sequences record a multiple vent, but caldera-dominated, low-relief volcanic region (Bryan et al., 2000).

## Geochemistry

Chemically, the igneous suite ranges continuously from basalt to high-silica rhyolite, with calc-alkali to high-K affinities (Ewart et al., 1992; Bryan, 2007). In detail, however, this range is defined by dyke compositions (Figure 4); lavas are bimodal (basalt–andesite and rhyolite to high-silica rhyolite), and ignimbrites are dominantly rhyolitic in composition. It should be noted that the geochemical data set is biased towards the volumetrically minor lavas and dykes (311 out of 585 analyses), and when volume considerations are taken into account, igneous rock compositions within the province are overwhelmingly silicic.

The range of compositions has been generated by two-component magma mixing and fractional crystallization superimposed to produce the rhyolites (Ewart et al., 1992). The two magma components are:

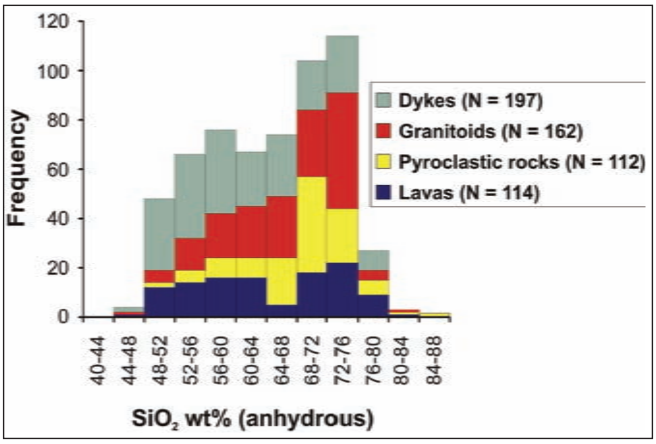


***Figure 4 Silica histogram showing the distribution of compositions within dykes, lavas, pyroclastic rocks, and coeval granitoid rocks from the Whitsunday Volcanic Province (585 analyses).***

(1) a volumetrically dominant partial melt of relatively young (Mesozoic–Paleozoic), non-radiogenic calc-alkaline crust; and (2) a within-plate tholeiitic basalt of E-MORB (Enriched-Mid Oceanic Ridge Basalt) affinity, and similar to the Cenozoic intraplate basalts of eastern Australia (Ewart et al., 1992; Bryan, 2007). A critical point to emphasise is that the calc-alkaline and arc-like signatures have been inherited from the crustal source, and do not provide any constraints on the (Early Cretaceous) tectonic setting in which the magmas were produced (e.g., Roberts and Clemens, 1993). The volumetric dominance of silicic igneous compositions reinforces the point that Early Cretaceous magmatism reflected a massive partial melting event of the continental crust.

## Geochronology

Relatively limited K/Ar and Rb/Sr isotopic dating of the Whitsunday Volcanic Province has established an age range of 132–95 Ma (Figure 5), but with a main period of igneous activity indicated between 120 and 105 Ma (Ewart et al. 1992; Bryan et al., 1997). Other K/Ar age data from eastern Queensland (e.g., Green and Webb, 1974; Allen et al., 1998) suggest precursory magmatic activity may have begun as early as 145 Ma, although U/Pb zircon dating has as yet, not duplicated these older ages. However, it is being increasingly

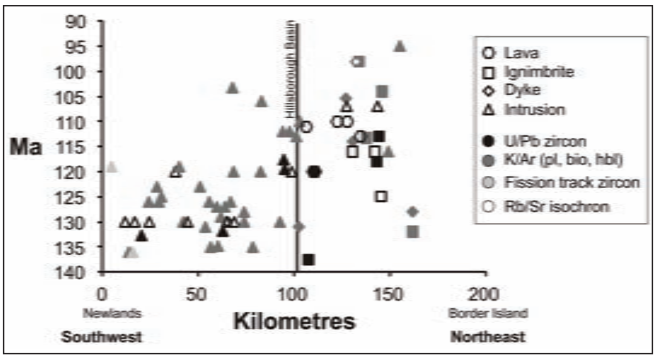


***Figure 5 Space-time presentation of age data from the northern Whitsunday Volcanic Province (updated from Bryan et al., 1997). Ages have been projected onto a SW-NE cross-section (see Figure 3), from the mainland (Newlands) to outer islands (Border Island). Age data are distinguished in terms of lithology and dating method.***

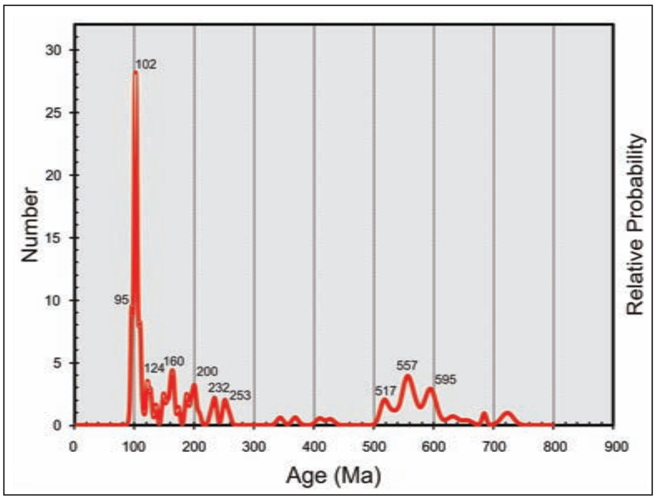


***Figure 6 Probability-density distribution plot of detrital zircon $^{208}$Pb corrected $^{206}$Pb/$^{238}$U ages <900 Ma from the latest Albian–Cenomanian Winton Formation of the Great Australian Basin system (Greentree, 2011). Sample is 700 km WSW from the centre of the Whitsunday Silicic LIP. Zircon ages <130 Ma are from euhedral zircons, highly rounded and abraded zircons yielded ages between 400 Ma and 2.8 Ga (Greentree, 2011). Plot contains 107 concordant ages.***

recognised that some Late Jurassic and Early Cretaceous K/Ar ages are reset ages (e.g., Allen et al., 1998; Uysal et al., 2001) due to the major thermal event associated with Silicic LIP magmatism (e.g., as evidenced by Early Cretaceous heating in the apatite fission track data for eastern Australia). Preliminary LA-ICP-MS U/Pb zircon dating of ignimbrites from the Whitsunday Volcanic Province (Bryan and Allen, unpublished data, 2005) supports the notion of a main period of activity between 120–105 Ma, and pulses are indicated between c. 118–113 and c. 110–105 Ma (Bryan and Ernst, 2008). However, a detrital geochronology study of the coeval (latest Albian–Cenomanian) volcanogenic Winton Formation in the Great Australian Basin system has recognised a dominant volcanic and euhedral detrital zircon component of c. 105–95 Ma (Figure 6). These ages, their general absence from the Whitsunday Silicic LIP, but widespread occurrence in age spectra of Mesozoic–Cenozoic basins and sediments (e.g., Sircombe, 1999; Cross et al., 2010), and the exhumation history of the Queensland margin (Figure 3) implies a substantial missing section exists in the Whitsunday Volcanic Province, which is now preserved in adjacent sedimentary basins. The new detrital zircon ages thus reinforce a young pulse of silicic explosive volcanism between 105–100 Ma, and the continuation of significant rhyolitic ignimbrite volcanism to c. 95 Ma.

The LA-ICP-MS studies have identified zircon xenocrysts with ages of c. 490 and 787 Ma. These inherited U/Pb ages are consistent with the Nd model $T_{DM}$ ages of 200–600 Ma calculated for the magma crustal sources (Ewart et al., 1992). Considerably more geochronological work is required to understand the eruptive history of the province, in particular to constrain better, crustal sources, melt production rates, and eruptive volume and composition relationships with time.

## Igneous rocks of southeast Australia

Minor volumes of Early Cretaceous igneous rocks are preserved along the SE margin of Australia (Figure 2) with the most prominent example being the Mount Dromedary ring complex (e.g., Smith et al., 1988; Nott and Purvis, 1995). Early Cretaceous intrusions, most likely related to the Mount Dromedary igneous complex have also been dredged from the continental slope off southern New South Wales (Hubble et al., 1992). The rocks along the SE Australian margin tend toward bimodal compositions, occurring as basalt–basanite lavas and dykes and monzonite–syenite intrusions and trachytic to rhyolitic volcanics. Overall, igneous compositions are slightly more alkaline than coeval igneous rocks from the Whitsunday Volcanic Province to the N (Smith et al., 1988; Middlemost et al., 1992). Igneous ages are mostly between c. 110–90 Ma (Figure 2; Jones and Veevers, 1983, Middlemost et al., 1992; Hubble et al., 1992). More substantial volumes of Early Cretaceous igneous rock existed along the SE margin, but have subsequently been rifted away and are now located in submerged rift basins on the Lord Howe Rise on the eastern side of the Tasman Basin. The strong increase in maximum Cretaceous paleotemperatures towards the SE margin as evidenced by apatite fission track data (e.g., Kohn et al., 2002) and thermally reset K/Ar ages of fault-related illite at c. 120 Ma in the Sydney region (Och et al., 2009) provide indirect evidence that a major igneous belt was located close to the SE margin during the Early Cretaceous.

## Great Australian Basin system

In addition to voluminous silicic volcanism, the Cretaceous geological history of eastern Australia is marked by the development of extensive sedimentary basins (Figure 1), most notably the Great Australian Basin system. The latest Triassic–Cretaceous Great Australian Basin (GAB) system covers 22% of Australia, but the preserved remnants of this intracratonic basin system and the stratigraphy of other, now isolated areas of Jurassic–Cretaceous outcrop suggest the GAB may have originally covered a much larger area of eastern Australia (Fielding et al., 1996). The various components of the GAB (Eromanga, Surat, Carpentaria, Laura and Clarence-Moreton basins) preserve Early Cretaceous (Aptian–Cenomanian), volcanogenic sedimentary rocks that cover >2 Mkm$^2$ to an average thickness of 500 m. The volcanogenic sedimentary rocks are dominantly mudstone, siltstone and sandstone that were deposited in environments ranging from fluvial/lacustrine to coastal plain and shallow marine (Smart and Senior, 1980; Hawlader, 1990; Fielding, 1992; Greentree, 2011). This volcanogenic sedimentation represents an abrupt and fundamental change in sediment provenance from underlying (Neocomian), basement-derived quartzose sandstones (Bryan et al., 1997). Volcanogenic sandstones form the Surat Basin are feldspathic–lithic, with an average Q:F:L ratio of 15:41:44, and volcanic lithic grains represent more than 90% of the total lithic component (e.g., Hawlader, 1990). Paleocurrent data indicate an easterly source (the Whitsunday Silicic LIP) for sediment (Hawlader, 1990; Greentree, 2011). The sheet-like external and internal geometries of the Aptian–Albian formations within the GAB suggest that predominantly passive, thermal subsidence (Fielding, 1996) coincided with volcanism along its eastern margin.

## Otway-Gippsland-Bass Basin System

The WNW-trending Otway, Gippsland and Bass basins in SE Australia (Figures 1 and 2) are transtensional rift basins that form the eastern extremity of a complex rift system (Southern Rift System of

Willcox and Stagg, 1990), which extended along the length of the southern margin of Australia during the Late Jurassic–Early Cretaceous. Their formation represents a precursor stage to the post-mid Cretaceous break-up of Australia and Antarctica (e.g., Falvey and Mutter, 1981). These basins are significant for containing >2.5 km thickness of Aptian–Albian age volcanogenic sandstone and mudstone, but significantly, the source volcanism was unrelated to Australia-Antarctic rifting.

As in the GAB, this volcanogenic sedimentation represented a fundamental change in provenance from underlying basement-derived quartz-rich sandstones, such that volcanogenic sedimentation was essentially to the exclusion of basement-derived material (Gleadow and Duddy, 1980). Paleocurrent data indicate the source of the volcaniclastic sediment lay outside and to the E of the Gippsland Basin (the easternmost basin; Bryan et al., 1997). The volcanogenic sandstones are dominantly volcanic lithic, with detrital minerals of predominantly plagioclase, with lesser quartz, hornblende, pyroxene (augite), apatite, titanite, and zircon. Fission track dating of apatite, titanite, and zircon supports a contemporaneous volcanic source for the sediment (Gleadow and Duddy, 1980), and is reinforced by palynological dating that indicate the volcanogenic sandstones are Aptian–Albian in age.

## Submerged volcanic and rift-fill sequences on marginal continental plateaux and troughs

It has previously been suspected by several workers that a substantial volume of Early Cretaceous igneous and volcanogenic sedimentary material may be present in submerged rift basins along the eastern margin of Australia. This was based largely on seismic studies (e.g., Falvey and Mutter, 1981), and the existence of c. 97 Ma rhyolites recovered from deep sea drilling of the Lord Howe Rise (Figure 1; site 207, DSDP Leg 21; McDougall and van der Lingen, 1974; Tulloch et al., 2009). The amount of geophysical data has greatly increased over the last 15 years, permitting a more detailed regional evaluation of these offshore regions. Three major rift basin systems are indicated to contain Early Cretaceous rift fill offshore from eastern Australia: the Central and Western Rift Provinces (Stagg et al., 1999) of the Lord Howe Rise off SE Australia, and the Queensland and Townsville Basins off NE Australia (Figure 1).

### *Lord Howe Rise*

The Lord Howe Rise (LHR) is a continental ‘ribbon’ or microcontinent 1,600 km long and 400–500 km wide, extending southward from the eastern Coral Sea to the western margin of New Zealand (Figure 1; Willcox et al., 2001). It became separated from the SE margin of Australia following sea floor-spreading in the Tasman Basin (e.g., Gaina et al., 1998). It has been divided into several basins and blocks by Stagg et al. (1999) based on satellite gravity and limited seismic profiles. On the western side of the LHR is an extensive rift basin system 150–250 km wide (Central and Western Rift provinces; Figure 1) that in plate reconstructions of the LHR against eastern Australia, would have extended along the SE margin of Australia, and adjacent to the Gippsland Basin (Willcox et al., 2001). Several seismic megasequences have been recognised and correlated with known megasequences of basins on the SE Australian margin (Willcox and Sayers, 2001). Significantly, the seismic data indicate N-S-trending graben development along the western LHR during the latest Jurassic–Early Cretaceous. Infill is inferred to have been initially by volcanics and then syn-rift sediments that average 1.5–3 km in thickness, but reach a maximum of 4+ km thickness in the deepest basins. This phase of syn-rift sedimentation has been correlated with the Early Cretaceous Otway/Strzelecki Groups and volcanogenic sandstones of the Otway-Gippsland basins (Willcox and Sayers, 2001). Furthermore, volcanic bodies are interpreted to form significant parts of the rift fill in some basins (Willcox et al., 2001).

### *Queensland and Townsville Basins*

The Queensland and Townsville basins are two major rift basins within a large, submerged extensional continental terrane off the NE margin of Australia (Figure 1). Prior to Late Cretaceous–Cenozoic break-up, a continental mass of c. 700 km width existed adjacent to the margin (Draper et al., 1997). The Queensland and Townsville basins are part of a complex rift system that probably began to form in the Late Jurassic–Early Cretaceous, and now coincide with major bathymetric troughs where water depths vary from c. 1–2 km (Wellman et al., 1997). The Queensland Basin represents a northern strike continuation of the Whitsunday Volcanic Province. In contrast, the Townsville Basin is distinctive in striking E-W across the margin (Figure 1), and is thought to have formed through overall oblique NW-SE extension (Wellman et al., 1997). Control by NE-trending lineaments (Figure 3) on basin opening is also likely. These rift basins are c. 100 km wide, and on the basis of seismic and gravity data, contain rift valley sequences up to 3 km thick, of which 1 km is Cretaceous in age (Taylor and Falvey, 1977; Falvey and Mutter, 1981). Presently, there is no information available on the character of the Cretaceous rift fills to the basins, but are likely to be volcanic and volcanogenic sedimentary rocks as in the adjacent Whitsunday Volcanic Province and Great Australian Basin system, as well as most likely recording the erosional unroofing of the Whitsunday Silicic LIP after 100 Ma.

### *New Caledonia*

A recent U/Pb detrital zircon geochronology study of Mesozoic greywacke terranes in New Caledonia (Figure 1) has revealed the significant presence of Early Cretaceous age zircons derived from a contemporary silicic volcanic source (Adams et al., 2009: Cluzel et al., 2012). Statistically significant detrital zircon age populations include: 102 ± 4 Ma (33% of dated population); 119 ± 2 Ma (18% of dated population); 108 ± 2 Ma (57%) and 123 ± 4 Ma (18%). The youngest age populations are remarkably similar to the detrital zircon age spectra from the Winton Formation (Figure 6).

# Discussion

## Revised volume estimates

An extrusive volume of >1.5 $Mkm^3$ was estimated for Whitsunday Silicic LIP magmatism by Bryan et al. (1997). This was based on the preserved volumes of volcanogenic sediment in the GAB system (>1 $Mkm^3$), the Otway-Gippsland Basins (>0.4 $Mkm^3$) and the Whitsunday Volcanic Province (>0.1 $Mkm^3$). A variety of new data are now available permitting a reassessment of extrusive volumes from the Whitsunday Silicic LIP. Revised volume estimates are given in Table 1.

*Table 1 Revised volume estimates for the Whitsunday Silicic Large Igneous Province.*

| Tectonic Element | Preserved volume of volcanic material (km$^3$) |
|---|---|
| Whitsunday Volcanic Province[1] | >5.4 x $10^5$ |
| Southeast Australia | ≤$10^2$ |
| Otway-Gippsland Basins | >4 x $10^5$ |
| Great Australian Basin | >1 x $10^6$ |
| Townsville-Queensland Basins[2] | >1 x $10^5$ |
| Lord Howe Rise[3] | >4.5 x $10^5$ |
| **Total** | 2,490,100 |

1. An average thickness of 3 km is used to estimate preserved extrusive volumes for the >900 x 200 km silicic volcanic belt along the Queensland coast.
2. Volume is based on basin extents of 1000 x 100 km and Early Cretaceous rift fill of 1 km.
3. Volume is calculated for the c. 1500 x 150–200 km wide rift basin system (Central and Western Rift provinces of Stagg et al., 1999) that contains on average, 1.5–3 km thickness (2.25 to 6.75 x $10^5$ km$^3$) of Early Cretaceous rift fill (Willcox et al., 2001). A conservative thickness of 2 km is used in volume calculations.

Initial volume estimates for the Whitsunday Volcanic Province were based on an average thickness of 1 km, but further stratigraphic work indicates preserved thicknesses to be in the order of 3–4+ km (Bryan et al., 2000). The detrital zircon age data from the Winton Formation, indicating a significant pulse of rhyolitic ignimbrite volcanism from 105–100 Ma, and the general absence of rocks of this age in the province indicates a substantial missing section to the Whitsunday Silicic LIP. Of note is that Smart and Senior (1980) estimated that a volcanic belt of 3,000 x 130 x 2 km was required to account for the volume of detritus preserved in the GAB system alone. The major addition to the revised volume estimate is that of volcaniclastic rock preserved in the 1,500 x 150–200 km rift basin system of the Lord Howe Rise (Stagg et al., 1999; Willcox et al., 2001; Willcox and Sayers, 2001). A similar volume to that preserved along the SE Australian margin (Otway-Gippsland basins) is indicated. Importantly, the occurrence of Early Cretaceous volcanogenic sediment in New Caledonia (Adams et al., 2009) raises the possibility that the dispersal area of volcanogenic sediment and volume of volcanic material produced are probably greater still than the revised estimates presented here. In summary, minimum igneous volumes of the Whitsunday Silicic LIP, including volcanic and coeval volcanogenic sedimentary rock, preserved along the eastern Australian margin are estimated to be >2.5 Mkm$^3$.

## Importance of the sedimentary record

The Whitsunday Silicic LIP is unusual in that a large proportion of its products are preserved as huge volumes (>1.4 Mkm$^3$) of coeval volcanogenic sediment in adjacent continental sedimentary basins (Figure 3; Bryan et al., 1997). The additional preservation of >0.55 M km$^3$ of volcaniclastic material is now indicated for offshore rift basins (Figure 4). Such substantial volumes of coeval volcanogenic sediment are not characteristic of other LIPs and volcanic rifted margins (cf. Menzies et al., 2002). Age dating of volcanic mineral grains (e.g., Gleadow and Duddy, 1980; Fig 6), the fresh nature of the detrital mineral grains and the sheer volume of volcanogenic sediment in the Otway-Gippsland and GAB systems preclude any arguments for an igneous basement-derived source for the sediment. An important issue then is how such huge volumes of predominantly sand-grade volcanogenic sediment were rapidly generated and transported over large distances (>500 km) and low-relief gradients with limited weathering (Smart and Senior, 1980) and textural maturation to fundamentally alter the basin fill history and reservoir potential of several widely separate and tectonically unrelated sedimentary basin systems.

In a comparison of volcanic phenocryst compositions from the Whitsunday Volcanic Province with detrital mineral grains from the coeval volcanogenic sedimentary formations of the Great Australian and Otway-Gippsland basin systems, Bryan et al. (1997) demonstrated that it was phenocryst compositions from the volumetrically dominant dacitic–silicic ignimbrites that matched the detrital volcanic mineral grain compositions. This overlap confirmed that silicic pyroclastic volcanism was the major expression of Whitsunday Silicic LIP magmatism in eastern Australia and sediment source. Furthermore, the overlap implied there was a remarkable consistency in mineral composition, and consequently, whole rock chemistry for Whitsunday Silicic LIP magmatism, particularly considering that volcanism and sedimentation were occurring over a distance of >2,500 km along the eastern Australian margin.

For the Whitsunday Silicic LIP, there were two important reasons why there was such an important sedimentary record of magmatism. The first was the fortuitous circumstance that two major pre-existing continental sedimentary basins were adjacent to the province, and able to accommodate huge volumes of volcanogenic sediment. A marine transgression during the Aptian–Albian further facilitated sediment accumulation in the GAB system. However, an increased flux of volcanogenic sediment in the late Albian helped drive a basin-wide regression as recorded by the Winton Formation (e.g., Gallagher and Lambeck, 1989). The intersection of the newly developing N-S volcanic rift system along eastern Australia with the E-W Antarctic-Australian rift to the E of the Gippsland Basin, allowed the shedding of volcanic material towards the W into the Otway-Gippsland basin system that was entering a sag phase of basin development. The second reason was that the pyroclastic mode of fragmentation and dispersal was a critical factor in producing large volumes of sand-grade sediment to be rapidly delivered into the continental basin systems. Large volumes of easily erodible, nonwelded pyroclastic debris were likely present along the flanks of the province that could be remobilised by efficient W-draining fluvial systems. The hydrology of drainage basins are substantially affected by explosive volcanism as opportunities for sediment erosion and transport are greatly increased resulting in prolonged and extreme sediment yields in rivers (e.g., Gran and Montgomery, 2005). In contrast, only minor volumes of coarse fragmentary material (e.g., a' a' lava breccias, scoria deposits) tend to be produced during subaerial flood basalt eruptions in continental flood basalt provinces. The quick burial of brecciated material by succeeding lava flows combined with the generally coherent (unfragmented) nature of the flood basalt lavas are important factors for why continental flood basalt provinces have a limited sedimentary response.

## Volcanic and compositional trends

Volcanic stratigraphic studies of the Whitsunday Volcanic Province (Bryan et al., 2000) highlight some important longer term trends in volcanism and erupted compositions. Although a range of igneous compositions are present (Figure 4), stratigraphic constraints indicate two important temporal compositional trends:

1) volcanism first evolving towards more bimodal compositions with time; and
2) the development of a late-stage, low-volume, largely intrusive phase dominated by intermediate (andesitic) magma compositions.

The compositional trend towards bimodalism is a more common characteristic of magmatism associated with continental rifting. Early phases of volcanism (c. 130–115 Ma) were dominantly explosive, erupting dacitic to rhyolitic magmas, whereas volcanism during later stages (<115 Ma) was both effusive and explosive, when mainly basaltic and rhyolitic lavas were erupted contemporaneously with predominantly rhyolitic to high-silica rhyolite ignimbrites (Bryan et al., 2000). Mafic magma compositions became less contaminated with time, with intrusion of primitive E-MORB gabbros occurring later in the igneous history. More crystal-rich and quartz-, biotite- and hornblende-bearing rhyolitic magmas are also more common in the upper parts of the stratigraphy. Further indication of this trend toward more mafic and alkaline compositions is evident from the c. 110–90 Ma igneous rocks along the SE Australian margin, and then the eruption of intraplate alkali volcanics beginning c. 80 Ma.

However, field relationships further indicate that at least within the province, bimodal volcanism was post-dated by a largely intrusive phase represented by more intermediate magma compositions. Andesitic compositions across the province are largely restricted to dykes (Figure 4) and represent an insignificant proportion of the total magma volume emplaced. Field relationships constrain much of the andesitic dyking to post-date the early silicic and later bimodal volcanic phases, and thus indicate a final phase of low-volume andesitic dyking focussed on the central part of the province. This final phase is interpreted to reflect an increased rate and focussing of rifting, which enhanced the disruption of magma systems such that magma mixing and emplacement of hybridised magmas was promoted by syn-volcanic faulting (e.g., Johnson and Grunder, 2000; Bryan et al., 2011).

## Summary

The Whitsunday Silicic LIP formed a linear silicic-dominated igneous belt several hundred kilometres inboard of the continental margin and thus resembles the Jurassic Ferrar LIP along strike to the S in terms of province geometry and relative position to the continental margin. LIP magmatism (c. 35 Myrs) and continental rifting (15–30 Myrs) was protracted before complete rupture of the continent and onset of sea-floor spreading at 84 Ma. Significant records of this Silicic LIP and rifting event are preserved in onshore and offshore sedimentary basins and complement the preserved onshore igneous components. Integrating the onshore and offshore records of volcanic rifted margins are vital to fully understand rift architecture, timing and evolution, the duration and magnitude of igneous activity, their potential environmental impact and for evaluating their resource potential.

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*by Alex. G. Cook*

# Cretaceous faunas and events, northern Eromanga Basin, Queensland

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*The stratigraphy, sedimentary history and paleontology of the northern Eromanga Basin are reviewed in the light of extensive field effort, searching for Cretaceous vertebrate fossils, in particular dinosaurs. Prolonged non-marine deposition throughout the Jurassic was followed by Lower Cretaceous marine incursions which extended to the late Albian. Whilst biostratigraphy is underpinned by microfloral assemblages there are three distinct marine faunas preserved from the late Aptian, early middle Albian and late Albian. Effective regression caused by sediment oversupply in the latest Albian heralded the final phase of non-marine deposition in the Eromanga Basin which continued into the Cenomanian. A distinct floral assemblage is accompanied by a modest fossil vertebrate assemblage.*

## Introduction

The Eromanga Basin is a large, continental basin which, with the interconnected Surat Basin in the southeast, and Carpentaria Basin to the north, forms the Great Australian Basin (GAB) that further connects with the Laura Basin in the northeast and the Clarence-Moreton, Nambour and Maryborough basins in the southeast to form the Great Australian Superbasin and dominate the continental geology of northeastern Australia (Figure 1). Each basin of the GAB contains a near flat-lying Jurassic to Cretaceous sequence which records basin subsidence and the varying roles of sea-level change and sediment supply throughout the GAB's history. Geological summaries are available for the Carpentaria Basin (McConachie et al., 1997), the Surat Basin (Green 1997), the southern and central Eromanga Basin (Gray et al. 2002) and the northern Eromanga Basin (Green et al. 1992).

The Eromanga Basin extends from South Australia and New South Wales into northern Queensland. It is separated from the Carpentaria Basin by the Euroka Arch, a basement high with structural drape, and by a similar feature, the Nebine Ridge, from the Surat Basin. The northern Eromanga Basin overlies the late Paleozoic Galilee Basin, and Paleozoic basement, and onlaps Proterozoic provinces to the north and northwest.

Lithostratigraphy of the Eromanga Basin has been elucidated by regional mapping and by water and petroleum exploration wells. In the north of the basin, these wells are less common, but have been supplemented by government sponsored stratigraphic wells.

The stratigraphy for the northern Eromanga is summarised in Figure 2. Identified depositional phases include Early Jurassic dominantly coarse-grained fluvial deposition, followed by more subdued Mid-Upper Jurassic–Lower Cretaceous fluvio-lacustrine sedimentation. In the Lower Cretaceous a succession of marine transgressions and regressions resulted in a series of shallow marine, varyingly fossiliferous deposits, the youngest of which is late Albian. Concomitant volcanism to the east which had been active since the Jurassic was more prevalent in the Lower Cretaceous, and was dominated by the Whitsunday Island Volcanic Province (Bryan et

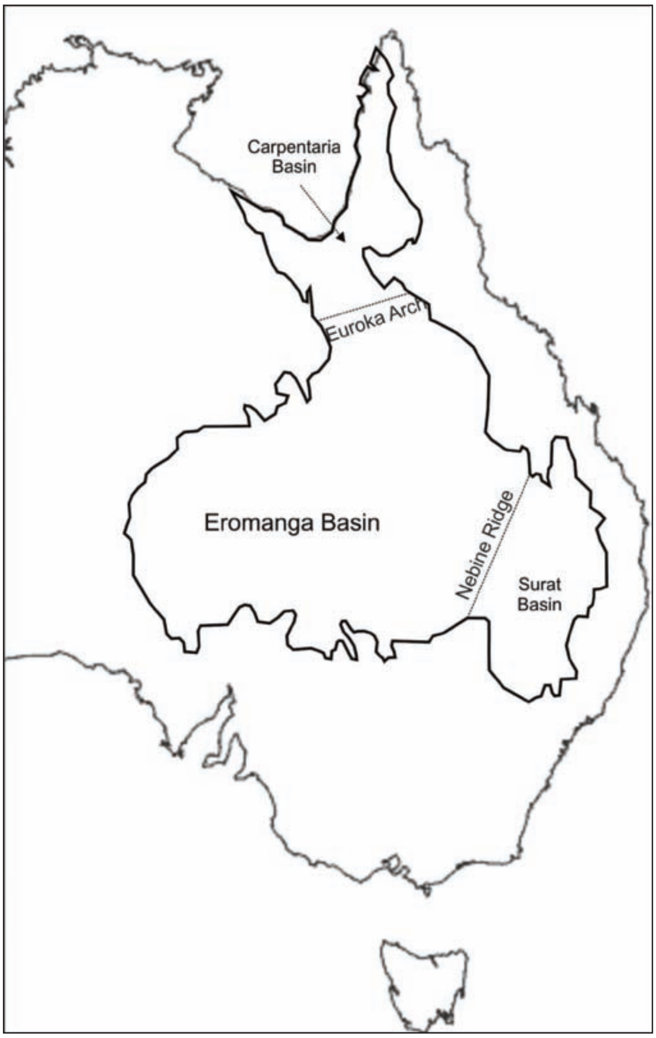


***Figure 1 The Great Australian Basin in northeastern Australia.***

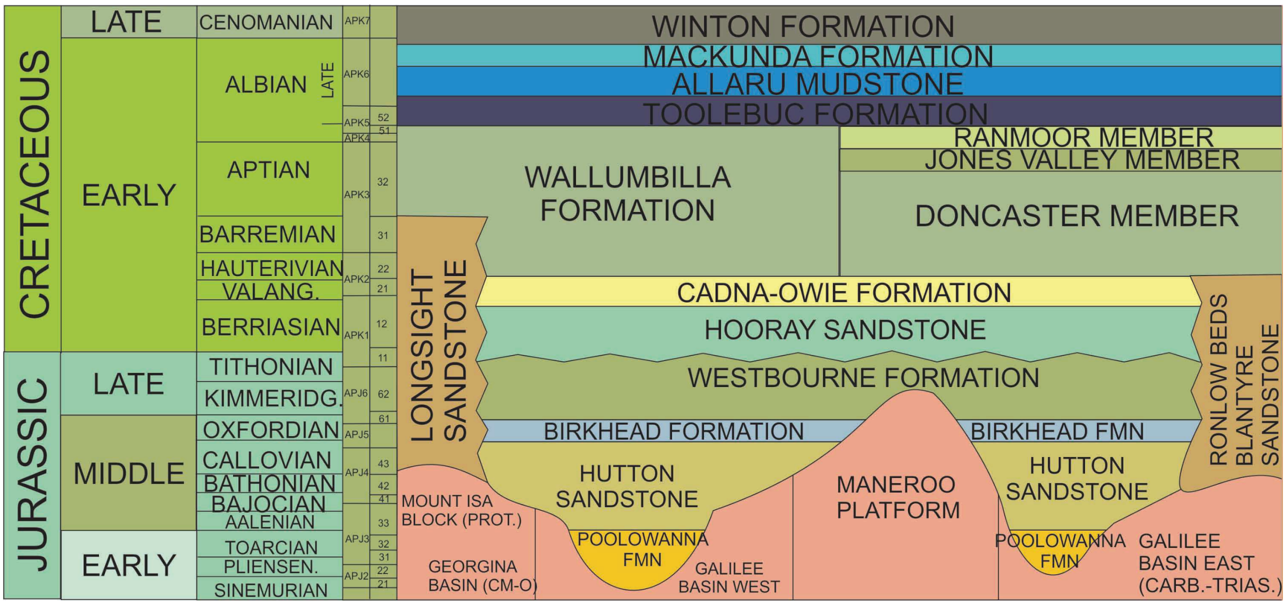


***Figure 2 Stratigraphy of the northern Eromanga Basin modified after Green et al. (1992) and Gray et al. (2002).***

al., 2000), situated near the east coast. Volcanogenic sediment oversupply peaked in the late Albian to Cenomanian, to the extent that it caused effective regression with nonmarine deposition. The globally recognised Cenomanian sea level rise did not manifest within the Great Australian Basin. Non-marine sedimentation ceased in the Upper Cretaceous, and was followed by three cycles of deep weathering and erosion which extended until at least the Miocene, with accompanying tectonic flexure along pre-existing structures.

## Principal stratigraphic divisions

Jurassic stratigraphy in the northern Eromanga Basin comprises a basal Jurassic coarse fluviatile unit, now referred to the Poolowanna Formation. This is overlain by the extensive coarse-grained fluvial, Hutton Sandstone and subsequent Birkhead Formation which was deposited in fluvio-lacustrine settings with strong supply of volcanogenic detritus from the east. The Adori Sandstone overlies this and represents braided fluvial systems. The fluviolacustrine Westbourne Formation, and overlying fluvial Hooray Sandstone are overlain by the Cadna-owie Formation which represents transition to marginal marine conditions during Valanginian–Barremian transgression. Rapid continued transgression in the Aptian resulted in deposition of the Doncaster Member of the Wallumbilla Formation, a widespread unit which represents maximum flooding of the continent. Effective but subdued regression-transgression in the early middle Albian is represented by deposition of the Ranmoor Member of the Wallumbilla Formation. Transgression renewed rapidly in the late Albian with deposition of the Toolebuc Formation, representing maximum flooding within a partly anoxic restricted basin, resulting in a mixed limestone-oil shale unit which is isochronous across the basin. The unit correlates with the global Oceanic anoxic event (OAE) 1c. Renewal of sediment supply from the east prompted deposition of the Allaru Mudstone and overlying Mackunda Formation which are late Albian regressive units. Sediment supply exceeded subsidence choking the basin during this time, and this continued into the Cenomanian, which is represented by an entirely non-marine sedimentary package.

Biostratigraphy within the basin is primarily controlled by rich palynofloras (Dettmann and Playford 1969; Price 1997), but also by ammonites, foraminifera and locally recognised molluscan assemblages (Day, 1969).

Palynomoprh stratigraphy developed for continental Australia has been summarised for the southern Eromanga and Surat basins by Price (1997), and Draper (2002). Marine and marginal marine units span Australian Palynological units APK2.1 to APK6, with the upper nonmarine Winton Formation belonging within APK7 (Figure 2). Dinocyst zones of Helby et al. (1987) indicate the Cadna-owie Formation extending to include an unnamed zone, and the basal *O. operculata* zone, with the youngest marine units in the *P. ludbrookiae* zone (see Branagan (2012) for more on Nell Ludbrook). Price (1997) replaced zones with dinocyst units, the upper Cadna-owie in ADK17, and the Mackunda Formation within ADK22. Foraminifera have been evaluated in some units (Haig, 1979; Haig and Lynch, 1993) and some investigations of nannofossils have been undertaken by Shafik (1985). The relation of palynology to local and global cyclicity has been addressed by Burger (1986), who could not identify all Jurassic global cycles within the Eromanga and Surat basins.

## Fossil Assemblages

Three principal marine fossil assemblages and an upper non-marine fossil faunal assemblage within the northern Eromanga Basin correspond to the late Aptian, early middle Albian, late Albian and late Albian to Cenomanian, respectively.

### Late Aptian assemblage

The Late Aptian marine incursion that brought maximum flooding of the continent, is represented by the Doncaster Member of the Wallumbilla Formation across the GAB, with equivalents in the Laura Basin. The diverse and distinctive fauna of the Doncaster Member that was locally named the Roma Fauna (Whitehouse, 1926) contains ammonites (Etheridge, 1880, 1892, 1909; Whitehouse, 1926; Day, 1974), predominantly heteromorphs including *Tropaeum*,


***Figure 3 Typical Late Aptian assemblage of macrofossils from the Doncaster Member, Wallumbilla Formation. (A).*** **Tropaeum undatum; *(B).* Australiceras irregulare; *(C).* Tonohamites taylori; *(D).* Aconeceras *sp.; (E).* Sanmartinoceras olene; *(F). A*ustraliceras jacki; *(G).* Fissilunula clarkei; *(H).* Eyrena palmerensis; *(I).* Maccoyella barkleyi; *(J).* Pseudavicula anomala; *(K).* Pinna *sp. (Scale bar = 2cm).***

*Australiceras, Tonohamites, Toxoceratoides, Sanmartinoceras* and *Aconeceras*. The more than 30 species of bivalves include locally stratigraphically important *Maccoyella barklyi, Fissilunula clarkei, Pseudavicula anomala* and *Eyrena palmerensis*. In addition *Panopaea*, *Pinna*, scaphopods, the crinoid *Isocrinus australis* and gastropods are locally abundant. *Purisiphonia clarkei* is the only known sponge (Jell et al., 2011). *Kronosaurus queenslandicus*, undescribed plesiosaurs and rare ichthyosaurs are known from the unit. In the northern Eromanga Basin algal build-ups resembling stromatolites form elongate ridges of tens of metres scale, or isolated domal, conical and lamellar forms. Commonly, build-ups are found around large marine reptile specimens. This is surprising given the high latitude position of the basin during the late Aptian, glendonites and dropstones elsewhere in the basin during this time.

## Early Middle Albian assemblage

The macrofaunas of the Ranmoor Member are depauperate, reflecting the its subdued outcrop patterns. Fauna recovered from the Ranmoor Member includes the ammonites *Beudanticeras flindersi*, *Falciferella* sp. and several undescribed species. Bivalves such as *Aucellina* and *Inoceramus* are in smaller numbers than in the overlying Toolebuc Formation.

## Late Albian assemblages (Figure 4)

The Toolebuc Formation, Allaru Mudstone and overlying Mackunda Formation have similar, but not identical faunas, which include vertebrates, ammonites, bivalves and other molluscs. The Toolebuc Formation is dominated by a restricted benthic assemblage of *Inoceramus sutherlandi* and *Aucellina hughendensis*, but a diverse nektic assemblage of fishes, marine reptiles and molluscs, including large ichthyosaur skeletons (Wade, 1990) of *Platypterigius australis*, the pliosaur *Kronosaurus queenslandicus*, elasmosaurs and at least one species of polycotylid, three turtles, being the giant *Cratochelone*, known from a single specimen, the common *Notochelone* and *Bouliachelys* Kear and Lee (2006). Abundant fish fossils include amiids, aspidorhynchids, chimaeroids, lamnids, dipnoans, (Lees, 1986, 1990, Lees and Bartholomai, 1987, Bartholomai, 1969, 2004, 2008). Ammonites include *Beudanticeras flindersi*, *Myloceras* spp., *Labeceras* spp., *Idanoceras*, and possibly *Worthoceras* with the nautiloid *Eutrephoceras hendersoni* also common. The teuthids *Trachyteuthis willisi*, *Meunsterella tonii* and *Boreopeltis soniae* were described from the Toolebuc Formation and overlying basal Allaru Mudstone (Wade, 1993). Abundant belemnites have received scant


***Figure 4 Typical Late Albian assemblage of macrofossils from the northern Eromanga Basin. (A)* Beudanticeras flindersi*; (B)* Goodhallites goodhalli*;* (C) Eutrephoceras hendersoni*; (D)* Myloceras ammonoides*; (E)* Myloceras intermedium*; (F)* Myloceras auritulum*; (G)* Labeceras compressum*; (H)* Homolopsis etheridgei *(Scale bar = 2cm).***

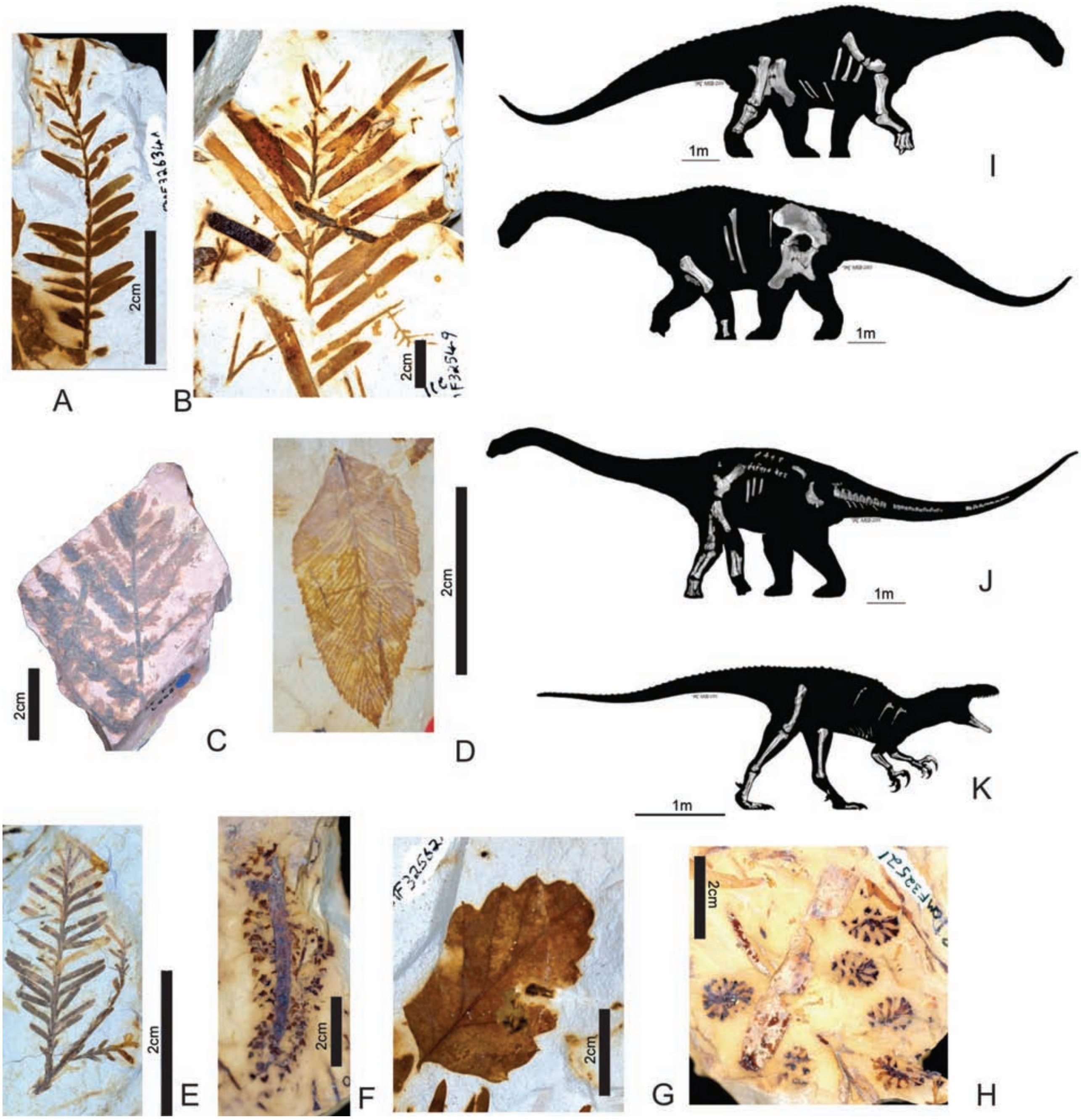


*Figure 5 Winton Formation floral assemblage and dinosaurs. (A)* **Araucaria** *cf.* **A. mesozoica;** *(B)* **Araucaria** *cf.* **A. mesozoica;** *(C)* **Cladophlebis** *sp.; (D)* **Phyllopteroides mcclymontae;** *(E) Conifer foliage; (F) Male sporangiate cone; (G) Angiosperm leaf; (H) Small cones cf.* **Austrosequioa wintonesis;** *(I)* **Diamantinasaurus matildae;** *(J)* **Australovenator wintonensis;** *(K)* **Wintonotitan wattsi** *(Dinosaur silhouettes after Hocknull et al. (2009)).*

attention. Terrestrial and associated faunal elements have also been recovered from the Toolebuc Formation including the dinosaurs *Minmi* sp., and *Muttaburrasaurus langdoni*, a small number of pterosaur elements, the bird *Nanatius eos* and sauropod bones.

The Allaru Mudstone has yielded common ammonites, *Beudanticeras*, *Myloceras* spp. and *Labeceras* spp. with lesser *Idanoceras* (Henderson and McKenzie, 2002), and *Goodhallites goodhalli* (Henderson and Kennedy, 2002), and a small fauna of undescribed micromorphs. Bivalve diversity is still low in the unit with *Inoceramus* dominant, alongside *Aucellina*. Crustacea are locally abundant. Rare ophiuroids and monasterid seastars are known. A single faviid coral recovered from the unit (Jell et al., 2011) attests to relatively warm conditions during deposition even at mid-high latitudes. There is some disparity between faunas in the lower parts of the unit compared to the upper, supported by seismic evidence of a break in the lower Allaru Mudstone, but this faunal disparity has yet to be investigated fully.

The Mackunda Formation is the uppermost marine unit and has a diverse benthic and nektic fauna. Benthic elements include bivalves with the ubiquitous *Inoceramus* and *Aucellina* accompanying *Maccoyella rookwoodensis*, *Panopaea*, *Acesta*, *Tatella*, and solemyiids among other taxa. Belemnites are abundant, but ammonites are much less common. Small naticiid and higher spired gastropods are common in some of the shell beds which are common in the formation. Teleosts are less common, but shark's teeth are abundant.

## Winton terrestrial assemblage

The Winton Formation contains a diverse and abundant fossil macroflora (Peters and Christophel, 1978; Dettmann et al., 1992, 2009; McLoughlin et al., 1995; Pole and Douglas, 1999; Pole, 2000a, b; McLoughlin et al., 2010; Dettmann and Clifford, 2000; Clifford and Dettmann, 2005) including conifers, ferns, cycadophytes, angiosperms and ginkgos. Undescribed insects are known in collections. A modest dinosaur fauna (Figure 5); Coombs and Molnar, 1981; Hocknull et al., 2009) includes the genera *Australovenator*, *Wintonotitan*, and *Diamantinasaurus*. *Australovenator* is a medium-sized allosauroid, neovenatorid taxon, whilst *Diamantinasaurus* is a lithostotian titanosaurian and *Wintonotitan* a titanisauriform. More discoveries of dinosaurian material in 2010 and 2011 extend the diversity. Other elements include crocodylians (Salisbury et al., 2006), fishes, lungfish (Kemp, 1991), a dolichosaur, turtles, and pterosaur remains. Of outstanding importance is the dinosaurian ichnofauna which is exemplified by the Dinosaur Stampede National Monument, Lark Quarry (Thulborn and Wade, 1979, 1984). The dinosaurian and marine reptile finds support an economically important tourism industry in the northern Eromanga Basin.

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# Plio-Pleistocene Climate and Faunal Change in Central Eastern Australia

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*Understanding the responses of Plio-Pleistocene terrestrial vertebrates to long-term trends in climate change in central eastern Australia has advanced considerably in recent years following the recovery and documentation of a series of remarkable fossil assemblages. The middle Pliocene Chinchilla Local Fauna of SE Queensland preserves a diverse suite of vertebrate taxa suggestive of a paleoenvironment consisting of wetlands, closed wet forest, open woodlands, and grasslands. Local extinctions of numerous arboreal and terrestrial woodland species suggest that significant faunal and habitat reorganization occurred between the Pliocene and Pleistocene, in part, reflecting the expansion of open woodlands and grasslands. Middle Pleistocene deposits in the Mt Etna region of central eastern Queensland contain extensive fossil assemblages of rainforest-adapted vertebrates dated >500–280 ka. Such faunal assemblages show remarkable long-term stability despite being subjected to numerous glacial-interglacial climatic shifts. However, sometime between 280–205 ka, a major faunal turnover/extinction event occurred, where the previously dominant rainforest-adapted faunas gave way to xeric-adapted forms. Independent paleoclimatic records suggest that this shift was a result of increased climatic variability and weakened northern monsoons. Late Pleistocene deposits of the Darling Downs, SE Queensland, provide an important temporal extension to the Mt Etna region. Recent studies have demonstrated minimally, a three stage extinction of local megafauna (giant land mammals, birds and lizards). Associated radiometric and optical dating indicates that the progressive loss of megafauna from the region was initiated at least 75 kyr before the continental colonisation of humans. The progressive changes in megafaunal community dynamics were most likely driven by intense climatic changes (i.e., increased aridity) associated with the last glacial cycle. The potential role of humans in the final extinctions (post-human colonisation) is unclear. However, if humans did have a detrimental impact on the last surviving megafauna, it is likely that they simply compounded upon longer-term climate-driven processes. Independent paleoclimate information suggests that Plio-Pleistocene climates were complex beyond glacial-interglacial cyclicity, and hence, faunal responses were similarly complex.*

## Introduction

Queensland, Australia, contains numerous, rich and diverse fossil deposits that chart the evolution and emergence of the continent's vertebrate faunas, from the first Carboniferous tetrapods (Parker et al., 2005), through to dinosaurs (Hocknull et al., 2009), and the appearance of the earliest marsupials (Beck et al., 2008). Arguably one of the most significant fossil sites, the Riversleigh World Heritage fossil area of NW Queensland is renowned for its abundance of deposits (>200) that document the evolution of the modern marsupial-dominated vertebrate faunas through the mid-late Cenozoic (Archer et al., 2006). Over 200 mammalian species have so far been identified including dasyurids, bandicoots, marsupial 'moles', possums, koalas, wombats, and kangaroos, as well as now-extinct groups such as diprotodontids and marsupial 'lions'. The older late Oligocene and early-mid Miocene fossil assemblages are typically represented by open woodland, through to wet forest and/or rainforest-adapted taxa, whereas the younger Plio-Pleistocene deposits contain more open, arid-adapted forms suggesting continental aridification during and after the late Miocene (Travouillon et al. 2009).

Significantly, Riversleigh is dominated by Miocene deposits of varying ages; the younger Pliocene and Pleistocene fossil record of Riversleigh is thus far known from only two sites, neither of which have been extensively documented (Travouillon et al., 2006). Thus, a firm understanding of faunal responses to more recent episodes of climate change, such as the establishment of glacial-interglacial cyclicity in the latest Cenozoic, remains unclear. Conversely, southern and central eastern Queensland contains numerous vertebrate fossil assemblages that span this critical period. Thus, it is possible to extend the important record from Riversleigh in order to track the emergence of the modern biota in response to Quaternary climatic regimes as well as long-term trends in the aridification of the Australian continent.

Three key sites, the middle Pliocene Chinchilla Local Fauna, middle Pleistocene Mt Etna region, and late Pleistocene Darling Downs faunas (Figure 1), have attracted significant attention in recent years. Each region contains numerous vertebrate faunal assemblages that have been the focus of intensive paleoecological and/or dating

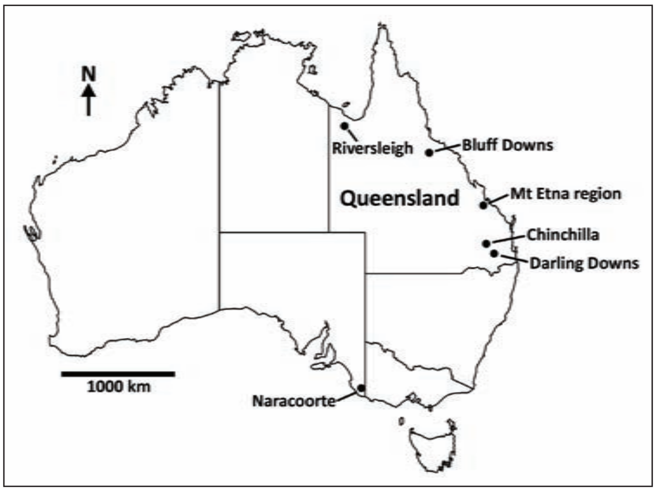


***Figure 1 Map of Australia showing major study sites and other mentioned localities.***

studies (e.g., Hutchinson and Mackness, 2002; Hocknull, 2005; Price and Webb, 2006; Hocknull et al. 2007; Price et al., 2009a, 2011).

## Middle Pliocene Chinchilla Local Fauna

The Chinchilla Local Fauna, SE Queensland, is derived from the laterally extensive Chinchilla Sand; an extensive sequence of clays, weakly consolidated grey to yellow sands, and ferruginised heterogeneous conglomerates exposed along the Condamine River (Woods, 1960). Strata are dominantly fluviatile in origin, exposed in thin surficial beds and deeply eroded sections (Mackness et al., 2010). The Chinchilla Local Fauna has not been directly dated analytically, however, on the basis of 'stage-of-evolution' and biocorrelation, it may be younger than the early Pliocene Bluff Downs Local Fauna of NE Queensland (Figure 1), but similar in age to the middle Pliocene (c. 3.5 Ma) Kanunka Local Fauna of central Australia (Tedford et al., 1992; Mackness et al. 2000).

Although the Chinchilla Local Fauna is particularly rich, containing a diverse range of taxa including molluscs (gastropods and bivalves), fish, lungfish, turtles, lizards, snakes, crocodiles, and birds (Hutchinson and Mackness, 2002), the mammals have been most intensively studied. The mammals are dominated, in terms of abundance, by large-bodied (i.e., >40 kg) semi-grazing/browsing kangaroos (such species of the extinct *Protemnodon*), and the grazing *Macropus* (extant, diverse and widespread across modern Australia). The abundance of grazing forms represents a major shift in dietary niche partitioning of Pliocene faunas from those of older assemblages (e.g., Riversleigh), which are mostly dominated by browsing forms.

Although the preponderance of grazing taxa suggests a dominatingly open paleohabitat, large-bodied browsing kangaroos such as "*Simosthenurus*", *Sthenurus* (both short-faced kangaroos of the Sthenurinae) and *Troposodon* (Lagostrophinae), and diprotodontoids (*Palorchestes parvus*) have also been recorded from the Chinchilla Sands (Prideaux, 2004). Medium-sized (10–40 kg) forest wallabies (*Silvaroo*), arboreal tree kangaroos (*Bohra*) and koalas (*Phascolarctos*) are also known (Dawson, 2004a, b; Price, 2008a; Price et al., 2009a). Today, tree kangaroos are restricted only to NE Australian and New Guinean tropical rainforests.

Taken together, the paleoenvironmental signal from the Chinchilla fossil vertebrates indicates a significant wetland component, as well as a mosaic of canopies of mature wet forest, possible open woodland, and open grassland (Hutchinson and Mackness, 2002; Price et al., 2009a). However, there remain some limitations on the accuracy of such an interpretation. Although the Chinchilla Local Fauna appears to exhibit exceptional taxic diversity, it is possible that numerous individual, temporally-disparate fossil assemblages have been mixed following collection, such that the 'Local Fauna' is actually significantly time- and habitat-averaged. If that interpretation is correct, existing collections represent at least some temporally disparate forms. Such a hypothesis is presently difficult to test as there is poor stratigraphic control for fossils collected from the region.

## Middle Pleistocene Mt Etna region

The Mt Etna fossil faunas, central eastern Queensland, represent a key temporal extension to not only the Pliocene Chinchilla Local Fauna, but also the older Riversleigh rainforest vertebrate fossil record. Most of the Pleistocene vertebrate fossils collected from the region (Mt Etna and the adjacent Limestone Ridge) represent a series of brecciated cave deposits that formed within Devonian limestones (Figure 2). Bone-yielding sediments are typically well-lithified and contain numerous interlayered datable speleothems (e.g., flowstones, stalagmites, straw stalactites) (Hocknull et al., 2007).

On the basis of biocorrelation, Hocknull (2005) originally considered the fossil faunas to be mostly early Pliocene, with some late Pleistocene and Holocene deposits. However, application of uranium/thorium (U/Th) dating of the speleothems demonstrated that the majority of deposits are in fact middle Pleistocene (Hocknull et al., 2007). Older fossil assemblages may be present in the region as several sites were dated to >500 ka, the chronological application limit of U/Th dating.

The >500–280 ka Mt Etna region fossil faunal assemblages contain a remarkable suite of rainforest-adapted taxa (Hocknull et al., 2007). This is exceptional because they represent the only unequivocal record of Pleistocene rainforest vertebrates on the continent. Moreover, the Mt Etna deposits extend the temporal range of many taxa previously considered to have been pre-Quaternary, into the middle Pleistocene (Hocknull, 2005; Hocknull et al., 2007). Similarly, the deposits are significant in filling in gaps in the fossil record for many extant lineages (Hocknull, 2005).

In terms of diversity, the Mt Etna region assemblages are dominated by small-bodied (<10 kg) vertebrates, many belonging to forms that are today restricted solely to rainforest environments (cannibal frogs, big-eyed tree frogs, forest dragons, striped possums, cuscuses, tree kangaroos, and white-tailed rodents). The diversity of arboreal mammals is high, with at least four species of tree kangaroo (*Dendrolagus* spp.), three species of greater gliders (*Petauroides* spp.), six species of rainforest ringtail possums (*Pseudochirops* spp. and *Pseudochirulus* spp.), and two tree rodents (*Uromys* sp.); each genus is found today either in northern Queensland and/or New Guinean rainforests. Most recently, a new genus and species of koala (*Invictokoala monticola*) was described from a 320 ka rainforest deposit at Mt Etna (Price and Hocknull, 2011). Identification of a rainforest form is significant because it was previously considered that koalas suffered extinction from such habitats by the late Miocene (Archer and Hand, 1987). Persistence of a vast range of rainforest-adapted vertebrates from >500–280 ka suggests relative faunal stability through several of the earlier interglacial-glacial cycles that characterised the Quaternary.

*Figure 2 Mt Etna illustrating middle Pleistocene fossil cave breccias (orange-brown sediments), exposed in cross-section as a result of mining activities (September 2008).*

Significantly, sometime between 280–205 ka, a major local faunal change occurred in the Mt Etna region where the formally dominant rainforest faunas were replaced by more xeric-adapted forms (e.g., grassland dragons, bilbies, pig-footed bandicoots). The change reflected the loss of 80% of small to medium-sized mammalian taxa, as well as numerous frog and lizard species. Similarly, approximately 75% of taxa identified in 205–170 ka deposits have not been recorded from older deposits in the region (Hocknull et al., 2007).

The change in rainforest-adapted, to xeric-adapted, faunas implies a significant reduction in precipitation after 280 ka, relative to previous interglacials. Similarly, U/Th dating of speleothems indicates that most grew during interglacials or interstadials, and that nearly 70% of all dated samples are older than 300 ka (Hocknull et al., 2007). Those data imply that conditions for speleothem genesis (i.e., a wetter environment) were more prevalent prior to 300 ka than after. Other independent paleoclimate data indicate a retraction of the northern monsoon and increased climatic variability after 300 ka (Hocknull et al., 2007).

Another local faunal change occurred in the Mt Etna region sometime after 205–170 ka, reflecting the loss of the xeric-adapted faunas, such that by the Holocene (c. 7 ka), a more mesic-adapted fauna occurred in the region (similar to today) (Hocknull et al., 2007). This youngest episode of faunal turnover saw an overall decline in species richness with the loss of 65% of small to medium-sized mammals (Hocknull et al., 2007). Approximately, 38% of mammals in the post-205–170 ka deposits are not recorded from older deposits in the region (Hocknull et al., 2007). Unfortunately, the late Pleistocene faunal record remains poorly dated. Thus, it is not possible to reliably examine the response of local faunas to climate change associated with the last glacial cycle.

Interestingly, the Naracoorte Caves World Heritage Area of SE South Australia provides a contrasting record of faunal history to that of the Mt Etna region. At Naracoorte, Prideaux et al. (2007) reported a series of middle–late Pleistocene (c. 500–50 ka) faunas from Cathedral Cave. The data suggest that there was period of marked stability in faunal composition through that time. On the basis of that evidence, Prideaux et al. (2007) argued that climate change prior to human arrival (c. 50–45 ka) was unlikely to be a mitigating factor in the extinction of late Pleistocene megafauna (large bodied terrestrial lizards, birds and mammals). In marked contrast, all megafauna recorded from the middle Pleistocene Mt Etna (marsupial 'lions', diprotodontoids, large-bodied kangaroos) have not been recorded in younger deposits within the region. Thus, the results cannot preclude climate change, in particular, enhanced aridity through the Pleistocene, as a factor involved in the final megafaunal extinctions, at least in northern Australia (Hocknull et al., 2007).

## Late Pleistocene Darling Downs

The eastern Darling Downs, SE Queensland, represents the most extensive deposits of Pleistocene-aged vertebrates on the continent. The deposits are predominately fluviatile in origin, with the sediments derived from erosion of underlying Mesozoic sandstones and mid Cenozoic basalts (Figure 3). Although over 50 distinct fossil sites have been identified in the region (Molnar and Kurz, 1997), the true number of deposits would likely exceed 200, for every single catchment within the region yields a remarkable number of fossils in abundance. Indeed, the Darling Downs may contain the continent's youngest records of articulated skeletons of now-extinct megafauna (Roberts et al., 2001).

Recent research in the region has focused on the rich Kings Creek catchment deposits of the southern Darling Downs. The Pleistocene Kings Creek represented a series of permanent and ephemeral streams within a geographically small catchment (Price, 2005). The geometry of the paleocatchment precludes long-distance transport of the fossils, thus, resulting paleoecological models should reflect only a local signature. The megafauna-bearing deposits typically represent a series of stacked highly fossiliferous sequences. The deposits reflect both high-velocity lateral accretion (channel) and low-velocity vertical accretion (overbank deposits) (Price and Sobbe, 2005; Price and Webb, 2006).

Vertebrate and invertebrate fossils are more common in the coarser grained high energy deposits than the low velocity vertically accreted sediments. The assemblages are exceptionally diverse, including freshwater and terrestrial gastropods, bivalves, fish, frogs, lizards (skinks, agamids and varanids), snakes, turtles, birds, and small and large mammals (Price, 2002, 2005, 2008b; Price and Hocknull, 2005; Price and Sobbe, 2005, 2011; Price et al., 2005, 2011; Price and Webb, 2006). The megafaunal taxa (e.g., Figure 4) include several forms descended from Pliocene ancestors as represented at Chinchilla (e.g., several species of the grazing *Macropus*) (Bartholomai, 1975). In most, if not all cases, the derived forms are significantly larger than their Pliocene counterparts, such as the kummerspeck *Diprotodon*, a 2,700 kg descendent of the more diminutive Pliocene *Euryzygoma*. The tendency for faunas to trend towards gigantism

*Figure 3 Late Pleistocene megafaunal fossil locality at Neds Gully, Darling Downs, illustrating high-velocity (channel) and low-velocity (overbank, vertical accretion) deposits (January 2011).*

over the Plio-Pleistocene may be a result of physiological responses to changes in the physical and/or biotic environment (Price and Piper, 2009).

Overall, the Pleistocene faunas differ significantly in composition from the older proximal Pliocene Chinchilla Local Fauna. Strictly arboreal forms, such as tree kangaroos (e.g., *Bohra* from Chinchilla), have not been reported from the younger Pleistocene Darling Downs deposits, nor the adjacent Texas Caves and Gore deposits (Price et al., 2009b). The Pleistocene deposits also lack the diverse small browsing forest wallabies (*Silvaroo* and related forms) characteristic of the Pliocene at Chinchilla. Crocodiles are much rarer elements of the Pleistocene deposits in comparison to Chinchilla. Proportionately, the Pleistocene faunas are composed of a greater number of grazing taxa and other mixed feeders that are typical of open habitats. Collectively, the Pleistocene assemblages reflect a significantly drier climatic signature than those of the Pliocene Chinchilla Local Fauna. The implication of such an interpretation is that major habitat and faunal reorganisation occurred locally between the Pliocene and Pleistocene, reflecting the expansion of open woodlands and grasslands, and contraction of dense woodlands (Price et al., 2009a).

*Figure 4 Lower left dentary of a giant forest wallaby, Protemnodon, from a late Pleistocene megafaunal fossil locality at Kings Creek, Darling Downs (January 2011).*

Recent radiometric (radiocarbon and U/Th) and optical dating has focused on developing reliable chronologies of deposition for the late Pleistocene Darling Downs megafaunal sequences (Price and Sobbe, 2005; Webb et al., 2007; Price et al., 2011). The dates are significant in that they show a long-term decline in local megafaunal community diversity during the period immediately leading up to the hypothesised time of human continental colonisation. Minimally, a three stage local extinction of late Pleistocene megafauna has been identified (Price and Webb, 2006). The data are most parsimoniously consistent with a temporally progressive, pre-human climate change model of megafaunal extinction. Other paleoecological information, gleaned from non-megafaunal taxa, suggests a long-term trend towards increasingly open, arid environments over the period of deposition (Price, 2005; Price et al., 2005; Price and Webb, 2006). Collectively, the data are not consistent with a nearly simultaneous extinction of megafauna as required for supporting the human-induced blitzkrieg overkill extinction hypothesis (Price et al., 2011). The results do not necessarily reject the overkill hypothesis for the later surviving forms, such as *Macropus giganteus titan* (giant grey kangaroo) and *Diprotodon optatum* recorded from the youngest megafaunal site at Neds Gully (dated to 46 ± 7 ka by Roberts et al. 2001). However, a current lack of younger, dated local sites does not allow for examination of extinction dynamics post-deposition of Neds Gully. It should also be noted that evidence for human-megafaunal interaction is lacking and that the oldest archeological record of humans locally post-date the youngest megafauna deposits by more than 30 kyr (Gill, 1978).

## Conclusions

Central eastern Australian vertebrate fossil deposits preserve evidence for long-term faunal change and attenuation in response to progressively arid climates during the Plio-Pleistocene. The overall trend is towards increasingly open habitat-adapted faunas as a response to the retraction of wet closed woodlands and tropical rainforests, and expansion of grasslands. Independent sedimentological, palynological, and paleontological datasets are beginning to demonstrate that major Quaternary climatic events (e.g., aridification) were temporally asynchronous in different regions of Australia (e.g., central vs coastal; northern vs southern). Thus, because late Quaternary climate changes were complex beyond glacial-interglacial cyclicity, faunal response(s) must have been similarly complex. In terms of understanding key biological events of the Quaternary, such as the extinction of Pleistocene megafauna, refinement and testing of existing hypotheses can only be achieved with far more extensive, complete and reliable datasets than is presently available. Crucially, existing datasets need to be better developed and contain regionally extensive, diverse and especially, well-dated faunal records, coupled with firm stratigraphic control. Regardless, the windows on Plio-Pleistocene Australian paleoclimate, paleohabitats and vertebrate assemblages afforded by Chinchilla, Mount Etna and the Darling Downs provide an exciting glimpse into Australia's recent past.

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# Accretion on the long-lived continental margin of northeastern Australia

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*The northern extremity of the late Neoproterozoic-Paleozoic Tasman Orogenic zone exposed in north Queensland forms a narrow belt of tectonised rock assemblages abutting Paleoproterozoic–Mesoproterozoic rocks of the North Australian craton. The craton-orogen contact (Tasman Line) is extensively exposed, a unique circumstance for Australia. Sedimentary protoliths of the cratonic rocks were mainly deposited between 1700–1600 Ma and multiply deformed between 1600–1500 Ma. The Lynd Mylonite Zone, one expression of the Tasman Line, separates rocks of the late Neoproterozoic–Ordovician Thomson Orogen from those of the craton. The succeeding Silurian–Devonian Mossman Orogen is generally faulted against the Thomson Orogen, but in its northern extent it may directly abut the craton along the Palmerville Fault, also an expression of the Tasman Line. These two orogenic systems are dominantly of active margin association and E-stepping but deep seismic imaging indicates that they are extensively underlain by crust of Archean or Paleoproterozoic age. The Tasman Orogenic Zone in its southern part represents a broad tract of crust c. 1,000 km across, added to the cratonic core of Australia in a phase of rapid accretion. In contrast, for its north Queensland development a much smaller volume of new crust was generated, expressing slow accretion. For this region the orogenic system laps extensively onto cratonic crust, a geometry which at least in part reflects overthusting during episodes of Paleozoic contractional orogenesis. As a consequence of little orogenic accretionary outgrowth of the north Queensland continental margin, three large-scale, successive igneous assemblages of active margin association generated throughout the Paleozoic form largely co-located and overprinting belts with plutonic suites stitching the Tasman Line and extending into the craton.*

## Introduction

The boundary between Paleoproterozoic–Mesoproterozoic rocks of cratonic Australia and Paleozoic rock assemblages of the Tasman Orogenic Zone (Tasmanides) is well exposed in north Queensland (e.g., Shaw et al., 1987; Fergusson et al., 2007c). To the S, this boundary, the Tasman Line of Hill (1951) and subsequent authors, is obscured by Mesozoic cover and has been mapped by magnetic and gravity images. Some authors (e.g., Gunn et al., 1997; Scheibner and Veevers, 2000) have interpreted the boundary to mark Rodinian breakup along which crust now represented by Australia separated from that of Laurentia in the Neoproterozoic (Hoffman, 1991; Karlstrom et al., 2001; Gibson et al., 2008). However, others (e.g., Direen and Crawford, 2003) have suggested that a complex late Neoproterozoic–Carboniferous history produced a range of geophysical responses and have rejected the Tasman Line as marking continental rupture. New deep seismic data suggest that cratonic rocks in north Queensland extend E of the Tasman Line at least 150 km as a lower crustal layer beneath Paleozoic orogenic systems in the upper crust (Korsch et al., 2012).

Paleoproterozoic and Mesoproterozoic rock assemblages of north Queensland form part of the North Australian Craton (Figure 1). As noted by Austin and Foster (2008) and Cawood and Korsch (2008), assemblages of the Georgetown Inlier located in the East adjacent to the Tasman Line have a striking similar history to those of the Mount Isa Inlier in the W, suggesting that these tracts were conjoint by 1600 Ma. Substantial marine sedimentary basins were developed in both tracts between 1700–1600 Ma with contactional orogenesis and plutonism between 1600–1550 Ma. Interpretation by Korsch et al. (2012) of two deep crustal suture zones imaged by seismic traversing, supported by magnetotelluric data, however, identifies episodes of Paleoproterozoic or older suturing involving crust of dissimilar origins.

The exposed north Queensland segment of the Tasmanides measures less than 300 km across compared to a span of c. 1,000 km for the coeval Lachlan Orogen in New South Wales and Victoria (Figure 1). It represents two sequential, E stepping orogenic systems expressed by voluminous igneous and sedimentary rock systems of continental margin association formed as belts over a specific interval of time and subsequently tectonised by discrete, intense episodes of contractional orogenesis. The inboard Thomson Orogen of Neoproterozoic–early Ordovician age, best represented by the Greenvale and Charters Towers provinces, is considered to be expansive in central and southern Queensland beneath Mesozoic cover (e.g., Betts et al., 2002; Glen, 2005). The outboard Silurian–Devonian

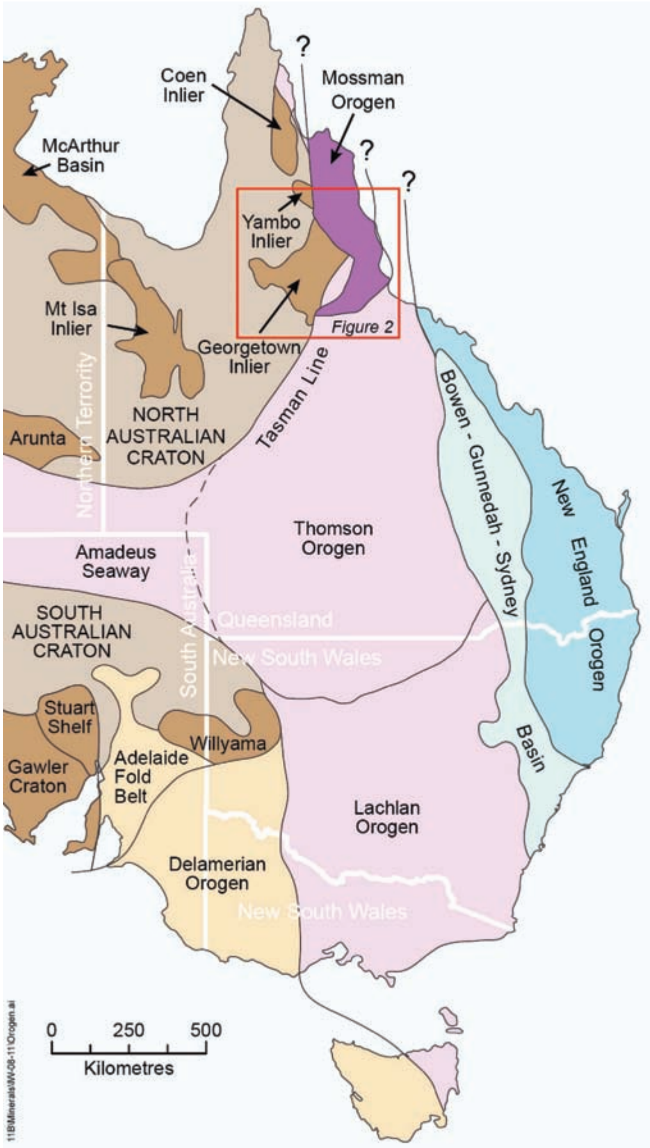


***Figure 1 Eastern Australia, showing the main Proterozoic elements (outcropping parts shown in dark brown) and Paleozoic orogens.***

Mossman Orogen, best represented by the Hodgkinson and Broken River provinces, terminates to the S on the Clarke River Fault Zone interpreted to be a sinistral strike-slip structure which has displaced southward continuation of the orogen offshore (Henderson, 1987, Henderson et al., 2011). The two orogenic systems have a meridional overlapping zone of contact considered in large part to consist of thrust dislocations. The New England Orogen, largely of Late Devonian–Permian age may form a third, easternmost orogenic belt in the region. It is mapped N along the coastal zone of central Queensland to Townsville where it passes offshore. Drill core records from the submerged Queensland Plateau (Mortimer et al., 2008) offshore from Cairns suggest it continues along the continental margin. late Carboniferous–Permian granitoids, hypabyssal suites including extensive dyke swarms and tracts of volcanics grouped as the Kennedy Igneous Association, considered to be related to the New England Orogen by Blevin et al. (1999), are widespread through the Thomson and Mossman orogenic belts extending W onto the craton.

In its southern expression, the Tasman Orogenic Zone is dominated by voluminous sedimentary systems associated with extensional tectonics (e.g., Foster and Gray, 2000; Kemp et al., 2009) shortened in a succession of contractional episodes, principally in the mid Silurian Benambran and Middle Devonian Tabberabberan orogenies (e.g., Glen, 2005). It its northern expression, extensional sedimentary systems of like scale are not in evidence but Benambran and Tabberabberan orogenesis applies, although the latter is offset in age to latest Devonian. Intrusions of two successive orogenic igneous belts, represented by the Siluro-Devonian Pama, and late Carboniferous–early Permian Kennedy Igneous Associations stitch the Tasman Line, a relationship which implies orogenic continental accretion of very limited scale. Deep seismic records have been interpreted as showing stacking of the Thomson and Mossman orogenic belts through large scale thrust transport (Korsch et al., 2012). Thus, as noted by Glen (2005), a major change in tectonic pattern from rapid to slow accretionary growth occurs along the Tasmanide belt between the Lachlan Orogen and north Queensland. The systematics of this shift and its underpinning mechanisms are presently unresolved.

This paper briefly reviews the nature of exposed crust in north Queensland (Figure 2) from the Precambrian craton to adjoining Paleozoic orogenic assemblages, in the broad context of crustal accretion along a long-lived continental margin. It probably first formed in the Neoproterozoic, although some authors (e.g., Betts and Giles, 2006) propose that it was a continental margin prior to the assembly of Rodinia. It was subject to active margin tectonics throughout the Paleozoic.

## Paleoproterozoic–Mesoproterozoic rocks

Three inliers with Paleoproterozoic–Mesoproterozoic rocks adjoin the Tasman Line (Figure 1). Of these the Georgetown Inlier is the most extensive and best known. It consists mainly of Etheridge Province, a discrete stratotectonic rock assemblage which also extends along the eastern parts of the Yambo and Coen inliers. In the Georgetown Inlier, the metasedimentary Etheridge Group, the preserved part of a basinal succession at least 6 km thick, is extensively developed in its western parts where the intensity of deformation and metamorphism has not precluded stratigraphic documentation (see Figures 2 and 3).

The lower part of the Etheridge Group consists of fine-grained calcareous-dolomitic sandstone and siltstone-mudstone, overlain by locally pillowed metabasalt, passing up to carbonaceous mudstone, intruded by sills of Cobbold Metadolerite. The environment of deposition has been interpreted as an upward deepening package from near shore to deep marine (Withnall et al., 1997). The mafic rocks are relatively evolved, low-K, Fe-rich continental tholeiites (Baker et al., 2010). The upper part is variably carbonaceous siltstone and mudstone, with sandstone composed entirely of mudclasts. Gypsum casts are of sporadic occurrence. Parts of the succession have been interpreted as shallow water in a mud-dominated tidal flat environment (Withnall et al., 1997).

Recent SHRIMP zircon dating has provided constraints on the age of the Etheridge Group. The lower part is characterised by detrital zircon populations with a significant Archean to earliest Proterozoic component (up to 75%), but with younger populations of 1850–1780 Ma and rare younger grains of c. 1695 Ma (Neumann and Kositcin, 2011). A metabasalt has an age of 1663±13 Ma (Baker et al., 2010) and a leucogabbro sill in the middle of the succession has a precise

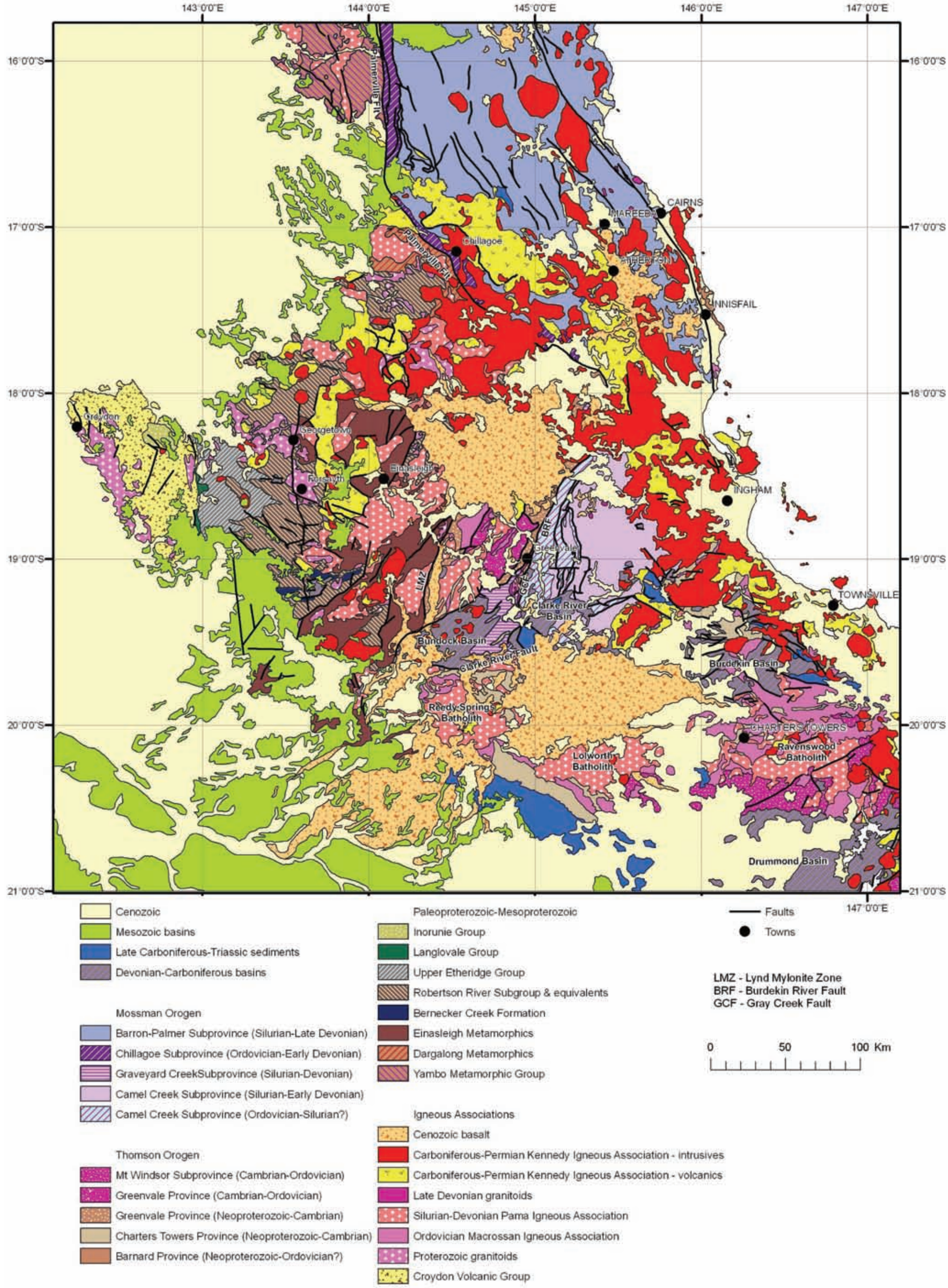


*Figure 2 Generalised geology of north Queensland.*

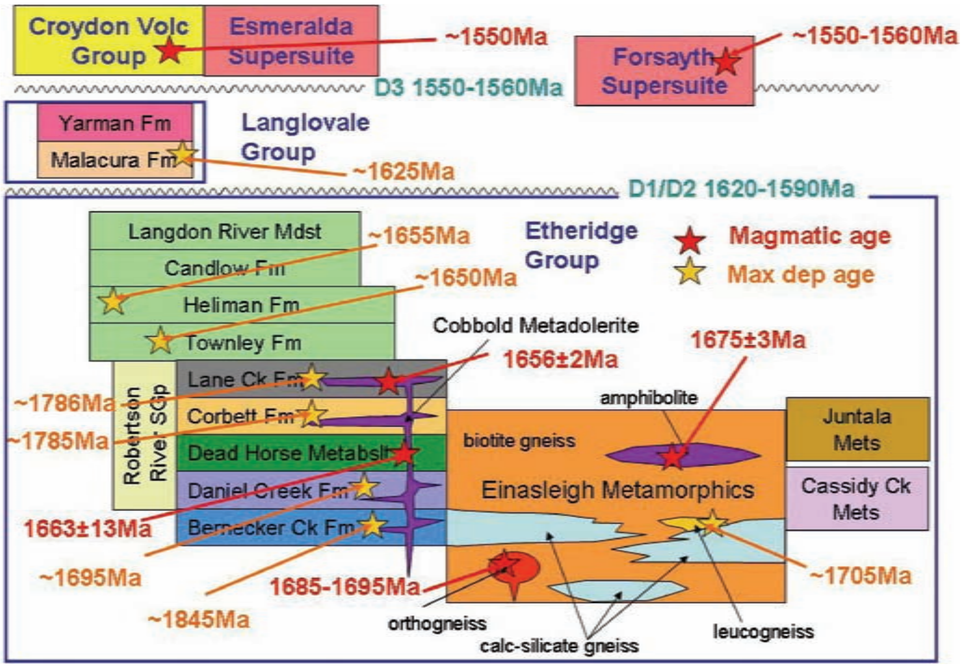


***Figure 3 Temporal relationships for Proterozoic rocks in the Georgetown Inlier.***

age of 1656±2 Ma (Black et al., 1998). These data constrain the protolith age as between 1700–1655 Ma.

Detrital zircons from its upper part have greatly diminished Archean to earliest Proterozoic components and significant populations at c. 1655–1650 Ma (Neumann and Kositcin, 2011), indicating a major change in provenance. The host rocks also have relatively primitive εNd values compared with those from the lower Etheridge Group, which are more evolved (Lambeck, 2011).

In the eastern part of the Georgetown Inlier, the Einasleigh Metamorphics consist of biotite and calc-silicate paragneiss, common amphibolite and rare leucogneiss and orthogneiss considered to overlap the Etheridge Group in protolith age (Figure 3). Extensive areas of migmatite and anatectic granite also are present. The main age constraints are based on zircon dates from felsic leucogneiss, considered to represent feldspathic psammite, for which simple zircon populations suggesting a maximum depositional age of c. 1705 Ma (Black et al., 2005). Rare granite gneisses give ages of 1695–1685 Ma (Black et al., 1998; Neumann and Kositcin, 2011). Amphibolite that cuts one of these gneisses has been dated at 1675±3 Ma (Black et al., 1998). The current data thus suggest deposition between c. 1700–1675 Ma.

The Langlovale Group unconformably overlies the Etheridge Group in the western Etheridge Province. It may be shallow marine in part but its but its upper part contains turbidites. Detrital zircon provides a maximum depositional age of c. 1625 Ma (Neumann and Kositcin, 2011).

The unconformably overlying Croydon Volcanic Group consists of almost flat-lying sheets of unmetamorphosed felsic ignimbrite intruded by comagmatic granites of the Esmeralda Supersuite. They contain abundant graphitic enclaves – presumably derived from the Etheridge Group at depth. They are peraluminous, reduced and considered to be S-types, and have provided a multi-grain TIMS U-Pb zircon age of 1552±2 Ma (Black and McCulloch, 1990).

The Etheridge Group is multiply deformed with the complexity of the deformation increasing eastwards, along with metamorphic grade which increases from greenschist to local granulite facies. A province-wide deformation history is apparent from regional structural analysis and relationships (Withnall et al., 1997), zircon dating (Black et al., 1998, 2005), and microstructural studies combined with EPMA monazite dating (Cihan et al., 2006). N-S shortening ($D_1$) of the Etheridge Province at c. 1620 Ma, prior to deposition of the Langlovale Group, was followed by E-W shortening ($D_2$) at c. 1590 Ma accompanied by medium P-T metamorphism with uplift and retrogressive metamorphism between 1590–1560 Ma. NW-SE shortening ($D_3$) and low pressure-high temperature metamorphism at 1560–1550 Ma, was associated with general metamorphic zircon growth and emplacement of the S-type Forsayth Supersuite. This deformational episode predated eruption of the Croydon Volcanic Group and emplacement of the Esmeralda Supersuite in the W at c. 1550 Ma.

## Neoproterozoic–Ordovician Thomson orogen

A substantial part of NE Australia, lying between the Tasman Line and the New England Orogen but obscured by Mesozoic cover, is regarded as a distinctive crustal tract which developed in the Neoproterozoic and early Paleozoic (Figure 1). Over most of its distribution it is known only from gravity and magnetic trends and basement cores recovered from petroleum drilling. In the northern Tasmanides it is exposed as the Anakie, Charters Towers, Greenvale and Barnard provinces which adjoin the Mossman Orogen (Figures 2 and 4). The Anakie Province lies immediately to the S of the area discussed in this paper and was described by Withnall et al. (1995) and Fergusson et al. (2001)

As presently understood, the Thomson Orogen is considered to consist of two elements (Fergusson et al., 2007a). Neoproterozoic to early Cambrian sedimentary systems and mafic igneous rocks formed on the passive margin along what is now eastern Australia following the breakup of Rodinia. This event is generally suggested as occurring at c. 700 Ma, although the abundance of rift-related magmatism of 600–570 Ma in eastern Australia led Crawford et al. (2003) to suggest that actual or renewed breakup began at c. 600 Ma. Delamerian contractional orogenesis followed, coincident with the inception of a convergent plate boundary located outboard of the east Australian margin. This older assemblage of the Thomson Orogen represents northward continuation of the Ross–Delamerian Orogen recognised for Antarctica and southern Australia. Rock assemblages of active margin association, including voluminous granitoids and volcano-sedimentary sequences developed in extensional basins, comprise the second element of the Thomson Orogen. Regional contraction during the mid Silurian Benambran Orogeny divides

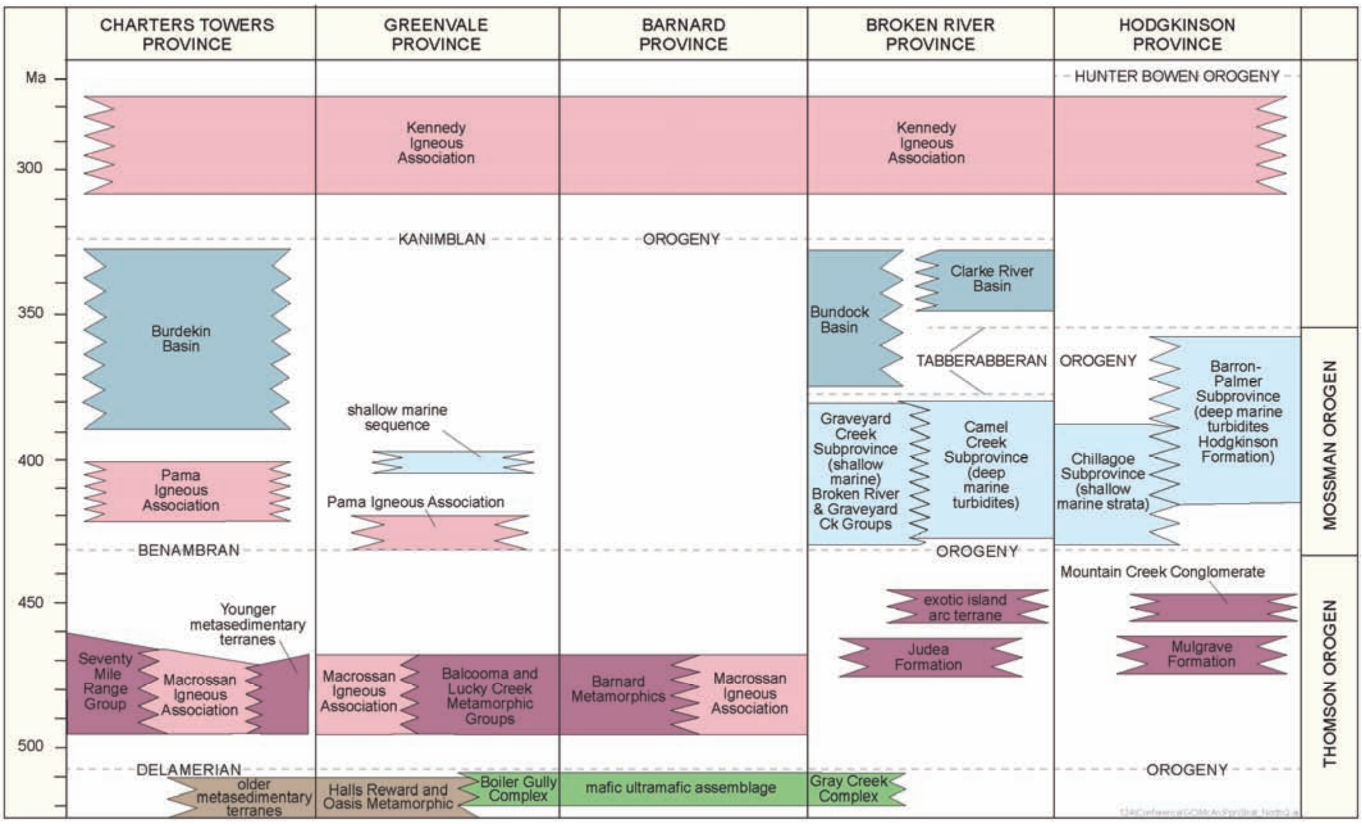


***Figure 4 Temporal relationships for Paleozoic rocks in north Queensland.***

Thomson orogenic assemblages from younger elements of the Tasmanides.

## Charters Towers Province

Remnants of Neoproterozoic sedimentary systems, now tectonised as disparate metamorphic terranes which host and are separated by voluminous granitoids of the Ravenswood and Lolworth batholiths, are the oldest elements. Ordovician parts of the batholiths are of Thomson Orogen association. The province also contains tracts of variably, but in general lightly, deformed volcanic and sedimentary basin fill of latest Cambrian–Early Ordovician age, the Mount Windsor Subprovince. The late Paleozoic Burdekin Basin is local cover to rocks of the Thomson Orogen in the eastern part of the province.

Metamorphic rocks form discrete, widely scattered terranes separated by granitoids (Figure 2), and consisting largely of multiply deformed pelitic and psammitic schist, quartzite, calc-silicate rocks, amphibolite and rare serpentinite. Gneissic granitoid is a common associate. Detrital zircon ages obtained from meta-arenite and a zircon age from an igneous meta-breccia clast indicate that protoliths includes both Neoproterozoic and early Cambrian strata (Fergusson et al., 2005, 2007b). Diorite intruding the Charters Towers Metamorphics has a zircon age of 508 Ma (Hutton et al., 1997).

Remnant foliations may record the late Cambrian Delamerian Orogeny, but are largely overprinted by an Early Ordovician episode of deformation that produced pervasive, shallowly dipping foliations accompanied by amphibolite grade metamorphism and considered to be extensional by Fergusson et al. (2007b). The dominant fabric is coplanar with that developed in granitic gneiss, emplacement of which is considered to be syntectonic. Zircon dates on the granitic gneiss are 493 Ma, whereas co-located, post-tectonic granitoid phases have ages of 461 Ma and 455 Ma (Hutton et al., 1997, Fergusson et al., 2007b). This extension is inferred to be the basin-forming mechanism related to deposition of protolith strata of the metamorphic terranes and to be coeval with basin development which accommodated the Seventy Mile Range Group volcano-sedimentary succession of like age. Late stage deformation of the metamorphic terranes involved upright folds and greenschist metamorphism dated by Ar/Ar as Silurian and considered to reflect Benambran orogenesis (Fergusson et al., 2005, 2007b).

The Seventy Mile Range Group is developed across the southern part of the Charters Towers Province (referred to as the Mount Windsor Subprovince) and represents the infill of an extensional basin, interpreted as a continental backarc feature (Henderson, 1986; Berry et al., 1992; Stolz, 1995). The base of the succession is stoped out by the Ravenswood Batholith and its upper part passes beneath younger cover. As preserved it has an apparent thickness of some 15 km. Its lower part consists of siliciclastic deep marine strata and hosts tholeiitic mafic dykes of intraplate association. Its upper part consists of intermediate to silicic volcanic formations of deep marine facies (e.g., Monecke et al., 2006) and volcaniclastic strata with horizons containing an early Ordovician (Tremadocian–Darriwilian) pelagic fauna (Henderson, 1984). In general, strata of the group dip gently S but cleavage and upright folding is developed in its western distribution.

Batholiths of the Charters Towers Province are composites that include elements of the Ordovician Macrossan Igneous Association (490–455 Ma) allied to the Thomson Orogen, the Silurian–Devonian Pama Igneous Association (418–382 Ma) allied to the Mossman

Orogen, and the late Carboniferous–early Permian Kennedy Igneous Association allied to the New England Orogen for which volcanic and sub-volcanic rocks are also represented. Hornblende and/or biotite bearing I-type granitoids dominate the Macrossan Igneous Association. They show persistent evidence of strain with common incipient fabric development and are cut by local shear zones with intense fabrics (Hutton et al., 1997).

The Burdekin Basin is nonconformable on crystalline rocks of the Thomson Orogen. It contains a Middle Devonian–early Carboniferous sedimentary and volcanic succession up to 6 km thick (Draper and Lang, 1994). It is mainly non-marine apart from the shallow marine basal part and later short-lived marine incursions. The basin fill shows broad, open folds of variable orientation and is cut by common faults, many of which have substantial displacements. It is considered to be of backarc, extensional association and is coeval with rocks of the Mossman Orogen to which it is allied. Basin inversion predates the emplacement of Permian plutons of the Kennedy Igneous Association and is thought to have occurred in the late Carboniferous, as part of the Kanimblan Orogeny which is widely recognised in eastern Australia.

## Greenvale Province

This part of the Thomson Orogen consists of a NE-trending domain of predominantly late Neoproterozoic–early Paleozoic sedimentary and igneous rocks now metamorphosed to greenschist or amphibolite grade (Withnall, 1989; Fergusson et al., 2007c). It adjoins the Paleoproterozoic–Mesoproterozoic Etheridge Province along a major dislocation marked by the Lynd Mylonite Zone (Figure 2), now recognised as the Tasman Line.

In the westernmost part of the province, paragneiss and amphibolite assigned to the Oasis Metamorphics are intruded by gneissic granitoid that can be equated to the Macrossan Igneous Association. The age of metamorphic rims to zircon in gneiss dates the dominant fabric as c. 475 Ma, and cores suggest a Neoproterozoic or Cambrian depositional age. A primary age of c. 485 Ma has been obtained from gneissic granitoid hosted by the metasedimentary suite (Fergusson et al., 2007c).

Between the Balcooma Mylonite Zone in the W and the NE-trending early Silurian Dido Batholith (Pama Igneous Association) in the E, the metavolcanic and metasedimentary succession of the Early Ordovician Balcooma Metavolcanic Group is of similar content to the Seventy Mile Range Group of the Charters Towers Province. Geochronological constraints indicate that the two sequences are broadly coeval (Huston, 1990; Withnall, 1989; Withnall et al., 1991, 1997). East of the Dido Tonalite, a second metavolcanic assemblage, the Lucky Creek Metamorphic Group is probably also of early Paleozoic age. It contains mafic–silicic volcanic and volcaniclastic rocks (Withnall, 1989), but its eastern part is dominantly metapelite.

The early Paleozoic domains have an intense early foliation/cleavage considered to have been of low dip when formed and subsequently steepened by up to two overprinting folding and cleavage events (Fergusson et al., 2007c). Protolith of the metasedimentary assemblage is interpreted to be the floor of an extensional basin, tectonised by continuation of the strain regime which sponsored extension. Overprinting upright cleavage and folding events are probably an expression of the Silurian Benambran Orogeny. Rare outliers of gently dipping, unstrained, Early Devonian marine rocks unconformably on the early Paleozoic metasedimentary and metavolcanic rocks (Withnall, 1989; Withnall and Lang, 1993). The Dido Batholith, dated as early Silurian (c. 430 Ma) shows strong local subvertical foliation (Withnall et al., 1997) consistent with syn-Benambran emplacement.

A metasedimentary tract (Halls Reward Metamorphics) with a Neoproterozoic or early Cambrian protolith age and associated with mafic–ultramafic rocks (Boiler Gully Complex) forms an eastern bounding strip to the province. It was deformed in the middle Cambrian (520-500 Ma; Nishiya et al., 2003). Deep seismic imaging shows a positive culmination in upper crustal structure coincident with outcrop of the Halls Reward Metamorphics (Korsch et al., 2012).

## Barnard Province

The Barnard Province (Figure 2) consists of multiply deformed schist, quartzite, gneiss, amphibolite, metamorphosed ultramafic rocks and granitoids. It occupies a narrow coastal strip extending from Mission Beach to E of Cairns. Strongly foliated S-type granitoid and less deformed felsic I-type granitoid were dated by SHRIMP at 486 and 463 Ma, respectively (Bultitude et al., 1997). The assemblage resembles metamorphic and plutonic suites of the Thomson Orogen represented in the Charters Towers Province.

## Silurian–Devonian Mossman Orogen

The Mossman Orogen is located in the coastal sector of north Queensland, embracing the Broken River and Hodgkinson provinces, which are separated by tracts of late Paleozoic igneous rocks. Its sedimentary assemblages abut those of the Thomson Orogen and the north Australian craton. A belt of Silurian–Devonian granitoids (Pama Igneous Association) hosted by older rock systems to the S and W are also assigned to the orogen (Figures 2 and 4). These rocks have been variously interpreted as a coeval magmatic arc (Henderson 1987; Henderson et al., 2011), and as a zone of magmatic underplating related to backarc extension (Bultitude et al., 1997).

Although tectonic interpretation remains contentious, the orogen is regarded as an active continental margin assemblage that formed adjacent to a plate boundary where W-directed subduction transported oceanic crust beneath the continental edge of Australia. Both backarc and forearc/accretionary wedge models have been applied in rationalising its tectonic setting (Arnold and Fawckner, 1980; Henderson, 1987; Withnall and Lang, 1993; Henderson et al., 2011). Deposition was terminated and the orogen was comprehensively tectonised by crustal shortening in the Late Devonian, broadly correlated with the Tabberabberan Orogeny of the Lachlan Orogen. Much of the Hodgkinson Province was later shortened by the late Paleozoic–Triassic Hunter-Bowen Orogeny (Davis et al., 2002) which is best known from the New England Orogen.

## Broken River Province

Two contrasting, spatially discrete assemblages are recognised. In the E, an Ordovician–Early Devonian, strongly tectonised assemblage of deep marine turbidites and local basalt and chert form the Camel Creek Subprovince. The smaller Graveyard Creek Subprovince in the SW consists largely of a thick, less deformed succession of marine to terrestrial late Silurian–early Carboniferous strata (Withnall and Lang, 1993). Rocks of the two subprovinces are separated by the Gray Creek Fault (Figure 2). The province is terminated in the S against the Clarke River Fault. Major dislocation on this structure in the Late Devonian juxtaposed the mid Paleozoic rocks of the Broken River Province against older assemblages of the Charters Towers Province. The western boundary is marked by the Burdekin River Fault.

### *Graveyard Creek Subprovince*

A basement assemblage of Thomson Orogen located immediately to the W of the Gray Creek Fault, and also as inliers in anticlinal cores, consists of a variably deformed and serpentinised mafic–ultramafic assemblage (Gray Creek Complex) of probable Cambrian age and cleaved quartzose turbidites and associated bimodal lavas of the Early Ordovician Judea Formation, both cut by small Ordovician tonalitic intrusions.

An angular unconformity separates this assemblage from an 8 km thick early Silurian–Late Devonian sedimentary succession. The Silurian section (Graveyard Creek Group) commenced with conglomerate up to 500 m thick and largely derived from the underlying basement, passing up into deep marine strata, locally volcaniclastic towards the base, and closing with shallow marine limestone. The Devonian succession (Broken River Group) is typified by shallow marine limestone subject to tight biostratigraphic control (Withnall and Lang, 1993; Talent et al., 2002).

Frasnian contraction, marking the terminal event in the Mossman Orogen, is registered in the Graveyard Creek Subprovince by a low angle unconformity of short duration in its western part. The succeeding Famennian–Visean Bundock Basin has some 7 km of infill. Its succession is largely fluviatile and commonly volcaniclastic, but has several shallow marine intercalations. Conglomerates in the lower part are derived largely from Camel Creek sandstones and jasper. In the eastern part of the subprovince, late Devonian contraction induced folding and inversion of Silurian and Devonian strata and major movement on the Clarke River Fault truncating the Mossman Orogen occurred at this time.

The Middle Devonian–Frasnian successions of the Graveyard Creek Subprovince are closely comparable to their correlatives in the Burdekin Basin of the Charters Towers Province. This relationship implies that they formed in closer proximity than that now shown, consistent with later separation by sinistral displacement on the Clarke River Fault (Henderson, 1987).

A younger, late Carboniferous contraction, assigned to the Kanimblan Orogeny, was the main deformation in the Graveyard Creek Subprovince. It produced broad scale, upright, NE-trending folds and high angle thrust faults.

### *Camel Creek Subprovince*

Several assemblages of sedimentary and volcanic rocks are recognised for the subprovince, but superpositional relationships have been prejudiced by widespread mélange development and faulting. Few fossils are known and age control is poor.

The westernmost assemblage is a distinctive, fault bounded and internally disrupted association of generally quartz-poor siliciclastic rocks, shallow marine fossiliferous limestone and predominantly mafic to intermediate volcanic rocks considered to be an exotic oceanic island arc terrane (Henderson et al., 2011). Corals and conodonts from the limestones indicate a Late Ordovician age (Withnall and Lang, 1993). Henderson et al. (2011) considered the assemblage to represent oceanic crust accreted during Benambran orogenesis.

Most of the subprovince consists of contrasting belts of lithofeldspathic and quartz-rich turbidite units, which locally include chert and basalt (Withnall and Lang 1993). Redeposited conglomerate, sporadically with limestone clasts, is represented in some lithofeldspathic units as are blocks of limestone, some several kilometres long, which are considered allochthonous (Sloan et al., 1995). Fossils from limestone are of Silurian and early Devonian age. The age of quartz-rich units is unknown and they could represent dismembered parts of the Thomson Orogen which deep seismic profiling suggests is extensively developed at depth in the crust (Korsch et al., 2012). Persistent westerly facing for individual beds is in opposition to the sparse age information which indicates an overall age progression from older strata in the W to younger in the E.

The turbidite assemblage was extensively deformed by mélange formation and by local folding in the early stages of lithification. Tight, asymmetric meso-scale folding with western limbs dominant and the imposition of a weak but very widespread slaty cleavage represents a succeeding event. Folds are characteristically steeply plunging. The folds trend NNE-SSW in the W but ENE-WSW in the SE, having been refolded, along with the cleavage, by a broad oroclinal bending with a NE-trending axial trace. This assemblage was weakly metamorphosed to sub-greenschist facies. Small syntectonic granitoid bodies hosted by the sedimentary assemblage are dated by K-Ar at 357 Ma (Withnall and Lang, 1993).

Such tectonism occurred prior to inception of the Late Devonian–Carboniferous (Famennian–Visean) Clarke River Basin which contains a succession, up to 1,500 m thick that is mainly fluviatile, but is sporadically shallow marine near the base, and becomes markedly volcaniclastic towards the top. It is coeval with, and of like character, to the Bundock Basin of the Graveyard Creek Subprovince. A final NE-SW trending fold phase in the Camel Creek Subprovince, representing Kanimblan orogenesis, is expressed by low to intermediate dips in the Clarke River Group.

## Hodgkinson Province

This element trends northwards and is bounded to the W by the Palmerville Fault. It contains two contrasting sedimentary tracts: the Chillagoe Subprovince, a narrow western sector characterised by limestone and basalt and segmented by intense faulting; and the Barron-Palmer Subprovince, a pervasively tectonised terrane of deep marine turbidites and subordinate basalt and chert (Bultitude et al., 1993, 1997). It has a metamorphic overprint which ranges from sub-greenschist in the W to upper greenschist (biotite zone) in the E.

### *Chillagoe Subprovince*

The element is less than 20 km across but extends along strike for over 150 km. Abutting the Palmerville Fault, a narrow belt of siliciclastic rocks affiliated with assemblages of the Thomson Orogen consists of quartzose turbidites with subordinate mudstone, chert and basalt (Mulgrave Formation) in fault contact with massive volcaniclastic conglomerate containing limestone lenses *(*Mountain Creek Conglomerate). A zircon date obtained from a dacite clast and conodonts from limestone provided a Late Ordovician age (Bultitude et al., 1993).

The main assemblage (Chillagoe Formation) consists of shallow marine limestone, chert, arenite, pelite, conglomerate and basalt, disrupted by intense thrust repetition. Corals and conodonts indicate ages from mid Silurian to late Early Devonian for limestone (Bultitude et al., 1993, 1997). The belt has extraordinary structure with strata dipping steeply E but facing to the W and thus overturned.

### *Barron-Palmer Subprovince*

Quartz-intermediate turbidites, commonly disrupted by mélange, dominate this subprovince, but basalt, chert, redeposited conglomerate and rare limestone bodies are also represented. No broad-scale lithological discrimination has proven possible and the entire assemblage has been mapped as Hodgkinson Formation. Corals and conodonts from limestone range from Silurian–Late Devonian (Famennian), but widespread plant macrofossils confirm that a significant part is Middle–Late Devonian.

Small tracts of late Carboniferous and Permian sedimentary cover successions are present. Most are little disturbed, but they are folded and cleaved in the E.

Structure of the Hodgkinson Formation shows a consistent, regional pattern. Ubiquitous mélange is an early phase and an early fabric ($S_1$) coplanar with bedding is also widely developed. Tight folding at both meso- and large-scale, with the generation of an incipient to intense cleavage ($S_2$) is characteristic of the Hodgkinson Formation throughout its distribution. It is dated at c. 355–360 Ma based on the age of syntectonic Mount Formartine Granite, and is only slightly younger than the Late Devonian age of the youngest rocks it affects (Zucchetto et al., 1999). Thrust faulting in the Chillagoe Subprovince predates Carboniferous plutonism. An overprinting shallowly inclined $S_3$ fabric and associated folds sporadically developed in the Barron-Palmer Subprovince are interpreted as extensional structures and considered to be early Permian (Davis and Henderson, 1999; Davis et al., 2002). A final deformational episode is widely expressed across the subprovince, particularly in the E, and is represented by the development of $S_4$ cleavage and associated mesoscopic folds that are coaxial with $D_2$. Dated syntectonic plutons indicate that the commencement of $D_4$ shortening was also early Permian and only slightly offset in time from $D_3$.

## Kennedy igneous association

The term Kennedy Province was coined by Bain and Draper (1997) for Carboniferous–Permian (c. 340–270 Ma) igneous rocks that extend throughout north Queensland. It is here referred to as the Kennedy Igneous Association.

In the area discussed in this paper, the rocks are concentrated in the Townsville-Mornington Island Belt (TMIB) (Figure 5), cropping out in a W-NW-trending band (800 x 300 km) that transgress all of the older provinces. Concealed rocks of the same geophysical character extend at least 300 km farther WNW to the coast of the Gulf of Carpentaria. Early Permian granite has been intersected by drilling on Mornington Island, a further 200 km to the W.

The TMIB includes major batholiths, smaller intrusions and numerous major volcanic fields. Most of these volcanic fields are preserved in large composite cauldron complexes, and are dominated by thick piles of dacitic–rhyolitic ignimbrite and some lavas with subordinate mafic lavas (Withnall et al., 1997; Bultitude et al., 1997; Mackenzie and Wellman, 1997).

The rocks are little disturbed by tectonism. Granitic intrusions are geochemically diverse, including S, I and A types. Two age clusters are represented: end-of-Devonian–early Carboniferous (357–335 Ma) and late Carboniferous–Permian (305–270 Ma). Basic–silicic dykes are widely represented, locally as swarms. The wide spread of magmatism in both space and time indicates broad-scale heterogeneous thermal input, or a series of inputs, over a period of 70 Myr.

The magmatic rocks are coeval with and merge southwards into late Paleozoic volcanic and plutonic rocks in the New England Orogen which are generally thought to have formed on an Andean margin related to a westerly dipping subduction zone. Carboniferous forearc and accretionary wedge assemblages are recognised in the E of the New England Orogen, although they are not evident for the Permian component, which shows some evidence for crustal thinning and

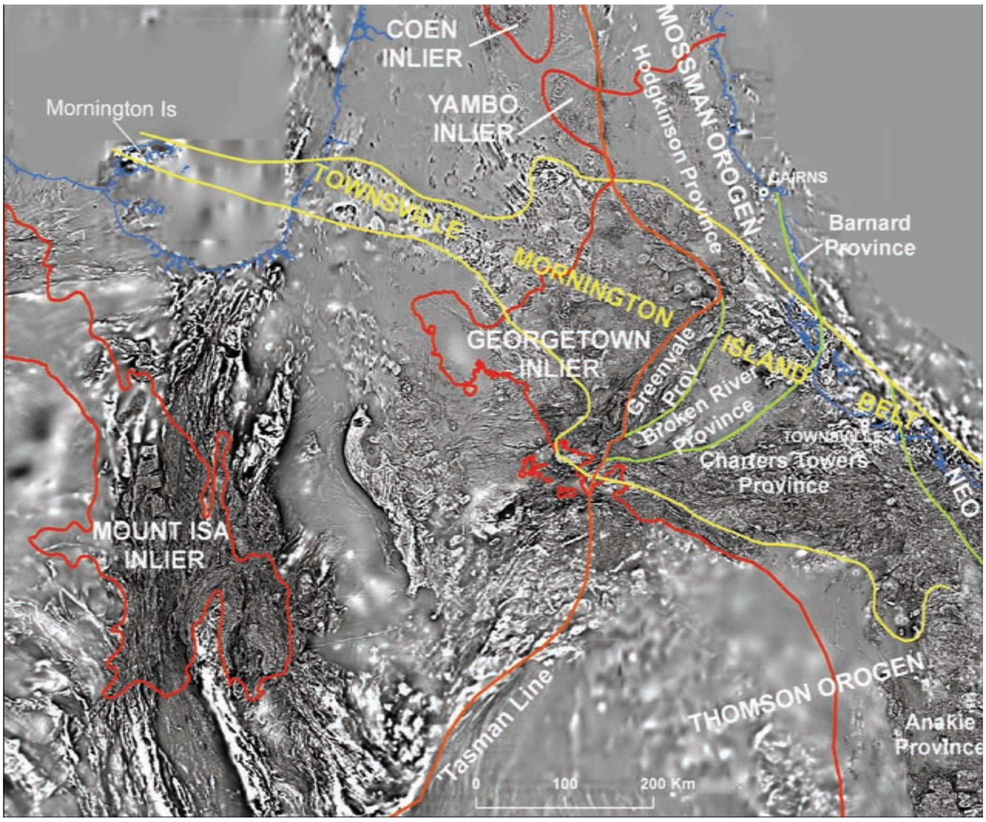


***Figure 5 First-vertical-derivative magnetic image for north Queensland showing the main structural elements and the Townsville-Mornington Island belt (yellow). The cratonic margin (Tasman Line) is delineated in orange, and the red line shows the limits of exposure of pre-Mesozoic rocks.***

transition to an extensional setting (Withnall et al., 2009). Most early models for the tectonic setting and origin of the rocks of the Kennedy Igneous Association also proposed a subduction-related setting (e.g., Henderson, 1980; Bailey et al., 1982). However, major caldera-related volcanic fields elsewhere in the world are clearly in extensional settings: e.g., southwestern USA (extension over a subducted spreading ridge), Taupo Volcanic Field, northern Turkey and parts of the Andes (backarc extension in continental crust). Oversby and Mackenzie (1995) interpreted the Carboniferous and early Permian volcanics in the Georgetown region in terms of E-W (Carboniferous) and subsequent (early Permian) NE-SW extension. Mackenzie and Wellman (1997) proposed that the Kennedy Province was essentially the result of crustal melting in an extensional (or transtensional), possibly backarc, tectonic environment. However, the reasons for the change of tectonic setting and trend of the rocks of the Kennedy Igneous Association compared with coeval magmatic rocks in the New England Orogen and their geodynamic linkages have yet to be satisfactorily explained.

## Discussion

Cratonic crust W of the Tasman line, as represented by adjacent Precambrian inliers, had no substantive addition through basin formation in the Neoproterozoic and Paleozoic. For the Thomson Orogen, older, pre-late Cambrian rocks are consistent with a passive margin origin but their contribution to surface exposure is very small. The more conspicuous late Cambrian–Ordovician assemblages with a strong plutonic and calc-alkaline volcanic representation, mafic–ultramafic rocks and sedimentary suites of deep marine character have the hallmarks of an active continental margin association. However they are generally of inboard, arc or backarc affiliation with potential outboard suites conspicuously under-represented. Representatives of the succeeding plutonic suites of the Siluro-Devonian Pama and late Carboniferous–Permian Kennedy Igneous associations, also of active margin association, are hosted by the Thomson Orogen and also by the craton implying very limited outward stepping of the continental margin by crustal accretion through the Neoproterozoic and early Paleozoic.

An ongoing active margin setting is apparent for the Mossman Orogen. The pairing of inboard shallow marine Silurian–Devonian successions (Graveyard Creek and Chillagoe Subprovinces) with outboard, broadly coeval, strongly deformed, deep marine turbidite-facies sedimentary belts has led some workers to advocate a forearc–accretionary prism model for their setting (e.g., Arnold in Arnold and Fawckner, 1980; Cooper et al., 1975; Henderson, 1987; Henderson et al., 2011). Others have suggested a backarc model (e.g., Fawckner in Arnold and Fawckner 1980; Bultitude et al., 1993 1997; Withnall and Lang, 1993). Regardless of interpretation, the Graveyard Creek and Chillagoe Subprovinces are disjunct, with their continuity broken by fault juxtaposition of the Camel Creek Subprovince against those of the Thomson Orogen. Substantial lateral transport by overthrusting across the shallow marine on this fault contact is implied.

Interpretation of deep seismic imaging by Korsch et al. (2012) suggests that shortening by very large scale thrust imbrications, induced by contractions of the Benambran and Tabberabberan orogenies, have been formative for the upper crustal architecture of the northern Tasmanides. The original palinspastic patterns of the Thomson and Mossman orogens have been severely disrupted by shortening, with very limited outward growth of the continental margin. Successive episodes of active margin melt generation were emplaced along zones of similar location.

Although a contractional setting is evident for the late Paleozoic of the New England Orogen to the S, this is not evident for the coeval Kennedy Igneous Association in north Queensland, which is transverse to the continental margin and appears to be related to an extensional or transtensional setting.

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# The Macquarie Arc, Lachlan Orogen, New South Wales: its evolution, tectonic setting and mineral deposits

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*The Lachlan Orogen of New South Wales, Victoria and eastern Tasmania is the best understood element of the eastern Australian Tasmanides. The Tasmanides encompass continental growth in a Neoproterozoic passive margin setting and a Paleozoic–Mesozoic active margin in east Gondwana, bounded to the east by the Pacific Ocean. In the Ordovician, the supra-subduction zone element in the Lachlan Orogen is the Macquarie Arc. This comprises one minor and three major belts of mafic to intermediate volcanic and volcaniclastic rocks, limestones and intrusions that, with two hiatuses in magmatism, span the Ordovician and extend into the Early Silurian. The three major belts in central New South Wales are separated by Silurian–Devonian rift basins and are therefore, inferred to have been rifted apart during crustal extension. Chemical and isotopic data suggest the Macquarie Arc is intraoceanic, developed on primitive oceanic crust. Despite this, key features such as its longevity, stacked magmatic phases, little deformation and flanking coeval craton-derived turbidites and black shales with no provenance mixing indicate differences from modern intraoceanic arcs. Paleogeographic setting and magmatic evolution of the Macquarie Arc provided perfect conditions for mainly porphyry-related, rich Au-Cu deposits, in the Ordovician, and especially in the Early Silurian after amalgamation of the arc with its flanking terranes.*

## The Tasmanides and internal subdivisions

The eastern third of Australia includes five orogenic belts plus a Permian-Triassic basin system that form the eastern Australian Tasmanides (Figure 1). The Delamerian Orogen occupies the southwestern part, the Lachlan Orogen is central and the northeastern part is the New England Orogen. The Thomson Orogen lies N of the Lachlan Orogen almost completely obscured by Mesozoic and Cenozoic cover. The Mossman Orogen occupies the far northeastern part in north Queensland (Figure 1).

The boundary between the Delamerian Orogen and Lachlan Orogen is contentious, even in western Victoria where their outcrops are closest together. Early workers (e.g., Glen, 1992; Wilson et al., 1992) placed the boundary at the Avoca Fault, between the Stawell and Bendigo structural zones (Figure 1) since, on a regional scale it truncates aeromagnetic marker units assigned to the Delamerian Orogen in the west. The Moyston Fault between the Stawell and Glenelg zones has been favoured also (e.g., Cayley and Taylor, 1997, 2001). Early deep seismic reflection data suggested a transitional zone W of the Moyston Fault (Murphy et al., 2006). New seismic data showing that the W-dipping Avoca Fault soles onto the more important crustal-scale E-dipping Moyston Fault was used by Cayley et al. (2011) to support the latter fault being the boundary. This interpretation lies at slight odds with an interpretation of the orogens based on time-space plots, in which a c. 500 Ma deformation characterises the Delamerian Orogen, but not the Lachlan Orogen. This changeover appeared to coincide with the Moyston Fault until Miller et al. (2005) demonstrated c. 500 Ma deformation in the Stawell Zone.

The boundary between the Lachlan Orogen and the Thomson Orogen is concealed by the Mesozoic cover. It is defined by major changes in regional trends, evident in geophysical datasets, with curvilinear WNW-NE trends in the southern part of the Thomson Orogen oblique to NNW-N trends in the northwestern part of the Lachlan Orogen (e.g., Murray and Kirkegaard, 1978; Wellman 1995; Stevens 1991; Glen et al., 2010; Hegarty, 2010). The WNW-trending boundary corresponds to the Olepoloko Fault (Stevens, 1991); the NE-trending part is the Louth-Eumarra Shear Zone of Glen et al. (1996), probably along its northern margin (Glen et al., 2010). Where imaged by deep seismic reflection profiling, the Oelopoloko Fault corresponds to a steep N-dipping reflector that intersects the Moho and separates thick dense crust to the N from more normal layered crust (Glen et al., 2007d).

The boundary between the Lachlan Orogen and the New England Orogen is obscured by the Permian–Middle Triassic Gunnedah and Sydney basins (Figure 1). These elongate basins evolved from early Permian rifts or transtensional basins into a foreland basin system yoked to the New England Orogen, which underwent conversion to a fold-thrust belt by progressive westward thrusting and folding.

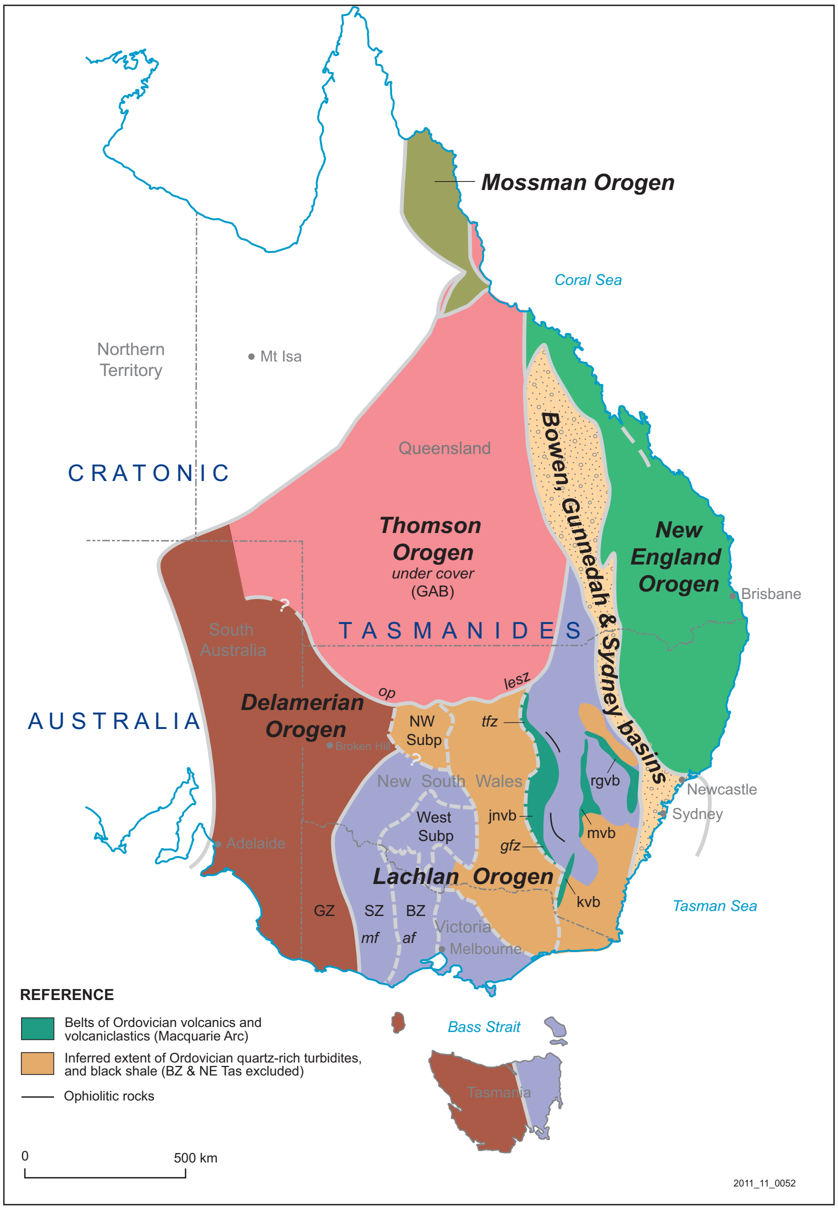



***Figure 1** Tasmanides of eastern Australia, showing in green Ordovician belts of Macquarie Arc in the Lachlan Orogen. Abbreviations: BZ = Bendigo Zone; GAB = Great Australian Basin; GZ= Gampians-Stavely Zone, western boundary omitted; SZ = Stawell Zone. Arc belts: jnvb=Junee-Narromine Volcanic Belt; mvb=Molong Volcanic Belt; rgvb=Rockley-Gulgong Volcanic Belt; kvb=Kiandra Volcanic Belt. Faults: af = Avoca Fault; mf = Moyston Fault; gfz=Gilmore Fault Zone; lesz = Louth-Eumarrra Shear Zone; op = Olepoloko Fault; tfz=Tullamore Fault Zone.*

The northeastern boundary of the Thomson Orogen against the Mossman Orogen (Henderson and Withnall, 2009; North Queensland Orogen of Glen, 2005) has been the subject of discussion (e.g., Day et al., 1978; Murray and Kirkegaard, 1978), but is now generally regarded to lie N of the Charters Towers Province of the Thomson Orogen.

## The Macquarie Arc in the Lachlan Orogen

The Lachlan Orogen is the central part of the Tasmanides and underlies most of NSW, Victoria and eastern Tasmania. It contains an intraoceanic supra-subduction zone system, the Macquarie Arc, key elements of which are now discussed. This summary draws strongly on results of a joint project undertaken by CODES and the GSNSW, published as Crawford et al. (2007a), and updated by recent detailed mapping in and around parts of the arc by Quinn and co-workers and analysis sampling of detrital zircons undertaken with GEMOC (Macquarie University) and reported in Glen et al. (2011). The work reported in Crawford et al. (2007a) was itself built on joint mapping by GSNSW and Geoscience Australia that produced 1:100 000 scale published maps. This mapping was then used to constrain geochemical sampling, leading to results that updated and built on earlier geochemical work by Pemberton and Offler (1985), Wyborn (1992) and Wyborn and Sun (1993).

Four belts of Ordovician volcanic, sedimentary (volcaniclastic plus limestone) and intrusive rocks in the Lachlan Orogen (Figure 2) have arc-like, calc-alkaline geochemistry and trace-element supra-subduction zone signatures. Three occur in central NSW – Junee-Narromine Volcanic Belt in the west; Molong Volcanic Belt in the centre; and the Rockley-Gulgong Volcanic Belt in the E (Figure 2). The Kiandra Volcanic Belt occurs in the Snowy Mountains, just N of the Victorian border. Andesitic volcaniclastic rocks interbedded with Ordovician cherts and black shales in the Limestone Creek area of NE Victoria (Figure 2) (Allen, 1988) have also been linked into the Macquarie Arc, but there is no continuity with the southern-most exposure of the Kiandra Volcanics in NSW (Quinn et al., unpublished data).

The three volcanic belts in central NSW have been variously interpreted. Interpretations range from unrelated to subduction, (e.g., the seamount model of Fergusson and Coney (1992) or Wyborn's (1992) model in which magmas were generated by melting of mantle lithosphere that was modified by Cambrian subduction. Subduction models vary from multiple island arcs (Fergusson and Colquhoun, 1996), to a single arc dismembered by strike-slip faulting (Packham, 1987; Fergusson, 2009), or by high-angle rifting during formation of the intervening younger Cowra and Hill End troughs and the Mumbil Shelf (Scheibner, 1989; Glen et al., 1998). Serial restorations based on matching key stratigraphic units across the belts preclude left or right-lateral strike-slip duplication, but support rifting and rigid block rotation (Glen et al., 2007a). Restoration of the arc with an irregular triangular shape (Molong Volcanic Belt and Junee-Narromine Volcanic Belt) projecting eastwards from a 800 km long baseline represented by the Junee-Narromine Volcanic Belt (inferred to approximate the magmatic core of the arc; Glen et al., 2007c) is also consistent with the matching of key geochemical units across the three belts (Crawford et al., 2007b; Glen et al., 2007b). Further support comes from the interpretation of gravity data and deep seismic reflection profiles that suggest blocks of the Macquarie Arc were

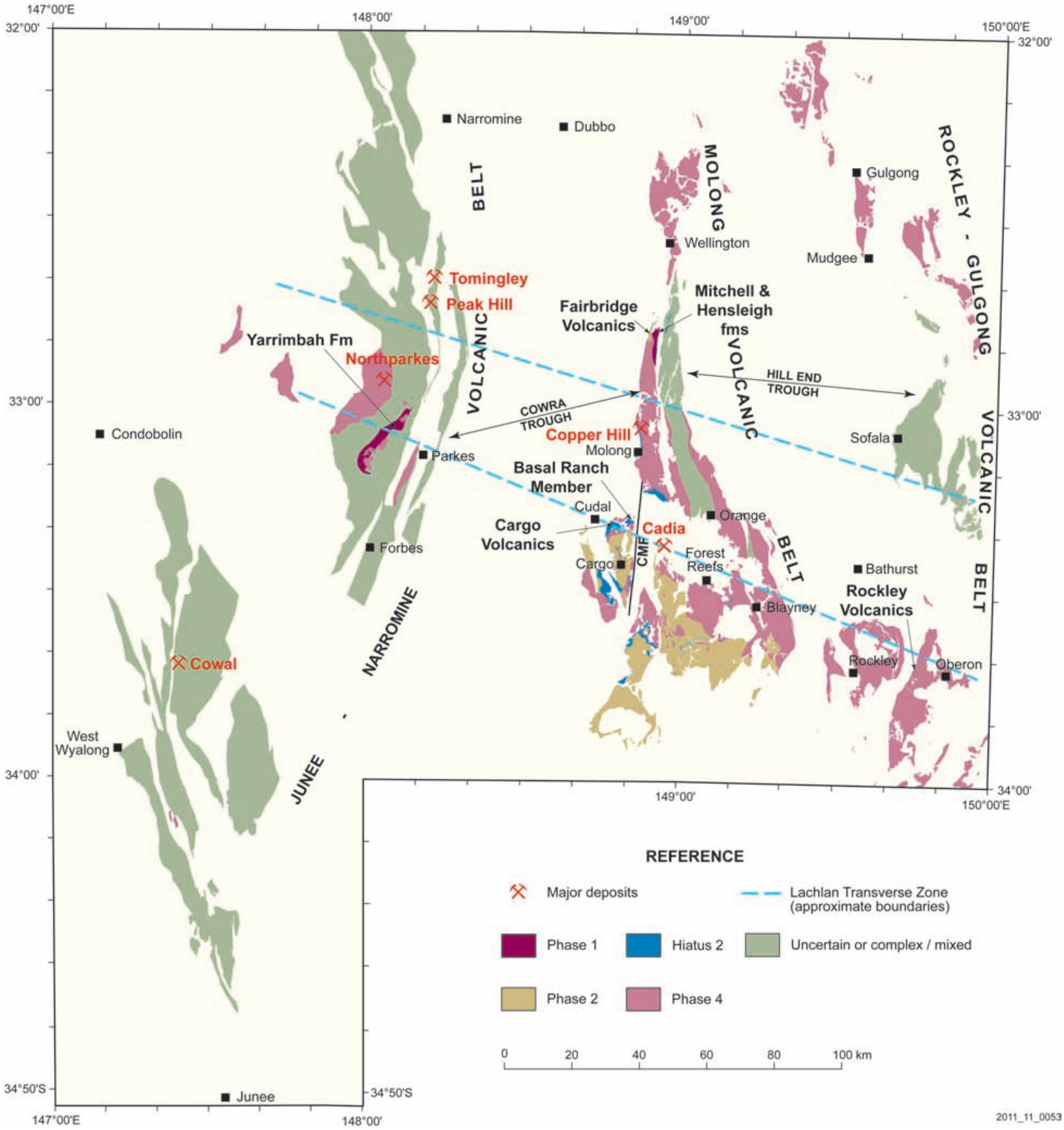



***Figure 2** Detail of three belts of Ordovician volcanic and volcaniclastic rocks, limestone and intrusions in central west NSW, showing phases where possible. Separation by Silurian–Middle Devonian rift basins reflects extensional rifting of the arc during these periods. Modified from Percival and Glen (2007).*

downthrown along reactivated extensional faults and underlie Silurian–Devonian basins (Glen et al., 2002).

## Arc evolution

Using detailed stratigraphy, Percival and Glen (2007) showed that evolution of the Macquarie Arc in central New South Wales involved three phases of volcanism, common to each of the preserved belts between >480 and c. 443 Ma (Figures 2 and 3). Relics of the first Early Ordovician phase occur in the two western belts. The minimum age of Phase 1 magmatism is constrained by overlying Bendigonian graptolites and conodonts (c. 480–476 Ma) and an intruding 481 Ma monzonite (I. Williams, pers. comm; Simpson et al., 2005; Butera et al., 2001). No definitive magmatic zircons have yet been recovered from Phase 1 volcanic rocks, and zircon populations from volcaniclastic rocks show dominantly inherited signatures (Glen et al., 2011). Evolved high-K calc-alkaline to shoshonitic andesitic to basalt-andesitic to basaltic lavas (Glen et al., 2007c) and volcaniclastics developed on emergent islands fringed by shallow-water carbonates and passing up into deepwater siltstones that reflect subsidence.

Phase 1 was followed by a volcanic hiatus of c. 8 Myr, c.474–466 Ma (late Bendigonian–early Chewtonian to early Darriwilian). Towards the end of this hiatus, some Early Ordovician rocks were uplifted, eroded, and redeposited into sediments associated with Phase 2 arc-wide magmatism. The best example is the clasts of Early Ordovician limestone in the lower Phase 2 volcanics in the northern Molong Volcanic Belt (Percival and Glen, 2007).

Phase 2 magmatism lasted from c. 466–455 Ma (Dariwillian–Eastonian), when it was overlain by an Eastonian 1–3 (455–450 Ma) carbonate platform that marks the second hiatus in arc evolution. Onset of Phase 2 magmatism is not well-defined, and is largely based on ages of allochthonous limestone bodies (Percival and Glen, 2007). Magmatic zircons from two volcaniclastic units provide mean ages of volcanism of 464–463 Ma, which are generally Dariwillian–Gisbornian (Glen et al., 2011). Volcanism during this second magmatic phase was largely submarine. However, in the northern Molong Volcanic Belt, locally emergent volcanic centres were flanked by shallow-water carbonate deposition rapidly passing offshore into deep-water environments marked by mass-flow and turbidite deposits (Simpson et al., 2005, 2007). Intrusive rocks of Phase 2 include 466 and 465 Ma monzodiorites and granodiorites in the Junee-Narromine Volcanic Belt.

Rocks with MORB–like geochemistry attributable to an ophiolite (sensu lato), occur between the Junee-Narromine and Molong volcanic belts (Ashley et al., 1979; Lyons and Percival, 2002) (Figure 2), where they are represented by the Coolac Serpentinite and redeposited fragments in the Silurian Jindalee Group near Cootamundra and Tumut (Quinn et al., unpublished data). Chert fragments yield conodonts from latest Middle Dariwillian (Da4) –earliest Late Ordovician (Gisbornian1, Gi1) – and may be restricted to Gi1. This age is slightly earlier than, or coincident with the onset of Phase 2 arc magmatism (Quinn and Glen, 2009; Quinn and Percival, 2010; Quinn et al., unpublished data).

Crawford et al. (2007b) recorded a stratigraphically upwards shift in geochemistry during Phase 2, from relatively primitive to medium-K calc alkaline to high-K and shoshonitic compositions.

A carbonate platform overlies Phase 2 rocks in the Junee-Narromine Volcanic Belt and western parts of the Molong Volcanic Belt, and marks the second (c. 455–450 Ma) hiatus in arc magmatism (Figure 3). During hiatus 2, (deepwater) volcanic rocks were uplifted, eroded and overlain by a stable carbonate platform that led to deposition of >360 m of limestone (Percival and Glen, 2007). Dating of magmatic zircons at the base of the hiatus suggests input not only from underlying Phase 2 volcanics, but also from Cambrian primitive mafic melts (Simpson et al., 2007; Glen et al., 2011).

Phase 3 of the arc is represented by the Copper Hill Suite of dacitic and diorite intrusives. At Copper Hill itself, Phase 3 dacite was emplaced during the second arc hiatus, with the best radiometric age being 450 Ma (D. Wyborn, Geoscience Australia OZCHRON database). Crawford et al. (2007b) suggested that the transient Copper Hill Suite marked a brief reversion to medium K-magmatic compositions that interrupted the increasingly shoshonitic trend from underlying to overlying lavas. Crawford et al. (2007b) drew attention to a spread of ages of phase 3 magmatic zircons; Glen et al. (2011) wondered whether some of these reflected older arc magmatism, a concept first suggested by Crawford et al. (2007b).

Phase 4 of arc magmatism (Crawford et al., 2007a) comprises widespread extrusive Late Ordovician (Bolindian, c. 449–443 Ma) lavas and associated sedimentary rocks that postdate the carbonate platform of hiatus 2 as well as richly-mineralised c. 440-437 Ma monzonitic to monzodioritic intrusions coeval with Llandovery sedimentation. Emplacement of these porphyries postdates deformation of Late Ordovician volcaniclastic units and is inferred to postdate the c. 443 Ma first stage of amalgamation of the arc with flanking sedimentary terranes (Glen et al., 2007b).

The onset of Phase 4 volcanism is dated by graptolites as early Bolindian (Percival and Glen, 2007), and occurred after an inferred rifting event marked by sudden depocentre deepening at the end of hiatus 2. This deepening is reflected in collapse of the carbonate platform, carving of submarine valleys, formation of mass flow bodies with clasts of underlying limestone, and deposition of deepwater graptolitic turbidites. Percival and Glen (2007) suggested that the onset of Phase 4 volcanism was diachronous (younging from W to E). Lava from the Rockley-Gulgong Volcanic Belt contains a zircon magmatic age of c. 454 Ma, just below the cessation of the carbonate platform deposition of hiatus 2 in the western Macquarie Arc.

This model of rift-generated uplift and subsidence in hiatus 2 to early Phase 4 replaces the model of Glen et al. (1998), wherein these features were attributed to the attempted subduction of a buoyant seamount.

Phase 4 intrusives are dated at c. 440–437 Ma (Glen et al., 2011) (Figure 3). According to Crawford et al. (2007b), Phase 4 magmatism was almost all shoshonitic, being dominated by more evolved compositions than the shoshonites erupted late in Phase 2. Crawford et al. (2007b) suggested that the more evolved trachyandesitic and trachytic lavas reflected the passing of these magmas through a thicker crust, thereby allowing more time for cooling and fractionation than experienced by phases 1 and 2 lavas.

The Bushman and Nash Hill Volcanics in the Junee-Narromine Volcanic Belt, and the Fifield Alaskan-type zoned complexes intrude Ordovician turbidites W of that belt. They are regarded as Llandovery, postdating the main development of the Macquarie Arc, and potentially overlapping with Phase 4 intrusives (Crawford et al., 2007b).

## Geochemical constraints on the Macquarie Arc

The lack of continental detritus (Glen et al., 1998; Crawford et al., 2007b; Meffre et al., 2007), the primitive Pb isotopes (Carr et al., 1995; Forster et al., 2011) and the positive whole rock εNd of lavas and intrusive rocks (Wyborn and Sun, 1993; Crawford et al., 2007b) are all consistent with formation of the Macquarie Arc as an intraoceanic arc built on primitive oceanic crust. Detrital zircon data from volcaniclastic rocks suggests input from the older Gondwana margin in Phase 1, but only to minor extent in phases 2 and 4. A key point is the similarity of ages of detrital zircons in Phase 1 volcaniclastics to those in sandstones in the flanking turbidites, suggesting an original position relatively close to the Gondwana margin (Glen et al., 2011).

## Deformation

All volcanic belts are cut by faults ranging from early Silurian–Carboniferous. Internal fault deformation and cleavage strain vary from intense to mild. The Junee-Narromine Volcanic Belt and the Molong Volcanic Belt are separated from Silurian–Devonian rocks by thrust faults. The Junee-Narromine Volcanic Belt is sigmoidal, reflecting control by the NNW-trending transpressional Gilmore Fault Zone in the W and unnamed blind fault in the NE. The west-dipping Gilmore Fault Zone separates the western Junee-Narromine Volcanic

Belt from coeval Ordovician turbidites to the west, with considerable underthrusting of the volcanics inferred from deep seismic reflection profiling (Glen et al., 2002). It passes northwards, via short unnamed buried faults into the blind Tullamore Fault Zone (Glen et al., 2002).

Despite this deformation, arc rocks are commonly characterised by low to moderate dips and thus low total strain, although localised high-strain zones occur. For example, the Junee-Narromine Volcanic Belt contains an eastern high-strain zone (c. 10 km wide), marked by subvertical white mica (sub or low grade greenschist) cleavage(s) and internal thrusts. The southern Molong Volcanic Belt contains meridional thrusts that die out approaching a corridor of WNW-trending faults near Cadia mine, that are part of the complex 'arc-normal' Lachlan Transverse Zone of Glen and Walshe (1999) – a low-strain zone marked by the preservation of late-stage porphyries that formed at depths of 2–4 km. Magnetic fabrics also show the Lachlan Transverse Zone to be a low-strain zone (Verard and Glen, 2008). A similar low-strain region, with little post-Ordovician uplift occurs around the Northparkes Mine in the northern Junee-Narromine Volcanic Belt.

## Mineral deposits

The Macquarie Arc hosts porphyry Cu-Au, epithermal, and structurally-controlled Au-Cu deposits. It is the world's second major alkalic porphyry province, largely because of the Au-rich porphyry deposits associated with alkalic intrusions in the Cadia district in the Molong Volcanic Belt (Cooke et al., 2004, 2007). The combined resources from Cadia make it the sixth largest Au-rich porphyry system on Earth (Cooke et al., 2005), with Ridgeway and Cadia East (underground) among the world's highest grade porphyry Au resources. Mines (Figure 2) include those of the Cadia Valley, those in the Northparkes area (n in Figure 3) and the Cowal mine. Significant resources also occur at Copper Hill (Figure 2), at Tomingley and at the high-sulfidation epithermal deposit at Peak Hill Mine (Figure 2).

The metallogeny of the Macquarie Arc is dominated by alkalic and calc-alkalic porphyry Au-Cu, skarn, high sulfidation Au-(Cu), and carbonate base metal epithermal Au deposits The array of deposit types is consistent with subduction-related metallogenic belts in Cenozoic island arcs (e.g., PNG, Indonesia, Philippines). The richest mineralisation is linked to oxidised, small volume, evolved shoshonitic magmatic systems in which Au, chalcophile metals, Cl, and importantly, $H_2O$, are strongly concentrated relative to precursor mafic and intermediate magmas (Blevin, 2002; Holliday et al., 2002; Lickfold et al., 2007; Cooke et al., 2007). The most obvious marker for such potentially mineralised magmas is their association with large volumes of broadly comagmatic shoshonitic lavas.

Glen et al. (2007c) divided porphyries into four age groups, all of which may be mineralised. Porphyry groups 1-3 are 'pre-accretionary'. They formed during critical events in the evolution of the arc related to interruptions and resumptions of arc activity: group 1 at the end of Phase 1, group 2 at the beginning of Phase 2 and group 3 (Copper Hill suite) during the Late Ordovician hiatus in volcanism. Group 4 porphyries, in contrast, were emplaced between c. 440–437 Ma, in the Llandovery and are 'syn-accretionary'.

Group 4 porphyries are the most highly mineralised. They include the porphyry Cu-Au deposits of the Northparkes area in the Junee-Narromine Volcanic Belt, and those in the Cadia Valley area in the southern part of the Molong Volcanic Belt. In both areas, mineralisation is centred in and around quartz monzonite porphyries. These composite intrusive complexes comprise pipes, dykes and stocks. Hydrothermal alteration in and around the intrusions produced a complex sequence of alteration assemblages, ranging from potassic, calc-potassic, sodic, propylitic and late-stage, typically fault- and fracture-controlled phyllic assemblages. Hematite dusting is a common alteration product. Several of the deposits have bornite-rich cores that grade outwards through chalcopyrite-dominant domains to an outer pyritic halo. Gold is well correlated with bornite in most deposits.

Group 4 porphyries were emplaced into folded and tilted volcaniclastic and volcanic rocks that were deformed in the first phase of arc deformation (Benambran Orogeny phase 1, c. 443 Ma). At Cadia, porphyry emplacement was synchronous with Llandovery rift basins during an extension or relaxation event. These basins were deformed in the late Llandovery (435–430 Ma), in the second phase of the Benambran Orogeny. The WNW-trending Lachlan Transverse Zone (Figure 2) is a major arc-normal corridor and structural direction that favoured emplacement of many porphyries, commonly, but not exclusively shoshonitic. Other WNW-trending faults occur at Tomingley (Alkane Exploration data) and near Cowal Mine, which occurs near the intersection of a N-trending, high-strain corridor.

## Plate tectonic setting and key issues

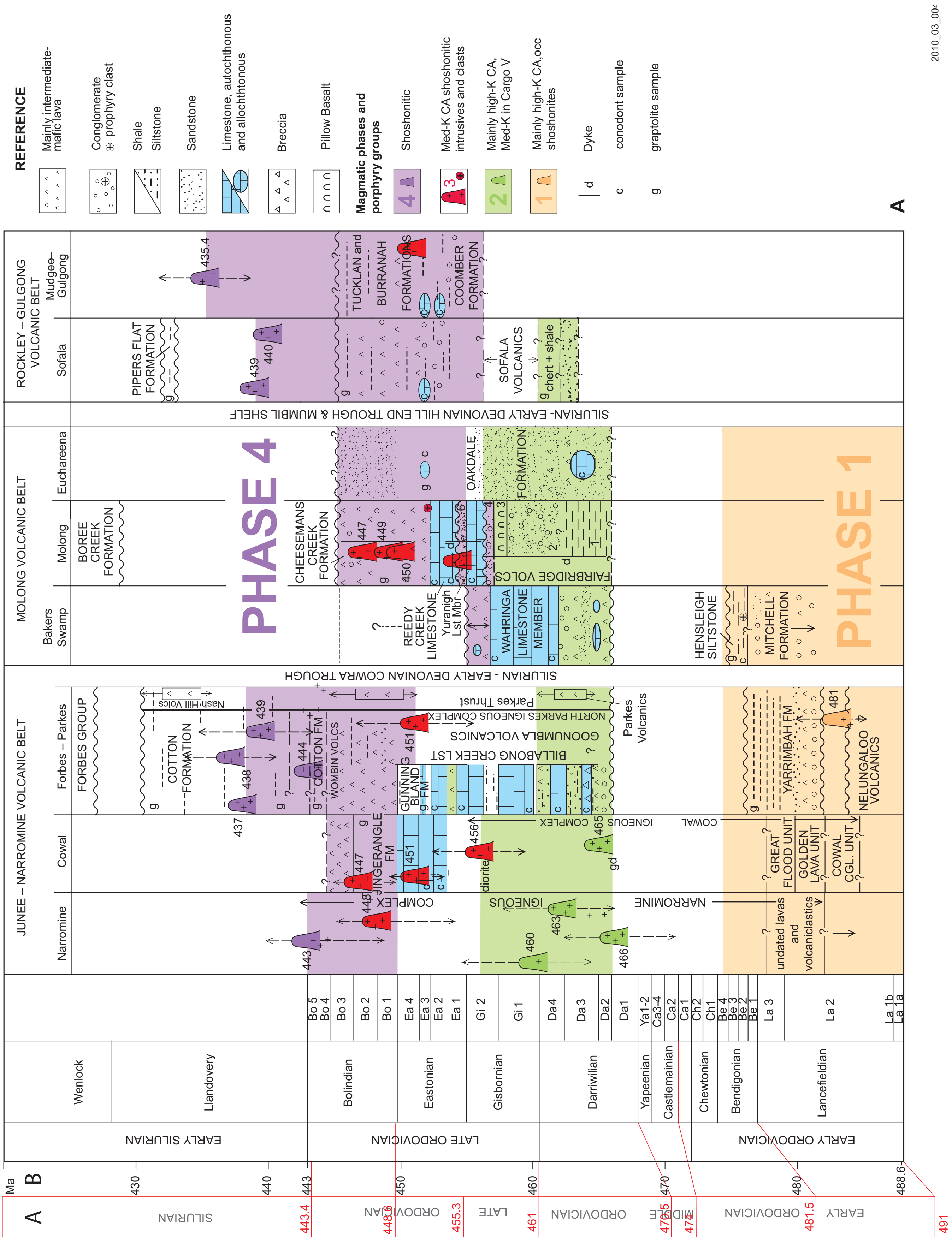
Glen et al. (2009, 2011) indicated several key features highlighting differences between the Macquarie Arc and 'standard' accreted intraoceanic island arcs. These complicate understanding of the tectonic setting and paleogeography of this part of the Ordovician east Gondwana margin. These are now briefly discussed, based on published work. Undoubtedly new mapping will help fill in the holes in existing data and allow the consideration of new tectonic models to explain the enigmatic Macquarie Arc.

### *Stacking of arc phases*

The long duration of the Macquarie Arc, c. 37 Myr of volcanism, separated by two hiatuses, plus up to 6 Myr of late intrusive activity contrast with most modern intraoceanic arcs. The vertical stacking of the arc phases requires consistent tracking of the subduction zone with no significant change in dip or direction after each hiatus. Older phases could represent relics of systems that have undergone intra-arc rifting and which formed basement for younger arc phases, although in this case, geochemistry is relatively constant throughout. Rifted extinct arcs have not as yet been found.

### *Relation to flanking sedimentary rocks*

The Macquarie Arc is flanked to the E and W by Early–Middle Ordovician turbidites containing two extensive chert horizons (Figure 2). Except in the NW, these turbidites are overlain by Late Ordovician black shales. Sandstones in the turbidite package are craton-derived (up to 90% quartz, 5% detrital plagioclase and mica.). Although Powell (1983) suggested that detrital feldspar and mica in Ordovician sedimentary rock were derived from the arc, this has been replaced by consensus that they were derived from plutonic/metamorphic sources. The absence of provenance mixing between turbidite and arc packages (Glen and Wyborn, 1997; Colquhoun et al., 1999; Meffre et al., 2007) is a key feature in


REFERENCE
Mainly intermediate-mafic lava
Conglomerate ⊕ prophyry clast
Shale
Siltstone
Sandstone
Limestone, autochthonous and allochthonous
Breccia
Pillow Basalt
Magmatic phases and porphyry groups
Shoshonitic
Med-K CA shoshonitic intrusives and clasts
Mainly high-K CA, Med-K in Cargo V
Mainly high-K CA,occ shoshonites
Dyke
conodont sample
graptolite sample
2010_03_004
A
ROCKLEY – GULGONG VOLCANIC BELT
Sofala
Mudgee–Gulgong
PIPERS FLAT FORMATION
TUCKLAN and BURRANAH FORMATIONS
COOMBER FORMATION
SOFALA VOLCANICS
chert + shale
SILURIAN- EARLY DEVONIAN HILL END TROUGH & MUMBIL SHELF
MOLONG VOLCANIC BELT
Bakers Swamp
Molong
Euchareena
BOREE CREEK FORMATION
PHASE 4
CHEESEMANS CREEK FORMATION
OAKDALE FORMATION
REEDY CREEK LIMESTONE
Yuranigh Lst Mbr
WAHRINGA LIMESTONE MEMBER
FAIRBRIDGE VOLCS
HENSLEIGH SILTSTONE
MITCHELL FORMATION
PHASE 1
SILURIAN - EARLY DEVONIAN COWRA TROUGH
JUNEE – NARROMINE VOLCANIC BELT
Narromine
Cowal
Forbes – Parkes
FORBES GROUP
COTTON FORMATION
Nash Hill Volcs
Parkes Thrust
NORTH PARKES IGNEOUS COMPLEX
GOONUMBLA VOLCANICS
BILLABONG CREEK LST
Parkes Volcanics
OSBORNE FM
WOMBIN VOLCS
GUNNINGBLAND FM
JINGERANGLE FM
COWAL IGNEOUS COMPLEX
NARROMINE IGNEOUS COMPLEX
diorite
gd
GREAT FLOOD UNIT
GOLDEN LAVA UNIT
COWAL CGL. UNIT
undated lavas and volcaniclastics
YARRIMBAH FM
NELUNGALOO VOLCANICS
Wenlock
Llandovery
Bolindian
Eastonian
Gisbornian
Darriwilian
Yapeenian
Castlemainian
Chewtonian
Bendigonian
Lancefieldian
EARLY SILURIAN
LATE ORDOVICIAN
EARLY ORDOVICIAN
Bo 5
Bo 4
Bo 3
Bo 2
Bo 1
Ea 4
Ea 3
Ea 2
Ea 1
Gi 2
Gi 1
Da 4
Da 3
Da2
Da1
Ya1-2
Ca3-4
Ca2
Ca1
Ch2
Ch1
Be 4
Be 3
Be 2
Be 1
La 3
La 2
La 1b
La 1a
Ma
B
430
440
443
450
460
470
480
488.6
A
SILURIAN
LATE ORDOVICIAN
MIDDLE ORDOVICIAN
EARLY ORDOVICIAN
443.4
448.6
455.3
461
470.5
474
481.5
491

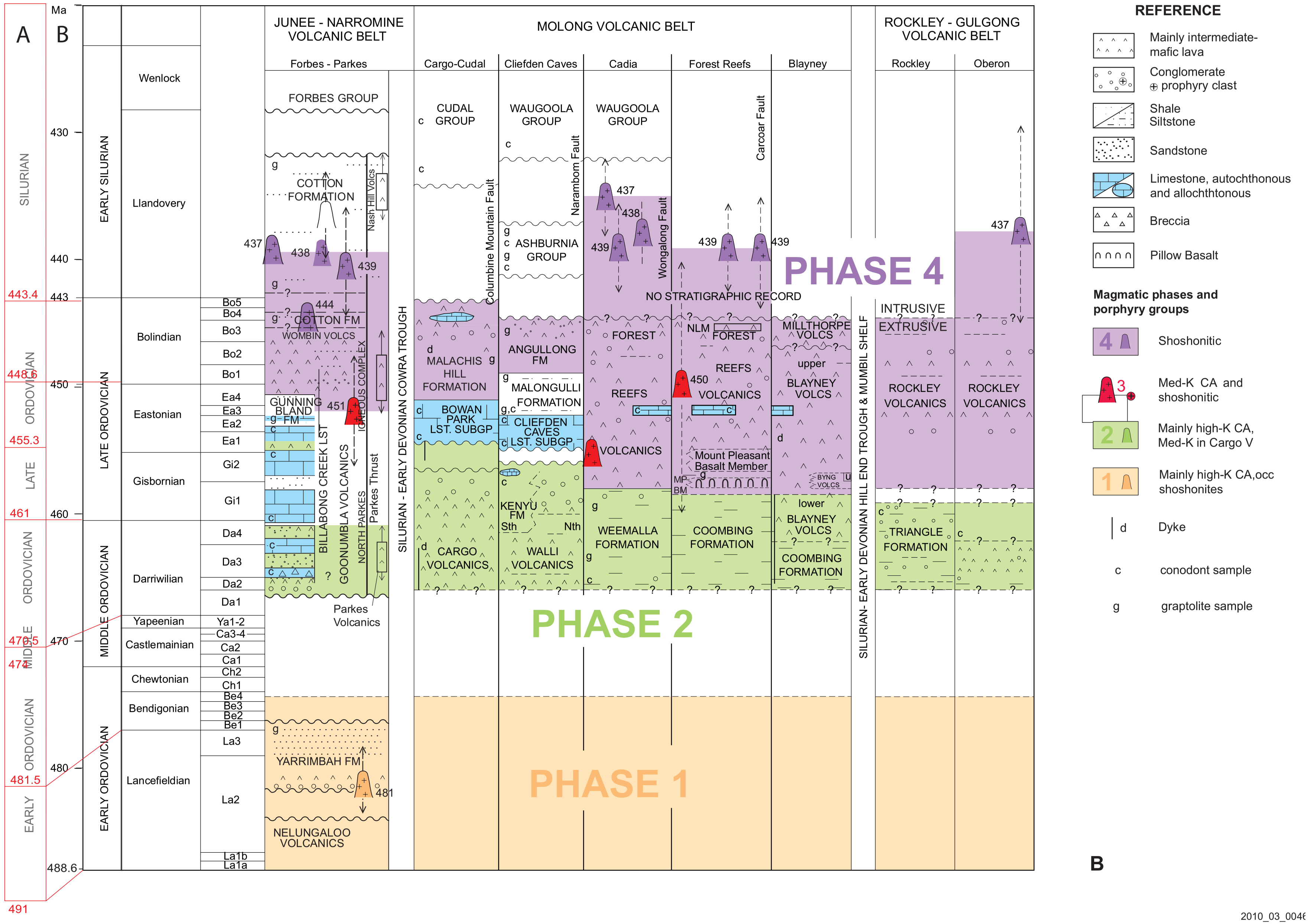



*Figure 3 Time space plots through the Macquarie Arc. (A). northern part of the arc; (B). southern part of the arc. Updated from Percival and Glen (2007). The time scale used for the original figure, that of Webby et al. (2004) shown in column B, has been augmented by the new time scale of Sadler et al. (2009) shown in column A, with the main differences occurring in the Early Ordovician.*

erecting paleogeographic models. As a result, the Lachlan Orogen lacks any forearc basin that can be related to the arc by provenance.

West of the arc, the arc-turbidite contact has been seismically imaged as the Gilmore Fault Zone (Glen et al., 2002). Its extension to the N comprises short unnamed buried faults that pass northwards into the blind Tullamore Fault System (Glen et al., 2007e). East of the arc there is no clearly defined fault zone and earlier workers suggested concordant or conformable contacts for the Kiandra Volcanic Belt (Owen and Wyborn, 1979) and for the northern Rocky-Gulgong Volcanic Belt (Colquhoun et al., 1999). In the southern Molong Volcanic Belt, suggestions of a 35 km long regional folded latitudinal fault between volcaniclastic arc rocks in the N and quartz-rich turbidites in the S (Glen, 1998) were based on the absence of provenance-mixing coupled with changes in regional trends (Glen and Wyborn, 1997) and seismic refraction studies (Glen et al., 2002). The progressive realisation that the Early and Middle Ordovician turbidites overlain by Late Ordovician black shales E of the arc are similar to those to the W of the arc led to the suggestion of duplication of the backarc package. By 2009, detailed biostratigraphic work showed that the Early and Middle Ordovician rocks E and W of the arc are identical (Glen et al., 2009), although the Hermidale Terrane, NW of the arc, lacked Late Ordovician black shales that typified the Albury-Bega Terrane E and W of the arc farther S.

Modifications to this strike-slip model are necessary as detailed remapping around the Kiandra Volcanic belt has shown that the contact between arc volcanics and turbidites is locally concordant, yet disturbed by major soft-sediment deformation, with limited provenance mixing (Quinn and Glen, 2009; Quinn et al., unpublished data; modifying Owen and Wyborn, 1979). These new data suggest an alternative model in which the sedimentary terranes E and W of the Macquarie Arc had rifted apart (perhaps obliquely) during Phase 2 of the arc. This rifting was synchronous with earliest Gisbornian abyssal cherts associated with the ophiolite between the eastern and western belts (Quinn and Glen, 2009; Quinn and Percival, 2010; Quinn et al., unpublished data).

### *Which plate was the Macquarie Arc on?*

In the absence of either forearc basins or subduction complexes, problems arise when trying to deduce which plate hosted the Macquarie Arc and the vector of any related subduction zone. Location on the Gondwana plate above a W(continent)-dipping subduction zone is inferred from the coincidence of changes in arc evolution (e.g. hiatuses) with changes in sedimentation patterns in flanking sediments (Glen, 2005; Glen et al., 2007d, 2009; Quinn and Glen, 2009; Quinn and Percival, 2010; Quinn et al., unpublished data). How long W-dipping subduction persisted into the Late Ordovician is a matter for discussion. Crawford et al. (2007c) and Squire and Crawford (2007) used a Fiji-type model to query whether the shoshonitic Late Ordovician Phase 4 of the arc was subduction-related or represented a change to rifting. Meffre et al. (2007) invoked a possible short period of E-dipping subduction to account for formation of the Copper Hill suite of intrusions.

Uplift, collapse and rifting preceding Phase 2 and Phase 4 magmatism have been referred to above. Linkages between flanking sediments and the Macquarie Arc show that late Middle Ordovician lithological changes immediately preceding Phase 2 arc magmatism were coeval with ophiolite development in a supra-subduction zone rift or marginal basin (Quinn and Percival, 2010; Quinn et al., unpublished data; after Lyons and Percival, 2002).

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*by Gideon Rosenbaum*

# Oroclines of the southern New England Orogen, eastern Australia

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*A series of tight bends (oroclines) are recognised in the late Paleozoic–early Mesozoic southern New England Orogen between Brisbane and Newcastle, but their exact geometry and tectonic evolution are still debated. This contribution provides an outline of the different tectonic elements within the orogen and the relationships of these elements with the oroclinal structure. Pre-oroclinal tectonic elements were derived from a Devonian–Carboniferous subduction zone, and include forearc basin terranes and accretionary wedge rocks that are separated from each other by a narrow belt of serpentinites and high-pressure rocks. Rocks are predominantly steeply dipping and their map-view pattern delineates three bends: the Z-shaped Texas and Coffs Harbour oroclines in the north and the U-shaped Manning Orocline in the south. During the early Permian (298–288 Ma), the area was affected by widespread, mostly S-type, magmatism that heralded a period of crustal extension accompanied by the formation of sedimentary rift basins. The spatial distribution of early Permian granitoids mimics the shape of the oroclines, which further defines a fourth bend, the Nambucca Orocline. Whether these curvatures formed by bending of a quasi-linear belt, or as primary early Permian arcuate features, is an unresolved question that warrants further paleomagnetic, geochronological and structural investigations.*

## Introduction

The Devonian–Early Triassic New England Orogen (NEO) in eastern Australia is the youngest orogenic component in the Australian continent. The northern part of the orogen, from northern Queensland to Brisbane (Figure 1a) is oriented NW-SE parallel to the continental margin. In contrast, the southern NEO, in the area between Brisbane and Newcastle (Figure 1a, b), is characterised by a series of sharp bends, herein referred to as the New England Oroclines.

This paper focuses on the structure and tectonics of the New England Oroclines. The exact geometry of these oroclines is controversial, and there are different interpretations to the structural grain of the orogen. The simplest structural model is of a Z-shaped double orocline, comprising the so-called Texas and Coffs Harbour oroclines (Murray et al., 1987; Offler and Foster, 2008) (Figure 1c). Evidence supporting the existence of these oroclines is the curved orientations of structural fabrics (Korsch, 1981; Lennox and Flood, 1997; Aubourg et al., 2004; Li et al., 2012) and the curvature of aeromagnetic lineaments (Figure 1c).

A different model, proposed by a number of authors (Cawood and Leitch, 1985; Korsch and Harrington, 1987; Glen, 2005; Cawood et al., 2011b), considers an additional orocline farther S, the Manning Orocline (Figure 1d). This orocline has been suggested based on (1) paleomagnetic data showing block rotations of forearc basin terranes around vertical axes (Geeve et al., 2002; Klootwijk, 2009; Cawood et al., 2011b); (2) the recognition of a contorted serpentinite belt (Korsch and Harrington, 1987); and (3) the curved structure delineated by Early Permian (298–288 Ma) granitoids (Rosenbaum, 2010; Rosenbaum et al., 2012). Nevertheless, structural information supporting the existence of this orocline is more ambiguous.

Based on the lateral continuation of early Permian granitoids, an even more complex structural model that involves four bends, has been proposed (Rosenbaum, 2010; Rosenbaum et al., 2012) (Figure 1e). The suggested fourth bend has been named the Nambucca Orocline (Rosenbaum, 2010).

The kinematics associated with the development of the New England oroclines is relatively poorly constrained, with only patchy structural, paleomagnetic and geochronological data. In this respect, 'orocline' is used here in a general sense, as referring to orogenic curvatures rather than the strict sense that implies a secondary bending of an originally linear belt (Weil and Sussman, 2004). This paper aims to highlight the structure of the oroclines by outlining the spatio-temporal distribution of different tectonic elements within the southern NEO. This will be followed by a discussion on the large-scale tectonic implications of this spectacularly contorted orogenic structure.

## Tectonic elements

### Devonian–Carboniferous subduction complex

The greater part of the exposed southern NEO is a Devonian–Carboniferous convergent margin complex (Figure 2) comprising forearc basin rocks of the Tamworth Belt and correlative terranes, and accretionary metasedimentary rocks of the Tablelands Complex. These domains are separated from each other by a tectonic contact, the Peel-Manning Fault System (PFZ and MFZ in Figure 1b), along which a lithological assemblage of serpentinites, blueschists and eclogites, is exposed. This assemblage, hereinafter referred to as the 'serpentinite belt' (Korsch and Harrington, 1987; Aitchison et al., 1994; Och et al., 2003), is most prominent along a long (c. 150 km),

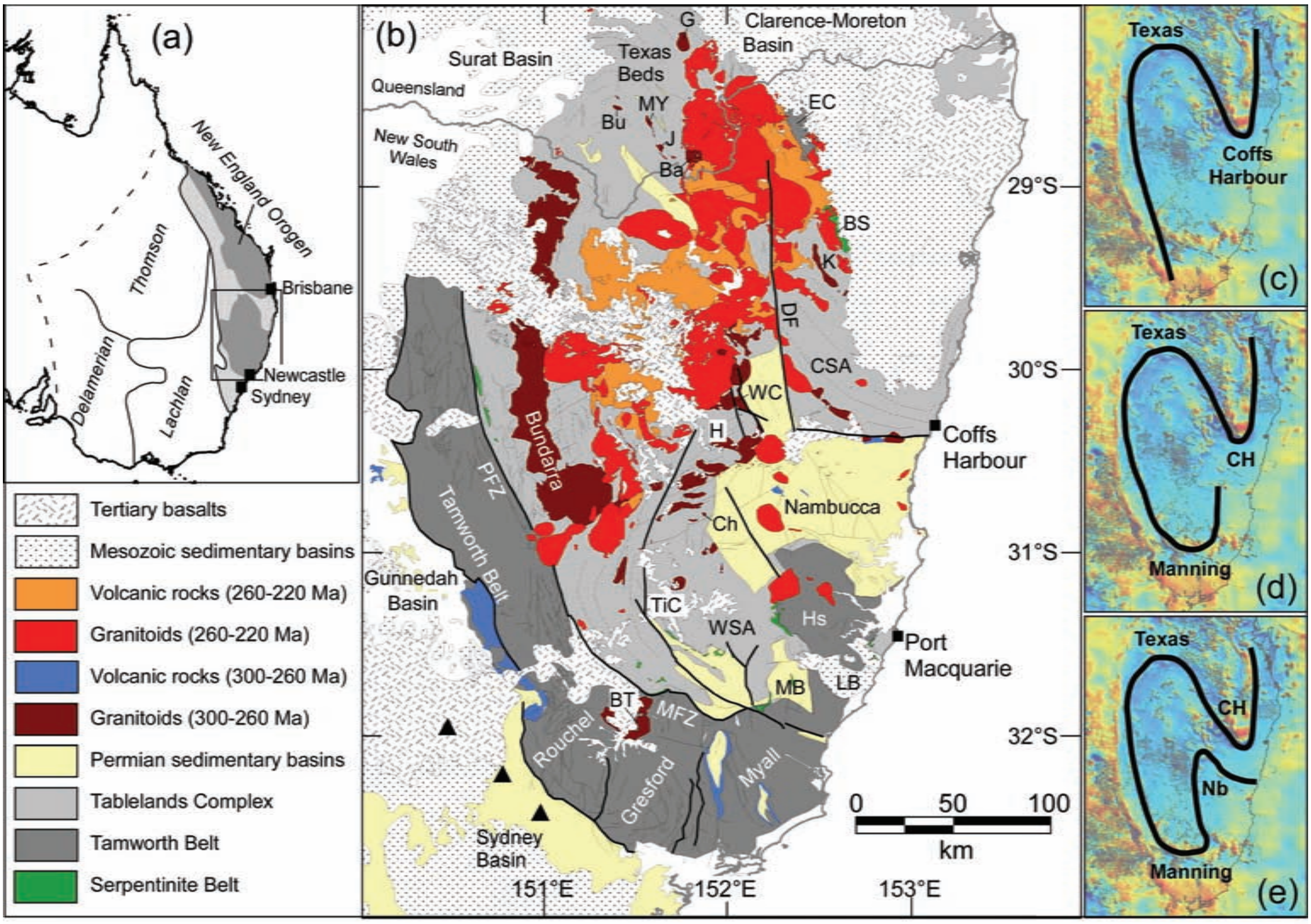


***Figure 1 (a) Location map. Dashed line = possible location of the Tasman Line (after Glen, 2005). (b) Geological map of the southern New England Orogen. Triangles = an Early Carboniferous volcanic arc (under cover). Ba, Ballandean Granite; BS, Baryulgil Serpentinite; BT, Barrington Tops Granodiorite; Bu, Bullaganang Granite; Ch, Cheyenne Complex; CSA, Coffs Harbour Association; DF, Demon Fault; EC, Emu Creek Block; Hs, Hastings Block; J, Jibbinbar Granite; K, Kaloe Granodiorite; LB, Lorne Basin; MB, Manning Basin; MFZ, Manning Fault Zone; MY, Mt You You Granite; PFZ, Peel Fault Zone; TiC, Tia Complex; WC, Wongwibinda Complex; WSA, Woolomin and Sandon Associations. (c), (d), (e) Alternative interpretations of the Texas, Coffs Harbour (CH), Manning and Nambucca (Nb) Oroclines on a background of the magnetic intensity image.***

narrow belt E of the Tamworth belt, but can also be recognised in numerous smaller outcrops in the southern part of the area, including the town of Port Macquarie (Figure 1b). The rocks are oceanic in nature, and are associated with a dismembered early Paleozoic ophiolite that records earlier (Lachlan orogeny?) subduction and accretion processes prior to its incorporation in the Devonian subduction complex (Aitchison et al., 1994; Fukui et al., 1995).

The coupled Tamworth Belt and Tablelands Complex represent a Devonian–Carboniferous W-dipping subduction zone, with a volcanic arc that existed farther W. Much of the volcanic arc is covered by younger sedimentary rocks of the Surat, Gunnedah and Sydney basins (see triangles in Figure 1b), but its existence can be inferred from facies relationships within detrital sedimentary rocks of the Tamworth Belt that show an eastward transport of arc-related detritus into the forearc basin, and a gradual change from intermediate to silicic magmatism during arc maturation in the Carboniferous (Crook, 1964; Leitch, 1974; Cawood, 1983; Morris, 1988).

Forearc basin rocks are represented in the Tamworth Belt and in a series of correlative blocks farther S (Rouchel, Gresford, Myall and Hastings; Figure 1b). The rocks are associated with sedimentation on a shelf that, in the Tamworth Belt, was gradually deepening from W to E (Crook, 1964; Roberts and Engel, 1987). The sedimentary facies in the Hastings Block has an opposite facing (i.e. deepening from E to W), and has been interpreted to be a displaced terrane of the Tamworth Belt (Korsch, 1977). Farther N, in northern New South Wales, there is another block with a comparable Carboniferous stratigraphy, the Emu Creek Block (Figure 1b), which likely represents a forearc basin terrane (Cross et al., 1987).

The Devonian–Carboniferous accretionary wedge rocks (Tablelands Complex) have been subdivided into a number of associations based on their sedimentary facies (e.g., Coffs Harbour, Woolomin and Sandon associations; Figure 1b) (Korsch, 1977). Typical lithologies include deep marine volcaniclastic turbidites, cherts and argillites, mafic volcanic rocks, and olistostromal deposits containing slabs of limestone, basalt, andesite and siltstone (Leitch and Cawood, 1980; Cawood, 1982; Fergusson, 1984). Rocks were subjected to varying degrees of metamorphism ranging from prehnite-pumpellyite/lower greenschist (e.g., Texas Beds and Coffs Harbour Association, Korsch, 1978) to amphibolite (Tia and Wongwibinda metamorphic complexes; Binns, 1966; Stephenson and Hensel, 1982;

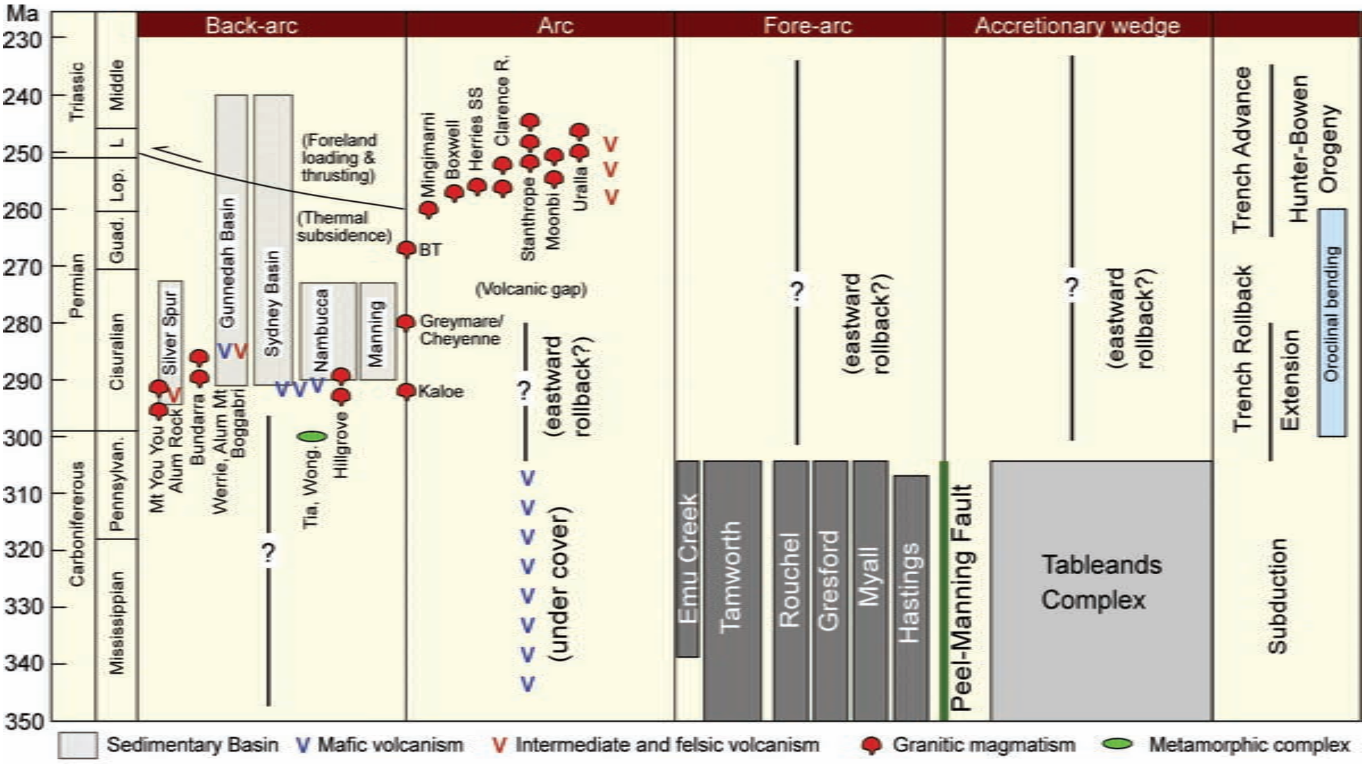


***Figure 2 Temporal relations and tectonic setting of rock units in the southern NEO (partly modified after Cawood et al., 2011b). Guad., Guadalupian ("middle Permian"), Lop., Lopingian (late Permian). BT, Barrington Tops Granodiorite; Wong. Wongwibinda Complex.***

Phillips et al., 2008; Danis et al., 2010; Craven et al., 2012) (Figure 1b). High-pressure rocks are relatively rare, but are recognised as an earlier blueschist assemblage in the Tia metamorphic complex (Phillips et al., 2008). Throughout the whole Tablelands Complex, rocks are commonly characterised by a penetrative structural fabric that most likely formed in the accretionary wedge prior to development of the oroclines (Fergusson, 1984; Murray et al., 1987).

## Early Permian rift basins and magmatism

Two major associations of early Permian (Cisuralian) rocks are found in the southern NEO. The first group is associated with magmatic rocks, predominantly S-type granitoids, which intruded the Tablelands Complex throughout the whole southern NEO (Flood and Shaw, 1977; Shaw and Flood, 1981; Hensel et al., 1985; Phillips et al., 2011; Rosenbaum et al., 2012). The second group includes clastic sedimentary rocks and mafic volcanic rocks, which were deposited in rift basins during the early Permian (Leitch, 1988; Roberts et al., 2006; Korsch et al., 2009a).

The majority of S-type granitoids belong to the Bundarra and Hillgrove suites (Figures 1b and 3b). The emplacement of these granitoids took place in a relatively short time interval, at 298–288 Ma, simultaneously with the intrusion of a number of other smaller S-type plutons in southern Queensland (Bullaganang, Mt You You, Jibbinbar and Ballandean granites) and the I-type Kaloe Granodiorite in northeastern NSW (Figure 1b) (Cawood et al., 2011a; Rosenbaum et al., 2012). The geochemistry of the Bundarra and Hillgrove suites was interpreted to indicate crustal melting in an extensional backarc setting (Jenkins et al., 2002). A backarc environment is accounted for by the fact that although these rocks are intruded into the Carboniferous accretionary wedge, by the early Permian, the subduction zone had retreated eastward bringing the accretionary wedge into a backarc position (Jenkins et al., 2002).

The intrusion of S-type granitoids during subduction rollback is also consistent with the development of coeval rift basins, which were developed throughout the whole New England Orogen in the early Permian and were accompanied by extensional deformation (Holcombe et al., 1997a; Korsch et al., 2009a). The early Permian sedimentary successions are well developed in the foreland area W of the NEO and are represented by the Sydney, Gunnedah and Bowen Basins (Totterdell et al., 2009). In the southern NEO, there are a number of structural blocks with a correlative early Permian stratigraphy (Leitch, 1988). The largest of which is the Nambucca Block, which appears in the "core" of the oroclinal structure E of the peripheral Sydney and Gunnedah basins (Figure 1b). Other early Permian sedimentary rocks are found in the Manning Block (also referred to as the Barnard Basin; Leitch, 1988) and in a number of smaller blocks in southern Queensland (Figure 1b).

Early Permian sedimentation was accompanied by bimodal magmatism (Asthana and Leitch, 1985; Jenkins et al., 2002), with volcanic material from the base of these sedimentary successions dated at 293–291 Ma (Roberts et al., 1996; Cawood et al., 2011a). These geochronological results indicate that sedimentation and volcanism in rift basins were contemporaneous with S-type magmatism, further supporting the idea that the early Permian rift basins developed in a hot backarc extensional environment, possibly linked to an eastward retreat of the subduction zone (Collins and Richards, 2008).

Following the pulse of widespread magmatism at 298–288 Ma, the southern NEO was subjected to a long period (20–25 Myr) with only scarce magmatism. S-type magmatism in the Bundarra Granite could have continued until c. 282 Ma (Phillips et al., 2011), and there are two reported c. 280 Ma ages from the small plutons of Greymare (Donchak et al., 2007) and Cheyenne Complex (Rosenbaum et al., 2012) (Figure 1b). The only known granitic magmatism in the 280–260 Ma time bracket is the I-type Barrington Tops Granodiorite (c. 267 Ma; Cawood et al., 2011a) located in the southernmost NEO (Figure 1b).

## Late Permian–Triassic magmatism

Magmatism in the southern NEO recommenced in the late Permian (Lopingian) at c. 260 Ma and involved voluminous magmatism associated with the emplacement of I-type granitoids and felsic volcanism (Shaw and Flood, 1981). This magmatic episode overlapped with a period of widespread E-W contractional deformation that affected the whole New England Orogen and is commonly referred as the Hunter-Bowen Orogeny (Collins, 1991; Holcombe et al., 1997b; Korsch et al., 2009b). Contractional deformation began 5–15 Myr prior to the onset of magmatism and continued until c. 230 Ma (Holcombe et al., 1997b). During the Hunter-Bowen Orogeny, the Sydney and Gunnedah basins, that developed as backarc basins in the early Permian, were transformed into foreland basins (Collins, 1991; Korsch et al., 2009b).

Late Permian–Triassic granitoids have been subdivided by Shaw and Flood (1981) into three major suites: (1) metaluminous and K-poor, mainly tonalite and granodiorite of the Clarence River Supersuite; (2) K-rich quartz monzonites of the Moonbi Suite; and (3) less metaluminous quartz monzonites and granodiorite of the Uralla Suite. The majority of these plutons were emplaced at 260–230 Ma (Shaw, 1994; Donchak et al., 2007), with the exception of the aforementioned I-type Barrington Tops (c. 267 Ma) and Kaloe (c. 292 Ma) Granodiorites (Cawood et al., 2011a). The geodynamic setting of the late Permian–Triassic magmatism is considered to be related to continental arc magmatism associated with the re-establishment of a W-dipping subduction zone (Bryant et al., 1997; Carr, 1998; Jenkins et al., 2002). A suite of younger (230–200 Ma) I-type granitoids are aligned N-S closer to the coast and are possibly post-orogenic (Figure 3c).

## Mesozoic sedimentary basins and Cenozoic basalts

Early–Middle Triassic sedimentary rocks were deposited in the Sydney, Gunnedah and Lorne basins, and Late Triassic–Cretaceous sequences accumulated in the Surat and Clarence-Moreton basins (Figure 1b and 2). The basal units of the latter basins unconformably overlie the rocks of the NEO and include a large volume of quartz-rich sand grains derived from the uplifted granitic basement (Donchak et al., 2007). Cenozoic volcanism in the southern NEO is predominantly basaltic. An earlier stage of volcanism, which took place mainly in the Eocene–Oligocene, resulted in extensive lava fields (e.g., Sutherland and Fanning, 2001). A series of younger (25–16 Ma) central volcanoes are generally younging from N to S (Vasconcelos et al., 2008), and are possibly marking a hotspot track that was superimposed on the northward moving Australian plate during the past 35 Myrs (Knesel et al., 2008).

# The structure of the oroclines

## Evidence from pre-Permian rocks

The curved structure of the Texas and Coffs Harbour oroclines is evident from variations in the orientations of the dominant steeply

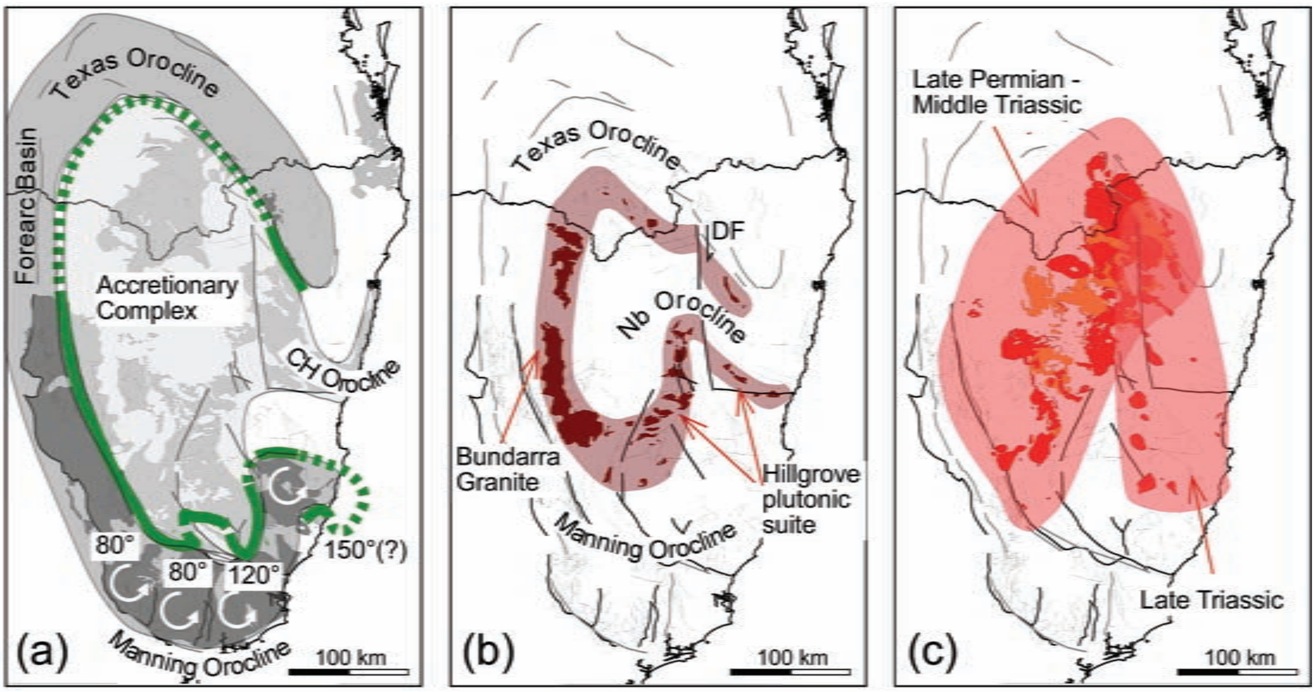


*Figure 3 Spatial distribution of (a) pre-oroclinal elements of the Devonian to Carboniferous subduction complex; (b) early Permian (298–288 Ma) pre- to syn-oroclinal magmatic belt (Rosenbaum et al., 2012); (c) post-oroclinal late Permian–Triassic magmatism. The background maps show lineaments inferred from aeromagnetic images and major faults. The inferred continuation of the contorted serpentinite belt is shown in the green line. Counterclockwise block rotations in (a) are after Geeve et al. (2002) and Klootwijk (2009). CH, Coffs Harbour; Nb, Nambucca. Dextral displacement along the Demon Fault (DF) in b) is after McPhie and Fergusson (1983).*

dipping structural fabric, which shows a map-scale curvature that delineates the oroclinal structure (Korsch, 1981; Lennox and Flood, 1997; Li et al., 2012). The structural fabric and fold plunges are steep or sub-vertical (Figure 4a), indicating that the curvatures recognised in a plan view are true bends around vertical axes.

The continuation of the Tamworth Belt around the Texas and Coffs Harbour oroclines is obscured by younger sedimentary rocks, but its existence under cover is supported by geophysical evidence (Wartenberg et al., 2003). Moreover, the appearance of the Emu Creek Block in the eastern limb of the Texas Orocline (Figure 1b) may represent the continuation of this belt (Figure 3a). Similarly, the position of the Baryulgil Serpentinite (Figure 1b) is consistent with the continuation of the serpentinite belt along the eastern limb of the Texas Orocline (Korsch and Harrington, 1987).

Farther S, the apparent arrangement of forearc basin terranes around the Manning Orocline, and the curvature of the Serpentinite Belt (Figure 3a), support the existence of this orocline (Korsch and Harrington, 1987). Constraints from paleomagnetic studies are limited and incomplete (Cawood et al., 2011b), but the available data are generally consistent with oroclinal bending, indicating counterclockwise rotations of the Rouchel (80°), Gresford (80°) and Myall (120°) Blocks during the latest Carboniferous or early Permian (Geeve et al., 2002). Paleomagnetic data from the Hastings Block, however, was interpreted to indicate a clockwise rotation of 130° around a vertical axis (Schmidt et al., 1994), which is not compatible with the expected rotations during oroclinal bending. The validity of this clockwise rotation, however, is debated (Klootwijk, 2009). An alternative interpretation that is consistent with oroclinal bending, is based on a comparison of the Namurian paleopoles from the Hastings Block and the northern Tamworth Belt, and is indicative of 150° counterclockwise rotations of the former relative to the latter (Klootwijk, 2009).

Further support for oroclinal bending in the Manning Orocline has been presented by Cawood et al. (2011b), who tested the agreement of available paleomagnetic data with two alternative tectonic models. The result of this analysis indicated that the previously suggested tectonic models, accounting only for the development of the Texas and Coffs Harbour oroclines by dextral strike-skip faulting (e.g., Offler and Foster, 2008), are not permissible. In contrast, a model involving a northward translation of forearc basin terranes during oroclinal bending in the Manning Orocline is consistent with the paleomagnetic data (Cawood et al., 2011b).

Notwithstanding the supporting evidence for oroclinal bending in the Manning Orocline, its structure remains elusive and

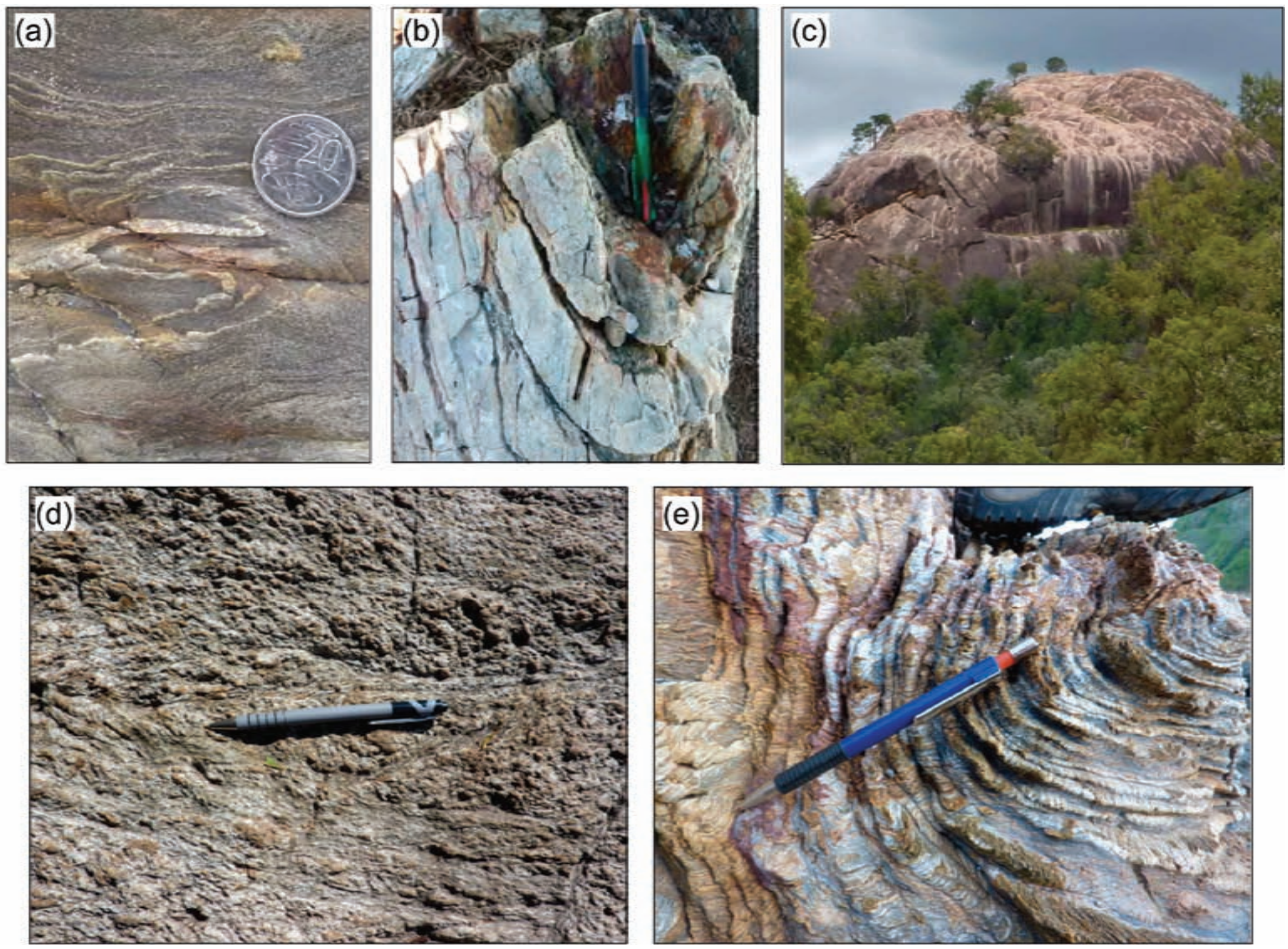


***Figure 4** a) Pre-oroclinal steeply dipping structural fabric in the Texas Beds and associated isoclinal folds; b) An isoclinal, steeply plunging, N-S fold in the hinge of the Manning Orocline; c) The Bundarra Granite, dated c. 289±3 Ma; d) a deformed granite of the Hillgrove plutonic suite (Henry River Granite) dated 293±5 (Rosenbaum et al., 2012); e) Strong ductile fabric and overprinting relationships in the early Permian Nambucca Block. Crenulation cleavage ($S_3$) with $S_2$ quartz-mica domains in the microlithons and a mesoscopic $F_4$ fold.*

controversial. One of the major reasons for this controversy is that, unlike the Texas and Coffs Harbour oroclines, the orientations of bedding and dominant secondary foliations, are not curved around the orocline, but are predominantly oriented c. N-S, roughly parallel to tectonic contacts and the axial planes of regional and minor folds (Figure 4b) (Collins, 1991). One explanation for that is that pre-oroclinal structural features were transposed to c. N-S orientation by late- or post-oroclinal deformation.

Deformation and metamorphism in the southernmost NEO is indeed much more intense in comparison to the relatively low strain deformation around the Texas and Coffs Harbour oroclines. In two localities within this area (Tia and Wongwibinda complexes, Figure 1), the originally accretionary wedge rocks were affected by multi-phase ductile deformation and high-grade (amphibolite) metamorphism (Stephenson and Hensel, 1982; Dirks et al., 1992; Landenberger et al., 1995; Phillips et al., 2008; Danis et al., 2010; Craven et al., 2011). Whether this deformation and metamorphism predated oroclinal bending, or occurred simultaneously with the development of the Manning Orocline, is an open question. Subsequent deformation associated with the Hunter-Bowen orogeny was responsible for further folding and thrusting along N-S structures, accompanied by sinistral strike-slip faulting (Collins, 1991).

## Evidence from Permian rocks

Recent U-Pb geochronology from early Permian rocks in the southern NEO has considerably improved our understanding of the oroclinal structure (Rosenbaum et al., 2012). In particular, the recognition that S-type granite emplacement occurred contemporaneously in the Bundarra and Hillgrove suites (Figure 4c, d), led Rosenbaum et al. (2012) to conclude that the lateral continuation of these plutons defines the hinge of the Manning Orocline (Figure 3b). Similarly, early Permian (298–288 Ma) granitoids are apparently curved around the Texas Orocline, and could possibly define a fourth bend (Nambucca Orocline; Figure 3b).

The remarkable curvature of the early Permian granitoids raises a major question with regards to the timing of oroclinal bending. If granite emplacement occurred at 298–288 Ma in a quasi-linear belt that was subsequently subjected to bending, then the timing of oroclinal bending must be younger than c. 288 Ma. However, paleomagnetic data seem to suggest that block rotations in the Manning Orocline took place prior to the Asselian (299–295 Ma) (Geeve et al., 2002). It is therefore possible that the curved shape of the early Permian Hillgrove and Bundarra Suites is at least partly a primary curvature rather than the result of oroclinal bending. In other words, S-type magmatism may have occurred in an arcuate fashion that mimicked the shape of the earlier curvature. The model proposed by Rosenbaum et al. (2012) considers a combination of a primary or progressive curvature, which was amplified and refolded by subsequent deformation. Evidence for such multi-phase deformation history is recorded in the superposition of strong ductile fabrics within the early Permian sedimentary rocks of the Nambucca Block (Figure 4e).

In contrast with the early Permian granitoids, which clearly mimic the shape of the oroclines, the spatial distribution of late Permian–Triassic granitoids is distributed along a broad NE-SW field that crosscut the oroclinal structure (Figure 3c). This phase of magmatism is interpreted to postdate oroclinal bending, thus providing a minimum age constraint of c. 260 Ma for their formation.

## Tectonic implications and concluding remarks

The southern NEO seems to represent one of the most contorted orogens in the world, comprising of three or four tight bends (Figure 1c, d, e and Figure 3). The oroclinal structure is supported by structural analysis (Korsch, 1981; Lennox and Flood, 1997; Li et al., 2012), available paleomagnetic data (Cawood et al., 2011b) and geochronological results (Cawood et al., 2011a; Rosenbaum et al., 2012), but all these datasets are still incomplete and require further investigations. In particular, structural information from the hinge zones of the Manning and Nambucca oroclines is relatively scarce, with existing data indicating complex deformational history involving high-grade metamorphism and multiple generations of folding (Dirks et al., 1992; Danis et al., 2010). The link between the development of these metamorphic complexes (Tia and Wongwibinda complexes, Figure 1b) and the oroclinal structure is still unclear.

The formation of the oroclines occurred in the time bracket of 300–260 Ma, i.e., in the early–middle Permian (Cisuralian–Guadalupian; Figure 2). Based on paleomagnetic constraints (Geeve et al., 2002; Klootwijk, 2009), it seems that the formation of the Manning Orocline initiated relatively early (c. 300–295 Ma). The Texas Orocline also records an earlier stage (>290 Ma) of oroclinal bending, which was followed by a second stage of deformation that took place after the deposition of early Permian sedimentary rocks in rift basins (Aubourg et al., 2004; Li et al., 2012).

Perhaps the most intriguing observation is the link between the first stage of oroclinal bending, the development of widespread extensional rift basins and the emplacement of S-type magmatism. This may suggest that during the early Permian, the whole southern NEO was in a backarc position relative to a retreating subduction zone. In such tectonic environments, orogenic curvature could be achieved by a combination of primary subduction curvatures and secondary oroclinal bending, triggered by variations along strike in rollback velocities and heterogeneous subduction (Rosenbaum and Mo, 2011). Whether this style of tectonic activity is applicable to the development of the southern NEO is a major question that should be addressed in future studies.

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*by Ron F. Berry[1] and Stuart W. Bull[2]*

# The pre-Carboniferous geology of Tasmania

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*Tasmania evolved over 900 Myr as a small fragment on the margin of Gondwana. There are three episodes of passive margin sedimentation from the Mesoproterozoic to the Devonian. Tasmania rifted from the Antarctic margin as part of Rodinia breakup and was then involved in "West-Pacific-Style" tectonics along the eastern margin of Gondwana through the Paleozoic. Arc-continent collision led to ophiolite obduction in the Cambrian. Much of this geological history can be recognised from the excellent coastal exposures along the north coast of Tasmania.*

## Introduction

This paper provides a broad overview of the complex geological history of Tasmania, with particular emphasis on the north coast (Figure 1). The paper relies upon supporting data that cannot be included in this short review and the reader should refer to Burrett and Martin (1989) for a pre-1998 view of the geological history. More recent papers on the geology of Tasmania are Black et al. (2004, 2005), Calver and Walter (2000), Meffre et al. (2000) and Berry et al. (2007) and a major detailed review of the geology is about to become available (Corbett et al., 2012). Biostratigraphic correlations for the Cambrian follow the review in Figure 3 of Corbett (2002). The latest review of the tectonics is in Cayley (2011).

Tasmanian is usually separated into a Western Tasmanian Terrane (WTT) and an Eastern Tasmanian Terrane (ETT). The Nd isotopes reported from the WTT indicate that the lower crust was extracted from the mantle in the late Paleoproterozoic (c. 1700 Ma; Berry et al., 2008). Only Paleozoic rocks are exposed in the ETT and the deep basement is probably late Neoproterozoic–Cambrian ocean crust (Black et al., 2010).

## Tectonic Overview

Australia and most of East Antarctica were combined into a continental fragment by 1100 Ma (Betts and Giles, 2006), which from 1100–780 Ma formed part of Rodinia (Li et al., 2008). The Neoproterozoic Australia/Antarctic continental fragment retained the same geometry until the Mesozoic (Veevers, 2000) and this geometry is well constrained from modern ocean structure. A major continent, possibly Laurentia or South China, rifted away from East Gondwana at c. 780 Ma (Li et al., 2008; Wingate et al., 2002). However, the separation of the WTT from the Australian craton probably occurred during a second rift phase at 580 Ma (Calver and Walter, 2000; Direen and Crawford, 2003; Meffre et al., 2004).

The position of Tasmania in Cambrian reconstructions remains controversial. Detailed Mesozoic reconstructions (Royer and Rollett, 1997) put Tasmania outboard of the Ross Orogen (Figure 2). Paleomagnetic data support the interpretation that Tasmania was near this position by the late Cambrian (Li et al., 1997). Despite this position as an isolated microcontinent E of the Ross-Delamerian Orogen, the Western Tasmanian Terrane shares many features of its Neoproterozoic depositional history with the Adelaide Fold Belt (Calver and Walter, 2000). Berry et al. (2008) concluded that the Western Tasmanian Terrane rifted away from the Transantarctic Mountains at c. 580 Ma (Figure 3). It remained isolated from Antarctica and from the whole evolving Gondwanaland margin from at least 540 Ma until the end of the Cambrian (Figure 4).

At the end of the early Cambrian, the WTT collided with an oceanic island arc (Figure 4b; Berry and Crawford, 1988; Crawford and Berry, 1992) and Tasmania was accreted back onto the craton margin in the middle to late Cambrian. The formation of an Ordovician arc to the E of Tasmania (Figure 4) trapped a section of ?Cambrian oceanic crust which formed the rigid lower crust for the Selwyn Block (Cayley, 2011) including NE Tasmania and the Melbourne Trough.

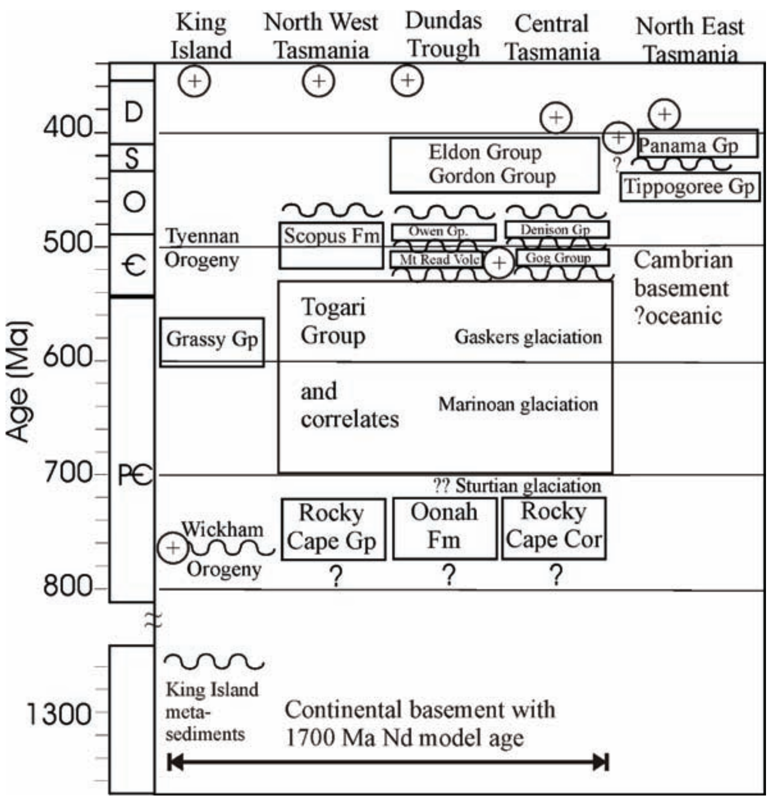


***Figure 1 Space time chart for Tasmania.***

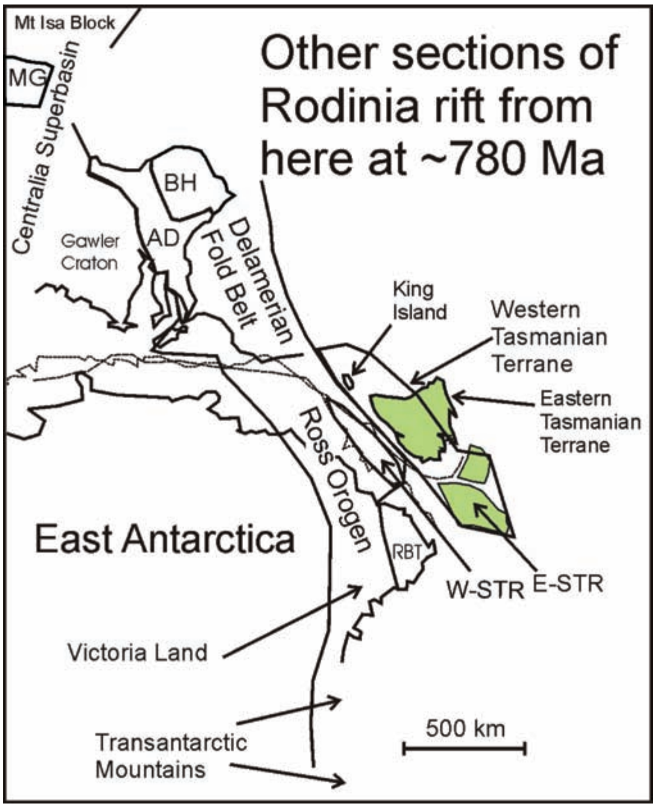


***Figure 2 Tasmanian position outboard of the Ross Delamerian Orogeny based on Mesozoic reconstruction of Royer and Rollett (1997).***

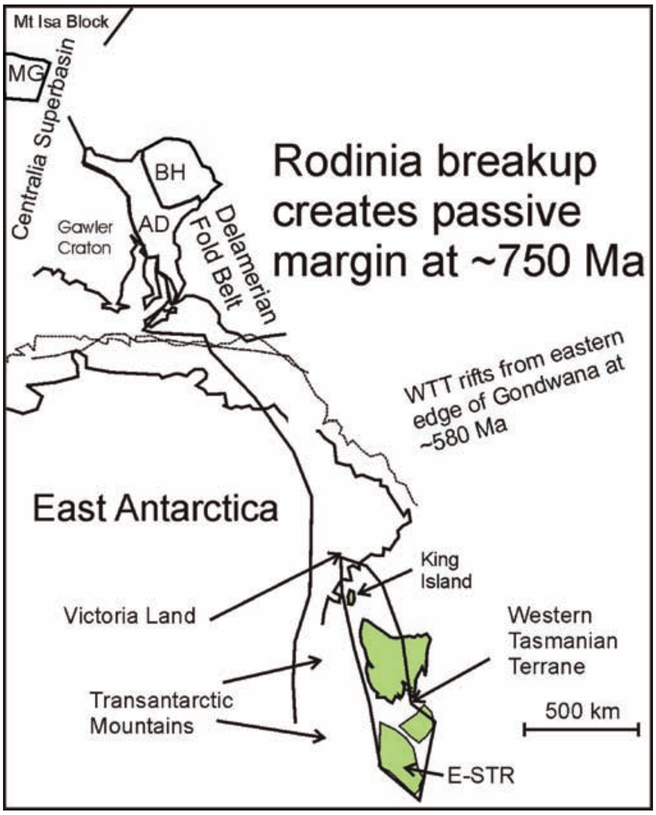


***Figure 3 Possible position for Tasmania near the Transantarctic Mountains before rifting at 580 Ma.***

The Paleozoic history of this block is tied to the evolution of the Tasman Fold Belt with a final orogenic event in the Devonian.

## Mesoproterozoic

The oldest rocks of the WTT are found on King Island. They are low amphibolite grade metasedimentary rocks and minor amphibolites. Chemical U-Th-Pb dating of monazite indicates a metamorphic event at 1290 Ma (Berry et al., 2005), whereas detrital zircon dating shows that these rocks were deposited after 1350 Ma (Black et al., 2004). Several types of mafic rocks intrude this succession (Cox, 1989). The oldest are discontinuous, concordant to subconcordant bodies of metamorphosed low K tholeiites emplaced prior to $D_1$. Polyphase deformation has affected the Mesoproterozoic metasediments. However only the first major deformation phase ($D_1$), tight folds with a penetrative axial surface cleavage defined by muscovite, is considered to have formed at 1290 Ma. This event is very restricted in distribution and its significance is still debatable. There are no obvious correlates in SE Australia.

Neoproterozoic (760–750 Ma; Black et al., 1997) biotite granites intrude the King Island succession (Cox, 1989; Turner et al., 1998). The granites have strongly deformed aureoles but are not linked to any larger scale deformation features. This event is known as the Wickham Orogeny. Holm et al. (2003) argued the King Island granitoids are related to rifting and the breakup of Rodinia.

## Neoproterozoic sedimentary rocks

This summary of the Neoproterozoic sedimentary record largely follows Calver et al. (2012a). The early Neoproterozoic stratigraphy of NW Tasmania (Figure 1) is dominated by shallow water siliciclastics (Rocky Cape Group in the NW, and correlates in the S and E) and turbidites (Oonah Formation in the N). The age of these units is poorly constrained. All have similar detrital zircon patterns (Black et al., 1997), with the youngest zircons dated at 1000 Ma. Calver and Walter (2000) argued that parts of the unconformably overlying succession are 750 Ma old. The Oonah Formation contains dolerite sills (Cooee Dolerite) that have a 725 ±35 Ma minimum age (Crook 1979). Calver et al. (2012a) argue this sequence is older than 760 Ma but no Neoproterozoic granites have been found in the widely exposed units. It is possible this sequence was deposited after the Wickham Orogeny and before 700 Ma.

The Rocky Cape Group is a thick sequence of siliciclastic sedimentary rocks deposited in a marine shelf (passive margin) environment (Calver et al., 2012a). The Pedder River Siltstone, Balfour Subgroup and Irby Siltstone were deposited in an open shelf environment between storm wave base and fair-weather wave base. They are dominated by mudstone and siltstone-sandstone. The sandstones are commonly wave rippled and cross-lamination and locally gutter casts and swaley and hummocky cross-stratification have been identified. Small synsedimentary clastic dykes are common in many pelite beds that probably formed as ‘diastasis cracks’. The Cowrie Siltstone and Emmetts Creek Shale are predominantly fine-grained and plane-laminated, indicating deposition in a low-energy, offshore shelf environment, shallowing to storm wave base at times. Supermature quartz arenites of the Lagoon River Quartzite, Detention Subgroup and Jacob Quartzite probably formed in a shallow marine, tide-dominated setting. Fining-upward cycles in the Lagoon River Quartzite may represent prograding tidal flats. Carbonate is rare in

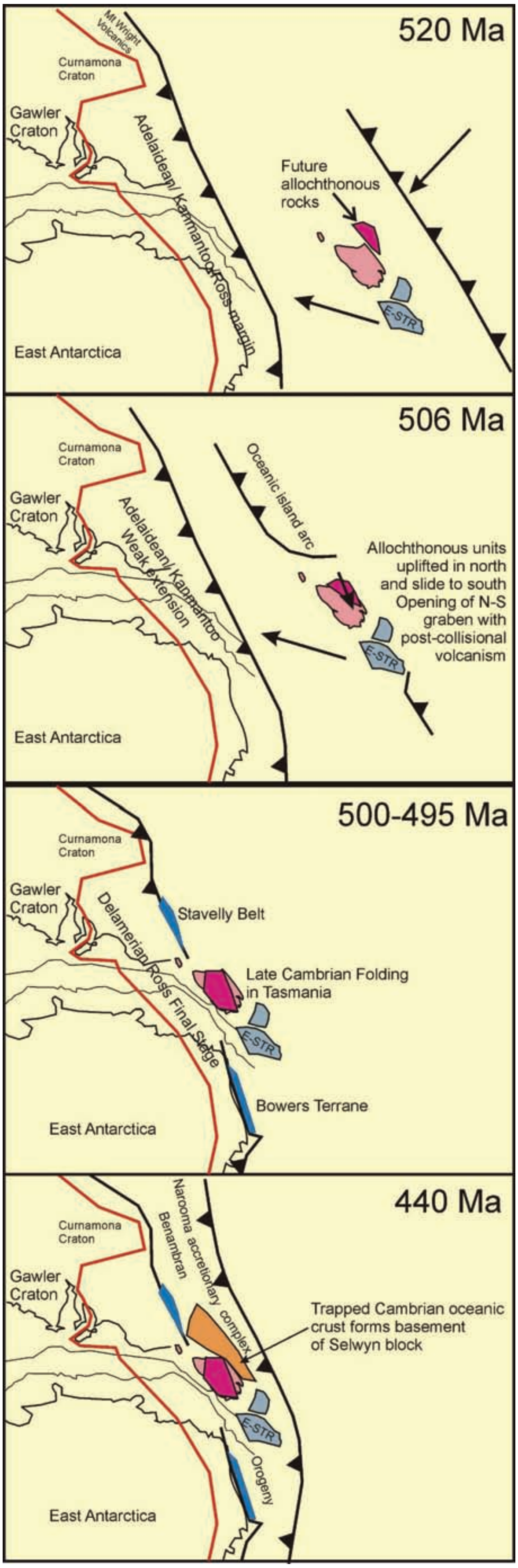


***Figure 4 Cartoon showing tectonic evolution of Tasmania, 520–440 Ma. Modified from Cayley (2011).***

the Rocky Cape Group, but local oolites and stromatolites, in combination with wave ripple cross lamination indicate a shallow marine depositional setting. Carbonate units are more common in equivalent sequences in SE Tasmania. All of these rocks have been metamorphosed to prehnite-pumpellyite grade (Chester 2006).

The Oonah Formation and correlates make up a number of inliers between the Rocky Cape and Tyennan regions and form the eastern part of the Arthur Metamorphic Complex. There is no known stratigraphic contact with the Rocky Cape Group or any of its correlates. The succession is prehnite-pumpellyite to low greenschist facies. Two lithological associations are included in the Oonah Formation.

Sedimentary structures in the dominant fine grained quartzwacke assemblage are typical of sandy turbidites (grading, cross-lamination, parallel lamination, convolute lamination, intraformational slumps, flute marks, load casts and rip-up clasts). At Sulphur Creek, coastal exposures thought to be stratigraphically high in the Oonah Formation display low-angle scours and possible hummocky cross-stratification, suggesting a shallowing of paleoenvironment to above storm wave-base. Sandy beds at or near the top of the Oonah Formation are particularly rich in coarse detrital muscovite (Turner, 1989). The turbidite package is intruded by the Cooee Dolerite and includes minor mafic pillow lavas. The second, relatively rare, lithological association is pelite and carbonate, with rare mafic volcanics and conglomerate. This association has been interpreted as a shallow water upper subdivision of the Oonah Formation (Brown, 1986).

The Togari Group (and correlates) forms the next phase of deposition. The basal units of the Togari Group rest on different stratigraphic levels of the Rocky Cape Group basement. The contact is a low angle (c. 20º) to disconformable surface (Brown, 1989). The metamorphic grades range up to low greenschist facies. The type section in NW Tasmania (Everard et al., 2007) is divided into four main lithostratigraphic units: (i) a lower dolomitic succession with basal siliciclastics and a diamictite near the top (?Marinoan glaciation), which represents a widespread phase of (c. 700–600 Ma) shallow marine shelf sedimentation; (ii) a phase of mafic rift volcanism (c. 600–570 Ma); (iii) shallow-marine carbonates (c. 570–545 Ma), and (iv) early Cambrian deep-water siliciclastics.

Stage (i) comprises a basal siliciclastic unit, the Forest Conglomerate, of variable thickness (0–120 m), followed by the Black River Dolomite (Brown, 1989) which consists of up to 800 m of interbedded dolostone, black shale and chert with a diamictite near the top. Calver (1998) correlated the stage (i) dolomitic unit with the Sturtian glaciation but, based on other constraints and the most recent dating for Neoproterozoic glaciations (Macdonald et al., 2010), a Marinoan age is assumed for the diamictite. The only direct numerical age constraint is a Re-Os date of 641 ± 5 Ma for black shale from the top of the Black River Dolomite (Kendall et al., 2009). Thus the ages of these sedimentary units remain contentious.

The second cycle of deposition, the Kanunnah Subgroup, is a 1.2 km thick succession of marine sedimentary rocks (siltstone, mudstone, volcaniclastic sandstone and diamictite) and basaltic lavas. On King Island, a ?glacial diamictite occurs near the base of this sequence (Direen and Jago, 2008; Hoffman et al., 2009). A rhyodacite lava in NW Tasmania extruded at 582 ± 4 Ma (Calver et al., 2004), and other slightly younger magmatic dates have been obtained on King Island (Meffre et al., 2004). This age suggests correlation of the diamictite with the Gaskiers glaciation (Macdonald et al.2010) but Hoffman et al.(2009) preferred a correlation with the Marinoan glaciation. The tholeiitic basalts, dolerite dykes and minor intrusions in this sequence are related to the 580 Ma rifting phase along the Gondwana margin.

The Kanunnah Subgroup is conformably overlain by a 1.5 km thick sequence of dolostone, the Smithton Dolomite. Strontium isotope chemostratigraphy suggests deposition at c. 570–545 Ma, consistent with the ages of units above and below. The whole sequence was deposited in shallow-marine, warm-water, occasionally evaporitic paleoenvironments (Calver et al., 2012a).

At the top of the Togari Group, the early Cambrian Salmon River Siltstone outcrops in the southern part of the Smithton syncline, where it is up to c. 350 m thick. The unit is composed of pale to dark grey, siliceous, thin-bedded siltstone. The contact with the Smithton Dolomite is not exposed but it appears to be conformable (Everard et al., 2007). By contrast, it probably has an unconformable contact with the overlying middle to late Cambrian Scopus Formation (Jago and Bentley, 2007).

## Tyennan Orogeny Stage 1: Ophiolite emplacement

The Tyennan Orogeny in Tasmania is a complex event with rapidly changing stress patterns. Other than the forearc rocks, remnants of which survive today as widespread Cambrian mafic-ultramafic complexes, some Proterozoic rocks were also probably obducted at this time. These components include blueschist and eclogite facies metamorphic rocks that are thought to have undergone partial subduction during the collision event, followed by exhumation to high crustal level after slab breakoff. Although probably derived from the outer parts of the same passive margin as the autochthonous and parautochthonous rocks, these units are difficult to correlate, in part because of their high strain. The only allochthonous element specifically identified by Berry and Crawford (1988) was the mafic/ultramafic complexes. Other possible allochthonous elements identified since are the Forth Metamorphic Complex, Badger Head Metamorphic Complex, Port Davey Metamorphic Complex, Franklin Metamorphic Complex, Mersey River Metamorphic Complex and the Arthur Metamorphic Complex (Meffre et al., 2000), and the Wings Sandstone (Black et al., 2004). These blocks are scattered across Tasmania lying structurally above the late Neoproterozoic rift facies and are unconformable overlain by middle Cambrian and younger sedimentary rocks. The Arthur Lineament (Figure 1) forms the western limit to allochthonous blocks in Tasmania and appears to mark the maximum extent of the thrust complex. The early part of the thrust emplacement is recorded in high temperature mylonites that indicate thrusting towards the Swest. A major phase of thrusting to the S is recorded in the greenschist facies metamorphic rocks and widespread cataclasite in western Tasmania (Holm and Berry, 2002).

There are, a large number of less deformed Neoproterozoic to early Cambrian rocks loosely associated with the mafic/ultramafic complexes that are interpreted to be parautochthonous to allochthonous fault blocks. Examples along the northern Tasmanian coast are the Motton Spilite and Barrington Chert (Calver and Everard, 2012). The Barrington Chert is low in terrigenous input and is probably of oceanic origin whereas the Motton Spilite consists of pillowed and massive metabasalt and minor associated breccia of possible MORB (Mid Oceanic Ridge Basalt) affinity.

The passive margin sequence was strongly deformed and metamorphosed during this early stage of the Tyennan Orogeny. In the past these rocks have been subdivided into medium grade metamorphic rocks, high strain low grade metamorphic rocks, low strain low grade metasedimentary rocks and mélange (e.g., Turner, 1989). The high strain rocks of medium and low grade are commonly intimately interleaved by faulting and these were grouped together as metamorphic complexes by Meffre et al. (2000). Mélange units are poorly exposed (Seymour and Calver, 1995).

It is possible to subdivide the WTT into four domains (Figure 5). NW Tasmania, W of the Arthur Lineament is largely outside the Tyennan Orogen with only weak deformation except near the Arthur Lineament. Passive margin deposition continued through the middle Cambrian. The ophiolite emplacement can be recognized in this area by a progressive increase in detrital chromite, sourced from the ultramafic bodies, in the Cambrian sedimentary record.

The second domain is the External Zone (Foreland) in which relatively intact ophiolites sit on top of a very low grade continental shelf sequence dominated by thin-skinned deformation. The external zone (Figure 5) occupies the area from the Arthur Metamorphic Complex to the edge of the Tyennan Complex (or the Forth Metamorphic Complex in the N). The highest metamorphic grades are exposed along the western margin of the zone. Marginal blueschist facies rocks (Turner and Bottrill, 2001) are strongly overprinted by greenschist facies metamorphism. The mylonites at the base of the ophiolites indicate transport to the SW but all other structures indicate transport to the S. The ophiolite is unconformably overlain by middle Cambrian (506 Ma) sedimentary rocks (Turner and Bottrill, 2001).

The third domain is the Internal Zone (Figure 5), which is dominated by strongly folded metasedimentary rocks of low to medium grade which are thrust into a complex structural sequence lacking recognisable systematic zonation. In most parts of the Tyennan and Forth regions the earliest deformation phase ($D_1$) produced isoclinal to tight folds and commonly a bedding-parallel foliation. The second phase ($D_2$) produced isoclinal folds in some areas and open folds in other areas (Turner, 1989). The differentiated crenulation cleavage associated with $D_2$ is the dominant surface in most pelitic rocks. Mylonitic fabrics and the common occurrence of faults which are subparallel to the axial surfaces of isoclinal to tight folds, or which form the boundaries of major lithological units, indicate that widespread low angle faulting occurred in the early deformation. The juxtaposition of metamorphic blocks drawn from different crustal levels in the orogenic pile almost certainly occurred during $D_{1\text{-}2}$ as part of the arc-continent collision. On structural and metamorphic grounds it is possible to separate the Tyennan region into three distinct subdomains (Figure 5):

i) a northern subdomain dominated by E-W fold trends;
ii) a SW subdomain characterised by fault bounded slices of medium and low grade metamorphic rocks;
iii) a SE subdomain of low grade rocks dominated by N-S folds.

The Eastern Province (Figure 5) of thin skinned deformation and very low-grade metamorphism is the fourth domain. The northern section, near Beaconsfield, has higher ductile strain and more similarities with the Internal Zone but is included here on the basis of the presence of ophiolites, a small fault slice of marginal blueschist facies rocks and the limited distribution of medium grade rocks. The southern area is very similar to the External Zone. In Oman, low

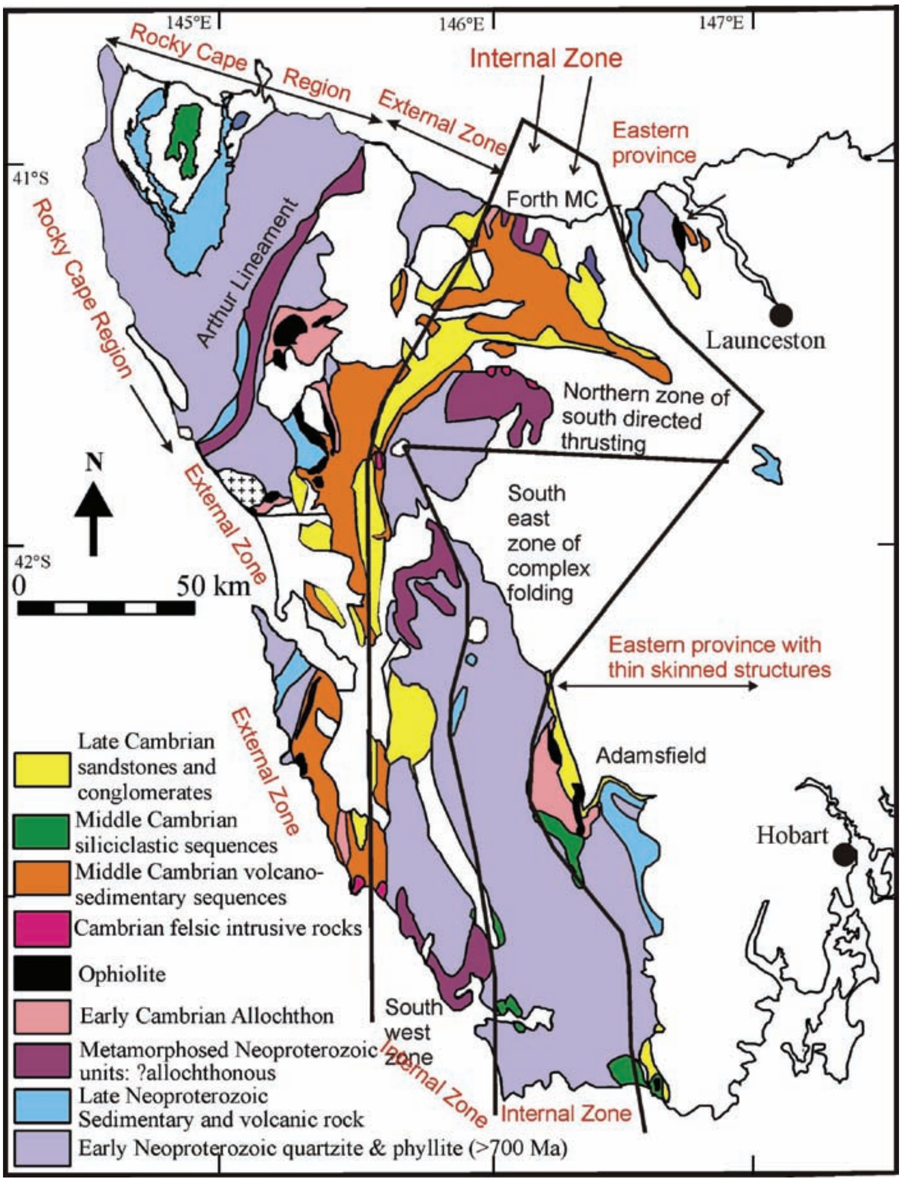


***Figure 5** Geology of the Western Tasmania Terrane relevant to the Tyennan Orogeny (simplified from Berry, 2012).*

grade areas E of the Internal Zone are interpreted to be the equivalents of the External Zone, which were trapped behind the rapidly uplifting metamorphic rocks of the Internal Zone after slab breakoff (Searle et al., 2004) and this interpretation matches the few constraints known in Tasmania.

All indications are that the first stage of the Tyennan Orogeny was very short. The ophiolite contains a gabbro that intruded at 514±5 Ma (Black et al., 1997). The age of the highest grade metamorphism is defined by U-Pb zircon and chemical U-Th-Pb monazite dating with the best estimate for the age of peak metamorphism at c. 511 Ma (Berry et al., 2007). The white mica ages reported by Foster et al. (2005) indicate that the Forth Metamorphic Complex cooled below the Ar/Ar white mica blocking temperature by 508 Ma. Detrital metamorphic minerals and metamorphic clasts in the post-collisional sediments suggest that the high pressure metamorphic rocks were unroofed by 506 Ma (Turner et al., 1998; Turner and Bottrill, 2001). The entire obduction stage of the Tyennan Orogeny took less than 10 Myr.

The rapid transition from high P metamorphism to exhumation indicates that collision occurred at plate tectonic speeds. The evolution is very similar to that modelled by Cloos et al. (2005) for West Irian (see also Davies, 2012). After slab breakoff, the buoyancy of the continental margin drives the metamorphosed continental margin back towards the surface and the slab falls away. The removal of the slab and rapid isostatic uplift drives extensional collapse at the surface. Lithospheric delamination can lead to an influx of asthenospheric mantle that controls the development of post-collisional volcanism. In western Tasmania, the post-collisional Mt Read Volcanics erupted into a Dundas Trough by 505 Ma and represent voluminous post-collisional volcanism (Crawford and Berry, 1992).

## Tyennan Orogeny Stage 2: Post collisional volcanism and extension

The late middle Cambrian (Guzhangian) is dominated by extension and post-collisional volcanism. The extension reached a maximum with the emplacement of the Henty Dyke Swarm towards the end of the middle Cambrian. Complex volcanic and sedimentary sequences, the Mt Read Volcanics, were deposited across much of Tasmania in the middle Cambrian (Corbett and Vicary, 2012). The main depocentre of Middle Cambrian rocks was a rift wrapping around the western and northern margins of the Tyennan Block. This zone is referred to as the Dundas and Fossey Mountain troughs (Figure 6). A new cycle of deposition (?sag phase) was coeval with the late middle Cambrian (Guzhangian) Tyndall Group. This extensional phase was very short lived (c. 5 Myr) and a new phase of tectonism began in the late Cambrian.

The rhyolitic to basaltic rocks are all marine and include coherent lavas, intrusives and volcaniclastics. The four major volcanic suites identified by Crawford et al. (1992) are:

- Suite 1: A voluminous group of transitional medium- to high-K calc-alkaline rocks of felsic to andesitic composition.
- Suite 2: Mainly hornblende phyric andesitic rocks, which are more P- and light REE- enriched than Suite 1 and have high-K calc-alkaline affinities.
- Suite 3: Basaltic to andesitic rocks with strong to extreme light REE- and P- enrichment indicating medium- to high-K calc-alkaline to shoshonitic affinities.
- Suite 4: Scattered tholeiitic basaltic and andesitic lavas and doleritic intrusives.

Crawford et al. (1992) suggested that volcanism evolved from the medium- to high-K calc-alkaline to the high-K calc-alkaline

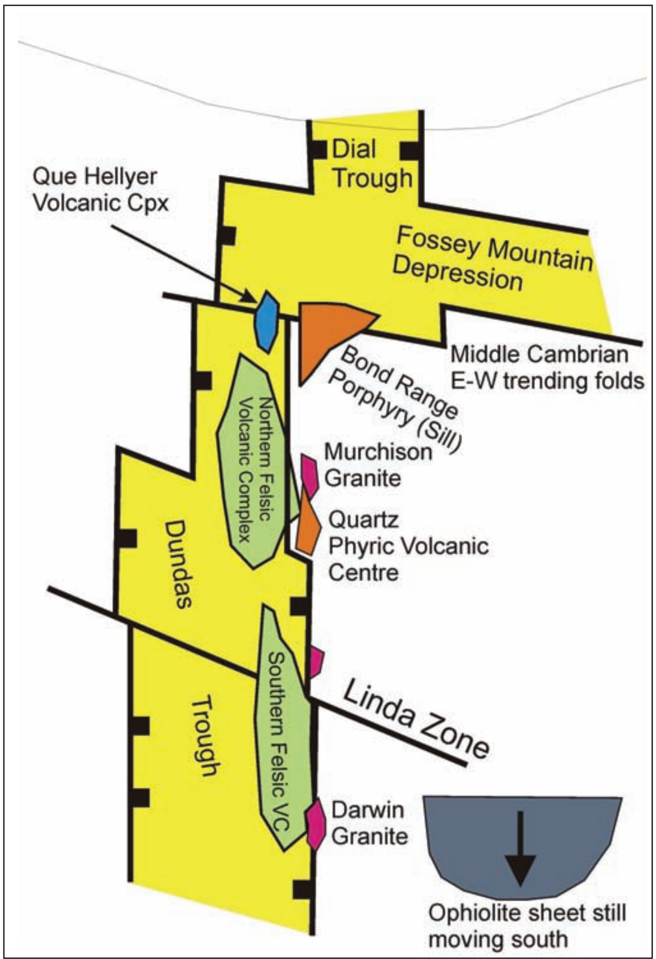


*Figure 6 Cartoon showing the extensional basin structure during the middle Cambrian (after Berry, 2012). Major volcanic complexes lie close to the eastern margin of the Dundas Trough.*

strongly light REE-enriched shoshonitic basalts, then rapidly to the Suite 4 tholeiites at maximum extension. Volcanism reverted to the Suite 1 felsic volcanism at c. 500 Ma.

East of the Dundas Trough, ?middle Cambrian generally E-W trending folding has been widely recognised. Along the southern margin of the Fossey Mountain Trough there is tight ENE folding in the volcanic sequence and more open folding in the late Cambrian Owen Group. There are sufficient unconformities exposed to indicate that these folds were initially formed in the Cambrian and have been tightened during the Devonian. The E-W folding is probably a continuation of the tight E-W folding in the Tyennan block, with Cambrian N-S compression of the area E of the Dundas Trough continuing to push the ophiolite sheet to the S in the Adamsfield area (Figure 6), while the Dundas Trough was already filling with post-collisional volcanic rocks 100 km further west.

## Tyennan Orogeny Stage 3: Late Cambrian Basin Inversion

Volcanism declined dramatically in the late Cambrian, and the Proterozoic rocks of the basin margin became the dominant source of sediments. The nature of the post-volcanic stage of the Tyennan Orogeny can be deduced from the widespread and complex unconformities within the Owen Group. The syn-orogenic Owen Group and correlates are highly variable and restricted to discrete tectonically active basins.

The nature of late Cambrian tectonism is controversial. There is a pre-Middle Ordovician N-S fold phase in the Dundas Trough which is probably synchronous with Owen Group deposition. However, there is also local evidence of syn-depositional normal faults (Noll and Hall, 2006) and of syn-depositional reverse faults (Arnold and Carswell, 1990). The bulk of existing evidence suggests that the late Cambrian was a period of compressional tectonics with open upright folds and W-dipping thrusts. Holm and Berry (2002) correlated the $D_3$ event in NW Tasmania with pre-Ordovician N-S folding in the Dundas Trough. Berry et al. (2008) correlated this event with the accretion of the WTT to East Antarctica at the end of the Ross-Delamerian Orogeny (Figure 4).

## Ordovician–Devonian

Following the late Cambrian basin inversion, western Tasmania was peneplaned and a new cycle of deposition began in the Middle Ordovician. Contrasting Ordovician–early Devonian histories are seen in the Western and Eastern Tasmania Terranes. In the WTT, a widespread intertidal to shallow marine tropical carbonate succession (Gordon Group) is overlain by a Silurian–early Devonian shallow marine siliciclastic sequence (Tiger Range Group). At the same time, the Mathinna Supergroup was deposited in NE Tasmania. A major fault system separates the two terranes. The summary below is based on the latest review of this sequence (Calver et al., 2012b).

In the WTT there is a thick post-orogenic succession of dominantly shallow-marine siliciclastics and carbonates, consisting of the Gordon Group (Ordovician–lowest Silurian) and the Eldon Group (Silurian–Lower Devonian). The predominantly limestone Gordon Group overlies the upper Cambrian–lower Ordovician syn-orogenic coarse-grained siliciclastics (Owen Group) or rests unconformably on older basement.

The lower siliciclastics of the Gordon Group in western Tasmania consist of a fining-upward succession, with siliceous conglomerate, bioturbated quartz sandstone and siltstone. This is a diachronous, transgressive succession, up to 750 m thick, that overlies the Owen Group disconformably or unconformably. In the early Middle Ordovician, limestone was deposited in shallow-marine conditions offshore of the basal siliciclastic unit. By the late Middle Ordovician micritic peritidal tropical carbonate sedimentation dominated most of the WTT. The youngest part of the Gordon Group is a 250 m thick siltstone-dominated marine sequence that coarsens upwards.

The Gordon Group is overlain conformably or disconformably by a shallow-marine siliciclastic succession with subordinate limestone, the Eldon Group that ranges from early Silurian–Early Devonian and is up to 2.3 km thick in the Queenstown area.

In the ETT, the Ordovician–Early Devonian succession is a thick sequence of deep-marine sandy turbidites known as the Mathinna Supergroup, with strong similarities to the lower Paleozoic turbidites of the Lachlan Fold Belt of Victoria and especially the Melbourne Trough (Cayley, 2011). No older rocks are exposed in the ETT, and basement to the succession is unknown. The geochemistry and zircon inheritance of Devonian granites in this area indicate that the basement is young and probably late Neoproterozoic to Cambrian oceanic crust unlike the Proterozoic continental crust of Western Tasmania (Black et al., 2010).

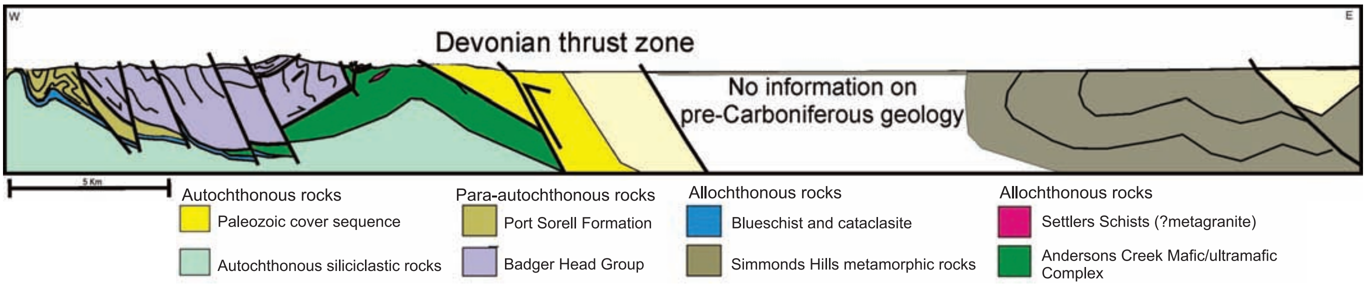


***Figure 7 Schematic cross-section across the WTT-ETT boundary. Based on Reed et al. (2002), Patison et al. (2001) and Seymour et al. (2011).***

The summary below comes from Seymour et al. (2011). The Mathinna Supergroup is separated into a lower Tippogoree Group and an upper Panama Group. Before 2001 these two groups were considered to be conformable but Reed (2001) argued the recumbent folding in the Tippogoree Group formed during the Benambran Orogeny (Silurian) and predated the deposition of the Panama Group. Recent mapping (Seymour et al., 2011) has demonstrated a faulted relationship between the two groups. An angular unconformity has not yet been found, but circumstantial evidence for the Benambran Orogeny in NE Tasmania is accumulating.

The lower section of the Tippogoree Group is dominated by thick (>1 m), graded beds of medium to fine-grained sandstone typical of proximal turbidites (Stony Head Sandstone c. 1 km thick). There is a fairly sharp transition at the top into the dark Turquoise Bluff Slate. Powell et al. (1993) interpreted the slate as pelagic and hemipelagic shale and marl. The Turquoise Bluff Slate contains Early–Middle Ordovician graptolites.

The Panama Group (Silurian–Early Devonian) is split into four formations. The Yarrow Creek Mudstone is a thin-bedded grey mudstone, with minor quartz-rich siltstone beds interpreted as distal turbidites. The Retreat Formation contains medium to fine-grained quartz-rich sandstone with minor mudstone. It was probably deposited in a series of submarine fan complexes. The Lone Star Siltstone (late Silurian) is a sequence of laminated siltstone and shale. Beds of quartz-rich sandstone become more common towards a transitional contact with the overlying Sideling Sandstone. The Early Devonian Sideling Sandstone contains quartz-rich, fine to medium-grained sandstone with minor siltstone. It was deposited in a marine passive margin setting, probably as sandy submarine fan complexes.

The Tabberabbean Orogeny, (390 Ma; Black et al., 2004) occurred throughout Tasmania and is characterised by the complexity of fold orientations, explained, in large part, by reactivation of older structures. In many areas the fold geometry is controlled by the Cambrian fold trends, which were tightened during the Devonian. This led to Devonian cleavage orientations that transect the axial planes of the folds with which they are associated. In the Fossey

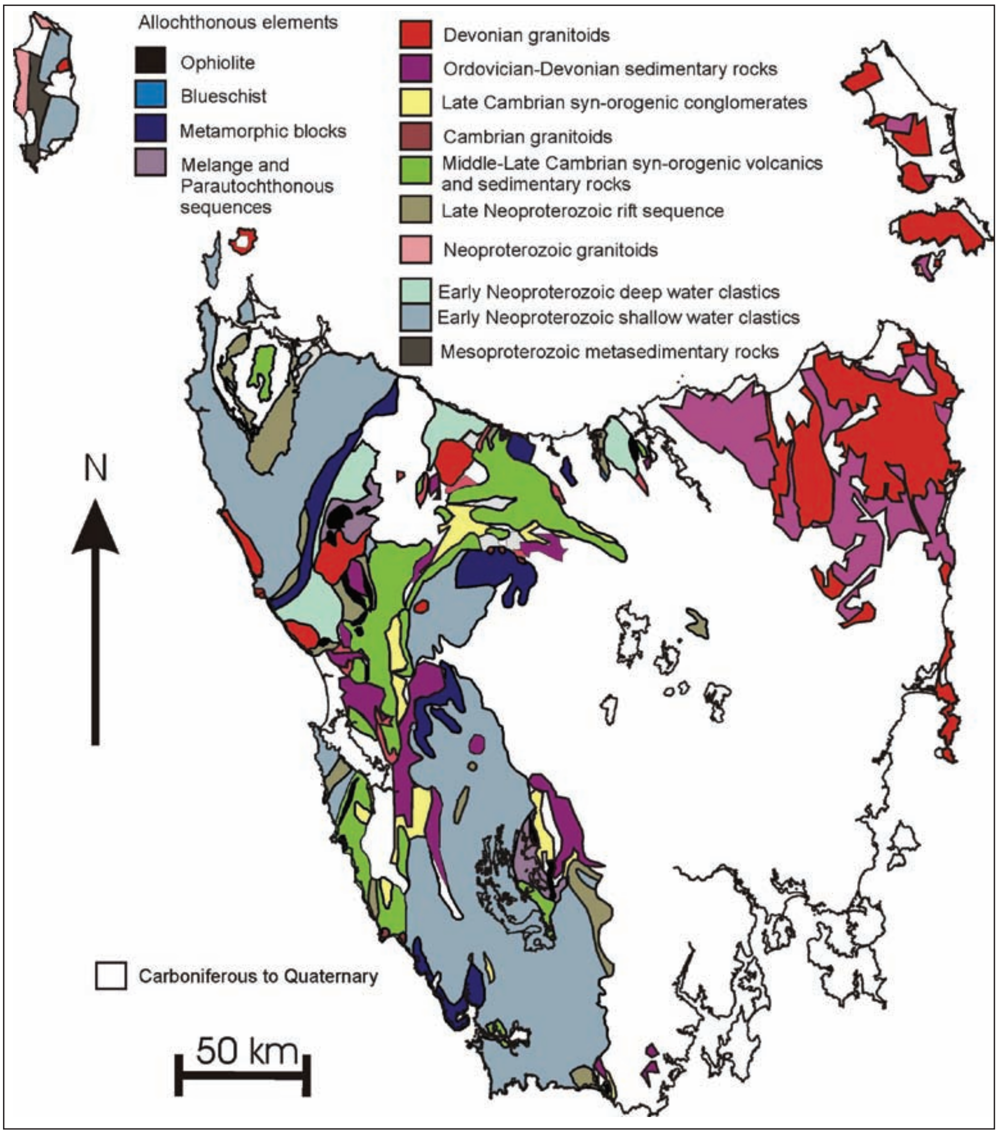


***Figure 8 Pre-Carboniferous geology of Tasmania. Simplified from 1:250,000 digital geological map of Tasmania (Brown et al., 1995).***

Mountain Trough, E-W Cambrian folds are tightened. In the Dundas Trough, N-trending Cambrian folds are tightened with an associated NNW-striking Devonian cleavage. The orientation of NNE-trending folds, N of Tullah, is controlled by reactivation of the Henty Fault.

A later N-S compression reactivated NNE-striking faults with sinistral strike slip movement. Detailed structural studies at the Renison Mine suggest that this event is the same age as granite-related mineralisation (Kitto, 1992). The closest dated granites to Renison have 360 Ma crystallisation ages (Black et al 2004), and this is the best evidence that this late deformation is much younger than the Tabberabberan Orogeny. Local reactivation of E-W thrust faulting occurred during this event.

In NE Tasmania the earliest phase of deformation thrust the passive margin E across the deep water section resulting in recumbent folds in the Georgetown area. However, this event is apparently restricted to the older units and may be due to the Benambran Orogeny. It was followed by back thrusting, which was especially strong in the Beaconsfield zone (Figure 7). A late stage of N-S compression produced strike slip movement on some faults and large scale kinks (Goscombe et al., 1994).

The regional metamorphic grade associated with the Devonian orogeny in the WTT is prehnite-pumpellyite with local zones of greenschist facies in the vicinity of late syn- to post-orogenic granites except in SE Tasmania where metamorphism was very low grade metamorphism (Burrett, 1992). The Mathinna Supergroup is very similar with very low grade metamorphism limited to the E coast at Scamander (Patison et al., 2001).

Voluminous granite intrusion in NE Tasmania (Figure 8) started before the Devonian deformation and continued until the early Carboniferous (400–375 Ma; Black et al., 2005). In western and NW Tasmania, their age range is 375–350 Ma, with the youngest known granite on King Island. Most of these granites are intruded at high level and have narrow contact aureoles.

## Summary

Tasmania forms an enigmatic province within the Neoproterozoic history of Australia. The WTT probably rifted from the East Antarctic margin at 580 Ma and was trapped outboard of the Cambrian Ross-Delamerian Orogen. It was accreted back to the Gondwana margin in the late Cambrian. The earliest rocks form a Mesoproterozoic shallow water sequence that was deformed at 1290 Ma. They were intruded by rift related granites at 760 Ma. The granites may be related to Rodinia breakup. A new cycle of passive margin sedimentation continued through the late Neoproterozoic but was interrupted by extensive rift tholeiites at 580 Ma.

An arc-continent collision in the early to middle Cambrian initiated the Tyennan Orogeny. This resulted in the emplacement of numerous allochthonous blocks, including obducted mafic and ultramafic slices in western and northern Tasmania. Post-collisional felsic volcanism and extension dominated the late middle Cambrian (Guzhangian). In the late Cambrian Tasmania was dominated by basin inversion and at this stage was accreted back onto the Gondwana margin.

Shallow water deposition dominated the WTT from the Ordovician to the Devonian. The oldest rocks of the ETT are Ordovician turbidites. These may have been deformed in the Benambran Orogeny before the deposition of more deep water clastic rocks. The first granites intruded this sequence at 400 Ma before the onset of the Tabberabberan Orogeny. This thrust the ETT over the WTT terrane. Granite intrusion continued until 350 Ma.

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*by Geoffrey R. Green*

# Ore Deposits and Metallogenesis of Tasmania

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*Tasmania contains a broad variety of economic mineral deposits, which includes several that have been known for over a century and are still operating today or were worked in the recent past. The Arthur Lineament, a belt of allochthonous amphibolite, carbonate rocks, psammite and pelite in northwest Tasmania hosts the Savage River magnetite deposit, which is now considered to be a Proterozoic carbonate replacement deposit with affinities to Kiruna-style iron-oxide Cu-Au deposits. The allochthon was formed during an early Cambrian collisional event between an east-facing passive margin sequence and an intraoceanic island arc. Post collisional, proximal submarine volcanism at c. 500 Ma in the Mount Read Volcanics followed and associated mineralisation includes world-class deposits. High grade Zn-Pb-Au-Ag-Cu massive sulfide deposits (e.g., Rosebery and Hellyer) were formed from seawater-dominated hydrothermal fluids. Disseminated Cu-Au-Ag deposits of the Mount Lyell field are associated with broad alteration zones that include phyllosilicate assemblages indicative of a component of oxidised magmatic fluid, as does the Henty Au deposit in the north. Orogenic Au is an important deposit style in northeast Tasmania and includes the Tasmania deposit, Australia's largest single Au reef. Largely post-orogenic granitic magmatism (Lower Devonian–Tournaisian) includes an important Sn-W-base metal± magnetite mineralising event associated with reduced, fractionated granite in northeast and western Tasmania. World class Sn±Cu sulfide skarn and vein deposits are the product of interpreted magmatic fluids exsolved from these granitic magmas. The highly unusual disseminated Avebury Ni deposit is associated with granite of this type. World class scheelite skarn deposits on King Island lie in the contact aureole of moderately oxidised, unfractionated Tournaisian granodiorite on King Island.*

## Introduction

Despite an area of only 68,000 km$^2$, Tasmania has a remarkable geological diversity and abundance of mineral deposits. Rocks from every period of the Earth's history from the Mesoproterozoic are present and there have been at least four major episodes of economic mineralisation. Significant mineral deposits include Proterozoic magnetite, silica, dolomite and magnesite deposits; Cambrian VHMS base metal-Au and ultramafic-related Pt-group minerals (PGM) and chromite deposits; Devonian orogenic and intrusion related Au deposits; Middle Devonian–Tournaisian granite-related Sn, W, fluorite, magnetite, Ag-Pb-Zn and Ni deposits; Triassic coal and Oligocene–Miocene lignite deposits; and Cenozoic alluvial Au, Sn and PGM deposits, and residual Ni-Co, Fe oxide, silica and clay deposits. Resource data are listed in Table 1.

This contribution is abbreviated from Seymour et al., (2006; revised 2007), but updated.

## Proterozoic

The oldest known rocks in Tasmania are metaturbidites on King Island, metamorphosed at c. 1300 Ma (Berry et al., 2005) and, with youngest detrital zircons dated at 1350 Ma (Black et al., 2004). In northwestern and central Tasmania quartz arenite, siltstone, shale (commonly carbonaceous and pyritic) and minor carbonate rocks of the Rocky Cape Group and Tyennan region respectively, thought to be of late Mesoproterozoic–Neoproterozoic age, were deposited in a passive margin marine shelf environment (Figure 1). Between these occurs a quartzwacke turbidite formation of roughly similar age (Burnie and Oonah formations), but slightly better constrained to c. 1070–750 Ma (Calver et al., 2012). These successions are overlain by Cryogenian–Terreneuvian carbonates, siliciclastics (including glacigene diamictite) and dominantly mafic volcanics. These volcanics include c. 580 Ma syn-rift tholeiites, which represent a probably E-facing continental margin (Ps in Figure 1; Calver et al., 2012). The carbonates are an important host to Devonian Sn, W, Cu and magnetite skarn mineralisation on King Island, and at Mt Bischoff, Mt Lindsay, Renison Bell and elsewhere in western Tasmania. High purity silica flour deposits are interpreted as disaggregated silicified Neoproterozoic dolomite. Gabbro dykes of probable Neoproterozoic age host minor, but locally high grade, Ni-Cu-Pt-Pd-Au mineralisation in the Cuni district, near Zeehan.

Separating the two sequences on King Island is an unconformity related to the Wickham Orogeny, with syn-orogenic granite dated at c. 760 Ma (Turner et al., 1998). On mainland Tasmania the boundary is marked by a mild deformational event and locally low angle unconformity.

## Cambrian Orogenesis, ultramafics, the Arthur Lineament and associated mineralisation

The major collisional Tyennan Orogeny occurred between c. 512–506 Ma (Turner et al., 1998), contemporaneous with the first phase

*Table 1 Non-alluvial deposits of Tasmania: pre-mining resources*

Note: These data rely on past production figures and various resource estimations. Some of the latter do not comply with the Joint Ore Reserves Committee Code standards.

| | |
|---|---|
| **Precambrian deposits in the Arthur Lineament** | |
| Savage River | 371 Mt @ 31.9% Fe |
| Arthur River | 29 Mt @ 42.8% Mg |
| Main Creek | 42.8 Mt @ 42.4% Mg |
| **Cambrian gabbro-hosted deposits** | |
| Nickel Reward (Cuni field) | 0.03 Mt @ 3% Ni |
| North Cuni–Genets Winze | 0.95 Mt @ 0.76% Ni, 0.94% Cu |
| **Cambrian deposits in the Mount Read Volcanics** | |
| Hellyer | 16.5 Mt @ 13.9% Zn, 7.2% Pb, 0.38% Cu, 169 g/t Ag, 2.55 g/t Au |
| Fossey Zone | 0.55 Mt @ 0.5% Cu, 7.1% Pb, 12.9% Zn, 134 g/t Ag, 2.6g/t Au |
| Que River | 3.3 Mt @ 13.3% Zn, 7.4% Pb, 0.7% Cu, 195 g/t Ag, 3.3 g/t Au |
| Mount Charter | 6.1 Mt @ 0.5% Zn, 25.5 g/t Ag, 1.22 g/t Au, 9.7% Ba |
| Rosebery | 46.70 Mt @ 12.46% Zn, 3.9% Pb, 0.50% Cu, 133 g/t Ag, 1.93 g/t Au |
| Hercules | 3.33 Mt @ 17.3% Zn, 5.5% Pb, 0.4% Cu, 171 g/t Ag, 2.8 g/t Au |
| South Hercules | 0.56 Mt @ 3.7% Zn, 1.9% Pb, 0.1% Cu, 157 g/t Ag, 3.0 g/t Au |
| Henty–Mt Julia | 2.83 Mt @ 12.5 g/t Au |
| Tasman and Crown Lyell | 0.138 Mt @ 10.0% Zn, 8.9% Pb, 0.54% Cu, 212 g/t Ag, 0.35 g/t Au |
| Mount Lyell | 311 Mt @ 0.97% Cu, 0.31 g/t Au |
| Garfield | 12 Mt @ 0.3% Cu |
| **Ordovician carbonate-hosted deposits** | |
| Oceana | 2.6 Mt @ 7.7% Pb, 2.5% Zn, 55 g/t Ag |
| Grieves Siding | ~0.7 Mt @ 8% Zn (primary); 0.15 Mt @ 5% Zn (secondary) |
| **Devonian orogenic Au deposits** | |
| Beaconsfield | 3.25 Mt @ 19.0 g/t Au |
| New Auen Gate | 0.51 Mt @ 15.6 g/t Au |
| Pinafore Reef (Lefroy) | 0.974 Mt @ 10.1 g/t Au |
| **Devonian granite-related Sn deposits** | |
| *Sulfide skarns* | |
| Renison Bell | 30.03 Mt @ 1.44% Sn; 1.93 Mt @ 0.35% Cu |
| Mount Bischoff | 10.54 Mt @ 1.1% Sn |
| Cleveland | 12.4 Mt @ 0.61% Sn, 0.25% Cu |
| Foley zone | 3.8 Mt @ 0.28% $WO_3$, 0.02% $MoS_2$, 0.05% Sn |
| Razorback | 0.34 Mt @ 0.9% Sn |
| Queen Hill | 4.36 Mt @ 1.1% Sn |
| *Silicate skarns* | |
| St Dizier | ~2.6 Mt @ 0.5% Sn, 0.05% $WO_3$ |
| *Magnetite skarns* | |
| Mount Lindsay | 18 Mt @ 0.3% Sn, 0.2% $WO_3$, 17% recoverable Fe |
| Nelson Bay River secondary hematite ore) | 12.7 Mt @ 36.1% Fe (includes |

*Table 1 Contd...*

| | |
|---|---|
| *Vein deposits* | |
| Aberfoyle | 2.1 Mt @ 0.91% Sn, 0.28% $WO_3$ |
| Pieman vein (East Renison) | 0.43 Mt @ 1.0% Sn |
| *Greisen deposits* | |
| Anchor | 2.39 Mt @ 0.28% Sn |
| **Other Devonian granite-related deposits** | |
| *Skarns* | |
| King Island field | 23.8 Mt @ 0.66% $WO_3$ |
| Kara | 5.2 Mt @ >30% Fe, by-product $WO_3$ |
| Avebury | 10.04 Mt @ 1.14% Ni |
| Moina | 18 Mt @ 26% $CaF_2$, 0.1% Sn, 0.1% $WO_3$ |
| Hugo (Moina area) | 0.25 Mt @ 5.5% Zn, 1 g/t Au, 0.1% Bi |
| Stormont (Moina area) | 0.135 Mt @ 3.44 g/t Au, 0.21% Bi |
| *Vein deposits* | |
| Storeys Creek | 1.1 Mt @ 1.09% $WO_3$, 0.18% Sn |
| Magnet | 0.63 Mt @ 7.3% Zn, 7.3% Pb, 427 g/t Ag |
| Salmons Vein (East Renison) | 0.83 Mt @ 3.2% Pb, 2.2% Zn, 104 g/t Ag, 0.19% Sn, 0.61% Cu |
| New North Mount Farrell and North Mount Farrell | 0.908 Mt @ 12.5% Pb, 2.5% Zn, 408 g/t Ag |
| Lakeside | 0.75 Mt @ 0.2% Sn, 0.2% Cu, 4.0% As, 2.1 g/t Au, 20 g/t Ag |
| **Paleocene residual deposits** | |
| Barnes Hill | 6.6 Mt @ 0.82% Ni, 0.06% Co |

of the Delamerian Orogeny on the Australain mainland, and is believed to by related to NE-directed subduction of the thinned Tasmanian Precambrian crust beneath an intraoceanic island arc (Meffre et al., 2000). Dismembered mafic-ultramafic complexes including oceanic forearc boninites, low-Ti tholeiites, gabbros, and orthopyroxene-rich ultramafic cumulates (Brown and Jenner, 1989) were obducted onto the craton (Berry and Crawford, 1988). A tonalite from one of the complexes has been dated at 513.6 ± 5.0 Ma (Black et al., 1997), providing an upper limit on the age of the alllochthon. Other sedimentary sequences, including deep marine chert, are considered part of the allochthonous units. Osmiridium is associated with the ultramafics rocks and derived alluvial deposits were the focus of a small industry during 1910–1959.

In central and western Tasmania there was locally strong deformation of the Proterozoic rocks with two phases of recumbent folding and metamorphism up to eclogite facies. The western boundary of the strong deformation and higher grade metamorphism is defined by the Arthur Lineament (Figure 2), a narrow, 110 km long, NE striking belt of tholeiitic amphibolites, metasediments and carbonates, including significant diagenetic magnesite deposits of the Arthur Metamorphic Complex. Several magnetite deposits are associated with the lineament, including the major Savage River deposit. The Arthur Lineament includes deformed albitised intrusive rocks (?granodiorite) of Wickham Orogeny age (777 ± 7 Ma; Turner et al., 1998). Similar rocks occur near the Savage River deposit and are considered to be part of the alteration assemblage associated with ore formation (Bottrill and Taheri, 2008). The central part of the Arthur Lineament, the Bowry Formation, is considered to be entirely allochthonous (Holm and Berry, 2002) and this is consistent with its unique mineralisation styles and the age of the intrusive.

The Savage River deposit consists of several lenses of magnetite-rich ore with the following associated minerals in various proportions

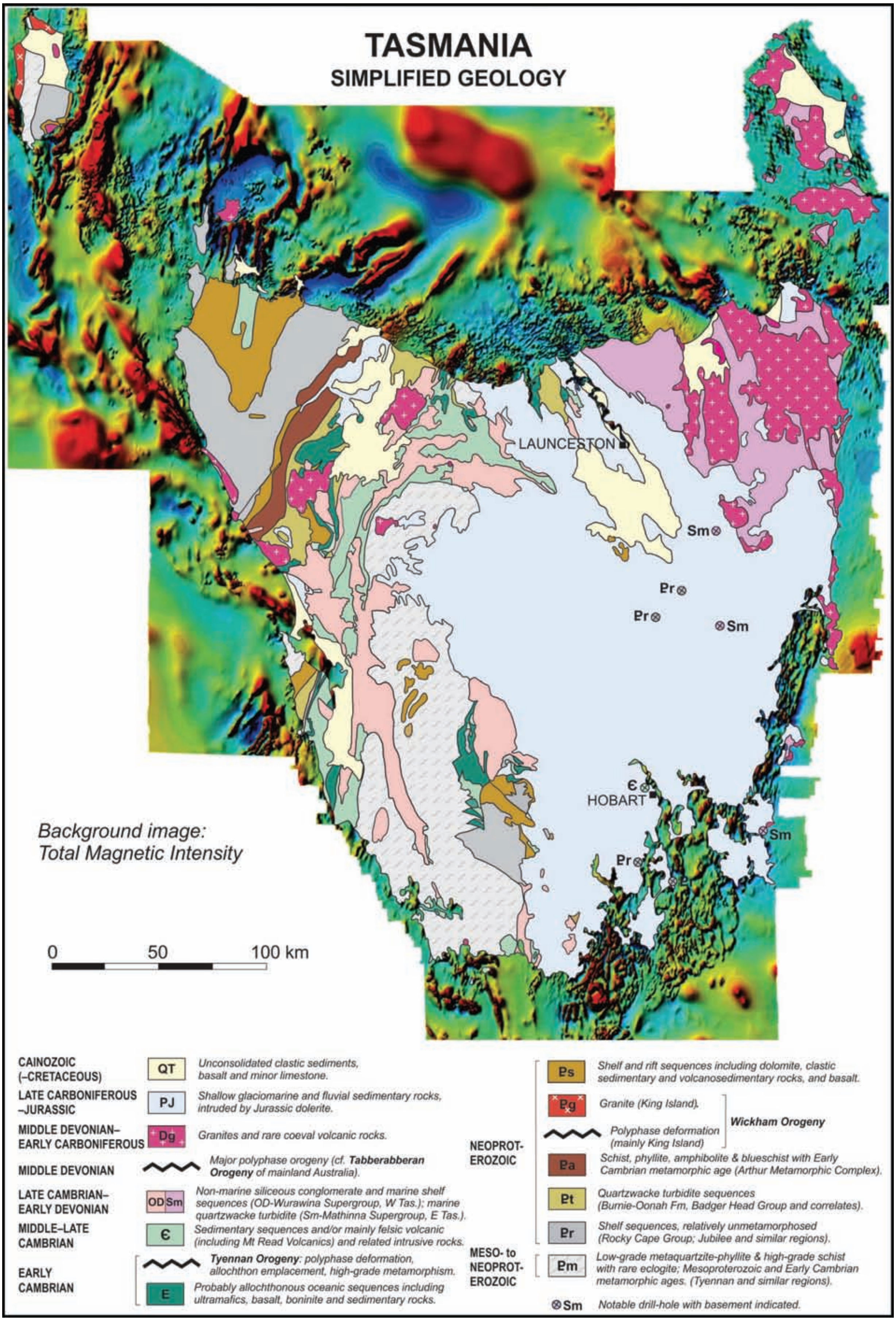



***Figure 1 Simplified geology of Tasmania, with total magnetic image by R.G. Richardson in offshore area (from Seymour et al., 2006).***

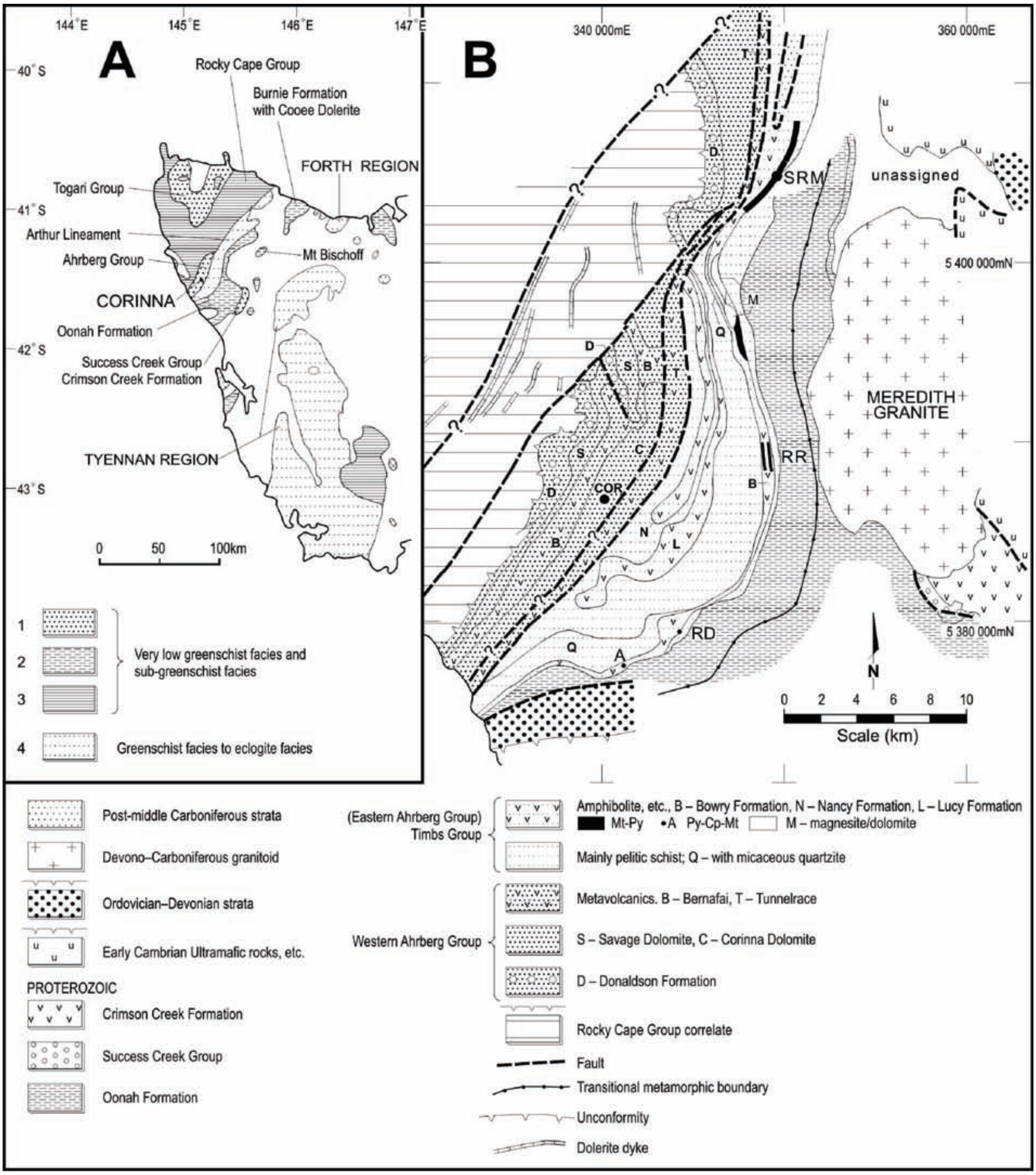



***Figure 2 (A) Western Tasmania showing Proterozoic rocks (1 – continental shelf and margin deposits with rift tholeiites near the top; 2 – sandy turbidite; 3 – quartz arenite, siltstone, minor dolomite) and c. 515–510 Ma metamorphic rocks derived from Proterozoic rocks (4) (after Bottrill and Taheri, 2008). (B) Detailed geology of the Corinna district after Turner et al. (1991). The Arthur Metamorphic Complex consists of the Timbs Group and the adjacent, metamorphosed part of the Oonah Formation. The boundaries are defined by the tectonic feature called the Arthur Lineament (Gee, 1967) shown in Figure 2A. Abbreviations: COR = Corinna; SRM = Savage River mine; RR = Rocky River; RD = Reece Dam; A = Alpine locality.***

in roughly decreasing maximum abundance: amphibole (dominantly tremolite-actinolite, but also relict hornblende and blue amphiboles including glaucophane and later barroisite-winchite-actinolite), serpentine, talc, dolomite, calcite, pyrite, chlorite, albite, quartz, apatite and hematite. Amphibolite, pyroxene-rich rocks, serpentinite, magnesite, talc-chlorite-quartz-dolomite schist, dolostone and mafic metasedimentary rocks are present within the host sequence and the ores and host rocks show strong deformational fabrics. Some minerals such as calcite and hematite are entirely retrogressive and post-date the two main Cambrian deformational events (Bottrill and Taheri, 2008).The serpentinite is notable in having insignificant Cr. The features of the deposit are consistent with hydrothermal replacement of carbonate and mafic host rocks predating or early in the deformational history of the area and have affinities with Kirunatype iron-oxide Cu-Au deposits (IOCG) deposits (Taheri and Bottrill, pers. comm., 2011). Trace quartz, rutile, titanite and chalcopyrite are present. Further south in the Bowry Formation, at the Alpine prospect, significant chalcopyrite is present within and near similar ironstone.

## Middle Cambrian post-collisional phase: Mt Read Volcanics

The most important metallogenic event in Tasmania coincided with the deposition of the Mount Read Volcanics (MRV). U-Pb zircon ages and numerous fossil occurrences constrain the bulk of the MRV to c. 506–494 Ma (Corbett and Vicary, 2012). The main mineralised belt of the MRV between Mount Darwin and Hellyer is the Central Volcanic Complex (CVC), which is dominated by proximal felsic volcanic rocks (rhyolite and dacite flows, domes and cryptodomes and massive pumice breccias) and andesite and rare basalt (lavas, hyaloclastites and intrusive rocks) deposited in a marine environment (Corbett 1992; 2002; Gifkins and Kimber, 2003). This belt is flanked to the west by the coeval Western Volcano-Sedimentary Sequence (WVS) of turbidites with lithic sandstone and mudstone (commonly rich in shards), siltstone, shale and subordinate intrusive rocks and lavas, commonly andesitic (Corbett, 2002).

These rocks are overlain by the Tyndall Group, a unit of quartz-bearing volcaniclastic sandstone and conglomerate, minor felsic and andesitic lavas and intrusive rocks and welded ignimbrite (White and McPhie, 1996). Considerable erosion took place locally before deposition of the Tyndall Group. Clasts of granite and altered volcanic rocks occur in the basal Tyndall Group in the Mount Darwin area (Corbett, 2002; Morrison, 2002).

Flanking the CVC to the east and abutting the metasedimentary rocks of central Tasmania is the Eastern Quartz-phyric Sequence, which consists mostly of quartz-feldspar-phyric lavas, intrusive porphyries and volcaniclastic sandstone, that are intruded by magnetite series granites. The base of this unit consists of Precambrian-derived sandstone and conglomerate which passes upward gradationally into volcaniclastic sandstone. There is uncertainty about whether the Eastern Quartz-phyric Sequence is a time equivalent of the CVC (Corbett, 2002) or part of the Tyndall Group (Murphy et al., 2004).

Tectonism was mostly near east-west extensional during Mount Read Volcanics deposition, as recorded by the orientation of hydrothermal veins (e.g., at the Hellyer deposit; Gemmell and Large, 1992) and basaltic dykes in the Henty Fault Zone. Further evidence of an extensional regime is provided by growth faulting and cauldron subsidence associated with the formation of the thick pumice-rich breccia underlying the Rosebery deposit (Green et al., 1981; Gifkins and Kimber, 2003). Solomon and Groves (2000) suggested that this episode of extensional tectonism and crustal thinning was related to eastward retreat of a west-dipping subduction zone following the 512–506 Ma collisional event.

Mineralisation was concentrated in a short time interval in the late middle Cambrian at the top of the CVC and in places in the immediately overlying Tyndall Group rocks. Major alteration zones are dominantly of quartz-sericite mineralogy and restricted to the proximal CVC volcanic facies (Gifkins and Kimber, 2003; Herrmann and Kimber, 2003), but chlorite-rich cores are apparent at Hellyer (Gemmell and Large, 1992) and Hercules (Green and Taheri, 1992). Despite this association with the CVC, the WVS, in areas of andesite proximally underlying Tyndall Group equivalents, represent under-explored targets (Corbett, 2002).

The Henty Fault Zone constitutes a fundamental metallogenic divide within the MRV. To the northwest, polymetallic Zn-Pb-Au-Ag-Cu massive sulfide deposits dominate (Hellyer, Que River, Rosebery and Hercules; Figure 3) together with disseminated deposits with low base metal, but relatively high Au and Ag, tenor (Mount Charter and South Hercules). Mount Charter is a barite-rich, low-grade precious and base metal deposit formed within quartz-sericite altered volcanic rocks in an alteration system that presumably underlay a lower temperature white smoker exhalative hydrothermal vent field. Ore fluids have been considered to have been derived by convective circulation of seawater and interaction with volcanic and basement rocks, but there is evidence for a magmatic contribution to the ore fluids at Hellyer (Solomon and Groves, 2000). There is also debate about the extent to which massive sulfide ore deposition took place in brine pools on the seafloor (Solomon and Groves, 2000), as opposed to in either Kuroko-type hydrothermally reworked seafloor mounds (e.g., Gemmell and Large, 1992) or sub-seafloor replacement or displacement (Allen, 1995). There is clear evidence for sub-seafloor ore formation locally, e.g., at the Hercules deposit and Fossey Zone, immediately south of Hellyer.

Southeast of the Henty Fault Cu-Au and Au deposits dominate, exemplified by the Mount Lyell field and the Henty Au deposit. The most economically important deposits in the Mount Lyell field are disseminated chalcopyrite-pyrite orebodies in alteration assemblages dominated by quartz-sericite or quartz-chlorite-sericite (Prince Lyell, Cape Horn, Lyell Comstock, in part, and Western Tharsis; Walshe and Solomon, 1981; Corbett, 2001). The Western Tharsis deposit consists of concentrically and vertically zoned alteration assemblages centred on the mineralisation, with a lower and central quartz-chlorite-sericite zone passing upward and outward in turn into pyritic quartz-pyrophyllite ± topaz ± fluorite ± zunyite ± woodhouseite with local bornite-bearing zones and a pyritic quartz-sericite assemblage (Huston and Kamprad, 2001). The highest grade ores are in the North Lyell area and consist of coarse-grained bornite with chalcopyrite and minor pyrite in brecciated chert ± pyrophyllite ± barite ± hematite or quartz-sericite schist at or near the faulted contact between the altered CVC and the Owen Conglomerate, which may be locally hematised. The origin of these high-grade ores has been a subject for debate. Solomon *et al.* (1987) suggested that they may have formed during Devonian deformation by mixing of metamorphic fluids from the volcanic rocks and conglomerate, but most current opinion is that the ores of the field formed at the same early Tyndall Group time (Corbett, 2001; Huston and Kamprad, 2001). A Re-Os date of 500.4 ± 2.3 Ma has been determined on molybdenite from Prince Lyell (Huston et al., 2009).

The first deposit mined, the Mount Lyell or Iron Blow deposit, consisted mostly of massive pyrite-chalcopyrite. There are also small lenses of polymetallic pyritic sphalerite-galena rich massive sulfide in the Lyell Comstock area at the northern end of the field. There are other disseminated Cu deposits in the CVC further south. Some, such as the Garfield prospect, appear geologically similar to Prince Lyell. A 475 $km^2$ HyMap airborne hyperspectral survey flown over Mount Lyell showed that the main Cu deposits area is associated with Al-rich mica and that there is a zone of intense pyrophyllite-topaz alteration that may represent a significant exploration target at Glen Lyell, about 1 km south of the Prince Lyell orebody (Huntington and Cocks, 2003).

The Henty Au mine (Figure 3) consists of a series of small high-grade lenses of Au mineralisation in quartz ± sericite-altered volcaniclastic and volcanic rocks that occupy a large sub-vertical quartz-sericite alteration zone that transects the CVC-Tyndall Group contact at a low angle. The deposit is regarded as a submarine epithermal system formed from a magmatic fluid (Callaghan, 2001). Halley (2007) identified quartz-topaz-pyrophyllite alteration in the footwall of the Mount Julia section of the deposit and invoked the mixing of $SO_2$-rich magmatic and seawater derived fluids in the formation of the deposit. Current opinion also favours a significant or dominant magmatic component in the fluids that formed the Mount Lyell deposits (Large *et al.*, 1996; Huston and Kamprad, 2001), with the fluids supposedly derived from Cambrian granitic magmas. The three dimensional geological model of Tasmania provides some empirical support for this, in that granite is interpreted to shallowly underlie the Cu-Au metallogenic region east of the Henty Fault, but is considered to be at far greater depths to the west.

Positive evidence of undiscovered VHMS deposits in western Tasmania exists in the form of debris flow deposits with rafts and clasts of high-grade ore (Figure 3).

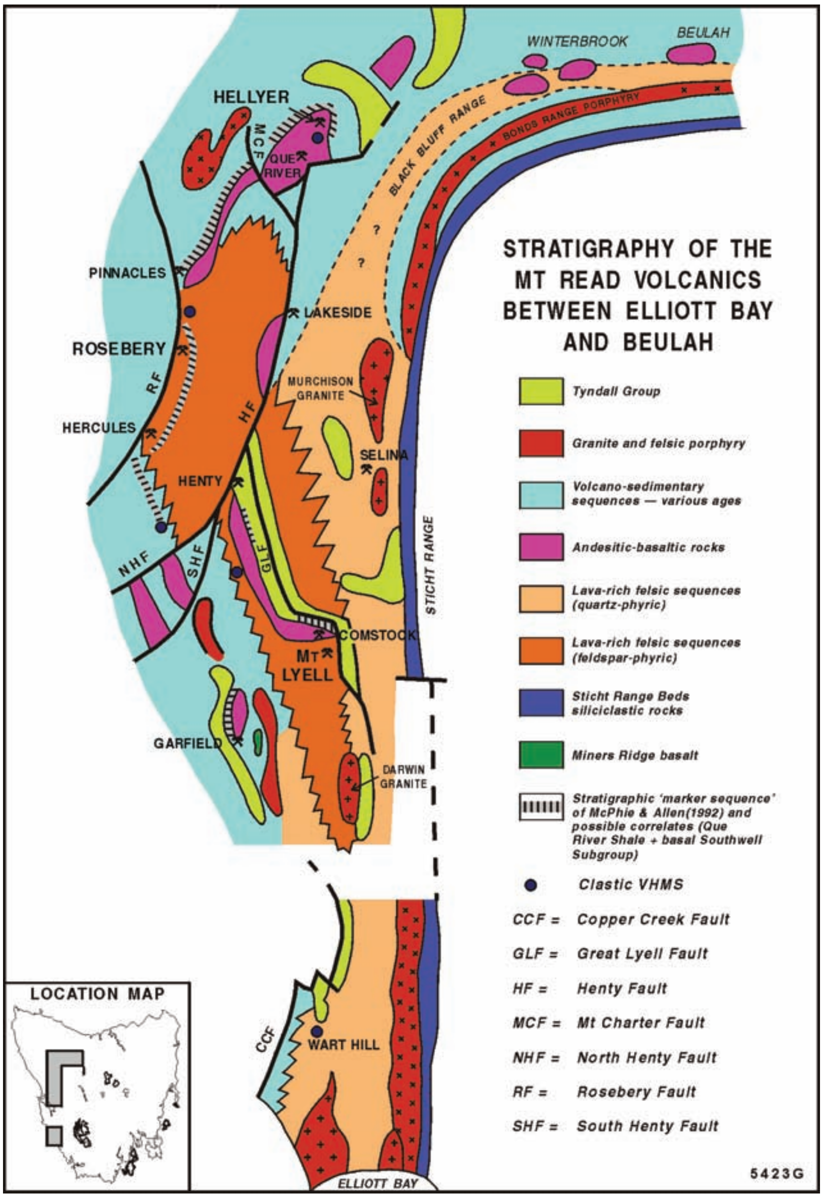



***Figure 3 Broad geological subdivisions of the Mount Read Volcanics (after Pemberton and Corbett, 1992).***

## Wurawina Supergroup: Upper Cambrian–Lower Devonian

In the late Cambrian, the final phase of the Tyennan Orogeny inverted earlier extensional faults (e.g., Henty Fault). Major reverse faults and upright open north-trending folds were formed in western Tasmania. This phase also caused uplift of the Tyennan region, with syn-orogenic sediments (Owen Group) accumulating in a half graben, commonly with angular or erosional unconformity on older units. It typically includes large volumes of coarse siliciclastic conglomerate composed dominantly of metaquartzite clasts derived from the Proterozoic rocks to the east, but also includes turbidite and shallow marine sandstone units (Noll and Hall, 2005; Seymour et al., 2006).

In western Tasmania, the Gordon Group rests on the Owen Group with angular unconformity and consists of basal sandstone followed by a shallow-marine to peritidal, platform succession of predominantly micritic, dolomitic limestone that is up to 1.8 km thick in central–southern Tasmania but considerably thinner in western Tasmania. The onset of carbonate sedimentation took place in the Middle Ordovician in western and northern Tasmania but was earlier (Early Ordovician) in the east (Banks and Burrett, 1980; Banks and Baillie, 1989; Laurie, 1995). Stratiform sulfide mineralisation and an associated breccia unit in the Zeehan area, notably at the Oceana deposit, indicate local synsedimentary faulting and possible exhalative activity (Taylor and Mathison, 1990). Gordon Group carbonate sequences became an important ore host for skarn mineralisation associated with intrusion of Late Devonian–Tournaisian granites. High purity limestone is mined at Mole Creek for metallurgical use and limestone from Railton and elsewhere is utilised for cement manufacture and other purposes.

The Gordon Group is overlain by the Silurian–Early Devonian Eldon Group comprised of shelf sandstone, shale and minor limestone (Banks and Baillie, 1989).

## Mathinna Supergroup: Ordovician–Early Devonian

Deposition of the Mathinna Supergroup in eastern Tasmania was approximately coeval with deposition of the uppermost parts of the Owen Group to the top of the Eldon Group in western Tasmania. The Mathinna Supergroup comprises a succession of turbiditic sandstone and mudstone and forms the pre-Carboniferous sedimentary basement to eastern Tasmania. It consists of Ordovician sandstone, mudstone and pyritic black slate in fault contact with Silurian–Early Devonian sandstone and mudstone (Seymour et al., 2011). The oldest deformation event affecting the sequence was Late Ordovician–early Silurian (Benambran) and only affected the Ordovician formations (Reed, 2001).

U-Pb analyses of detrital zircon indicate the source of the Mathinna Supergroup was not the Western Tasmanian terrane (Black et al., 2004). The current models suggest that northeastern Tasmania may have been substantially separated from the rest of Tasmania at the time of its deposition, and that it was finally docked with western Tasmania during Middle Devonian orogenesis (e.g., Black et al., 2010).

## Middle Devonian orogenesis, granite emplacement and mineralisation

During the Middle Devonian, most of Tasmania was affected by polyphase deformation, characterised by a complexity of fold orientations, due in part to reactivation of older structures. The fold geometry was commonly controlled by the trends of Cambrian folds which were tightened during the Devonian, and as a result, in places Devonian cleavage orientations transect the axial planes of associated folds (Seymour et al., 2006).

The folding occurred in two main phases in western Tasmania. The early phase produced NNW-trending folds in areas where reactivation effects were not significant, and was followed by a second phase that produced NW- to WNW-trending folds and thrusts (Seymour et al., 2006).

The Mathinna Supergroup of NE Tasmania also shows evidence of two Devonian compressional deformation events, the last of which was WSW directed thrusting that accompanied orogenic Au mineralisation in the region (Powell, 1991; Reed, 2002, 2004), at c. 400 Ma (Bierlein et al., 2005).

Sandstone, siltstone and conglomerate in the Beaconsfield area and the Mathinna Supergroup, further east, host economically important orogenic vein-style Au mineralisation, In the western part of the Eastern Tasmanian Terrane, notably in the Beaconsfield and Lefroy districts (Figure 4), steeply dipping reefs have an E–ENE orientation and formed near-parallel to the axis of maximum principal stress (Powell, 1991; Reed, 2002), whereas to the east along the Mangana-Mathinna-Waterhouse Au lineament and elsewhere, the dominant orientation is NNW and orthogonal to this axis. The latter structures form as a response to failure on the steep eastern limbs of anticlines formed during the earlier fold event. In detail, Au-rich shoots within the reefs tend to pitch steeply, as a response to either favourable lithology for reef formation (Beaconsfield) or due to Au mineralisation during late transcurrent movement on the structures (Mathinna; Keele, 1994).

An extended period of large-scale granitic intrusion at relatively high crustal levels commenced in eastern Tasmania at c. 400 Ma with the emplacement of unfractionated I-type granodiorites. This occurred prior to the Devonian deformation events, and continued after the close of Devonian deformation, with the youngest intrusions at c. 350 Ma on King Island (Black et al., 2005). There is also one body of thick welded tuff (the St Marys Porphyrite), dated at 388 ± 1 Ma (Turner *et al.*, 1986). Most of the western Tasmanian granitic rocks post-date Devonian folding events. On mainland Tasmania the granitic rocks form three large complexes; one in the east, another in the northwest (Figure 5), and a third largely concealed beneath the southwest corner of Tasmania. Dating reveals westward younging regionally across Tasmania, and there is also a compositional trend towards felsic, fractionated I-type and S-type granite and monzogranite with decreasing age (Black *et al.*, 2005).

Important deposits of Sn, W and lesser occurrences of Ni, magnetite, Cu, Ag, Pb, Zn and Au are associated with Devonian granite emplacement. All significant granite-related deposits lie within the 4 km granite isobath (Figure 5; Leaman and Richardson, 2003). Contrast in the host rocks in western and northeastern Tasmania is responsible for fundamental differences in the Late Devonian–early Carboniferous metallogenesis of the regions. There is a more restricted range of deposits in the northeast due mostly to a lack of reactive host

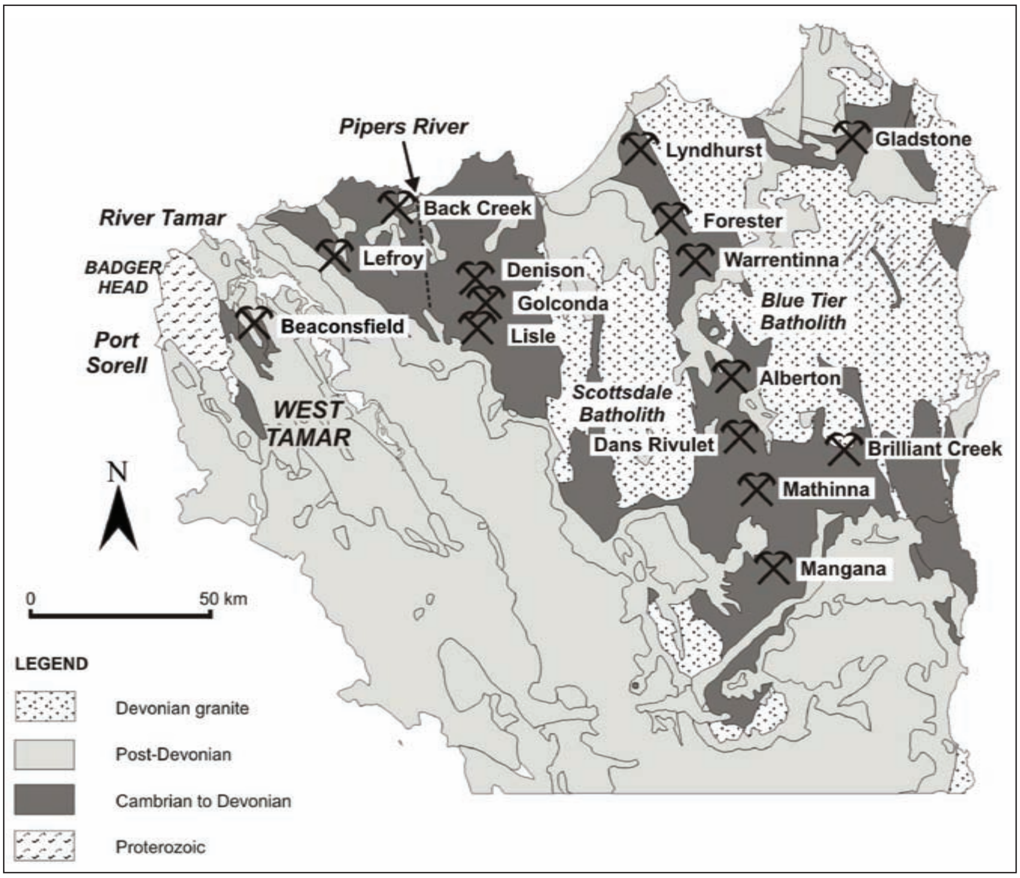
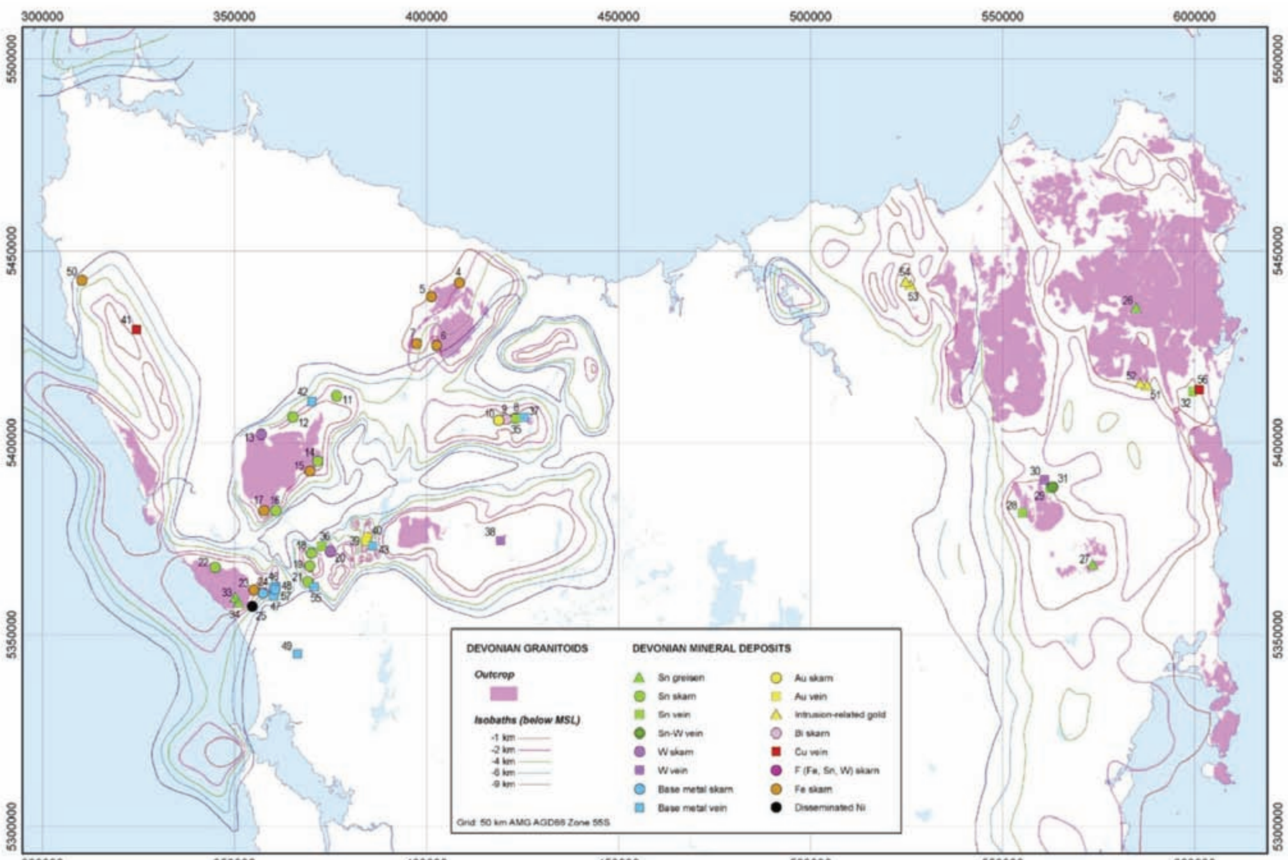



***Figure 4 Location of goldfields in NE Tasmania. Brilliant Creek and Golconda are intrusion-related Au; Denison shows feature of both intrusion-related (IRG) and orogenic Au deposits; Lisle was a significant alluvial field probably derived from IRG; remainder of fields are orogenic Au (after Reed, 2002).***




4: Natone (Fe),
5: Highclere (Fe),
6: Kara no. 2 (Fe),
7: Kara (Fe-W),
8: Moina ($CaF_2$-Fe-W-Sn),
9: Ti-Tree Creek (Bi-Au-Ag),
10: Stormont (Bi-Au-Ag),
11: Mount Bischoff (Sn),
12: Cleveland (Sn-Cu),
13: Mount Youngbuck (W),
14: Mount Ramsay (Sn-Cu),
15: Laurel Creek (Fe),
16: Mount Lindsay (Sn-W-Fe),
17: Stanley River (Fe),
18: Renison Bell (Sn-Cu-Ag),
19: Pine Hill (Sn),
20: Colebrook Hill (W-Cu),
21: Razorback (Sn),
22: St Dizier (Sn),
23: Tenth Legion (Fe-Zn),
24: Avebury (Ni),
25: Comstock (Zn-Pb-Ag),
26: Anchor (Sn),
27: Royal George (Sn),
28: Rex Hill (Sn-Zn-Cu-Pb),
29: Aberfoyle (Sn-W),
30: Storeys Creek (W-Sn),
31: Lutwyche (Sn-W),
32: Great Pyramid (Sn),
33: Federation (Sn),
34: Sweeneys (Sn-Zn-Ag),
35: Shepherd and Murphy (Sn-W-Bi),
36: East Renison (Zn-Pb-Ag-Sn-Cu),
37: Narrawa Creek (Higgs, Au-Zn-Pb-Ag),
38: Oakleigh Creek (W),
39: Lakeside (Au-Sn-Cu-Ag),
40: Lorrigans Luck (Sterling Valley Sn, Cu-Au-Ag),
41: Murrays Reward (Cu),
42: Magnet (Zn-Pb-Ag),
43: New North Mount Farrell (Pb-Zn-Ag-Cu),
46: Zeehan Montana (Ag-Pb-Zn),
47: Silver Spray (Pb-Au-Zn),
48: Oonah (Sn-Cu-Ag-Pb),
49: Queensberry (Zn-Pb-Ag-Cu),
50: Nelson Bay River (Fe),
51: Queen of the Earth (Au),
52: Brilliant (Au),
53: Enterprise (Au),
54: Potoroo (Au),
54: Queen Hill (Sn),
55: Comet (Pb-Ag-Zn),
56: Orieco (Cu),
57: Queen Hill (Sn).

***Figure 5 Location of significant Devonian granite related deposits (after Green et al., 2012) and Devonian granite isobaths (after Leaman and Richardson, 2003).***

lithologies, and perhaps also to a greater degree of unroofing (Solomon and Groves, 1994). There is current exploration interest in intrusion-related Au that occurs as part of the Au-As-Bi-Mo association within, and in the aureoles of, unfractionated I-type moderately oxidised to moderately reduced granodiorite in the Lisle-Golconda area. Significant mineral deposits (Figure 5) are cassiterite- and wolframite-bearing vein deposits (e.g., Aberfoyle and Storeys Creek) and Sn-bearing greisen deposits associated with post-tectonic reduced, strongly fractionated, S-type granite (e.g. Anchor and Royal George). Base metal vein deposits are known but are insignificant. Kaolinite is found in the haloes of greisen Sn deposits and within alluvial Sn deposits. Kaolinite of debatable hydrothermal or supergene origin from Tonganah, near Scottsdale, has been exploited as paper filler.

In contrast, western Tasmania contains a wide variety of reactive host rocks including dolomite, limestone and ultramafic rocks. This has led to a much more diverse suite of deposit styles, the most important of which are sulfide and silicate skarns.

World-class calcic scheelite skarns on King Island (Dolphin and Bold Head) are in the contact aureole of I-type, moderately oxidised unfractionated granodiorite (Kwak, 1987; Solomon and Groves, 1994), whereas at the Kara deposit calcic magnetite-scheelite-bearing skarn abuts moderately fractionated, moderately to strongly oxidised granite (Zaw and Singoyi, 2000). Calcic magnetite-fluorite skarn with minor Sn and W is associated with strongly fractioned, moderately reduced I-type granite at Moina, where there are also distal Au ± Bi ± Cu skarns (Figure 5).

Distal sulfide skarn Sn deposits (Renison Bell, Mount Bischoff, Cleveland, Queen Hill and Razorback; Figure 5) are the most economically important and the former two, both world-class deposits, are associated with reduced, moderately to strongly fractionated I-type granite. Fracture filling quartz-arsenopyrite-pyrrhotite-cassiterite-fluorite veins occupy faults that were active during ore formation at Renison Bell. These were the hydrothermal feeders to the dolomite-replacement pyrrhotite cassiterite horizons (Kitto, 1994; Patterson *et al.*, 1981) that have been the dominant source of ore. At Mount Bischoff, cassiterite-bearing greisenised quartz porphyry dykes were the conduits for ore formation and form a significant part of the Sn resource (Halley and Walshe, 1995; Solomon and Groves, 1994). At the Cleveland mine the subeconomic dyke-hosted deposit (Foley's Zone) contains significant Mo, W and Bi as well as Sn (Collins, 1981; Jackson, 1992) and also appears to have been a fluid conduit.

Tin silicate skarns lie in the immediate contact zones of reduced, fractionated I-type and indeterminate-type granites. These deposits may be metallurgically difficult depending upon the partitioning of Sn between cassiterite and a variety of silicate, oxide and borate minerals (Kwak, 1987). The most important proximal deposits are the Mt Lindsay (Figure 5) and nearby magnetite-cassiterite-scheelite-bearing skarns, on which a full feasibility study is currently being undertaken.

Granites associated with Sn mineralisation contain evidence of a component of mantle melts and volatiles (e.g., Sun and Higgins, 1996; Solomon and Groves, 1994; Walshe et al., 1996), which may have been related to mafic underplating of the crust associated with post-collisional slab break-off (Black et al., 2010).

The most unusual deposit is the Avebury Ni deposit, which lies in the aureole of the strongly fractionated, reduced Heemskirk white granite (Figure 5). There is debate on whether it is a skarn or a hydrothermally-altered more conventional type of magmatic Ni deposit. Disseminated Ni mineralisation, mostly as pentlandite, occurs within Cambrian serpentinite adjacent to structurally overlying basalt and mudstone-lithicwacke-minor carbonate sequences. There are two main gangue assemblages; antigorite-magnetite-chromite and tremolite-diopside-magnetite, both with pentlandite- pyrrhotite-millerite-arsenides (Callaghan and Green, cited in Green et al., 2012). Keays et al. (2009) showed that there was positive correlation of Ni with Pd and Au contents in mineralised ultramafics, but there was an apparent negative correlation between Ni and Ir and no correlation between Pt and Ni. These features are consistent with Au and Pd being added to the ores together with Ni in the same Devonian metasomatic event. Lygin et al. (2010a, b) provided further evidence for a hydrothermal origin, including a lack of PGE inclusions in pentlandite and pyrrhotite and contamination of these minerals with Pb, Bi and As, both in fine grained mineral inclusions and incorporated in the sulfide structures. Chrome spinels are strongly altered and are veined and partly rimmed and replaced by magnetite and chrome magnetite. Boundaries between Cr-spinel cores and Cr-magnetite or magnetite overgrowths are commonly marked by the presence of Si, Pb, Sb and locally Cu. Magnetite is intimately associated with pentlandite and is relatively enriched in Sn, pointing to a hydrothermal origin. Altered Cr-spinel has lower Cr/(Cr + Al) ratios and may be enriched in Zn and Mn. Whole rock geochemistry indicates that, while incompatible trace element contents in the ultramafic are very low, there are distinct positive anomalies for W, U, Pb, Bi, Mo, Sn and Sb when data are plotted on a primitive mantle-normalised spidergram.

Tin greisen (e.g., Sweeneys and Federation) and vein deposits of Sn (Pieman vein at East Renison), W (Oakleigh Creek), Cu (Murrays Reward), and Pb-Ag-Zn ± Au (Magnet, Zeehan and Mount Farrell fields; Figure 5) were historically important producers in western Tasmania and are currently being explored. Conspicuous haloes occur around the reduced, fractionated granites associated with Sn mineralisation, but are relatively rare around the moderately or unfractionated granites with marginal scheelite skarns.

The movement of mineralising fluids outward from the granitic intrusions to favourable sites for ore formation was assisted by the complex pre-existing network of major faults which was largely the net result of Cambrian and Devonian orogenesis. Mineralised structures were initiated or reactivated at the time of granite emplacement and occur in a variety of settings including fractures formed in a regional stress field above granite cupolas (Zeehan Pb-Ag-Zn veins) or ridges (Lakeside and Mount Farrell field), faults tangential to margins of granite intrusions (Aberfoyle and Renison Bell) and porphyry dykes that also acted as hydrothermal conduits and form a radiating array above an inferred granite cupola (Mount Bischoff).

## Post Devonian

Large-scale erosion followed the close of Devonian orogenesis. Deposition restarted in the late Carboniferous with 1.5 km of generally flat-lying sedimentary rocks of late Carboniferous–Late Triassic age deposited in the Tasmania Basin. These consist of a lower glacial and glaciomarine and subordinate terrestrial sedimentary rocks, overlain by fluvial and lacustrine sedimentary rocks, largely of Triassic age. Both units contain subordinate coal measures (Seymour et al., 2006). Large volumes of Jurassic dolerite, in the form of slightly discordant sheets and dykes crop out over a large part of Tasmania and contribute to its scenery. A Cretaceous felsic alkaline porphyry complex at Cygnet in southeast Tasmania is associated with minor Au mineralisation (Taheri and Bottrill, 1999).

Paleocene weathering produced Ni laterite deposits near Beaconsfield, which are the subject of a feasibility study. Placer Sn deposits have been a significant contributor to the economy of northeastern Tasmania. The bulk of production has come from upper Oligocene braidplain deposits around the southeastern edge of the Ringarooma Valley (Morrison, 1989).

Paleogene and Neogene basalt flows are widespread. Alluvial heavy minerals, including rutile and zircon, have been mined from Pleistocene strandlines on King Island, and production is currently being revived.

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Thoughtful reviews by David Huston and Ken McQueen resulted in substantial improvements to the manuscript.

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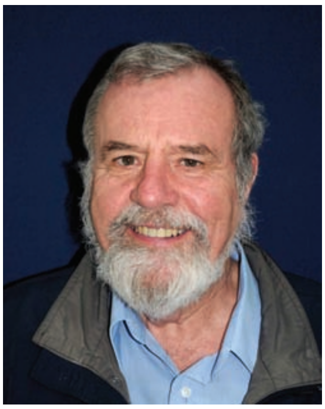

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# Mesoarchean to Mesoproterozoic evolution of the southern Gawler Craton, South Australia

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*The Gawler Craton preserves a complex and prolonged tectonic history spanning the interval c. 3200–1500 Ma. Reworking of Paleoarchean, c. 3400–3250 Ma crust led to the formation of c. 3150 Ma granites now exposed within a narrow belt in the eastern Gawler Craton. Following this, there is no known record of significant tectonic activity until the onset of bimodal magmatism during the Neoarchean to earliest Paleoproterozoic, c. 2560–2470 Ma. This magmatism was terminated by high temperature metamorphism and deformation during the 2465–2410 Ma Sleafordian Orogeny. Magmatic events associated with widespread sedimentation over the interval c. 2000–1740 Ma largely sources this older crust. The c. 1730–1690 Ma Kimban Orogeny reworked these Paleoproterozoic basins and the Neoarchean basement in a pre-dominantly transpressional orogenic system. Juvenile mantle input followed by wide-spread crustal melting occurred over the interval c. 1620–1570 Ma. This period of intense magmatism initiated with emplacement of the relatively juvenile c. 1620–1608 Ma St Peter Suite. This was followed by the economically significant c. 1600–1570 Ma Gawler Range Volcanics/Hiltaba Suite magmatic event, which resulted from widespread mid-crustal melting. Synchronous deformation and high temperature metamorphism accompanied the Gawler Range Volcanics/Hiltaba Suite magmatic event indicating it occurred in an orogenic environment. Far field stress was distributed around a central core zone of largely undisturbed Gawler Range Volcanics with deformation localised in the northern and southern Gawler Craton. The Gawler Range Volcanics/Hiltaba Suite magmatic event resulted in formation of a province of major economic significance that includes the giant Olympic Dam Cu-Au-U ore body.*

## Introduction

The Gawler Craton preserves a complex and prolonged tectonic history spanning the interval c. 3200–1500 Ma and includes Mesoarchean gneisses which are the oldest rocks in Australia outside of the Western Australian shield (Fraser et al., 2010a). The evolution of the Gawler Craton is dominated by Neoarchaean to Mesoproterozoic magmatic and mineralising events, and includes the formation of the giant Olympic Dam Cu-Au-U deposit. Olympic Dam occurs within an extensively altered and mineralised belt that is host to several other deposits and prospects of the iron-oxide copper gold (IOCG) style and related copper-gold mineral systems (Skirrow et al., 2007). This Cu-Au province along the eastern margin of the Gawler Craton receives considerable attention from mineral explorers and economic geologists, being a type-locality of the breccia-hosted IOCG deposit class (Groves et al., 2010).

Prior to the Jurassic-Cretaceous breakup of Australia and Antarctica, the Gawler Craton was part of a larger continental entity, the Mawson Continent (Figure 1; Fanning et al., 1996). Counterparts occur on the coast of Terre Adélie and George V Land (Peucat et al., 1999; Goodge and Fanning, 2010) and in the Nimrod Group of the Miller Range, which have been correlated with the Kimban Orogeny in the Gawler Craton (Goodge et al., 2001). More generally, similarity of the satellite-derived geophysical imagery of the Gawler Craton and the correlative region under the Antarctic ice sheet (Finn et al., 2006) together with geochronology from Antarctica indicate a Proterozoic crustal province of considerable extent (Figure 1; Fitzsimons, 2003; Payne et al., 2009). The relationship between the Mawson Continent and Laurentia is also of interest with numerous reconstructions of Rodinia placing the Mawson Continent proximal to western Laurentia (Goodge et al., 2001), providing a spatial relationship for the contemporaneous c. 1590 Ma IOCG breccias of the Gawler Craton and c. 1590 Ma IOCG breccias of the Wernecke Supergroup, northwestern Laurentia (Thorkelson et al., 2001).

Central to developing paleogeographic reconstructions that involve the Gawler Craton (e.g. Myers et al., 1996; Cawood and Korsch, 2008), or to generating predictive models for metallogenesis, is a detailed understanding of the stratigraphic and tectonic events preserved within it. In this paper we briefly review the geology of the Gawler Craton, focusing on the southern portion of the province in order to provide a framework for its evolution. We consider the lithostratigraphic composition and examine the cycles of orogenic and magmatic reworking evident within the Gawler Craton.

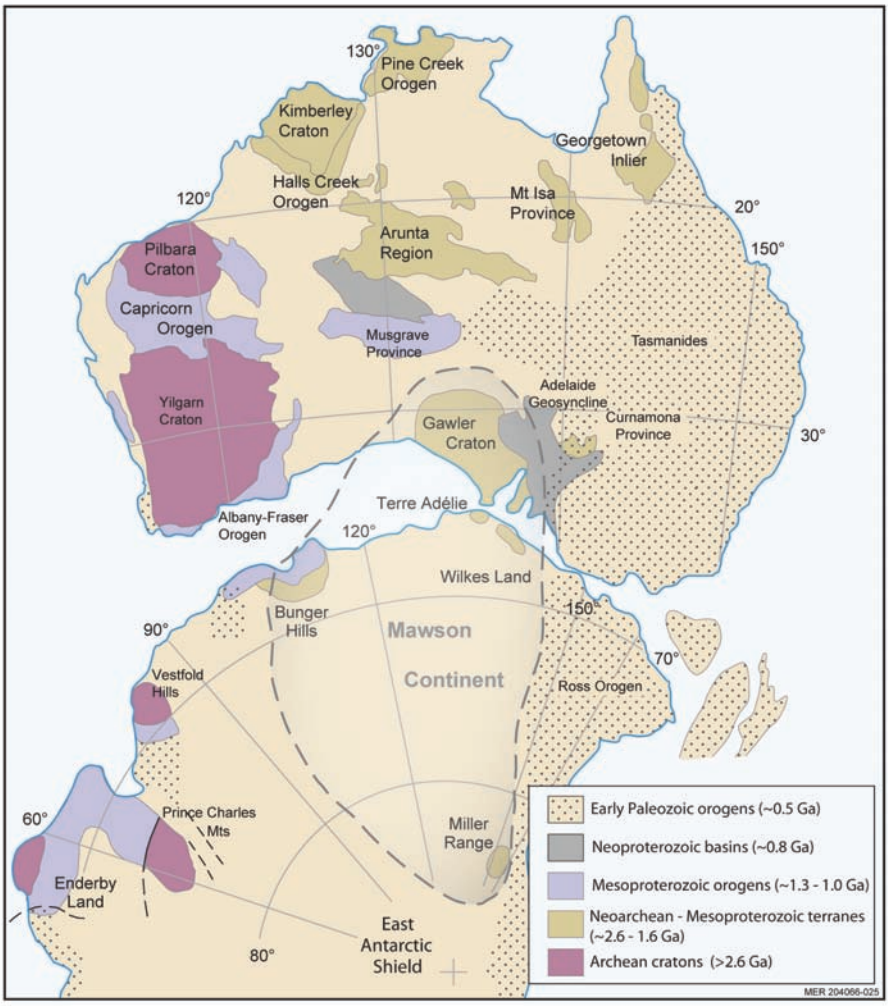


*Figure 1 Location of the Gawler Craton in the context of Gondwana reconstruction –160 Ma. After Fitzsimons (2003) and Myers et al. (1996).*

## Macroscale architecture of the Gawler Craton

The southern boundary of the Gawler Craton is the continental margin developed during rifting of Australia from Antarctica. The other boundaries are poorly constrained, being largely defined by the thickness of Neoproterozoic–Neogene cover sequences. The eastern boundary coincides with the Torrens Hinge Zone, the transitional zone between thick, folded Neoproterozoic sediments of the Adelaide Rift Complex and flat lying cover sequences of equivalent age that cover large region of the eastern Gawler Craton (Parker, 1990). It is probable that part of the Gawler Craton is the basement beneath the Adelaide Rift Complex. Likewise, the northern and western boundaries coincide with deep burial by Neoproterozoic–Paleozoic successions of the Officer Basin (Korsch et al., 2010). The nature of the boundary between the Gawler Craton and the adjacent c. 1600–1080 Ma Musgrave Province to the north (Figure 1) is poorly understood. Magnetotelluric experiments over the transition zone reveal a conspicuous lack of any major electrical discontinuity, which might be expected were there to be some type of ancient suture between the two provinces (Selway et al., 2011). Nevertheless, the Musgrave Province is composed of isotopically more juvenile material (Wade et al., 2008) and cannot simply be the northern continuation of the Gawler Craton. Deep crustal seismic data reveal crustal-scale north-dipping structures in the northern Gawler Craton, which may form part of a transition zone between the two provinces (Korsch et al., 2010).

The southern Gawler Craton is exposed on Eyre and Yorke peninsulas (Figure 2). The dominant strike-direction of the deformed rocks is N-S and is largely due to the structural grain imposed during the c. 1730–1690 Ma Kimban Orogeny. The Kalinjala Shear Zone corresponds to a major discontinuity in geophysical data sets (Fraser et al., 2010b) and appears to separate zones of differing lithostratigraphic composition (Figure 3). This shear zone may be of fundamental significance in understanding the amalgamation of the proto-Gawler Craton (Hand et al., 2007).

## Lithostratigraphic packages of the southern Gawler Craton

### Mesoarchean–Neoarchean of northeastern Eyre Peninsula

Mesoarchean granitoids, emplaced between c. 3200–3150 Ma (Fraser et al., 2010a; Jagodzinski et al., 2011b), are exposed in the northeastern Eyre Peninsula (Figure 2). Inherited zircons, with ages up to c. 3400 Ma, occur within these granitoids suggesting still older crustal material is present at depth. This is also suggested by the geochemistry of the c. 3150 Ma Cooyerdoo Granite, which has characteristically elevated LREE (light REE) contents and low Na/K, and may be post-tectonic in origin, the product of melting a pre-existing tonalite–trondhjemite–granodiorite (TTG) crust (Fraser et al. 2010a).

A number of the c. 3150 Ma samples from this region contain c. 2500–2510 Ma metamorphic zircons and are associated with similarly aged leucogranites (Fraser et al., 2010a; Jagodzinski et al., 2011b). The gneissic fabric within the Mesoarchean granitoids may have developed during this Neoarchean event, and we note that c. 2510 Ma is an interval of metamorphic zircon growth for which there is no equivalence in the Neoarchean rocks that dominate the central and south-western portion of the craton (Figure 3). Possible Neoarchean sedimentary rocks also occur to the east of the Kalinjala Shear Zone, within the Middleback Ranges. Detrital zircons from these rocks yield maximum depositional ages c. 2560 Ma (Jagodzinski et al., 2011a; Szpunar et al., 2011).

### Neoarchean–early Paleoproterozoic complexes, western Eyre Peninsula and central-northern Gawler Craton

Most of the Archean units in the Gawler Craton occur in two belts of latest Neoarchean–earliest Paleoproterozoic rocks, the

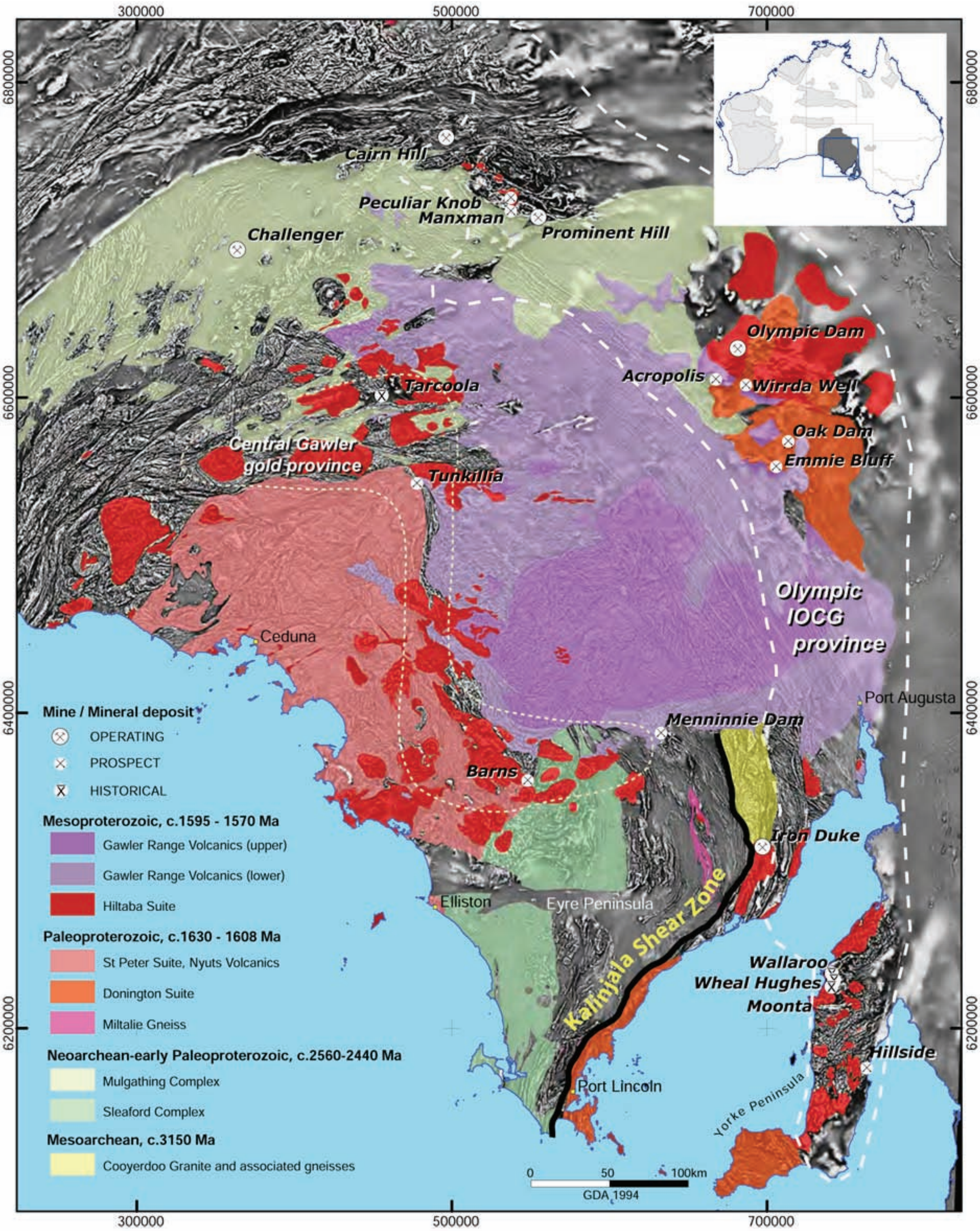


***Figure 2 Interpreted solid geology of the southern Gawler Craton, shown over a 1st vertical derivative magnetic intensity image (PIRSA data).***

Mulgathing Complex and Sleaford Complex (Figure 2). Although they are probably contiguous, any structural continuity between the two is concealed by the Gawler Range Volcanics and younger cover (Figure 2). The age of basement to the Mulgathing and Sleaford complexes is uncertain. It is possible that correlatives of the Cooyerdoo Granite underlie some parts, as rare inherited zircons and whole rock Nd isotopes indicate the presence of Paleo- to Mesoarchean crust, c. 3400–2800 Ma (Daly and Fanning, 1990, 1993; Fanning et al., 2007; Jagodzinski et al., 2009; Fraser and Neumann, 2010). The oldest rock within these complexes is the protolith of the Coolanie Gneiss, which was emplaced at 2823 ± 37 Ma (Fraser and Neumann, 2010) although the dominant rock forming interval in both the Mulgathing and Sleaford complexes was between c. 2555–2480 Ma.

The oldest Neoarchean unit in the Mulgathing Complex is the

c. 2555 Ma bimodal Devils Playground Volcanics, which have a calc-alkaline composition, moderately juvenile $\varepsilon Nd_{2555Ma}$ values (-2.3–3.1; Figure 4), Nb-, Ta-, Ti-depletions, and elevated LREE signatures (Reid et al., 2009). They have been interpreted to have formed in a continental magmatic arc setting (Swain et al., 2005). Correlates of the Devils Playground Volcanics have not been identified in the Sleaford Complex.

The c. 2520 Ma Hall Bay Volcanics are dominatly felsic, with some basaltic and komatiitic members (Teale et al., 2000). Coeval komatiitic volcanics also occur in the c. 2520 Ma Harris Greenstone Belt in the central Gawler Craton. These are typical Al-depleted Archean komatiites derived from a depleted mantle source (Hoatson et al., 2005). Clastic and chemical sedimentation, including carbonates and banded iron formations, took place between c. 2520 Ma until c. 2485 Ma, in both the Mulgathing and Sleaford complexes (Swain et al., 2005; Fanning et al., 2007; Jagodzinski et al., 2009). Syn-sedimentary intrusives (e.g., the 2519 ± 8 Ma Coulta Granodiorite of the Sleaford Complex; Fanning et al., 2007), are also known.

The abundance of bimodal magmatism and continuity of sedimentation across the Archean–Paleoproterozoic boundary suggests that the overall tectonic setting during the Neoarchean–earliest Paleoproterozoic of the Gawler Craton is most likely to have been extensional. The resulting high heat flow may have primed the crust for the subsequent high temperature metamorphism and deformation associated with the c. 2465–2410 Ma Sleafordian Orogeny, an event which terminated deposition within the volcano-sedimentary basin.

The relationship between the Mulgathing and Sleaford complexes to any possible Neoarchean sedimentary rock in the Middleback Ranges is not clear. Metasedimentary rocks in the Middleback Ranges record only the effects of the Kimban Orogeny (Jagodzinski et al., 2011a), whereas the Neoarchean c. 2510 Ma event is only recorded in the orthogneisses to the west of the Middleback Ranges (Jagodzinski et al., 2011b). It is possible that the metasedimentary units in the Middleback Ranges were in fact deposited sometime after the 2510 Ma metamorphic event, or that there is a structural discontinuity between the gneisses and the metasedimentary rocks.

## Paleoproterozoic magmatism

Recurrence of magmatism in the Gawler Craton took place at c. 2000 Ma (Fanning et al., 2007) when a series of intrusives were emplaced in the southern Gawler Craton producing the felsic protoliths to the Miltalie Gneiss. Since the Miltalie Gneiss occurs structurally below metasedimentary units of the Hutchison Group (Parker et al., 1993), the protoliths to the Miltalie Gneiss have been interpreted to represent stitching granites formed during extension that signalled the onset of basin formation (Daly et al., 1998). However, very little is known of the geochemical and petrogenetic affinities of the Miltalie Gneiss.

At c. 1850 Ma, the Donington Suite was emplaced along the eastern margin of the Gawler Craton, extending some 600 km from the southern-most coastal outcrops to the vicinity of Olympic Dam (Figures 2 and 3) in the north. The Donington Suite ranges in composition from granite to charnockite and includes a mafic unit, the Jussieu Metadolerite. The Donington Suite is enriched in elevated incompatible elements and LREE and has $\varepsilon Nd_{1850Ma}$ values between -4 and -2 (Figure 4). It is interpreted to have evolved from a contemporary mantle source substantially contaminated by Archean lower crust (see Reid et al., 2008a and references therein). The Donington Suite was associated with a brief compressional orogenic phase recorded only in Donington Suite units on Yorke Peninsula, east of the Kalinjala Shear Zone and known as the Cornian Orogeny (Reid et al., 2008a).

## Paleoproterozoic volcano-sedimentary basins

Several volcano-sedimentary packages in the southern Gawler Craton formed during the interval c. 1865–1740 Ma (Figure 3). While the lithostratigraphic units referred to above occur either east or west of the Kalinjala Shear Zone, basin development during the c. 1865–1740 Ma period occurred at different times on both sides of the shear zone. The basal group of this basin system is the Hutchison Group (Parker et al., 1993). Recent zircon studies suggest that the Hutchison Group is an amalgamation of units of differing provenance and maximum depositional ages (Szpunar et al., 2011). They recognised two depositional packages: the c. 1860 Ma Darke Peak Group and the c. 1790 Ma Cleve Group. Detrital zircons in the Darke Peak Group have age-components at c. 2520–2440 Ma and c. 2000 Ma (see also Warrow Quartzite samples in Fanning et al., 2007), reflecting a predominantly local source, an inference supported by the presence in southern Eyre Peninsula of an unconformity between the Warrow Quartzite (basal package to the erstwhile Hutchison Group), and the underlying c. 2440 Ma Kiana Granite (Fanning et al., 2007). The Warrow Quartzite is overlain by metamorphosed dolomitic and politic units and iron formations, all indicative of deposition on a stable shelf (Parker, 1980b). Interlayered amphibolites within the Cleve Group have continental thoelitic affinities, consistent with emplacement on a passive margin (Parker, 1993). Detrital zircons from the Cleve Group are dominated by c. 1850 Ma and c. 1790 Ma populations, suggesting the sediments may have been derived, at least in part, from the c. 1850 Ma Donington Suite (Szpunar et al., 2011). This implies that the Cleve Group was deposited proximal to the Donington Suite (Figure 3).

Volcano-sedimentary packages deposited at c. 1790 Ma include the bimodal Myola Volcanics and associated Broadview Schist in northern Eyre Peninsula. These packages were succeeded by compositionally similar volcanics and interlayered sediments of the Wallaroo Group (Cowley et al., 2003). Volcanic sequences in the Wallaroo Group include the 1772 ± 14 Ma Wardang Volcanics, 1753 ± 8 Ma Moonta Porphyry and the 1740 ± 6 Ma Mona Volcanics (Fanning et al. 2007). Detrital zircons from Wallaroo Group equivalent units, including those present to the north in the vicinity of Olympic Dam, typically contain c. 1850 Ma and c. 1790 Ma detrital zircons (Jagodzinski, 2005; Reid et al., 2011) suggesting a local source for the sediment and that the Donington Suite, or equivalents, was exposed during the deposition of the c. 1790–1740 Ma packages. Other temporally equivalent sedimentary packages include the c. 1770 Ma Price Metasediments (Oliver and Fanning, 1997) on southwestern Eyre Peninsula and the metasedimentary rocks in the northern and western Gawler Craton (Payne et al., 2006; Howard et al., 2011). Detrital zircon age data suggest the sequences in the eastern Gawler Craton have different source regions to those in the western and northern parts of the craton (Payne et al., 2006; Fanning et al., 2007; Howard et al., 2011). Sm-Nd isotopic data suggests the c. 1790 Ma and c. 1740 Ma bimodal volcanics were largely derived from pre-existing continental crust, with $\varepsilon Nd_{1740Ma}$ values for the McGregor Volcanics between -3 and 0 (Turner et al., 1993; Szpunar and Fraser,

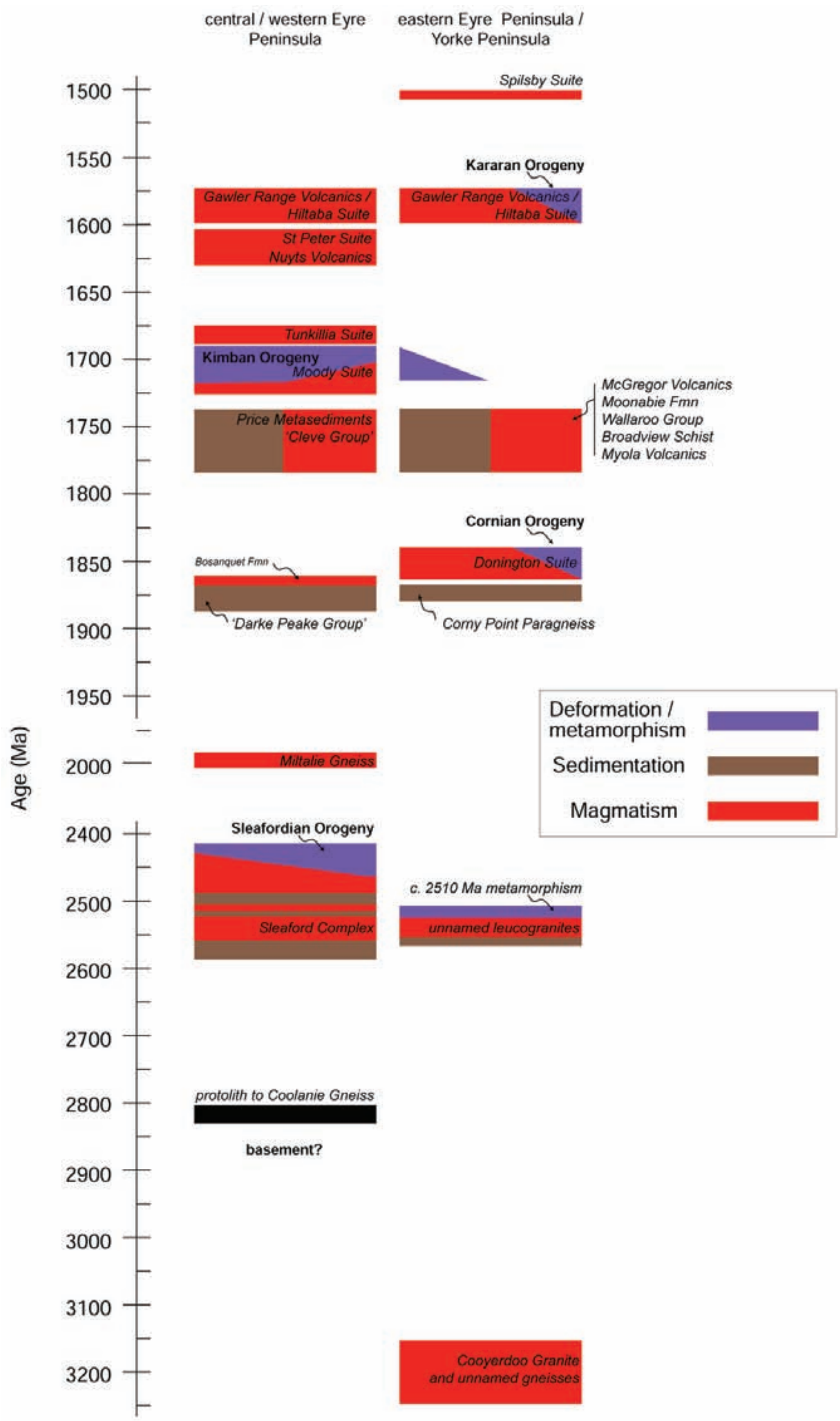


***Figure 3** Time-space plot for the southern Gawler Craton. The division into eastern Eyre Peninsula-Yorke Peninsula vs central-western Eyre Peninsula reflects the approximate location of the crustal-scale discontinuity represented by the Kalinjala Shear Zone.*

2010). Sedimentation in the southern Gawler Craton was terminated by the c. 1730–1690 Ma Kimban Orogeny.

## Paleo- to early Mesoproterozoic magmatism

Following the Kimban Orogeny, the lithostratiography of the Gawler Craton changed from dominantly volcano-sedimentary basins to the emplacement of voluminous magmatic suites (Figure 3). The c. 1690–1670 Ma Tunkillia Suite was emplaced during the waning stages of the Kimban Orogeny and represents a late to post-orogenic magmatic event (Payne et al., 2010), as the age of deformation in the western Gawler Craton overlaps with the age of the Tunkillia Suite (Howard et al., 2011). Limited sedimentation and volcanism (Tarcoola Formation) occurred in the central Gawler Craton at c. 1660 Ma in a local rift setting (Daly et al., 1998).

The c. 1630 Ma rhyodacitic–rhyolitic Nuyts Volcanics occur in the southwestern Gawler Craton (Rankin et al., 1990) and are succeeded by a more-extensive, bimodal, c. 1620–1608 Ma intrusives (St Peter Suite; Flint et al., 1990; Fanning et al., 2007). Felsic and mafic rocks from the St Peter Suite are characteristically juvenile, with pronounced Nb and Ti anomalies, marked Y depletion, moderate to high Sr along with $\varepsilon Nd_{1620Ma}$ values between -2 and +2 (Swain et al., 2008), reflecting the formation of new continental crust from a mantle source. They argue that the St Peter Suite represents a continental magmatic arc and that the Gawler Craton was, therefore, the hinterland of a plate margin at this time. However, as Hayward and Skirrow (2010) have suggested, the St Peter Suite could simply reflect I-type magmatism derived in part from metasomatised subcontinental lithospheric mantle, in which case there is no need to invoke contemporaneous subduction to explain the petrogenesis of this suite.

St Peter Suite magmatism was followed by the c. 1592 Ma Gawler Range Volcanics and co-magmatic c.1600–1570 Ma Hiltaba Suite. The Gawler Range Volcanics forms part of a felsic large igneous province, with correlative units in the Curnamona Province (Benagerie Volcanic Suite), estimated to occupy some 100,000 km$^3$ (Wade et al., 2012). There may also be correlatives in Antarctia (Peucat et al., 2002). The Gawler Range Volcanics are dominantly felsic, although minor basalts are also present (Allen et al., 2008), indicating widespread crustal melting was associated with mantle melting. Gabbroic intrusives are also present within the Hiltaba Suite such as the Curamulka Gabbronorite from central Yorke Peninsula (Zang et al., 2007). Hiltaba Suite granites (Flint et al., 1993) occur throughout the central and southern Gawler Craton, and are implicated as sources of heat (± metals?) for the IOCG mineral system of the eastern Gawler Craton (Skirrow et al., 2007; Groves et al., 2010).

Sm-Nd isotopic data indicate that the Hiltaba Suite contains a significant component of pre-existing crust (Stewart and Foden, 2003). The composition individual plutons varies with their country rocks. For example, Hiltaba Suite granites that intruded the St Peter Suite are significantly more juvenile ($\varepsilon Nd_{1590Ma}$ = 0.1–1.2) than those that intruded the Mesoarchean gneisses of north-eastern Eyre Peninsula (e.g., Charleston Granite, $\varepsilon Nd_{1590Ma}$ = -13.7 to -7.3). This isotopic variation is mirrored in the associated early Mesoproterozoic mineral

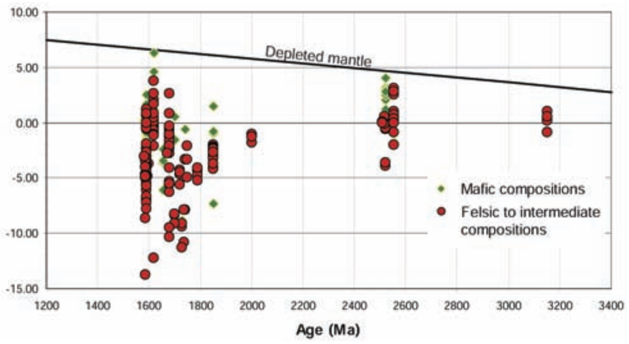


***Figure 4 Compilation of εNd values for igneous rock suites of the Gawler Craton. Data derived from published references and unpublished PIRSA and University of Adelaide data and that the data compilation is available upon request from the authors.***

systems of the Gawler Craton. The Olympic IOCG province occurs in the eastern part of the craton (Skirrow et al., 2007), whereas Au-dominated systems occur in the central part (Ferris and Schwarz, 2003). The spatial distribution of IOCG versus Au-dominated mineral systems is controlled, to some extent by the composition of associated Hiltaba Suite granites. The more evolved, U- and Th-richer, and oxidized granites occur in the Olympic IOCG province (Budd et al., 2001). Modern-day heat flow in the IOCG province is also significantly higher (90 ± 10 $mWm^{-2}$) than that of the Au-dominated province (54 ± 5 $mWm^{-2}$). This suggests lithospheric compositional differences between the eastern and western/central Gawler Craton. These differences may reflect an older phase of craton assembly (Hand et al., 2007). The presence of the Kalinajala Shear Zone along the western boundary of the IOCG province suggests that strain was partitioned along a fundamental lithospheric boundary during the Kimban Orogeny.

The volumetrically minor c. 1500 Ma Spilsby Suite occurs in the southern Gawler Craton on a number of islands in the Spencer Gulf and at Corny Point on Yorke Peninsula (Fanning et al., 2007; Jagodzinski et al., 2007); however, their petrology has not been studied in any detail.

## Orogenic framework

Several orogenies have affected the Gawler Craton. The oldest recorded metamorphic zircon growth, at c. 2510 Ma occurs within the Mesoarchean gneisses of northern Eyre Peninsula and is associated with the emplacement of leucogranites and the formation of a gneissic fabric within the Mesoarchean granitoids (Fraser et al., 2010a). It is not clear whether this deformation was part of a distinct, widespread tectonothermal event.

The c. 2465–2410 Ma Sleafordian Orogeny resulted in high-temperature metamorphism, isoclinal folding, and transpressional deformation of the supracrustal sequences within the Mulgathing and Sleaford complexes (McFarlane, 2006). The Sleafordian Orogeny is best expressed in the Mulgathing Complex, as the effects of the Kimban Orogeny have largely overprinted Sleafordian-aged fabrics elsewhere (Dutch et al., 2010).

The c. 1855–1840 Ma Cornian Orogeny is defined within the Donington Suite cropping out on Yorke Peninsula. It is characterised by migmatites and orthogneisses, and syn-kinematic granites reworked by late-stage, broadly south-directed extensional fabrics (Reid et al., 2008a). No record of the Cornian Orogeny is found west of the Kalinjala Shear Zone.

The principal orogenic event of southern Gawler Craton is the c. 1730–1690 Ma Kimban Orogeny (Parker et al., 1993). On Eyre Peninsula, the Kimban Orogeny is characterised by transpressional deformation in a belt up to 100 km wide, including the Kalinjala Shear Zone, a subvertical high-strain zone, 4–6 km wide, along eastern Eyre Peninsula (Parker, 1980a; Vassallo and Wilson, 2002; Dutch et al., 2008, 2010). The Kalinajala Shear Zone and its flanking structures mark the eastern-most limit of Kimban-aged deformation in the Gawler Craton, with rock to the east (such as the Wallaroo Group) showing little or no effects of Kimban deformation. In contrast to the high grade metamorphism evident in southern Eyre Peninsula, the metamorphic grade in northern Eyre Peninsula shows more variability, with amphibolite facies shear zones (Reid et al., 2008b) interspersed with granulites (Fraser and Neumann, 2010). Deformation in northern Eyre Peninsula is characterised by fold-thrust systems that verge to the east, away from the orogenic core (Parker et al., 1993). Thus, in the southern Gawler Craton, the structural architecture of the Kimban Orogeny forms an obliquely exposed crustal-scale positive flower structure (Hand et al., 2007).

The Kimban Orogeny is also recorded in strongly deformed Paleoproterozoic metasedimentary sequences in the northern and western Gawler Craton (Payne et al., 2008; Howard et al., 2011; Jagodzinski and Reid, 2010) indicating that, aside from regions to the east of the Kalinjala Shear Zone, the Kimban Orogeny was virtually craton-wide (Fanning et al., 2007). Syn-Kimban sedimentation is recorded in the central Gawler Craton, where the c. 1715 Ma Labyrinth Formation contains clastic material derived from local sources (Daly et al., 1998).

The next major phase of reworking occurred during the interval c. 1600–1550 Ma and is broadly termed the Kararan Orogeny, although Hand et al. (2007) recognise a slightly more complex orogenic history over this period than is summarised here. The apparent lack of deformation in the Gawler Range Volcanics and Hiltaba Suite in central Gawler Craton has led many workers to infer an anorogenic setting for these high-temperature felsic igneous rocks (Flint et al., 1993; Allen and McPhie, 2002). However, it is clear that deformation and high-temperature metamorphism did occur across the Gawler Craton during the time and that it continued for several tens of millions of years (Hand et al., 2007). Examples of Kararan-aged tectonism include greenschist facies fabrics within the Wallaroo Group and Hiltaba Suite granites on Yorke Peninsula (Conor, 1995), syn-Hiltaba deformation accompanied by cooling of mid-crustal rocks to below c. 500°C on Eyre Peninsula, (Foster and Ehlers, 1998), and overprinting of Kimban-aged fabrics adjacent the Kalinjala Shear Zone (Hand et al., 2007). Further, high- to ultra high-temperature metamorphism in the Coober Pedy Ridge and adjacent Mt Woods Domain, of the northern Gawler Craton, also occurred c. 1585 Ma (Cutts et al., 2011; Forbes et al., 2011). This suggests that this orogenic phase was widespread, yet partitioned into zones of deformation that increase in intensity to the north and east, away from the central Gawler Craton, which may have acted as a strain-buffer during this event (Figure 5; Hand et al., 2007, 2008). The formation of the Gawler Range Volcanics internal to zones of active deformation has led to the suggestion that the Gawler Range Volcanics may be part of a foreland basin fill (Hand et al., 2008).

Post-Kararan reworking in the Gawler Craton is restricted to the growth of muscovite in major structures in the western Gawler Craton

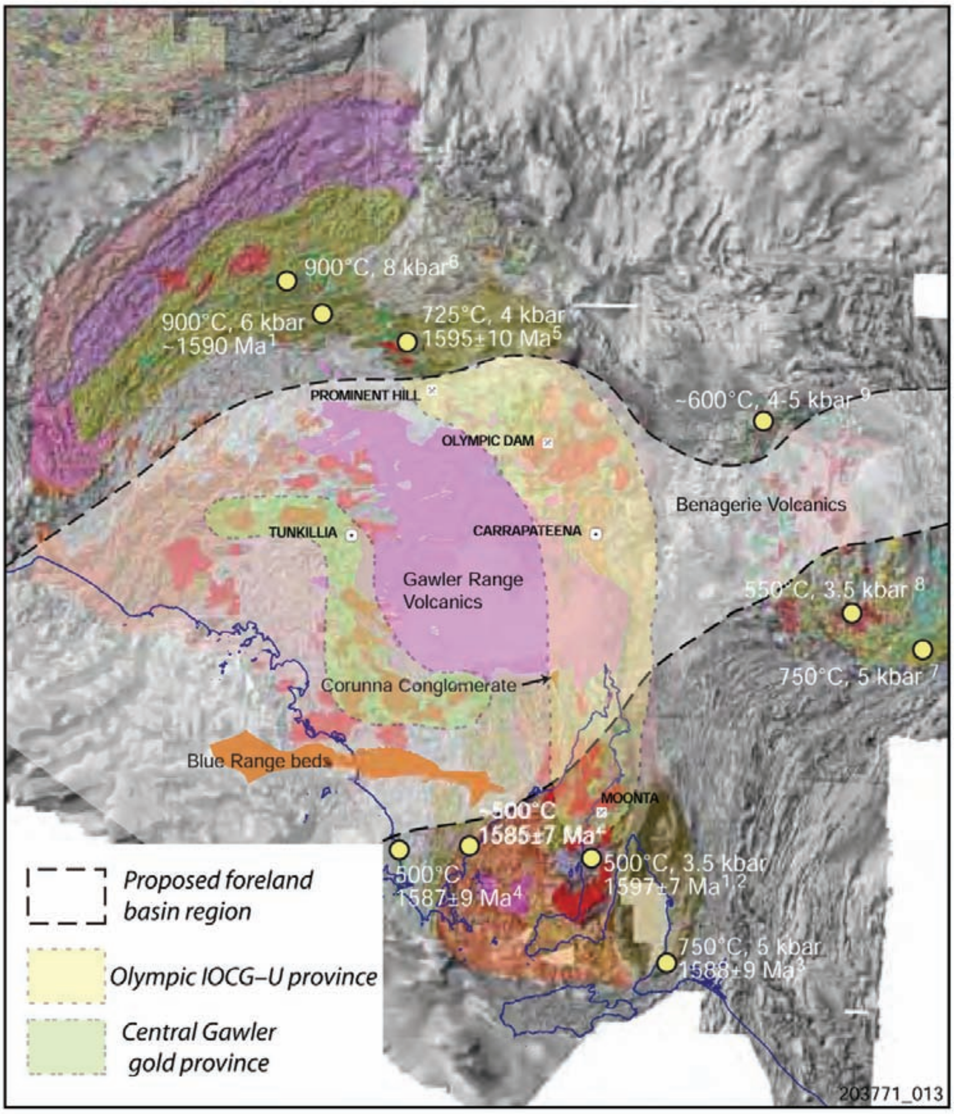


***Figure 5 Inferred early Mesoproterozoic NW-SE contractional tectonic system in southern Australian Proterozoic terranes with a foreland region occupied by the Gawler Range Volcanics and Benagerie Volcanic Suite flanked by regions of deformation. Also shown are outlines of the Olympic IOCG and central Gawler Au provinces. Data sources are given in Hand et al. (2008).***

at c. 1450–1400 Ma (Fraser and Lyons, 2006). Regional cooling of the craton to below closure of biotite in the K-Ar system, c. 300°C, occurred by 1400 Ma (Webb et al., 1982).

## Magmatic reworking through time

The early period of magmatism within the Gawler Craton is distinctly punctuated, with Mesoarchean and Neoarchean–early Paleoproterozoic events that were followed by a period of some 250 million years, between c. 1850–1570 Ma, of frequent and abundant magmatism (Figure 4). The Nd isotopic signatures of the Mesoarchean Cooyerdoo Granite are somewhat evolved, suggesting they derived from pre-existing felsic crust (Fraser et al., 2010a). The Neoarchean magmatism, c. 2555–2500 Ma, displays a significantly more juvenile character, particularly related to widespread mafic magmatism, involving a significant mantle component in addition to the likely remelting of pre-existing crustal material (Figure 4).

Magmatism occurring between c. 2000–1670 Ma has εNd values that define a trend expected if their source was reworked Neoarchean crust (Figure 4) e.g., younger units, such as the c. 1850 Ma Donington Suite have εNd values varying from -7.4 to 1.45 whereas units aged between c. 1790–1730 Ma have εNd values from -11.3 to -0.58. This trend suggests an increase in the proportion of crustal melting driven by thermal and material input from the mantle that occurred in the c. 60 million years leading up to the onset of the Kimban Orogeny. The post-Kimban Tunkillia Suite has εNd values from -10.4 to 2.6, implying both mantle and crustal input, as would be expected for post-orogenic magmatism occurring in an extensional setting.

Emplacement of the St Peter Suite signaled a change in the pattern of εNd values, with a marked shift towards more juvenile compositions (Figure 4). The associated mantle input immediately pre-dates the voluminous magmatism of the c.1600 – 1570 Ma Gawler Range Volcanics and Hiltaba Suite. They show a large range in εNd values (-13.5–2.5) and trends towards a significant crustal component, consistent with studies that show the Gawler Range Volcanics/ Hiltaba Suite are dominantly derived from crustal melting (e.g., Creaser, 1995). The major juvenile input evident in the St Peter Suite likely indicates the influence of mantle-scale processes, possibly a plume (Flint et al., 1993) or magmatic arc (Swain et al., 2008). Whatever the origin, if the increased contribution from the mantle, it was likely important in generating the widespread lower crustal melting during the 1600–1570 Ma Gawler Range Volcanics/Hiltaba Suite magmatic event. Indeed it may be that the temporal gap between the onset of the c. 1620–1608 Ma, St Peter Suite magmatism and the c. 1595 Ma, evolved Gawler Range Volcanics/Hiltaba Suite magmatism reflects a period during which the thermal regime of the lower to middle crust across the Gawler Craton was re-heated to levels required for widespread anatexis, resulting Gawler Range Volcanics/Hiltaba Suite magmatic event, a felsic large igneous province of global significance.

## Conclusion

The interplay between crustal melting, episodic mantle inputs, and orogenesis is a feature of the Mesoarchean–Mesoproterozoic evolution of the Gawler Craton. Of the three major orogenic events, the 2465–2410 Ma Sleafordian Orogeny, the 1730 – 1690 Ma Kimban Orogeny, and the c. 1590–1560 Ma Kararan Orogeny, both the Sleafordian and the Kararan were proceeded by significant mantle-derived heating of the crust. Mafic magmatism in the Neoarchean is manifest as ultramafic volcanics (including komatiites) as well as mafic and mafic-related felsic igneous rocks. The juvenile mafic and felsic intrusives of the c. 1620–1608 Ma St Peter Suite may have intiated the c. 1595–1575 Ma Gawler Range Volcanics/Hiltaba Suite magmatic event and associated high-temperature metamorphism and distributed deformation. Crustal melting and thermal and material inputs from mantle resulted in rheological changes that facilitated the localisation of deformation. In each case, high-temperature metamorphism accompanied deformation indicative of the steep geothermal gradients that both preceded and resulted from the crustal reworking. In contrast, the Paleoproterozoic c. 2000–1740 Ma magmatic events that predate the Kimban Orogeny largely sourced older crust, consequently, the Kimban Orogeny does not appear to have been instigated by significant mantle melting. The drivers for this craton-wide event may have operated at a length scale beyond the present boundaries of the Gawler Craton.

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# The Flinders Ranges and surrounds, South Australia: a window on astrobiology and planetary geology

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*The Flinders Ranges and its surroundings in South Australia comprise an impressive rugged terrain that rises abruptly from piedmont plains to the east and west and merges into the plains of the Cenozoic Lake Eyre Basin to the north. Folded and faulted Neoproterozoic–Cambrian clastic and carbonate sedimentary rocks of the Adelaide Geosyncline (Adelaide Rift Complex) form the predominant geology of the ranges and record varied depositional environments and metamorphic overprints and have had a complex landscape history, resulting in a diverse regolith. This ancient, arid terrain represents some of the best analogue landscapes and settings in Australia to observe features and processes fundamental to the evolution of the Earth. The strata of the Flinders Ranges record the evolution of terrestrial surface environments and the biosphere through the Cryogenian, Ediacaran and Cambrian periods, including evidence for Neoproterozoic glaciations, orbital and rotational dynamics and asteroid impact. The diverse assemblages of stromatolites, ancient and modern hydrothermal systems, and alteration assemblages provide field laboratories for astrobiological and hyperspectral research and training. For these reasons the northern Flinders Ranges near Arkaroola have been selected as a site for multi-disciplinary Mars analogue research and space education.*

## Introduction

Analogue research involves an integrated set of disciplines investigating terrestrial features and processes which can offer insights into understanding landscapes and processes observed on other planets. Analogue features also serve as training grounds for planetary scientists and astrobiologists and as field test sites for equipment and instruments used in planetary exploration. Analogues should be distinguished from simulation, for example the dunes of the Strzelecki Desert are potentially good morphological and process analogues for martian dunes (Bishop, 1999) but they are not good simulations of martian dunes because of the different gravity, composition and atmospheric conditions between the Strzelecki Desert and Mars.

Australian geology and landscapes provide numerous opportunities for analogue research, in particular, Mars analogue research (West et al., 2009) because of the preservation of its long geological history, including extensive hydrothermal activity and a predominantly arid climate. The diversity of processes and landforms found in regions such as the Flinders Ranges and the surrounding area in central South Australia, in particular the features found within c. 100 km of Arkaroola, offer powerful insights into extraterrestrial and especially martian stratigraphy and geomorphology, along with astrobiology sites and examples. The Arkaroola region has been used for a range of Mars-related research and education activities including remote sensing (Thomas and Walter, 2002; Brown et al., 2004), landscape evolution (Waclawik and Gostin, 2006), engineering research related to planetary exploration (Waldie and Cutler, 2006) and education and outreach in all these disciplines (Laing et al., 2004; West et al., 2009).

This paper provides a brief description of some features in the Flinders Ranges and surrounds of interest to those studying astronomical influences on terrestrial geology, Mars analogues, and astrobiology that will be visited during a field trip associated with the 34th International Geological Congress (34 IGC). These are:

- Proterozoic and Phanerozoic rocks of the northern Flinders Ranges
- Regolith and landscapes of the northern Flinders Ranges
- Astrobiological features in the northern Flinders Ranges and surrounds
- Neoproterozoic stromatolites at Arkaroola
- A possible Neoproterozoic deep hot biosphere
- Fossil hydrothermal systems of the Mount Painter Inlier, Arkaroola
- Paralana Hot Springs, Arkaroola
- Cryogenian glacials, Arkaroola
- Late Cryogenian tidal rhythmites and Earth's paleorotation, Pichi Richi Pass
- Ediacaran Acraman impact ejecta horizon, Bunyeroo Formation, Central Flinders Ranges

## Geology of the northern Flinders Ranges

The geology of the Neoproterozoic Adelaide Geosyncline (also termed the Adelaide Rift Complex) is reviewed by Preiss (1987, 1993, 2000), with Coats and Blissett (1971) focusing on the Mount Painter area and its surrounds. The Paleozoic, Mesozoic and Cenozoic successions of the region are reviewed by Drexel and Preiss (1995). These works remain unsurpassed in providing a framework to understanding the geology of the region.

Arkaroola is located in the northern Flinders Ranges (Figure 1), where rocks of the Mount Painter Inlier form a basement nucleus over which a younger Neoproterozoic succession was deposited. The Mount Painter Inlier comprises Mesoproterozoic metasediments and metavolcanics (including the Radium Creek Metamorphics) intruded by granites, pegmatites and minor amphibolite dykes. The highly radiogenic nature of the Mesoproterozoic granites has resulted in a long-lasting history of hydrothermal activity that has continued to the present.

The Neoproterozoic succession includes numerous horizons of stromatolitic carbonates, evidence for two Cryogenian glaciations, and an Ediacaran impact ejecta horizon derived from the Acraman impact structure in the adjacent Gawler Ranges. The main phase of deformation was the Ordovician Delamerian Orogeny at 514–490 Ma (Drexel and Preiss, 1995; Foden et al., 2006). Ordovician granites and pegmatites form a younger granite suite which intrudes the Proterozoic basement (Hore et al., 2005).

Mesozoic and early Cenozoic sediments unconformably overlie the Proterozoic and Paleozoic rocks. This cover has been almost completely removed by later Cenozoic reactivation and uplift of the folded strata to form the Flinders Ranges. The youngest Cenozoic sediments form piedmont alluvial aprons derived from the ranges.

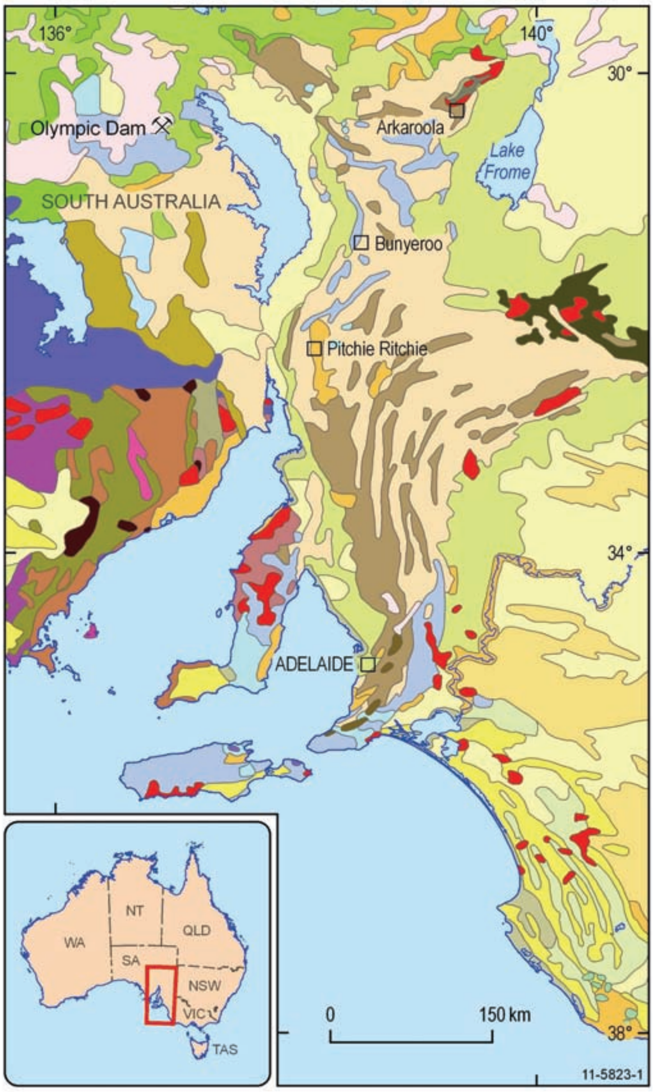


*Figure 1 Regional geological context of the study area.*

## Regolith and landscapes of the northern Flinders Ranges and surrounds

Regolith and associated paleolandscapes within the Mesozoic Eromanga Basin include a paleodrainage deposition (fluvial and minor glacial) in a landscape of moderate topographic relief and bedrock exposure (Davey and Hill, 2007; Hill and Hore, 2009). Cretaceous marine transgressions across low-lying parts of the landscape deposited chemically reduced clays and silts. The Mesozoic sediments have since been partially oxidised and tectonically disrupted, particularly during the Neogene (Davey and Hill, 2007). Early Cenozoic regolith and landscape evolution is associated with the development of the Lake Eyre Basin region and Paleogene fluvial deposits. Ephemeral fluvial, lacustrine, colluvial and aeolian deposition, as well as tectonism and pedogenesis, have been spatially and temporally variable in the more recent geological evolution (Hill, 2008).

The youngest Cenozoic landscapes of the plains to the east and north of the Arkaroola region record a complex evolution under changing climatic conditions (Twidale and Bourne, 1996; Twidale and Wopfner, 1996; Hill, 2008). Major alluvial fan systems occur along the range front and drain into the adjacent salt lakes (Frome, Grace, Blanche and Callabonna) or into the dune fields on the margin of the Strzelecki Desert. The fans are variably duricrusted and dissected and record a Quaternary history of fan deposition and incision related to both climate change and tectonics. Examples of these Cenozoic deposits flank the Flinders Ranges at the Paralana Hot Springs (see below). The various surfaces, duricrusts and sediments provide an analogue for the complex landforms to be expected on Mars. Some of these deposits, such as the mobile barchanoid sand dunes at Gurra Gurra Waterhole, have previously been studied as Mars analogues (Bishop, 1999).

## Astrobiology in the Flinders Ranges

*Astrobiology is the study of the origin, evolution, distribution, and future of life in the universe. This multidisciplinary field encompasses the search for habitable environments... the search for evidence of prebiotic chemistry and life on Mars and other bodies in our Solar System, laboratory and field research into the origins and early evolution of life on Earth, and studies of the potential for life to adapt to challenges on Earth and in space. From:* "About Astrobiology". (*NASA Astrobiology Institute*, 2008).

### Neoproterozoic stromatolites at Arkaroola

There are numerous Neoproterozoic stromatolitic carbonates and associated chert formations within the Adelaide Geosyncline, which may contain microfossils. These include the oldest carbonates of the Callanna Group and the Skillogalee Dolomite of the Burra Group, (both pre-Cryogenian), the Cryogenian interglacial carbonates of the Balcanoona, Etina and Trenzona formations, the Ediacaran Wonoka Formation, and numerous horizons in the Cambrian (Coats, 1972; Preiss, 1987, 1993, and references therein). They attest to a flourishing photosynthetic bacterial microbiota that was widespread in marine (and probably also lacustrine) environments ranging from peritidal to the base of the photic zone (c.100–150 m depth) during favourable conditions throughout the region. In the Arkaroola area, the best preserved stromatolites are found in the Balcanoona Formation as large reefs up to 1.1 km thick, with associated enigmatic structures that may be sponges (Giddings et al., 2009). Fore-reef, reef-margin and back-reef facies are all well represented.

The astrobiological significance of stromatolites lies in their ubiquity, as they are the oldest known macroscopic evidence of life on Earth and are found in diverse environments including saline and freshwater lakes, intertidal flats, springs, hydrothermal vents, and both shallow and deep marine. While not all stromatolite-like structures are biogenic, not all biogenic stromatolites preserve microfossils or biomarkers (Brasier et al., 2002; Schopf, 2006), and both microfossils and biomarkers can be found in non-stromatolitic lithologies (Marshall, 2007). Recognition of stromatolite-like features on Mars would be a major discovery and would provide a focus for subsequent investigations. Therefore various researchers have suggested that recognition for stromatolitic morphologies be included in the search for life on Mars (e.g., Walter and Des Marais, 1993; McKay and Stoker, 1989), as well as drawing parallels between terrestrial stromatolites and what might be found on Mars (e.g., Allwood et al., 2007; Van Kranendonk, 2006).

## A possible Neoproterozoic deep hot biosphere

The Arkaroola region is also notable for the serendipitous discovery of possible evidence for a deep hot biosphere that inhabited Neoproterozoic sedimentary successions during peak burial and metamorphism. While examples of such organisms are known from deep wells today (Fyfe, 1996) and have been postulated to have lived as long as 3.8 billion years ago (Pinti et al., 2001), fossil examples are poorly known. Structural investigations by Bons and Montenari (2005) examined fibrous anataxial calcite veins in the Tindelpina Shale Member of the Tapley Hill Formation. These formed at c. 585 Ma at an estimated 3–6 km depth. Scanning electron microscope observations revealed about 1-micron-sized structures within the veins. Further studies showed that these structures were composed of calcite but contained higher sulfur than the surrounding material. Fluid inclusions in the calcite indicate a temperature of formation of c. 60–80°C, and not exceeding 100°C. While a nonbiogenic origin of the objects is possible, it was considered unlikely (Bons et al., 2007). The weight of evidence from morphology, chemistry and size distribution indicates that the objects are fossilised microbes that lived in the veins at the time and depth of vein formation. Further work is needed on the structures to test their biogenicity. If these features are indeed biogenic they are a potential analogue for a possible martian habitat (Hoffmann and Farmer, 2000).

## Mount Painter/Mount Gee

The Mount Painter area contains a complex collection of breccias, which are characterised by large dyke-like bodies of siliceous hematitic breccia and are interpreted to represent hydrothermal systems (Hore and Hill, 2009). This hydrothermal activity is most evident in the area centred on Mount Painter and Mount Gee (Sprigg, 1945; Drexel and Major, 1987) with lesser expressions to the east at Livelys Find Au prospect (Collier, 2000) and to the northeast at the Hodgkinson U prospect (Smith, 1992). Many of these breccias contain U and minor sulfide mineralisation. The breccias occur as irregular bodies within the basement complex, adjacent to a zone of extensive faulting, which also contains Ordovician granites and pegmatites (Lambert et al., 1982). The Mount Gee system includes a range of features including crustiform and colloform textures, bladed and replacement carbonate, fluorite, zeolites and siliceous fluids resulting in cavity-fill crystallisation and jasperoidal formations as seen in Figure 2 (Hore, 2008). The jasperoidal unit at Mount Gee could have been deposited as a gel (e.g., Eugster and Jones, 1967), and many textures in the Mount Painter and Mount Gee silica deposits are similar to the botryoidal silica surfaces observed at the Sleeper Au deposit in Nevada (Saunders, 1994) and the recrystallised silica gels from Yellowstone National Park (Fournier et al., 1991). While the preservation of microfossils would appear likely in the Mount Painter hydrothermal deposits, preliminary attempts to find them have been unsuccessful (Carlton, 2002).

***Figure 2 Mount Gee epithermal quartz.***

Ancient hot-spring deposits and associated hydrothermal alteration exemplify key environments in the exploration and study of earliest life on Earth (e.g., Bock and Goode, 1996). These systems could have similar significance in the search for extraterrestrial life, and particularly on our nearest neighbour, Mars, where hydrothermal activity is likely to have occurred in the past and may even continue somewhere underground today (e.g., Walter and Des Marais, 1993; Catling and Moore, 2003). Martian hematite-precipitating spring deposits are high-value targets for martian astrobiology missions, in particular sample return missions (Catling and Moore, 2003; Allen et al., 2001, 2004). The hematite-rich shallow hydrothermal systems of the Mount Painter Inlier are therefore of considerable astrobiological interest as Mars analogues (Thomas and Walter, 2002, 2004; Brugger et al., 2011) because of the high preservation potential for both microfossils and organic matter in the silica-hematite precipitates. The high radiation environment of these systems, of which Paralana Hot Springs (see below) is the current example, provides another similarity to the surface of Mars, while the association of hydrothermal activity with an icy surface environment in the Mount Painter hydrothermal system appears unique for a fossil terrestrial hydrothermal system (Brugger et al., 2011) and is a further parallel for what might be found on the surface of Mars.

Alteration (including hydrothermal) can be detected in hyperspectral data in regions like the Mount Painter Inlier (Thomas and Walter, 2002; Brown and Thomas, 2004; Brown et al., 2004). Studies indicate that careful band selection and testing in terrestrial field analogues are required if spacecraft systems are to detect hydrothermal alteration on Mars and differentiate it from more regional alteration signatures.

## Radioactive hot springs of the Paralana Fault Zone

The Paralana Fault Zone, consisting of northeast-trending faults, runs along the eastern margin of the Mount Painter Inlier, northeast of Arkaroola (Figure 1). The fault zone is associated with an area of anomalously high heat flow attributed to high concentrations of radioactive elements in Mesoproterozoic granites of the adjacent Mount Painter Inlier and is generally believed to be the main conduit for hydrothermal fluid dispersal in the area (Coats and Blissett, 1971; Thomas and Walter, 2002; Brugger et al., 2005). Hydrothermal activity during the Paleozoic has produced large volumes of uraniferous breccias, siliceous deposits (possibly sinters), and Cu-Fe and Fe-U deposits. Leaching of these deposits may have contributed to the secondary U deposits in Cenozoic sandstones east of the Inlier (Brugger et al., 2005). The Paralana Hot Springs occur to the north east of Mount Painter on the Paralana Fault Zone (Figures 1 and 3). The springs are evidence of continuing hydrothermal activity along the fault zone. The source of this fluid and of past hydrothermal fluids is difficult to ascertain, although it is most likely to be predominantly meteoric in origin (Foster et al., 1994) representing the surface expression of a cyclic, low-temperature, non-volcanic hydrothermal system (Brugger et al., 2005).

Water from the Paralana Hot Springs is neutral (pH 7–8) and the springs discharge 16 L/s at a temperature of 57°C. The 5 ppm F, 33 ppb Mo, 11 ppb W, 16 ppb Cs and 200 ppb Rb concentrations in the spring water are comparatively high. $\delta^{13}C$ values of $CO_2$ (g) emanating from the springs and dissolved $HCO_3^-$ suggest that the carbon source is organic matter, e.g., soil or plants. Radon concentrations at the springs are very high (radiation of 10,952 Bq/$m^3$), suggesting a localised radiogenic source at shallow depth (Brugger et al., 2005). These radiogenic springs are not associated with any active volcanism, and therefore provide an opportunity to study the effects of amagmatic heat sources, which could drive a hydrothermal system for an extended timeframe, perhaps a billion years or more (Brugger et al., 2011).

***Figure 3 Paralana Hot Springs source pool and microbial communities.***

Studies of hot springs have shown they support a high diversity of bacteria and Archaea. The hot spring environs at Paralana have been studied as an analogue for early life, and microbial adaptation to extreme temperatures and radioactivity levels (e.g., Anitori et al., 2002). The composition of microbial ecosystems around subaerial hot springs and submarine vents is significant to the understanding of early life on Earth and tracing evolutionary phylogenic trees, and interpreting presumed biomarkers (microfossils, lipids, isotopes etc.) found in these ancient systems (Walter and Des Marais, 1993). Analysis of bacterial communities by Anitori et al. (2002) indicated that a diverse community of bacteria is supported by the waters at the Paralana Hot Springs. The presence of Ra gas, which bubbles up intermittently in the thermal source pool, makes the Paralana Hot Springs highly suited as an analogue for ionising radiation environments, which may have been common on the early Earth and Mars.

# Cryogenian glacials

Sedimentological and paleomagnetic studies of Cryogenian glacial deposits in South Australia have been a stimulus for worldwide multidisciplinary research on the nature and extent of Cryogenian glaciations. Two major glaciations are recorded in the succession, the earlier Sturt glaciation (Preiss et al., 2011) and the later Elatina glaciation (Lemon and Gostin, 1990; Williams et al., 2008, 2011), often incorrectly referred to as the Marinoan glaciation.

The late Cryogenian (c. 630 Ma) Elatina glaciation in South Australia is represented by a spectrum of facies (Figure 4). The periglacial–aeolian Whyalla Sandstone on the Stuart Shelf passes eastwards into glaciofluvial, deltaic, littoral and inner marine-shelf deposits of the Elatina Formation and outer marine-shelf diamictite and mudstone-siltstone with dropstones in the Adelaide Geosyncline. Important features of the Elatina glaciation include the widespread and persistent rainout of fine-grained sediment and ice-rafted debris, grounded ice on diapiric islands within the basin and on cratonic regions in the east (but absent on the Gawler Craton in the west). Other key features include permafrost near sea level with a strongly seasonal periglacial climate, glaciofluvial deposition, several glacial advances and retreats, wave-generated ripple marks, and the annual (seasonal) oscillation of sea level (Preiss, 1987; Lemon and Gostin, 1990; Williams et al., 2008, 2011). The plaque for the Global Stratotype Section and Point (GSSP) for the recently established Ediacaran System and Period (Knoll et al., 2006) is placed between the Elatina Formation and the overlying Nuccaleena Formation in Enorama Creek in the central Flinders Ranges.

The late Cryogenian glacials in South Australia were deposited at low paleolatitudes (Figure 4, inset), as shown by high-quality paleomagnetic data for red beds from the Elatina Formation that demonstrate the early acquisition of magnetic remanence and deposition within 10° of the paleoequator (Schmidt et al., 1991, 2009;

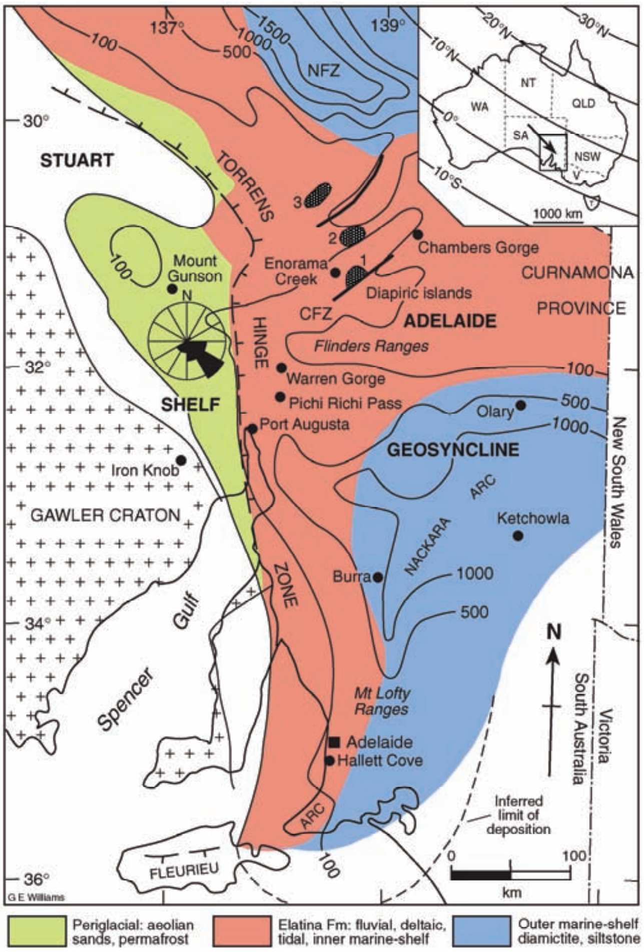


***Figure 4** Facies of the late Cryogenian Elatina glaciation. Isopachs in metres. CFZ = Central Flinders Zone, NRZ =North Flinders Zone. 1, 2, 3 = diapiric islands. Paleowind rose diagram for the Whyalla Sandstone. Inset shows paleolatitudes and paleowind direction (arrow). After Williams et al. (2008).*

Schmidt and Williams, 1995; Sohl et al., 1999). However, the indicated paleogeography, with extensive and long-lived open seas, unglaciated continental regions and an active hydrological cycle, conflicts with the "snowball Earth" hypothesis of Hoffman and Schrag (2002).

The older, Sturt glaciation is equally well represented by diverse facies throughout the Flinders Ranges (Preiss, 1987, 1993; Young and Gostin, 1989). At Stubbs Waterhole, 5 km from Arkaroola, spectacular thick-bedded sandy boulder and pebble conglomerates with diverse lithologies were deposited from vigorous glacial meltwater streams. It is believed that during the middle Cryogenian an elevated large ice cap covered the whole area between the Mount Painter Inlier and Broken Hill (250 km to the southeast). Major thickness variations of the c. 500 m thick Bolla Bollana Formation reflect rapid subsidence in rifted basins.

## Tidal rhythmites and Earth's paleorotation, Pichi Richi Pass

The Elatina Formation near the margins of the Adelaide Geosyncline includes tidal rhythmites of siltstone and fine-grained sandstone deposited on a series of ebb-tidal deltas and estuarine tidal flats that formed during a high stand of sea level during temporary glacial retreat (Williams, 1991, 2000; Williams et al., 2008). The fine-grained sediment load of ebb tidal currents is related directly to tidal range (or maximum tidal height), and deposition offshore from ebb-tidal jets and plumes forms neap–spring cycles comprising semidiurnal and diurnal (lunar day) graded laminae mostly of fine sand and silt, with mud bands deposited during slack water at neaps (Figure 5). The rhythmite unit is 18 m thick at Warren Gorge (Williams, 1996) and somewhat thinner at a more distal setting in Pichi Richi Pass (Figure 4) where exposure is limited.

Detailed study of cores from three vertical holes drilled through the rhythmite unit in Pichi Richi Pass have provided an internally-consistent paleotidal data-set comprising numerous tidal cycles ranging from semidiurnal to the lunar nodal cycle (9.4 m log of 1580 successive fortnightly neap–spring cycles recording 60 years of continuous deposition; Williams, 1991). The neap–spring cycles contain 8–16 diurnal laminae, with many cycles abbreviated at neaps (Figure 5b); semidiurnal increments occur locally, and are conspicuous in thicker neap–spring cycles from tidal rhythmites in the correlative Reynella Siltstone Member near Hallett Cove, 300 km to the south (Figures 5a and 6). Paleotidal cycles resulting from variation in the thickness of successive neap–spring cycles compare closely with modern tidal patterns for Townsville, Queensland (Figure 7): features common to both data sets include first-order peaks marking the solar year and the annual oscillation of sea level, second-order peaks marking the semiannual tidal cycle, and a sawtooth pattern reflecting alternate high and low spring tides due to the eccentric lunar orbit. The duration of Elatina rhythmite deposition matches the c.70 year

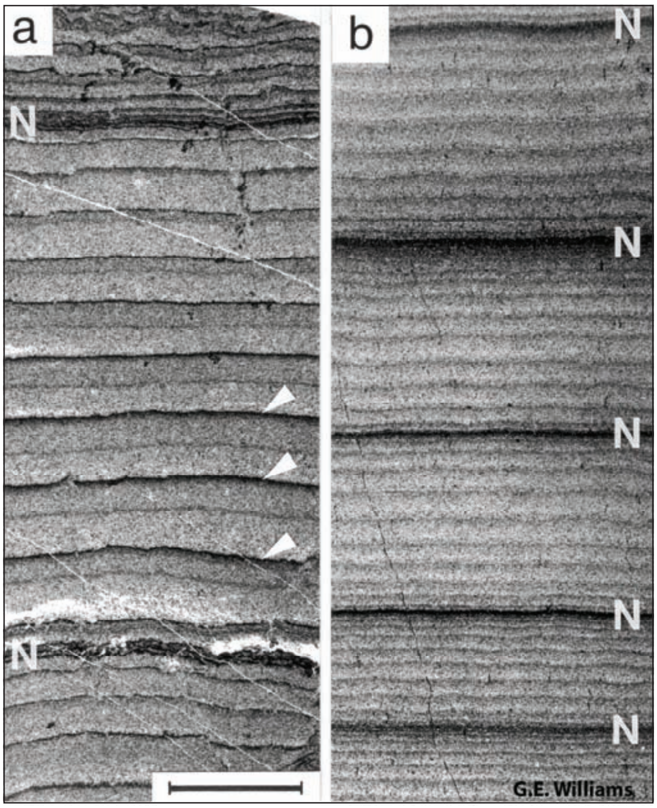


***Figure 5** Neap–spring cycles of late Cryogenian tidal rhythmites. Scale bar 1 cm, N = neap-tidal muddy bands. (a) Reynella Siltstone Member; diurnal laminae have mudstone tops (arrows) and comprise semidiurnal increments. (b) Elatina Formation, Pichi Richi Pass. From Williams (2004).*

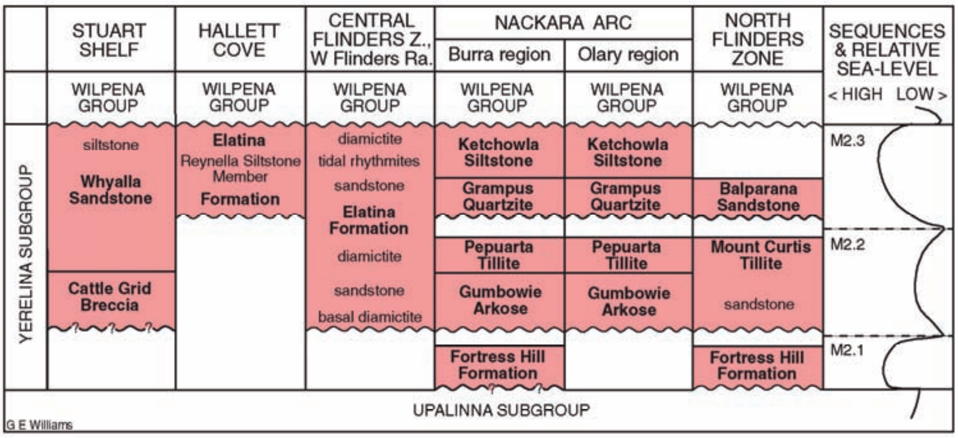


***Figure 6 Stratigraphy of the late Cryogenian Yerelina Subgroup, which encompasses all the units of the Elatina glaciation in South Australia. After Preiss et al. (1998) and Williams et al. (2008), with Marinoan sequence sets M2.1–M2.3 and relative sea-level curves from Preiss (2000).***

morphologic cycle for a modern ebb-tidal delta (Ping, 1989). Comparable tidal rhythmites occur in modern glaciomarine settings (Smith et al., 1990; Cowan et al., 1999).

Information on the Earth's paleorotation and the Moon's orbit in late Cryogenian times revealed by time-series analysis of the Elatina core log, supplemented by data for the Reynella Siltstone Member, includes 13.1 ± 0.1 lunar months/year, 400 ± 7 solar days/year, 21.9 ± 0.4 hours/solar day, and a mean Earth–Moon distance of 96.5 ± 0.5% of the present distance (Williams, 1991, 2000). The rhythmites also record the non-tidal, annual oscillation of sea level (Figure 7), which is a response mostly to seasonal changes in water temperature as well as variation in winds and atmospheric pressure, indicating extensive open seas during late Cryogenian glaciation (Williams et al., 2008, 2011).

## Acraman impact ejecta horizon in Bunyeroo Gorge

The Acraman impact ejecta horizon (AIEH) in the mid-Ediacaran (c.580 Ma) Bunyeroo Formation (Gostin et al., 1986, 1989) was the first Precambrian impact ejecta deposit to be linked to a specific astrobleme, the Acraman impact structure on the Gawler Craton, several hundred kilometres to the west (Williams, 1986). The AIEH consists of a solitary thin layer of red dacitic fragments and sand, enveloped by a reduction halo of grey-green shale, and is well exposed within marine-shelf maroon shale 80 m above the base of the 400 m thick formation (Figure 8). The volcanic fragments show features of shock

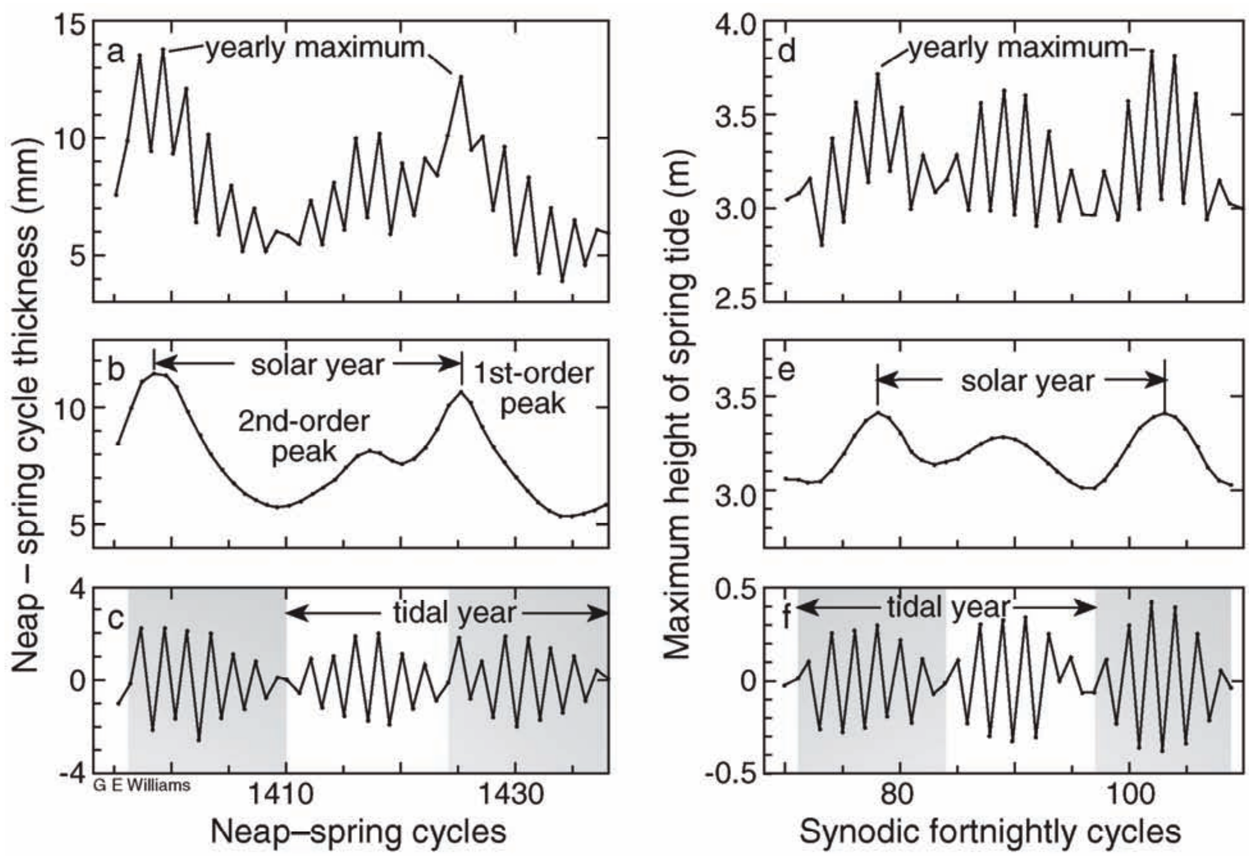


***Figure 7 Tidal patterns for (a–c) the late Cryogenian Elatina rhythmites and (d–f) Townsville, Queensland, for 19 October 1968 to 3 June 1970. (a, d) Unsmoothed curves. (b, e) Smoothed curves. (c, f) Residuals (a minus b; d minus e), showing 180° phase changes in the sawtooth patterns. From Williams (1991, 2000).***

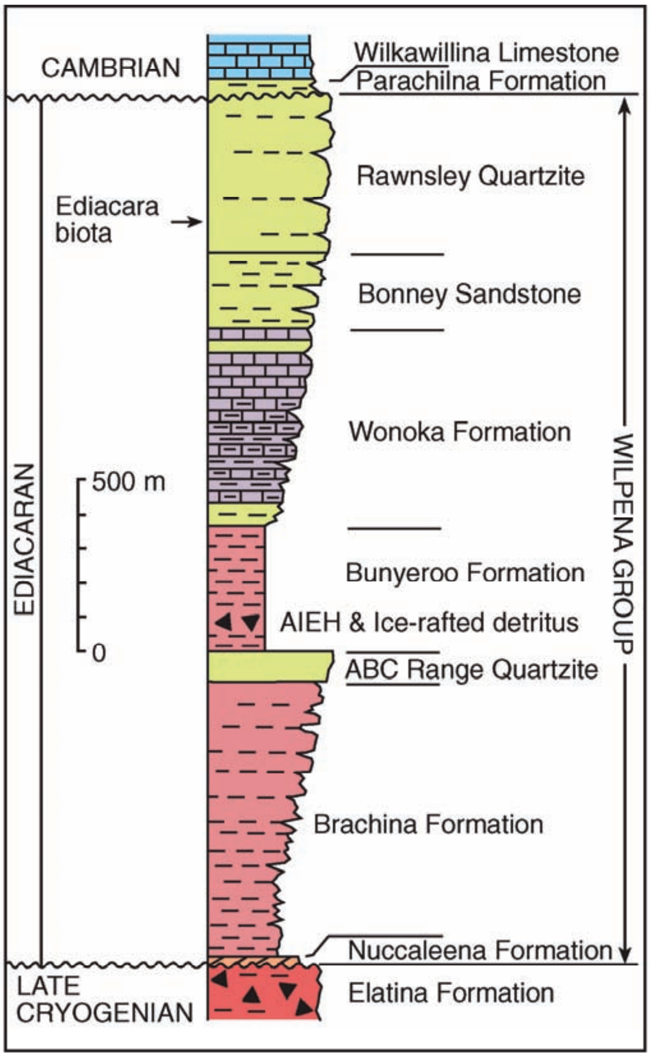


*Figure 8 Stratigraphic column for Ediacaran strata in the central Adelaide Geosyncline. AIEH = Acraman impact ejecta horizon. The famous metazoan fossil assemblage (Ediacara biota) occurs near the base of the Rawnsley Quartzite (Preiss, 1987, 2000). After Williams and Gostin (2005).*

metamorphism and the horizon is anomalous in cosmogenic siderophile elements, including Ir (Gostin et al., 1989). U-Pb zircon dating, paleomagnetic data and the regional distribution of the ejecta confirm derivation of the volcanic fragments from the Acraman impact structure in the 1592 Ma Yardea Dacite on the Gawler Craton (Compston et al., 1987; Schmidt and Williams, 1996; Williams and Wallace, 2003; Williams and Gostin 2005). Typical sections of the ejecta are shown in Figure 9 and ejecta regional distribution in Figure 10.

The magnitude, age and potential environmental effects of the Acraman impact and the nature of the AIEH were reviewed by Williams and Wallace (2003) and Williams and Gostin (2005). The Acraman impact structure is complex, with a transient cavity 40 km in diameter and a final structural rim (90 km diameter) that is eroded ≥2.5 km below the original crater floor. The AIEH in the Adelaide Geosyncline occurs at radial unfolded distances of up to 400 km from the centre of the Acraman structure. The impact occurred at a low (c. 12.5°) paleolatitude and probably perturbed the atmosphere in both the northern and southern hemispheres. The dimensions of Acraman

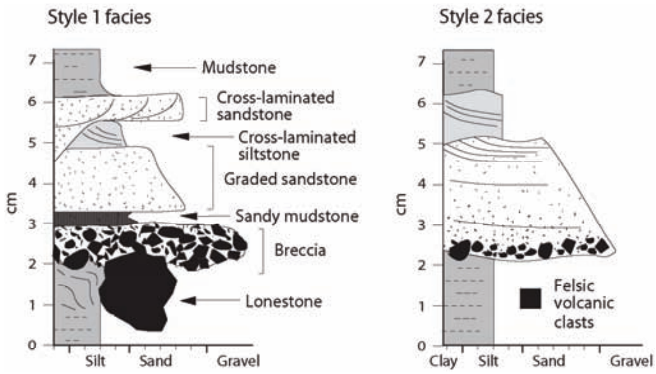


*Figure 9 Principal styles for the ejecta horizon in the Bunyeroo Formation. Style 1 (Bunyeroo Gorge–Brachina Gorge area) represents primary fallout deposited from suspension. Style 2 (western Adelaide Geosyncline) represents ejecta reworked by impact-induced tsunamis. After Wallace et al. (1996).*

structure indicate that the impact would have caused earthquakes, tsunamis, atmospheric ozone loss, and the global dispersal of a dust cloud that lowered light levels below those required for photosynthesis.

The presence of ice-rafted dropstones, granule clusters, frozen aggregates and till pellets in mudstones of the lower Bunyeroo Formation in the central Adelaide Geosyncline (Figures 8 and 10) and in the correlative Dey Dey Mudstone (Officer Basin) implies a cold climate and nearby areas of freezing and glaciation both before and after the Acraman impact (Gostin et al., 2010). Following

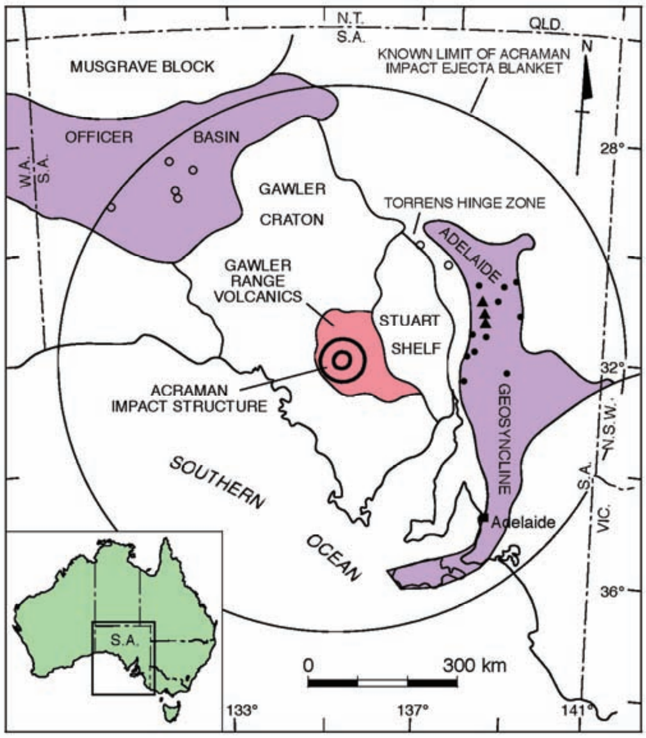


*Figure 10 Location of the Acraman structure and the Acraman impact ejecta horizon in mid-Ediacaran strata. Solid circles indicate ejecta seen in outcrop, and open circles ejecta recorded in drill core. Triangles indicate ice-rafted detritus and ejecta seen in close stratigraphic proximity. From Gostin et al. (2010).*

glaciation and impact, the overlying Ediacaran succession shows a negative carbon isotopic excursion, rapid diversification of acanthomorph acritarchs, and biomarker hydrocarbon anomalies, climaxing with the advent of metazoans. Release from the combined environmental stresses of glaciation and impact may have expedited biotic evolution in the later Ediacaran (Grey et al., 2003; McKirdy et al., 2006; Gostin et al., 2010).

## Conclusions

This paper highlights a selection of features in the central and northern Flinders Ranges that illustrate important events and processes in the physical evolution of the Earth as a planet, including the widespread Neoproterozoic glaciations, the rotational and orbital evolution of the Earth-Moon system, and asteroid impact. Stromatolitic successions provide examples of their sedimentary role and morphological evolution during the Cryogenian through to the Cambrian. Also highlighted are some of the features that provide analogues for possible sites on Mars that would be prime targets in the search for extraterrestrial life. These features exemplify the considerable scope for study and understanding of astrobiology and planetary science that exists in the Flinders Ranges and surrounds.

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*by James G. Gehling[1] and Mary L. Droser[2]*

# Ediacaran stratigraphy and the biota of the Adelaide Geosyncline, South Australia

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*The Neoproterozoic and Cambrian sequences of the Mount Lofty and Flinders ranges of South Australia record a crucial interval in the evolution of life and environments on Earth. This paper summarises the key characteristics of the Ediacaran succession of the Adelaide Geosyncline. Recent interest in the Neoproterozoic and Cambrian successions on all continents has been driven by a quest to better understand the causes and effects of the size revolution that saw the advent of multicellular organisms in the ocean. The principal role of the Neoproterozoic Subcommission has been to determine unique time boundaries that reflect evolutionary changes in the Earth system. Utilisation of a combination of geochemical, paleobiological and stratigraphic events, to offset the limitations of the Neoproterozoic paleontological record, is required to bracket this interval of Earth history. Fossils of the Cryogenian, Ediacaran and Cambrian document a series of unexpected quantum leaps in the size and diversity of life.*

## Introduction

The folded Neoproterozoic and early Cambrian successions of the Mt Lofty and Flinders Ranges in South Australia form a continuous N-S-trending belt from Kangaroo Island in the S to more than 600 km N and NE into the interior of the continent (Figure 1). Much of this succession is correlative with intracratonic Amadeus and Officer basins of central Australia. Plate reconstructions suggest continuations into the Ross Orogen of Antarctica. To the W the succession thins rapidly across the meridional Torrens Hinge Zone. Here some Adelaidean formations lap onto the Stuart Shelf, a platform with an undeformed, flat-lying succession of Neoproterozoic and Mesoproterozoic sediments and volcanics, underlain by Paleoproterozoic and Archaean metasediments and intrusives of the Gawler Craton.

Evidence of fault-controlled sedimentation on the margins and within the basin led von der Borch (1980) to propose the phrase “Adelaide Rift Belt”, or “aulacogen” (Rutland, 1973), while others have used “Adelaide Fold Belt’ (Scheibner, 1973; Plummer, 1990; Mancktelow, 1981). The historical usage of “Adelaide Geosyncline” (Sprigg, 1952; Thomson, 1969, 1970) does not adequately describe the combination of passive margin and rift-based basin, but it does acknowledge the tectonics involved in accumulation of >20 km of strata, in 300 Myr (Preiss, 1987; Williams et al., 2008). Preiss (2000) subdivided the successions into the Warrina, Heysen and Moralana supergroups. These record, in turn, at least three phases of rifting during the break up of Rodinia and the Sturt glaciation, followed by evolution of the Centralian Superbasin, and the Delamerian Orogeny with rifting E and SE of the Adelaide Geosyncline in the Cambrian.

Diapiric intrusions of massive and flow-banded carbonate cemented breccias with disorganised megaclasts of sedimentary and igneous rocks occur in the cores of anticlines and as apophyses along fault systems throughout the Flinders Ranges (Preiss, 1985). Diapiric breccias appear to be sourced from the Callanna Group, the basal succession of sedimentary strata in the Adelaide Geosyncline (Figure 2). Where outcropping, these carbonates, siliciclastics and volcanics show evidence of intercalated shallow marine, evaporitic and terrestrial sedimentation that accompanied the early rifting stage in the evolution of the Adelaide Geosyncline.

## Neoproterozoic Stratigraphy and Paleontology

The Heysen Supergroup (Figure 2), consisting of the Umberatana and Wilpena groups, records the mid to late Cryogenian Sturt and Elatina glaciations (Preiss et al., 2011; Williams et al., 2008, 2011) succeeded by the Ediacaran intermittent return to warmer climates and the rise of multicellular organisms in the Ediacara biota. The base of the Cryogenian has yet to be defined.

### *Heysen Supergroup*

This supergroup encompasses the entire succession in the central Flinders Ranges other than the diapiric rocks sourced from the base of the underlying Warrina Supergroup (Preiss, 1985, 1993). In the central Flinders Ranges the 4.5 km thick Umberatana Group encompasses the two main phases of glacial deposition (see Thomas et al., 2012). The carbonaceous, calcareous and pyritic Tindelpina Shale Member, of the interglacial Tapley Hill Formation, caps the Fe-rich diamictite and tillite formations of the Sturt glaciation. The upper Cryogenian glacials of the Elatina Formation are truncated by the Nuccaleena Formation at the base of the Wilpena Group and the Ediacaran System.

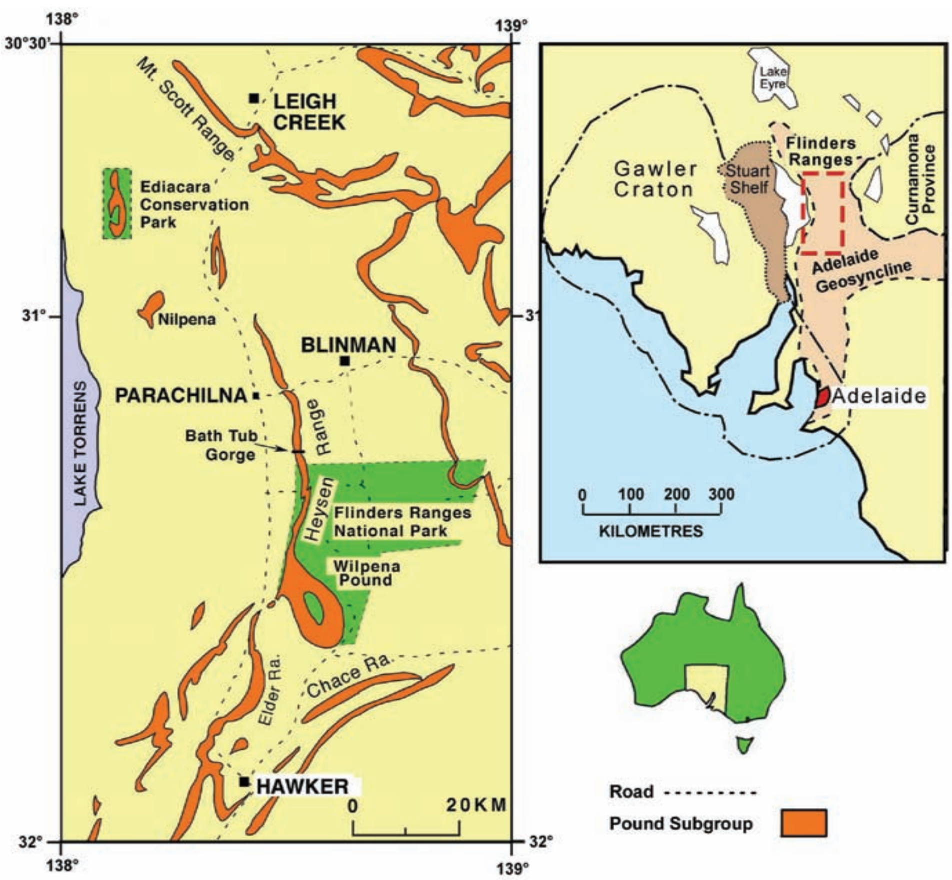


***Figure 1 Location map of the Adelaide Geosyncline and geological provinces in South Australia, with distribution of the late Ediacaran Pound Subgroup in the central Flinders Ranges.***

## *Global Stratotype Section and Point: Base of the Ediacaran System*

In 2004, the Global Stratotype Section and Point (GSSP) for the terminal Proterozoic was placed near the base of the Nuccaleena Formation in Enorama Creek in the central Flinders Ranges (Figure 1), thus establishing the Ediacaran System and Period (Knoll et al., 2006). As the Nuccaleena Formation has not been accurately dated, a date of c. 635 Ma from near-correlative levels in Namibia and China is presumed for the base of the Ediacaran (Hoffmann et al., 2004; Condon et al., 2005; Zhang et al., 2005).

## *Wilpena Group*

The Wilpena Group represents siliciclastic sedimentation that continued into the earliest Cambrian. Repeated erosional incisions followed by a rapid local rise in sea level characterise both the Ediacaran and Early Cambrian successions. The Wilpena Group (Figure 3) began with the immediate post-glacial Nuccaleena Formation 'cap carbonate' which as been correlated with post glacial 'cap carbonates' on other continents (Kennedy et al., 1998).

Regionally, the Nuccaleena Formation, which rests conformably on the Enorama Formation but with local disconformity, is an important lithological marker. It consists of pink to yellowish dolostone, with purple shale interbeds and lateral sandy facies. In many places, beds are truncated by 50–150 cm high "tepee" structures (see Schmidt et al., 2009, Figure 2), ramping with crests continuous for up to 50 m and NNW-SSE directed. Plummer (1979) regarded the Nuccaleena Formation as intertidal to supratidal. Allen and Hoffman (2005) interpreted "tepee" structures from correlative "cap-carbonates" as giant standing wave ripples generated by sustained wind fields in the wake of late Cryogenian glaciation. Schmidt et al. (2009) provided paleomagnetic evidence that the tepee-like structures in the Nuccaleena Formation developed during early diagenesis. However, draping of these well-oriented structures, in alternate directions across crests, suggests formation as wave structures, but exaggerated by diagenetic expansion and disruption or gravity sliding. The lack of microbial textures suggests rapid chemical sedimentation under conditions unsuited to preserving microbial activity. Calver (2000) described an upward decrease in carbonate $\delta^{13}C$ values but a discordant increase in organic $\delta^{13}C$ values explained as a possible consequence of continued stratification of ocean waters following deglaciation. Retallack's (2011) re-interpretation of the Nuccaleena Formation as a product of post-glacial loess deposition and soil formation requires exclusive exposure of a carbonate source rock after glacial meltdown, which cannot be justified. Cap-carbonates, like the Nuccaleena Formation, and seep-like carbonate structures in the Seacliff Sandstone Member have been interpreted as products of destabilisation of gas-hydrates during near-equatorial deglaciation (Kennedy et al., 2001, 2008).

The Nuccaleena Formation grades into red siltstone and shale representing a maximum flooding surface at the base of the overlying Brachina Formation. In the S of the Adelaide Geosyncline, the Nuccaleena Formation is interbedded with the Seacliff Sandstone in a series of incised valley fill deposits (Dyson and von der Borch, 1994). Dyson (1985) reported parallel, ribbed structures resembling frond-like fossils from sandstone in the Elatina Formation in the southern Flinders Ranges. Soft sediment deformation in the host sediment creates some doubt remains about the biological origin of these structures.

The Brachina Formation is an upward shallowing and coarsening regressive sequence of siliciclastics, >1.3 km thick. Rhythmically bedded laminated siltstones grade up into silty, cross-laminated fine-grained sandstones. After a brief lowering of base level, the succession becomes sandier to the top where it grades into the cleaner and coarser ABC Range Quartzite (Plummer, 1990). To the N and E, the Brachina Formation represented by green-grey, finer grained Ulupa Siltstone. In coastal cliffs, S of Adelaide, turbiditic facies at the base grade up into a thick section with hummocky cross-stratified sandstone. In addition to facies with ball and pillow, as evidence of soft-sediment deformation in fore-delta settings, certain facies suggest microbial sealing. Common crescentic synaeresis cracks, "Kinneyia" ripples

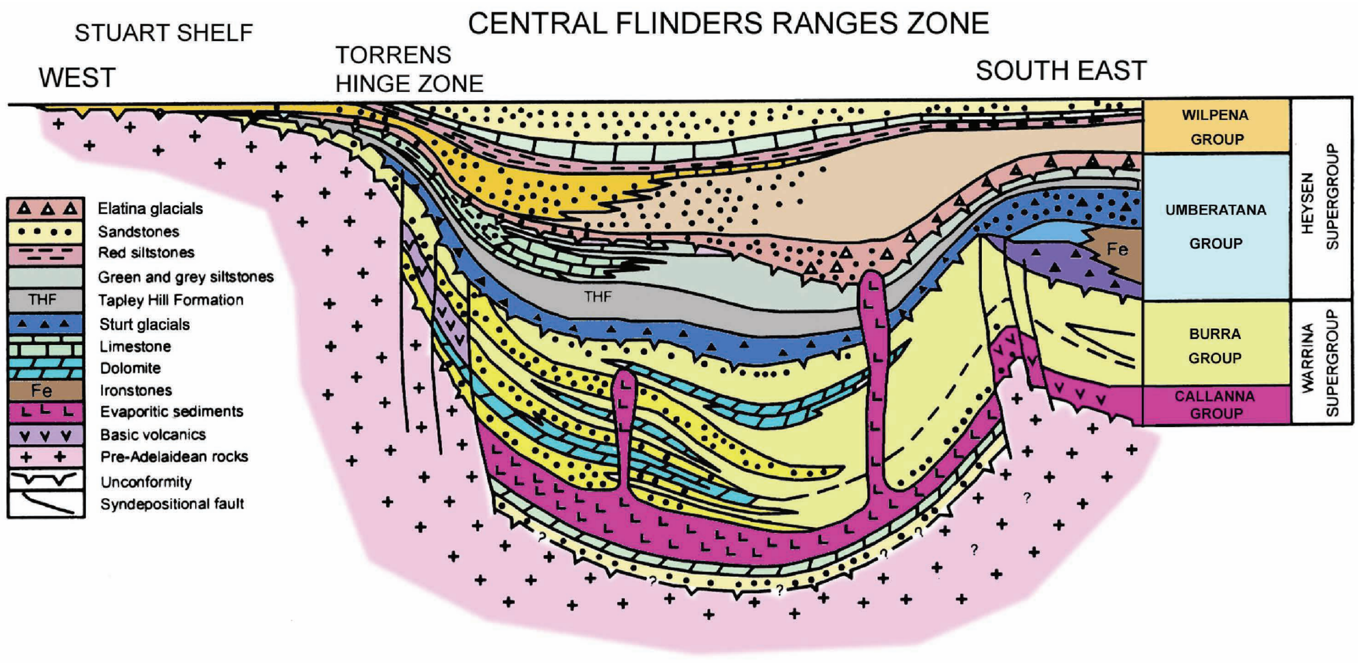


***Figure 2 Schematic cross-section of the Adelaide Geosyncline through the Flinders Ranges, showing facies relationships (adapted from Lemon, 1996).***

and syndepositional microfaulting record mechanical responses in microbial-bound sediment to varying degrees of wave-action, dewatering and gravity creep (see Pflüger, 1999). Pothole casts (Jenkins et al., 1981), confirm early stabilisation of soft sediment in reaction to storm events. A single specimen of an arcuate structure, *Bunyerichnus dalgarnoi*, originally interpreted by Glaessner (1969) as a molluscan trail, is probably a pseudofossil produced by a tethered biomat fragment. The Brachina Formation lacks body fossils.

Outcropping on the western side of the Adelaide Geosyncline, the ABC Range Formation is a medium to thickly bedded sandstone with herringbone cross bedding and desiccation cracks. From 2 km thick, S of Port Augusta, it thins N and S to <100 m (Plummer, 1990). The upward shoaling Brachina Formation and ABC Range Quartzite represents a highstand sequence tract where siliciclastics prograded from the Gawler Craton into the Adelaide Geosyncline with N-S tidal redistribution (Plummer, 1990).

The base of the 500 m thick Bunyeroo Formation, is a basin-wide sequence boundary, with the Wilcolo Sandstone Member, a low-stand veneer of conglomerate and pebbly sandstone, filling local incisions into the top of the ABC Range Quartzite (Figure 3). The Bunyeroo Formation is a maroon, laminated siltstone with apparent cyclic bedding. Except for its lower member and local onlapping breccias near diapirs, it was deposited in deep water during maximum sea level. About 80 m above the base of the Bunyeroo Formation, a single layer of angular fragments of red, porphyritic dacite is embedded in shale and overlain by a 1–10 cm of graded, sand and granules with swaley cross-lamination (Figure 4a). The shale 10–20 cm above and below has been altered to a green colour. Gostin et al. (1986) interpreted this persistent layer as debris from a bolide impact. The clastic debris is comparable in lithology and U–Pb zircon geochronology with the Mesoproterozoic Gawler Range Volcanics that lie over a large area of the Gawler Craton. Williams (1986) identified the circular Lake Acraman, 300 km W of the central Flinders Ranges, as the probable impact site, supported by accordant paleomagnetic data for Acraman melt rock and Bunyeroo red beds (Schmidt and Williams, 1996). The importance of this debris layer as a chronostratigraphic marker is emphasised by its tracing throughout the Flinders Ranges and in the Dey Dey Mudstone in Officer Basin drill core, 600 km NE of Lake Acraman (Wallace et al., 1989). The Acraman horizon in the Officer Basin marks both a negative C-isotope excursion and a transition from the older, simple leiosphere-dominated palynoflora (SLP) to the larger Ediacaran complex acanthomorph-dominated palynoflora (ECAP) (Arouri et al., 2000; Walter et al., 2000; Grey et al., 2003; Grey, 2005). Gostin et al. (2010) described ice rafting based on dropstones and pebble clusters both above and below the Acraman horizon, pointing to a cold mid-Ediacaran phase. Since Schmidt and Williams (1996) indicated a paleolatitude of c. 15°, the evidence for marine ice rafting offers a tentative link to the Gaskiers glaciation of 582–580 Ma (Bowring et al., 2003).

The Wonoka Formation begins with the Wearing Dolomite Member, a cupriferous, concretionary dolostone, with intraformational and edgewise conglomerates, indicates a hiatus followed by transgressive carbonates. In the central Flinders Ranges, the Wonoka Formation is an upward shallowing and coarsening sequence of 500 m of calcareous shale, siltstone and fine sandstone (Haines, 1990). In the N and S Flinders Ranges, canyons incised 1–1.2 km through the basal Wonoka Formation and underlying formations suggesting submarine cut and fill (von der Borch et al., 1982), a fluvial cut followed by marine fill (von der Borch et al., 1985, 1989), a Messinian-style evaporitic drawdown of local sea level (Christie-Blick et al., 1990), and a mantle-plume uplift (Williams and Gostin, 2000). Giddings et al. (2010) suggest submarine cut and fill due to salt withdrawal tectonics (Dyson 1998, 1999, 2003; Dyson and Rowan, 2004). The stable isotope geochemistry of the Wonoka Formation features the "Wonoka-Shuram" anomaly with values of $\delta^{13}C$ as low as -12 (Pell et al., 1993). These have been used to suggest a reservoir of dissolved organic carbon depleted in $^{13}C$ for Ediacaran oceans

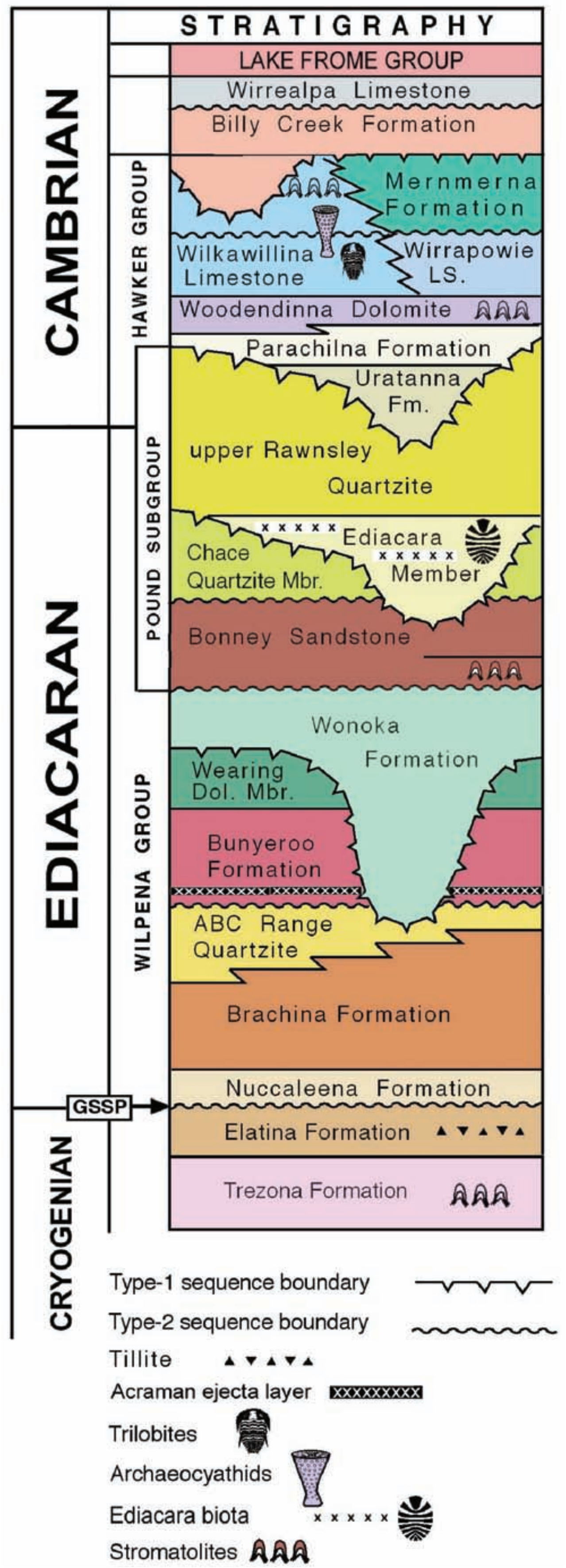


***Figure 3 Depositional Sequences of the Ediacaran and Cambrian, Flinders Ranges, South Australia.***

(Swanson-Hysell et al., 2010) after 580 Ma. Away from the canyons, the Wonoka Formation consists of turbidites grading up into hummocky cross stratified, silty limestone, followed by green, stylolitic, cryptomicrobial limestone alternating with sandy limestone, featuring soft-sediment deformation and intraformational breccias (Haines, 1990). The formation is capped by two parasequences of red, clayey sandstones overlain by a carbonate unit. Reid and Preiss (1999) recognised an erosive base and redefined these last cycles as part of the overlying Bonney Sandstone. The upper most carbonate is composed of black cryptalgal laminae, peloidal and oncolitic limestone, with stromatolites resembling *Tungussia julia* that support correlation with the Julie Formation in the Amadeus basin. From the upper Wonoka Formation, Haines (1990, 2000) described a serial, arcuate tubular fossil, *Palaeopascichnus* (Figure 4b) that occurs in most other Ediacara assemblages. It is the oldest identifiable member of the Ediacara biota in South Australia. Claims of other Ediacara fossils from the Wonoka Formation (Jenkins, 1983, 1995) are limited to a frondose structure, faint discs and radial arrays of spicular structures.

The Pound Subgroup, consisting of the red siliciclastics of the Bonney Sandstone disconformably overlain by the cleaner sandstones of the Rawnsley Quartzite (Figure 3), varies from 1 km in Wilpena Pound (Figure 1) to >3 km in the NE Flinders Ranges (Gehling, 1982) and dominates the topography (Figure 5a). The red Bonney Sandstone, which is 300–400 m thick in the central Flinders Ranges, has transgressive shallow marine sandstone and carbonates at the base, overlain by deeper water siltstone and fine sandstone shallowing and coarsen upward as a highstand systems tract. These flat bedded and low-angle crossbedded sandstones, feature ebb and flood cycles in tidally reworked estuaries. The upper third of the formation consists of alluvial, red, poorly sorted, sandy mudstones (Gehling, 1982). Claimed occurrences of trace fossils and small discoidal body fossils in the Bonney Sandstone (Jenkins, 1995) are more likely products of wrinkled and desiccated microbial mats.

The Rawnsley Quartzite is composed of pale, medium to coarse grained, felspathic sandstone. In deeply incised sections, the Rawnsley Quartzite is reddish with Fe-stained sand grains. The light colour and indurated ridges and outcrops are likely remnants of early Cenozoic leaching and surface silicification. The disconformable base of the Rawnsley Quartzite is a low relief erosional surface marking a change, without interfingering, from the Bonney Sandstone to better sorted and lighter coloured sandstones of the Chace Quartzite Member. The Rawnsley Quartzite is a shallow marine, wave and tide reworked deltaic succession that marked a phase of uplift and erosion on the Gawler Craton (Gehling, 2000) and the onset of the Petermann Orogeny 600 km to the NW between the Officer and Amadeus basins (Flöttmann et al., 2004). Comprising as much as 250 m of the Rawnsley Quartzite, the unfossiliferous Chace Quartzite Member is a yellow weathering, pink coloured, felspathic, medium- to coarse-grained sandstone with granule trains. Sets of low angle, trough cross-bedding grade up into wavy and disrupted, thinly bedded sand. This distinctive 'petee' lamination, featuring domed and disrupted, polygonal and sinuous ridges, overturning and roll-ups of sand lamination (Figure 6b), is interpreted as the product of gas-doming and desiccation of microbial bound sand laminae on tidal flats (Gehling, 1999, 2000). The near absence of silt or clay suggests that fine material bypassed these tidal flats due to aeolian sorting of alluvial sediment before it reached the ocean. Higher in the section, sand-

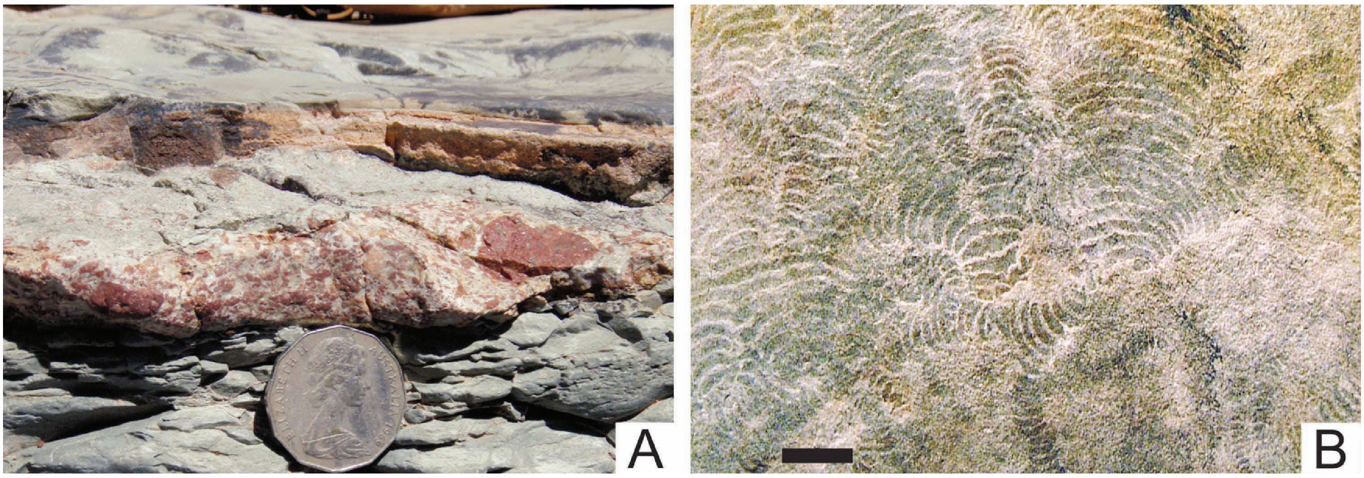


***Figure 4** (A) Acraman ejecta layer, 80 m above base of Bunyeroo Formation, Flinders Ranges National Park; coin 3.5 cm. (B)* **Palaeopascichnus** *in the upper Wonoka Formation; scale 1 cm.*

pseudomorphs after sulfate crystals, associated with petee lamination (Figure 6a), were likely formed on supratidal sand flats (Gehling, 1999, Figure 4; Gehling, 2000).

Ediacara fossils are mainly confined to the deeper water 5–350 m thick Ediacara Member (Jenkins et al., 1983; Gehling, 1982, 2000). The base of the Ediacara Member is a sequence boundary with 5–10 m deep channels and regional, 1–5 km wide, 50–250 m deep canyon incisions into the Chace Quartzite Member and, in some cases, well into the Bonney Sandstone (Figure 3). This erosional surface, traceable throughout the central Flinders Ranges, represents a marked fall in base-level. The lack of evidence of subaerial exposure and fluvial sedimentation in canyon floors indicates that they were cut and filled subaqueously. Syndepositional faulting on canyon margins, in the Chace Range and at Nilpena, suggests that the disconformable base of the Ediacara member was also a product of salt withdrawal and subsidence of mini-basins, as proposed by Dyson and Rowan (2004) at other stratigraphic levels.

The Ediacara Member commenced with a lowstand systems tract and a massive unit sandstone with basal, angular sandstone breccias filling local, 50 m wide and 1–20 m deep, step-sided channels within the valley incisions. In the deepest canyons, a sudden change to well laminated clayey siltstones represents a maximum flooding surface. Fossils appear on the first thin, storm sandstone beds in the upward coarsening and thickening beds in the upper half of the Ediacara Member. The Ediacara Member is a deltaic succession that prograded from the NW, filling local depocentres via incisions 50–250 m deep. The accommodation space provided by the canyons enabled a number of facies to be developed that are unknown from the canyon shoulders. On the western side of the Flinders Ranges, at Nilpena, Ediacara Conservation Park and the Mt Scott Range, the Ediacara Member is represented by a series of facies associations, the lowest of which fill smaller-scale channels and narrow canyons. While these are correlated with thicker parasequences in the main ranges, which Gehling (2000) included within the amended Ediacara Member, Jenkins and Nedin (2007) redefined these as separate members. Since these subsidiary units contain key taxa of the Ediacara biota, they represent a single generic package, unless a significant time separation can be established. Two to three parasequences cap the Ediacara Member, each with fossil horizons in the zone between fair-weather and storm wave base (Figure 5b). The top is marked by crossbedded, shore-face

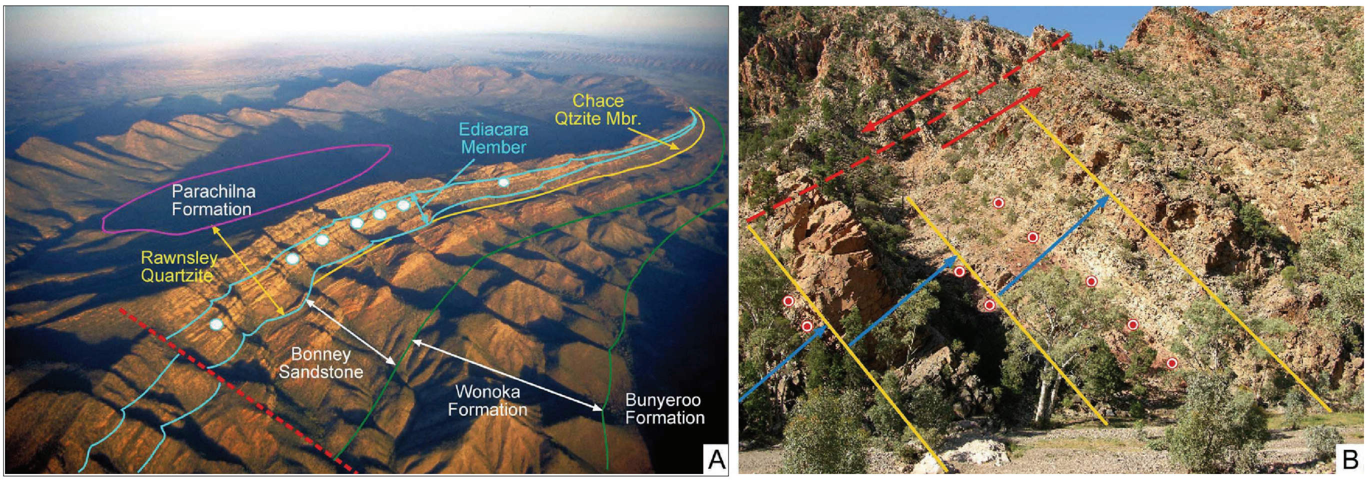


***Figure 5** (A) Oblique aerial photo of Wilpena Pound, a raised synform of the Pound Subgroup capping the Ediacaran succession in the central Flinders Ranges. The fossiliferous Ediacara Member represented by discs at the base of the topographic rim. Formations age to the right. An erosional remnant of the early Cambrian Parachilna Formation remains in the centre of the synform. (B) Three upward shallowing parasequences of the Ediacara Member in Brachina Gorge, Flinders Ranges National Park, with fossil horizons represented by discs.*

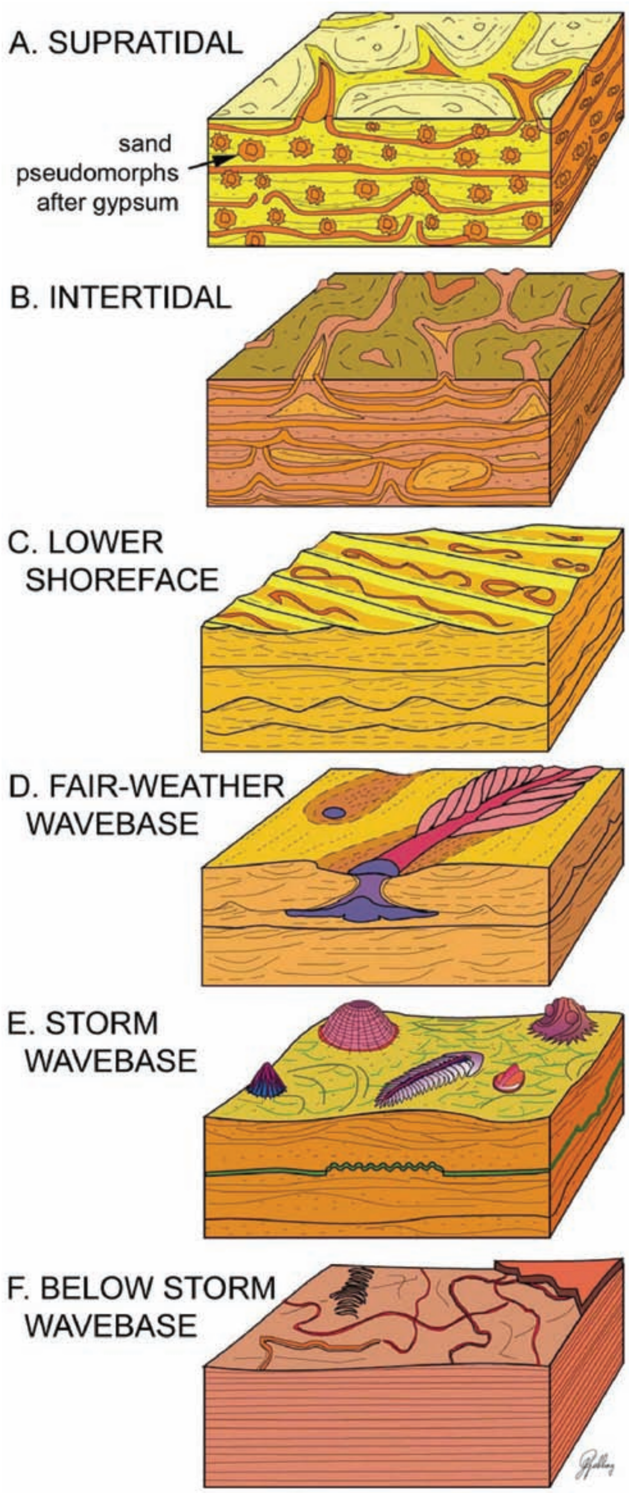


***Figure 6 Facies of the Rawnsley Quartzite with depth-related associations of microbial mats and fossils. (A). Supratidal sandflats with desiccation of microbial mat bound sand lamination and sand-pseudomorphs after sulfate. (B). Intertidal sandflats with domed and disrupted microbial mats. (C). Lower shore-face wave rippled sand with synaeresis cracks in microbial bound ripple troughs. (D). Fair-weather wave base with preservation of anchoring holdfasts of fronds, either flattened or torn away by storm surge. (E). Storm wave base with communities of benthic Ediacara organisms. (F). Below wave base preservation below distal flows of storm surge sand.***

sandstone units grading into the upper 500 m of the Rawnsley Quartzite. The upper, Rawnsley Quartzite continues with the same association of crossbedded and petee sandstones found in the Chace Quartzite Member. Near the top of the Rawnsley Quartzite there is a transition back into wave-ripple bedded and low angle crossbedded, shore-face sandstone, with rare beds of discoidal fossils (Gehling, 2000).

## *Ediacara Biota*

Ediacara fossils were first discovered by Sprigg (1947, 1949) at the Ediacara minefield on the western rim of the Flinders Ranges. There the Ediacara Member is restricted to c. 30–50 m of thinly bedded sandstones in a condensed and truncated Rawnsley Quartzite (Goldring and Curnow, 1967). Early study of the beds suggested beach stranding of 'jellyfish' (Glaessner and Daily, 1959; Glaessner and Wade, 1966), largely because ripple marks and synaeresis cracking of microbially bound sand were mistaken for exposure on tidal flats. Fossils of the Ediacara biota are most common on soles of thin to medium bedded sandstones from storm to fair-weather wave base (Figure 6d, e, f; Gehling, 1991, 1999). However, transported organisms occur also as deformed impressions and sand-filled casts in massive sandstone beds within canyon filling facies. Most common in lower shoreface beds, discoidal holdfasts (Figure 7b, h), with radial and concentric characters, originally named *Cyclomedusa, Ediacaria, Medusinites* etc. (Sprigg, 1947, 1949; Wade, 1969), are now regarded as preservational states of *Aspidella* (Gehling et al., 2000). In most cases they represent as attached, bulbous holdfasts of frond-like organisms (Figure 7b), where the frond has been torn off or decayed before preservation, leaving a stem-like impression, escape marks or a ghost of a frond on the opposite side of the bed (Tarhan et al., 2010). *Charniodiscus* (Figure 7a), the most common frondose form, resembles pennatulid soft corals in that the branches bear serial, polyp-like subdivisions. Although widespread in deep and shallow marine and environments (Narbonne, 2005), their phylogenetic affiliations remain uncertain.

*Helminthoidichnites*, groove and levee traces, were made by tiny motile benthic organisms too small to be recognised as external moulds (Droser et al., 2005). Gehling et al. (2005) interpreted burial contraction marks, bedding plane distribution and serial resting traces made by *Dickinsonia* (Figure 7l) and *Yorgia* (Figure 7k), as evidence of muscularity, tactophobic behaviour, and direct absorption feeding on benthic mats, made by of motile, bilaterian-grade organisms. The fan-shaped sets of radular scratch marks associated with *Kimberella* (Figure 7s, t) suggest affinities with molluscan grade organisms (Gehling et al., 2005).

Taphonomy of fossils of the Ediacara biota is a vital prerequisite to assessment of their paleobiology and paleoecology (Wade, 1968; Gehling et al., 2005). Preservation of moulds of soft-bodied organisms on bed bases resulted from smothering of microbial mats by storm-sand surges strong enough to orient some frondose fossils, but not so erosive as to have stripped off unattached benthic organisms. Bacterial decay and reduction of Fe and sulfate produced pyritic death masks in a sole veneer that enabled coherent external moulds to form (Gehling, 1999; Gehling et al. 2005). The finest preservation of body fossils generally coincides with "textured organic surfaces", suggesting a mature mat community involving bacterial films, dead organisms and sometimes, close packed benthic organisms (Droser et al., 2006; Gehling and Droser, 2009).

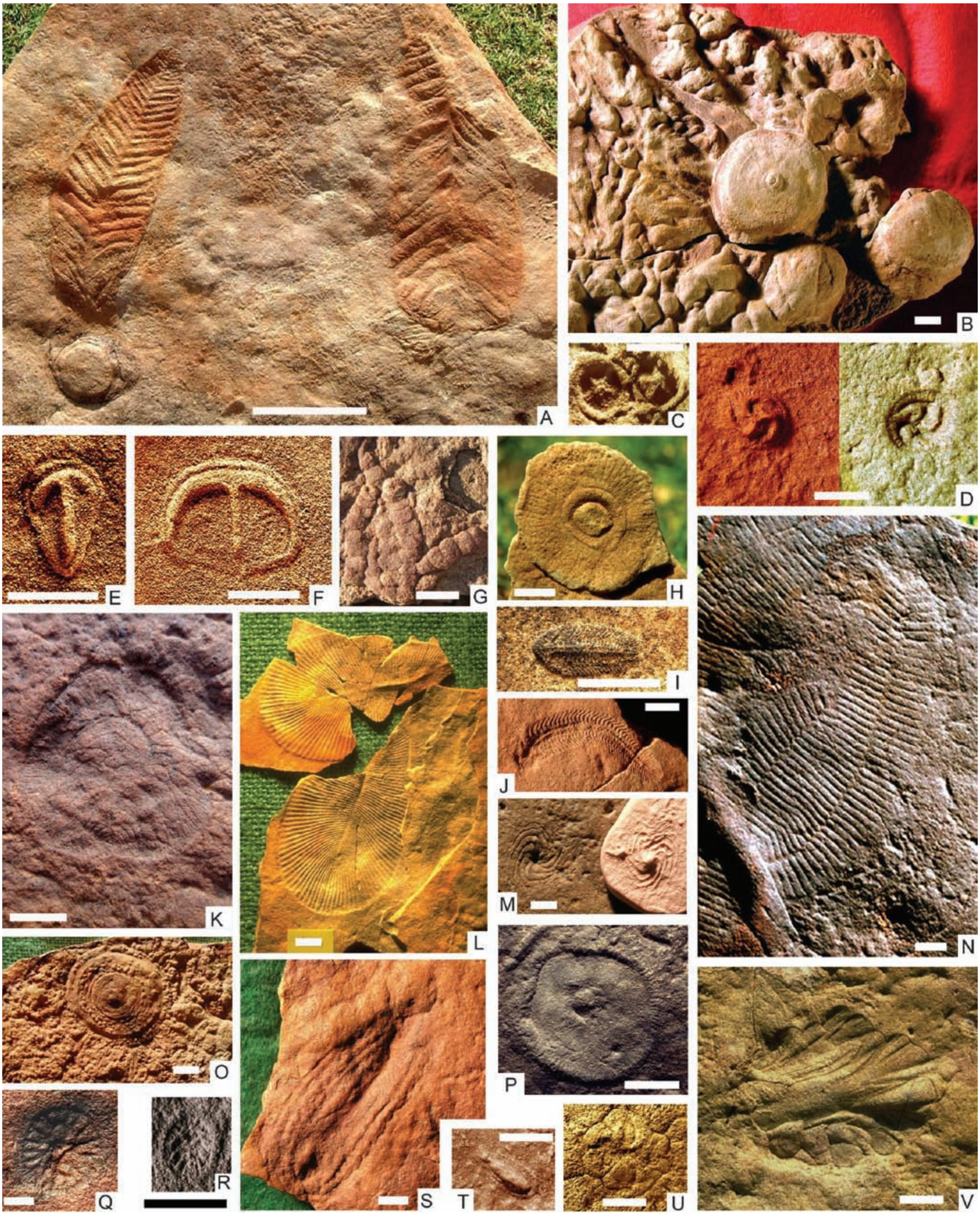


*Figure 7 Fossils of the Ediacara biota, Ediacara Member, Rawnsley Quartzite, central Flinders Ranges. (A).* **Charniodiscus arboreus.** *(B).* **Aspidella** *with stalk, with load casts. (C).* **Arkarua.** *(D).* **Tribrachidium,** *rubber cast and external mould. (E-F).* **Parvancorina,** *rubber casts. (G).* **Funisia.** *(H).* **Aspidella** *(***Cyclomedusa radiata***). (I).* **Spriggina ovata.** *(J).* **Spriggina floundersi.** *(K).* **Yorgia.** *(L).* **Dickinsonia costata,** *part and counterpart. (M).* **Eoandromeda,** *natural cast and rubber mould. (N).* **Phyllozoon.** *(O).* **Eoporpita.** *(P). Palaeophragmodictya with spicule impressions. (Q).* **Rugoconites.** *(R).* **Praecambridium.** *(S).* **Kimberella** *(holotype). (T).* **Kimberella** *with proboscis, rubber cast. (U). Conomedusites. (V).* **Inaria** *(holotype). Scale bars 1 cm, except (A) where it is 10 cm.*

Narbonne (2005) assessed the taphonomic, environmental and age constraints of all known assemblages of the Ediacara biota. However, regional distinctions between assemblages are now less distinct. Serial excavation of fossil beds has revealed that paleoenvironments were as strong an influence on the taxonomic composition of preserved assemblages as is their stratigraphic position. Typically, fossiliferous beds are dominated by one taxon, sometimes in very large numbers. Droser and Gehling (2008) described entire bedding surfaces covered with matted specimens of the chambered, tubular animal, *Funisia* (Figure 7g). Local comparisons show that distinct sedimentary facies correlate with taxa present. Diverse assemblages of Ediacara fossils in the Flinders Ranges include many elements of the Mistaken Point and the Nama assemblages of Waggoner (2003). Many smaller taxa are preserved only as external moulds on sole surfaces. Most discoidal and frondose taxa are preserved in positive hyporelief (Figure 7). These contrasting preservational modes and different body plans suggest Ediacaran phyletic divergence (Gehling 1991) arguing against of claims that most Ediacara taxa represent a single extinct clade (Seilacher, 1984, 1989, 1992; Brasier and Antcliffe, 2004, 2009). Globally, most assemblages of the Ediacara biota have at least some taxa in common. For example, although *Eoandromeda* (Figure 7m), a clockwise spiraling, eight-armed disc, and the close-packed discs of *Nemiana,* are the only taxa common to both China and South Australia (Zhu et al., 2008), most taxa of the White Sea assemblage are common to South Australia (Gehling et al., 2005). However, similar fossil assemblages, on widely divergent continental fragments in the Ediacaran, are not explained by the present distribution of the source rocks alone (Wagoner, 2003).

## Basal Cambrian Stratigraphy

The earliest Cambrian stratigraphy of South Australia (Jago et al., 2012) continues the late Ediacaran sedimentary character with siliciclastic units at its base. In the northern Flinders Ranges (Figure 3) the Uratanna Formation is a shallowing upward, incision-fill succession, up to 400 m thick (Mount and McDonald, 1992; Jensen et al., 1998), resembling the Ediacara Member. In the Mount Scott Range, the Uratanna Formation cuts down into the Rawnsley Quartzite to the thinned Ediacara Member (Daily, 1973), and passes up into the shallow marine, shore-face and lagoonal Parachilna Formation, which everywhere onlaps the Rawnsley Quartzite. Regionally, the Ediacaran-Cambrian boundary is marked by un-burrowed sandstone of the Rawnsley Quartzite disconformably overlain by the Parachilna Formation with dense accumulations of *Diplocraterion* burrows and, further up section, diverse trace fossils including *Plagiogmus* and small helcionellids (Zang et al., 2007).

In a valley-fill succession, 40 km E of Leigh Creek, *Treptichnus pedum* and *Phycodes coronatum* first appear in sandy facies more than 200 m above the base of the Uratanna Formation (Jensen et al., 1998). Considering the facies problem, the base of the Cambrian succession is locally mapped at the basal disconformity (Nedin and Jenkins, 1991). The subtidal, storm bedded sandstones in the upper part of this succession contain a diverse Cambrian trace fossil assemblage together with a single horizon of frond-like fossils of Ediacaran aspect (Jensen et al., 1998). Daily (1972) described *Rusophycus* burrows from this sandy facies of the Uratanna Formation in the Mundy Waters Syncline, that demonstrate the existence of arthropods with paired burrowing appendages in the earliest Cambrian, well before the first mineralised trilobites. *Sabellidites cambriensis* and acritarchs of the Redkinia-Cymatiosphaera Assemblage Zone in the Uratanna Formation (Mount and McDonald, 1992; Zang et al., 2007) indicate a Nemakit-Daldyn age.

## Summary

The Ediacaran succession in the Flinders Ranges preserves the transition from a biological system of small organic walled eukaryotic and prokaryotic microfossils that were responsible for stromatolitic reefs and raised free oxygen levels. The Ediacaran marked a series of biological changes in the wake of Cryogenian ice ages that precipitated the rise of multicellular life. In the Flinders Ranges, the separation between Ediacaran and early Cambrian siliciclastic sequences is defined by the onset of deep burrowing that is closely associated with the first small shelly fossils. This indicates an ecological revolution that extinguished many of the life-styles of the Ediacara biota and ushered in the Phanerozoic. The behavioral evolution that produced deep burrowing, whether in response to predation or to access buried organic matter, effectively destroyed the microbial mat "film" that recorded impressions of soft-bodied organisms in the Ediacaran Period.

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*by James B. Jago[1], James G. Gehling[2], John R. Paterson[3], Glenn A. Brock[4] and Wenlong Zang[5]*

# Cambrian stratigraphy and biostratigraphy of the Flinders Ranges and the north coast of Kangaroo Island, South Australia

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*The lower Cambrian sediments of the Flinders Ranges, South Australia can be divided into three sequence sets. They rest unconformably on the Ediacaran succession. Sequence set Є1 comprises lower clastic units overlain by a carbonate dominated marine succession that shows marked lateral and vertical facies changes. Sequence sets Є2 and Є3 together comprise a largely clastic dominated succession of marginal marine to non-marine sediments with subordinate shallow marine carbonates. Sequence set Є1 is richly fossiliferous at some levels with biostratigraphy established for trilobites, archaeocyaths, brachiopods, small shelly fossils, acritarchs and molluscs. The Emu Bay Shale Lagerstätte (Cambrian Series 2) from the north coast of Kangaroo Island occurs within a clastic-rich shelf succession dominated by conglomerate and sandstone. The fossil content is dominated by trilobites in terms of relative abundance and currently over 50 taxa are known including,* Anomalocaris*, the bivalved arthropods* Isoxys *and* Tuzoia*, the nektaspids* Emucaris *and* Kangacaris*, the megacheiran* Oestokerkus *amongst a variety of other arthropods. Other common taxa include palaeoscolecid worms,* Myoscolex*, sponges, hyoliths, brachiopods, a vetulicolian and several other enigmatic forms. The oldest known well preserved complex arthropod eyes occur in this biota.*

During the Cambrian, Australia was near the equator and in the tropical carbonate development zone (Brock et al., 2000, Jago et al., 2006). In South Australia, Cambrian sediments occur in the Stansbury, Arrowie, Warburton and eastern Officer basins (Figure 1). In the Stansbury and Arrowie basins, Cambrian sediments rest unconformably on Neoproterozoic rocks, probably reflecting the influence of the late Ediacaran Petermann Orogeny of Central Australia. Cambrian deposition was strongly influenced by regional tectonism, starting with rifting and finishing with the 514–490 Ma Delamerian Orogeny. The following information is based on Zang et al., (2004, 2006). More detailed descriptions and interpretation of individual stratigraphic units are found in Gravestock (1995). Figure 2 summarises the Cambrian stratigraphic relationships of the Arrowie and Stansbury basins. The Cambrian of the N coast of Kangaroo Island is treated separately.

## Cambrian succession in the Arrowie Basin

Gravestock (1995) divided the Cambrian of South Australia into four sequence sets, Є1, Є2, Є3 and Є4. Set Є4, from the Warburton Basin, is not considered herein. The lowest Cambrian sequence,Є1.0, Uratanna Formation, has limited outcrop in the northern Flinders Ranges. It comprises channel deposits overlain by prodeltaic and deltaic siltstones and sandstones and was deposited in valleys cut into the Ediacaran Rawnsley Quartzite. Where the Uratanna Formation is absent, the Parachilna Formation disconformably overlies the Rawnsley Quartzite (Gravestock, 1995); where both the Uratanna and the Parachilna Formations are missing, the younger Hawker Group carbonates directly overlie the Ediacaran sediments. Mount and McDonald (1992) suggested that the change from the clastic dominated Ediacaran, Uratanna and Parachilna successions indicated a decrease in clastic sediment supply due to drowning of source areas in central and southern Australia.

Sequence Є1.1 (Parachilna Formation, Woodendinna Dolomite, lower Wilkawillina Limestone) contains an unconformity that divides the sequence into subsequences Є1.1A and Є1.1B; the unconformity does not extend beyond the shelf edge. The Parachilna Formation comprises transgressive upwards fining sandstones and siltstones with minor carbonates. The basal sandstones are strongly bioturbated with abundant *Diplocraterion* in places. The maximum of this initial Cambrian transgression was reached at the top of Є1.1A (c. Tommotian) with deposition of the Woodendinna Dolomite that comprises oolitic, stromatolitic, dolomitic shale with desiccation

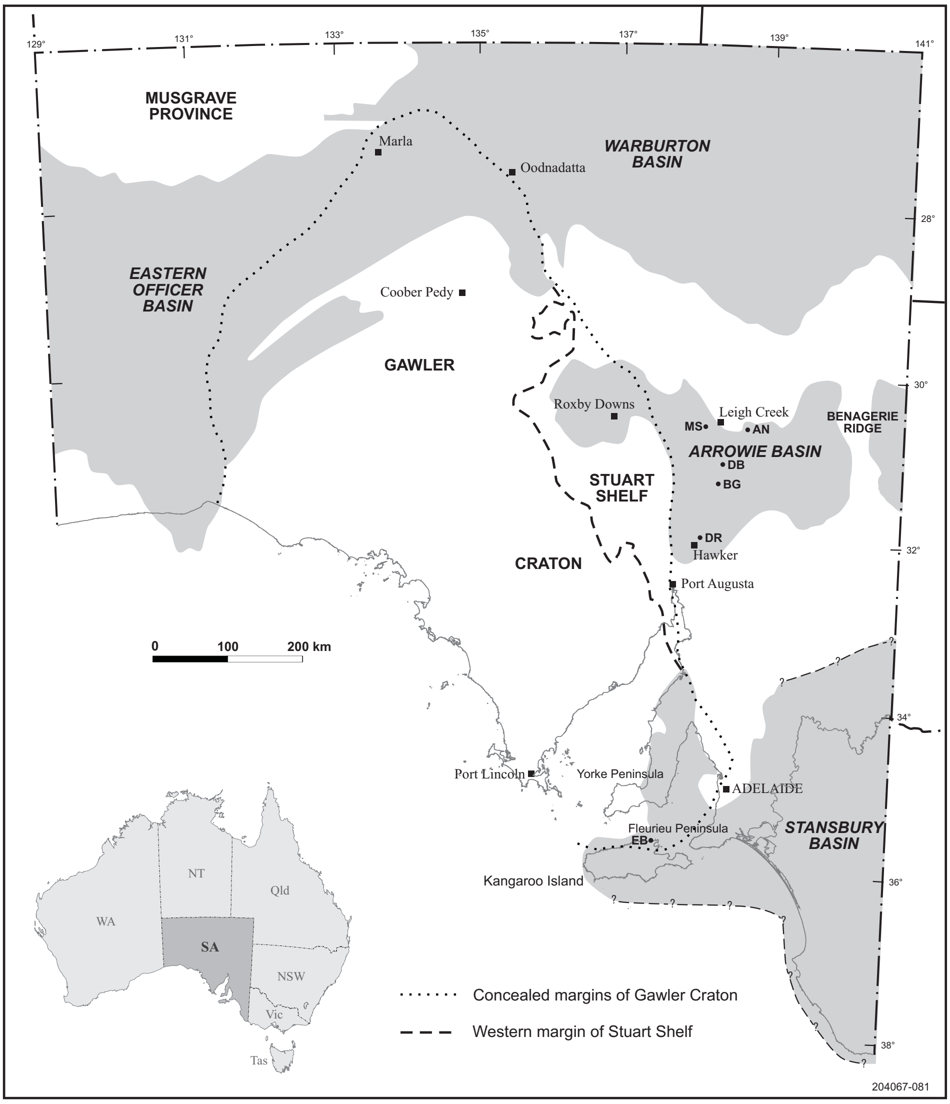


***Figure 1 Cambrian basins of South Australia (after Jago et al., 2006). AN, Angepena-Nepabunna area; BG, Bunkers Graben; DB, Donkey Bore Syncline; DR, Druid Range; EB, Emu Bay; MS, Mt Scott Range.***

cracks and halite casts plus abundant quartz sand horizons (Haslett, 1975). Sequence Є1.1B commences with the lower Wilkawillina Limestone representing an Atdabanian shelf limestone that shows marked lateral facies changes and passes basinwards (eastwards) into the Wirrapowie Limestone which contains the oldest known archaeocyaths and trilobites in Australia. On shelf areas, the Woodendinna Dolomite is overlain disconformably by the lower Wilkawillina Limestone; basinwards there is no corresponding break within the Wirrapowie Limestone.

Sequences Є1.1 and Є1.2 are separated by the Flinders Unconformity (James and Gravestock, 1990) with the lower Wilkawillina Limestone disconformably overlain by either the middle Wilkawillina Limestone (platform carbonates) or the Mernmerna Formation (dark slope facies carbonates and calcareous siltstones). There are numerous lateral and vertical facies changes (Daily, 1976). The limestone shows considerable thickness variations with 100–500 m in many sections, up to 800 m in the Bunkers Graben (Clark, 1990) and over 2,500 m in the Nepabunna Trough (Gravestock, 1995). In the northern Flinders Ranges, in the Angepena–Nepabunna area, there are two siltstone units (Midwerta Shale, Nepabunna Siltstone) that intertongue with limestones of the Mernmerna Formation.

Sequence Є1.3 commences with the Bunkers Sandstone, which in the Bunkers Graben comprises up to 300 m of cross bedded, quartz sandstone, but elsewhere it is mainly either quite thin or missing; however, in the Donkey Bore Syncline, to the NE, it comprises deeper water basin floor fan deposits (Reilly and Lang, 2003). The distinctive khaki Oraparinna Shale represents a transgressive unit; in the eastern Nepabunna Trough the Oraparinna Shale intertongues with fine to medium grained feldspathic sandstones of the Narina Greywacke (Gravestock, 1995). In places, the Oraparinna Shale abuts reef complexes of the upper Wilkawillina Limestone (Clark, 1990). The dark silty limestones of the upper Mernmerna Formation comprise the main transgressive unit beyond the shelf edge. The upper Wilkawillina Limestone and the Moorowie Formation both contain archaeocyath-rich reef and biohermal complexes. These include some of the earliest known corals (e.g., Fuller and Jenkins, 2007) and were deposited during the worldwide Botoman transgression (Gravestock and Shergold, 2001).

Above the Hawker Group in the Flinders Ranges there is much less lateral and vertical lithological variation. Sequence Є2 was deposited during the world wide Toyonian regression (Gravestock and Shergold, 2001); it comprises the dominantly iron-rich red Billy Creek Formation deltaic sandstones and siltstones (Moore, 1980) and the overlying shallow marine carbonates of the Wirrealpa Limestone (Youngs, 1977). In the Mt Scott Range area, the Aroona Creek Limestone is equivalent to the Wirrealpa Limestone.

Gravestock (1995) suggested that Sequence Є3 comprises the clastic dominated Lake Frome Group, although it could be argued that sequences Є2 and Є3 are a single clastic dominated succession briefly interrupted by the Wirrealpa Limestone. The lowest member of the Lake Frome Group is the Moodlatana Formation that comprises marginal marine shale, siltstone (including evaporite casts and abundant trace fossils) and cross bedded sandstones with two or three thin carbonate units near the top, one of which contains the youngest known Cambrian trilobites in the Flinders Ranges. The conformably overlying Balcoracana Formation comprises a shallow to marginal marine cyclic succession of shallowing upwards packages of red mudstones with abundant trilobite tracks that pass up into thin horizons of stromatolitic carbonates. The Pantapinna Sandstone comprises mainly planar to trough cross bedded fine to medium grained sandstones with minor siltstone and shale; much of the Pantapinna is probably of marginal marine to fluvial origin, although there is some open marine influence. The Grindstone Range Formation has a shallow marine lower part comprising well bedded fine sandstone with abundant planar cross bedding. The upper half, the Dawson Hill Member, is a fine to medium grained sandstone with abundant tabular cross bedding; it has numerous pebble and cobble clast horizons (Jago et al., 2010). Gravestock (1995) suggested a fluvial sandy braidplain depositional environment. The top of the Grindstone Range Formation is not exposed.

## Cambrian biostratigraphy of the Stansbury and Arrowie Basins

The Cambrian biostratigraphy of South Australia has formed the basis of early Cambrian correlations between Australia and elsewhere (Gravestock, 1984; Jell in Bengtson et al., 1990; Zhuravlev and Gravestock, 1994; Paterson and Brock, 2007). Summaries are found in Gravestock et al. (2001), Jago et al. (2006) and Kruse et al. (2009). In the Arrowie and Stansbury basins, fossils occur mainly in the lower part of the succession, i.e., Hawker Group and correlates. Jell in Bengtson et al. (1990) erected four trilobite zones: *Abadiella huoi* (base), *Pararaia tatei*, *P. bunyerooensis* and *P. janeae* Zones (Figure 2). Paterson and Brock (2007) correlated the *A. huoi* Zone with the Chinese *Parabadiella* Zone and with the latest Atdabanian–earliest Botoman in Siberia.

The *Pararaia tatei* Zone correlates with the *Eoredlichia-Wutingaspis* Zone of China and the early Botoman *Bergeroniellus micmacciformis–Erbiella* Zone of Siberia (Jell in Bengtson et al., 1990, Paterson and Brock, 2007). Paterson and Brock (2007) correlated the *P. bunyerooensis* Zone with the *Yunnanocephalus* assemblage subzone (upper *Eoredlichia–Wutingaspis* Zone) of South China. The *Pararaia janeae* Zone contains the most trilobite taxa; Paterson and Brock (2007) suggested correlation with the Tsanglangpuan Stage of China and the *Bergeroniellus gurarii* to *Bergeroniaspis ornata* Zones (Botoman) of Siberia.

Higher in the succession in the Flinders Ranges (Arrowie Basin), trilobites are known from only three stratigraphic levels. The emuellid, *Balcoracania dailyi,* occurs in gregarious clusters in the Billy Creek Formation (Paterson et al., 2007a). This is probably late Botoman and may correlate with the Emu Bay Shale and Marsden Sandstone from Kangaroo Island (see below). *Redlichia guizhouensis* and *Onaraspis rubra* occur in the Wirrealpa Limestone and Moodlatana Formation, respectively; these last two formations have been correlated with the late early Cambrian Lungwangmiaoan Stage of China (Jell in Bengtson et al., 1990), or the latest Toyonian of Siberia (Paterson and Brock 2007).

The shallow water carbonates of the Arrowie and Stansbury basins contain abundant archaeocyaths. Gravestock (1984) and Zhuravlev and Gravestock (1994) recognised three Atdabanian archaeocyath zones from the Wilkawillina Limestone beneath the Flinders Unconformity, *Warriootacyathus wilkawillensis* (base), *Spirillicyathus tenuis* and *Jugalicyathus tardus* Zones (Figure 2). The informal term *Syringocnema favus* beds was erected by Zhuravlev and Gravestock (1994) for the abundant assemblage of archaeocyaths and radiocyaths that occurs in both the Arrowie and Stansbury basins within the Wilkawillina Limestone and correlates above the Flinders Unconformity. Paterson et al. (2007b) suggested that the archaeocyathan fauna referred to as the '*Syringocnema favus* beds' is early Botoman (pre-*Pararaia janeae* Zone) in age. Kruse (1991) reported the youngest known archaeocyath fauna in Australia from the Wirrealpa Limestone.

Four informal brachiopod assemblages are recognized (Jago et al., 2006). The oldest assemblage, dominated by a new species of *Askepasma,* is mid-Atdabanian in age from the basal Wilkawillina Limestone and Wirrapowie Limestone of the Arrowie Basin. Undescribed calciate taxa from the lower Wirrapowie Limestone co-occur with this paterinid taxon. The succeeding *Eoobolus* aff *viridis* Assemblage Zone, first erected by Gravestock et al. (2001), ranges from late Atdabanian to mid Botoman in both the Arrowie and Stansbury basins; the most abundant taxon in this zone is *Kyrshabaktella davidi.* The overlying *Eoobolus priscus* Assemblage Zone is latest Botoman and occurs within the upper Parara Limestone in the Stansbury Basin and the upper Mernmerna Formation in the Arrowie Basin. The youngest brachiopod zone is the *Vandalotreta djagoran* Assemblage Zone which is characterised by the acrotretid *Vandalotreta djagoran* in both the Arrowie and Stansbury basins, plus the obolellid *Trematobolus wirrialpensis* in the Toyonian Wirrealpa Limestone.

There is a diverse and relatively common molluscan fauna in the early Cambrian carbonate successions of the Arrowie and Stansbury basins. Gravestock et al. (2001) erected four informal molluscan assemblages: the *Pelagiella subangulata* (base), *Bemella communis, Stenotheca drepanoida* and *Pelagiella madianensis* "zones".

Topper et al. (2011) described the oldest bradoriid fauna in Australia from the Ajax and Wirrapowie limestones of the Flinders Ranges. The oldest bradoriid occurs c. 20 m below the first appearance of *Abadiella huoi*. Skovsted et al. (2006) and Topper et al. (2007, 2011) described biostratigraphically significant bradoriid assemblages correlating with each of the younger (post *A. huoi*) trilobite zones noted above.

Abundant small shelly fossils (SSF) within the carbonate units of the Stansbury and Arrowie basins have been documented by Bischoff (1976), Bengtson et al. (1990), Brock and Cooper (1993), Gravestock et al. (2001), Skovsted et al. (2008, 2009 a, b, 2011) and Topper et al. (2009). Skovsted et al. (2011) demonstrated a consistent biostratigraphic disjunct between the tommotiid *Eccentrotheca helenia* and isolated sclerites of *Kulparina rostrata* and *"Eccentrotheca" guano*. *Eccentrotheca helenia* occurs in the Ajax and Wilkawillina Limestones with a stratigraphic range coincident with *Abadiella huoi*. In contrast, the stratigraphic ranges of *K. rostrata* and *'E.' guano* are largely coeval, with the LAD of both sclerite types occurring significantly below the FAD of *A. huoi* and *E. helenia*. Gravestock et al. (2001) erected three informal SSF assemblage zones: *Hippopharangites dailyi* (base), *Australohalkieria parva* and *Kaimenella reticulata,* although the utility of these zones remains to be tested outside the Stansbury Basin.

Seven acritarch assemblage zones have been reported from the Lower Cambrian of the Arrowie and Stansbury basins (Zang et al., 2004, 2007). These extend from near the Ediacaran–Cambrian boundary within the Uratanna Formation to the Toyonian. Details are given by Zang et al. (2007).

## Cambrian stratigraphy of the north coast of Kangaroo Island

Flöttmann et al. (1998) suggested that the Cambrian successions on Kangaroo Island comprise essentially unmetamorphosed shelf

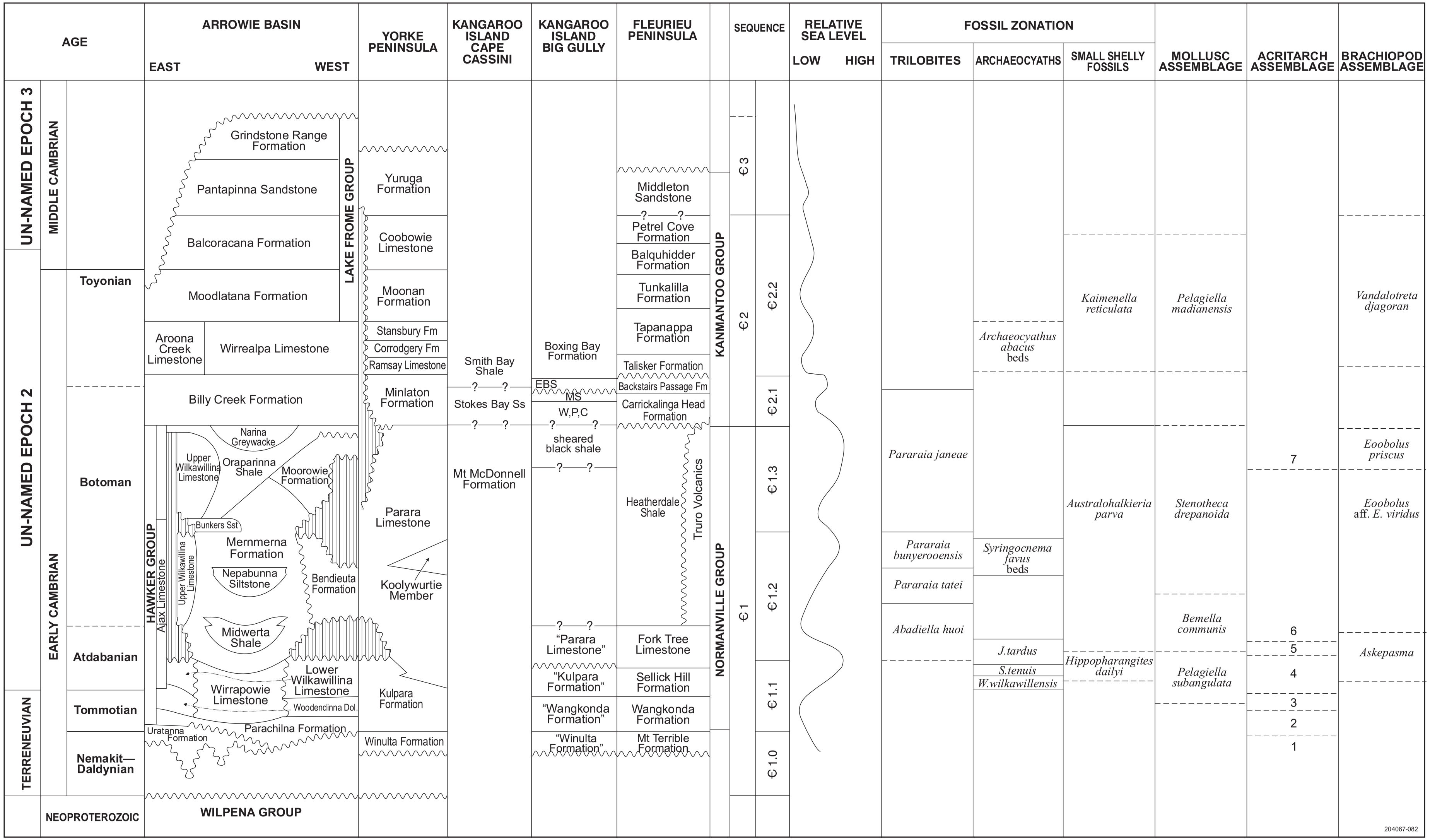


***Figure 2  Generalised Cambrian stratigraphy, sequence stratigraphy and biostratigraphy of the Arrowie and Stansbury basins, South Australia (after Jago and Zang, 2006). EBS, Emu Bay Shale; MS, Marsden Sandstone; WPC, White Point Conglomerate.***

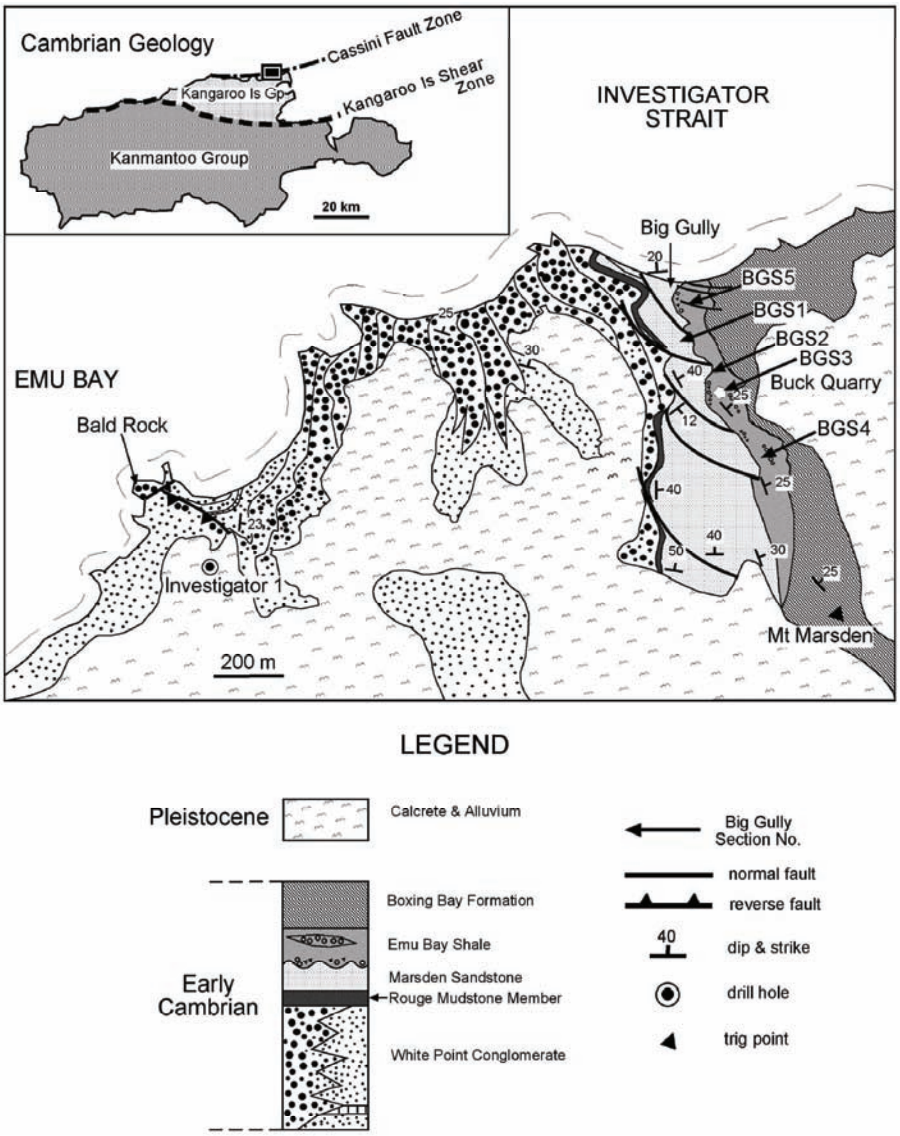


***Figure 3 Geological map of the Big Gully area, N coast of Kangaroo Island (from Gehling et al., 2011). BGS, Big Gully Section.***

sediments, the Kangaroo Island Group (Daily, 1956), to the N of the Kangaroo Island Shear Zone (Figure 3) and the thick metamorphosed turbidite dominated Kanmantoo Group deposits to the S. Daily et al. (1979) considered that on the N coast of Kangaroo Island, the Kangaroo Island Group comprised six formations: Mt McDonnell Formation (base, containing Botoman archaeocyaths), Stokes Bay Sandstone, Smith Bay Shale, White Point Conglomerate, Emu Bay Shale (containing the Emu Bay Shale Lagerstätte) and Boxing Bay Formation. The top of the succession is not exposed. The three lowest formations outcrop between Smith Bay and Snelling Beach to the W of Emu Bay; from Emu Bay to the E the top three formations outcrop between Cape D'Estaing and Point Marsden, with the only suggested link being the Smith Bay Shale reported by Daily et al. (1980) at the base of a measured section that continues up through the White Point Conglomerate, Emu Bay Shale and Boxing Bay Formation. Gehling et al. (2011) regarded the "eastern" Smith Bay Shale of Daily et al. (1980) as part of the White Point Conglomerate and thus there is no proven link between the two areas of Cambrian sediments along the N coast of Kangaroo Island. Gehling et al. (2011) separated out the top part of the White Point Conglomerate as described by Daily et al. (1980) as a new formation, the Marsden Sandstone, which includes the thin Rouge Mudstone Member at the base. Stratigraphic correlation of the Cambrian units on the N coast of Kangaroo Island with those on Yorke and Fleurieu peninsulas is shown in Figure 2.

Gehling et al. (2011) noted that the tectono-sedimentary setting of the Cambrian successions to the N of the Kangaroo Island Shear Zone is unclear. Nedin (1995b) suggested that the Kangaroo Island Group may have been deposited in a series of sub-basins in the area of syndepositional tectonic activity. This tectonic activity is what Daily and Forbes (1969) termed the Kangarooian Movements and may represent an early phase of the Delamerian Orogeny, the timing of which is discussed by Foden et al. (2006).

## Geology in the vicinity of the Emu Bay Shale Lagerstätte

Gehling et al. (2011) mapped the area around the known occurrences of the Emu Bay Shale Lagerstätte (Figure 3). The lowest unit is the White Point Conglomerate that, in coastal sections, comprises cobble to boulder polymict conglomerate with subordinate sandstone and mudstone horizons. Clast types include archaeocyath-bearing limestones, dolomites, granite, gneiss and quartzite, all consistent with having been derived from the Proterozoic and Cambrian rocks of Yorke Peninsula to the N. The clast size decreases rapidly to the S with the conglomerate horizons becoming lenticular and disappearing entirely about 400 m S of the coast. The White Point Conglomerate is overlain by the Marsden Sandstone which is essentially a fine to medium-grained feldspathic sandstone at the base of which is the Rouge Mudstone Member that comprises 3 m of bioturbated argillaceous limestone containing the emuellid trilobite *Balcoracania dailyi*. The Marsden Sandstone is overlain unconformably by the Emu Bay Shale (Gehling et al., 2011). Detailed mapping of the area indicates that there was considerable syndepositional folding and faulting during deposition of the White Point Conglomerate and the Marsden Sandstone (Figures 3 and 4).

The Emu Bay Shale comprises dark grey to black laminated mudstone, with subordinate lenticular cross bedded fine sandstone beds with at least one conglomerate horizon with a clast content similar to that found in the White Point Conglomerate. The mudstone containing the Lagerstätte is not bioturbated. The water column above the sediment-water interface was oxic while the sediment below the sediment-water interface was anoxic with a cyanobacterial mat at the interface playing an important part in the preservation of the Lagerstätte (McKirdy et al., 2011; Hall et al., 2011). The Emu Bay Shale coarsens upwards with large arthropod tracks including

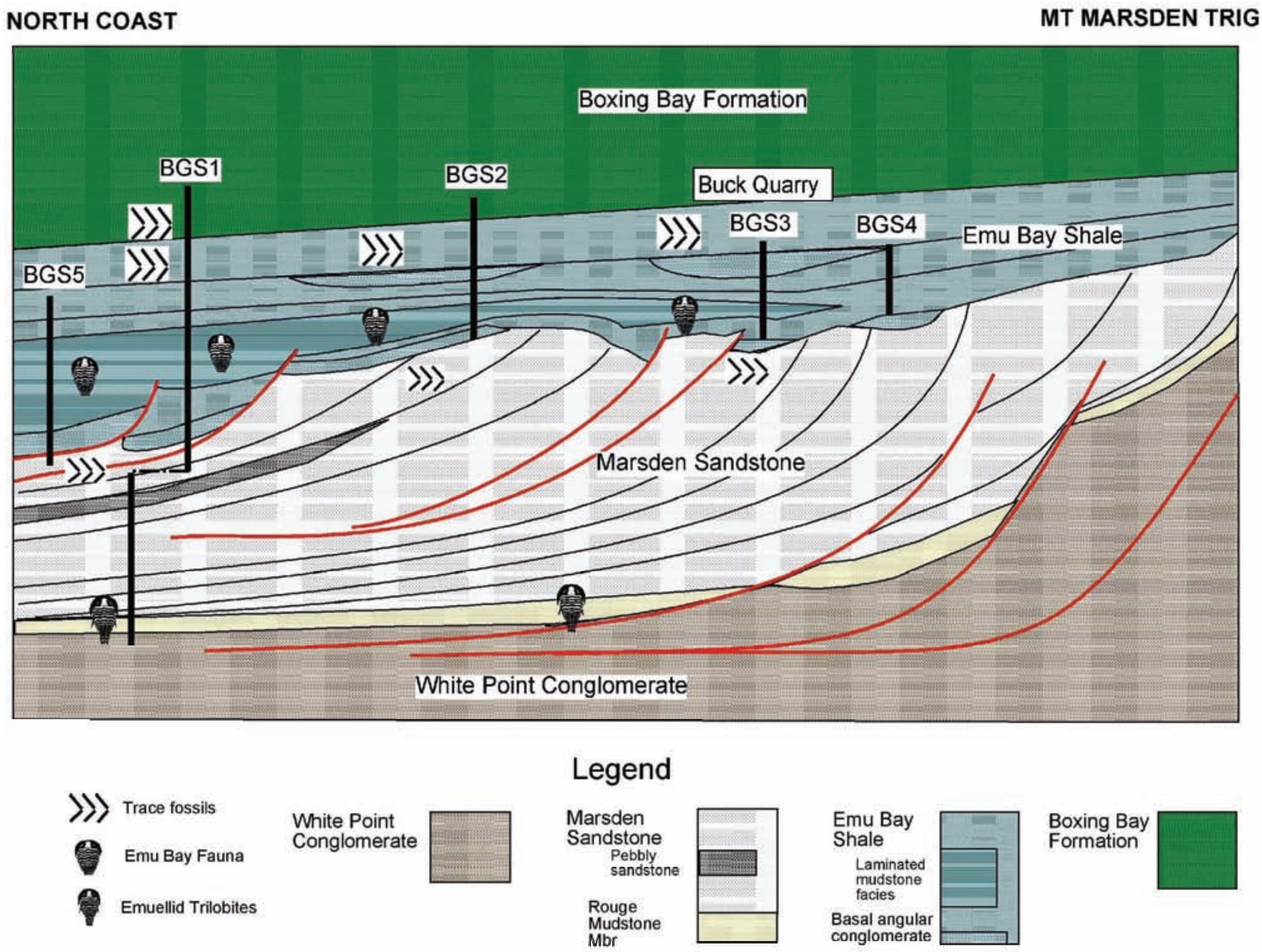


***Figure 4 Schematic N-S cross-section through the Big Gully area (from Gehling et al., 2011). BGS, Big Gully Section.***

*Cruziana* and *Monomorphichnus* occurring in fine sandstone towards the top of the unit. The Emu Bay Shale is overlain by the subtidal Boxing Bay Formation that comprises a red-brown feldspathic sandstone with abundant arthropod tracks.

## Fossil content of the Emu Bay Shale Lagerstätte

The Emu Bay Shale Lagerstätte was first reported from coastal outcrops 3 km E of Emu Bay by Daily (1956), with the first descriptions of soft-bodied taxa published by Glaessner (1979). Summaries are in Paterson and Jago (2006) and Paterson et al. (2008). Excavation at a new locality, Buck Quarry, several hundred metres inland commenced in 2007 (Paterson et al., 2008). This has yielded a greater variety of taxa than the coastal locality. At this new locality the fossil content of the Emu Bay Shale Lagerstätte is dominated by the trilobite *Estaingia bilobata* that makes up well over 75% of the biota. *Redlichia takooensis* is relatively common, but other trilobites are rare. Glaessner (1979) and García-Bellido et al. (2009) described species of the bivalved arthropods *Isoxys* and *Tuzoia*. Paterson et al. (2010), Edgecombe et al. (2011) and Paterson et al. (2012) described new taxa of nektaspid, megacheiran and other artiopodan arthropods, respectively. Investigation of other arthropods is continuing. Glaessner (1979) described the palaeoscolecid *Palaeoscolex antiquus* plus the enigmatic forms *Vetustovermis planus* and *Myoscolex ateles*. McHenry and Yates (1993) reported the Cambrian predator *Anomalocaris*. *Anomalocaris briggsi* was erected by Nedin (1995a) who left a second species under open nomenclature. Other fossils include sponges, chancelloriids, hyoliths, brachiopods, a vetulicolian, an *Odontogriphus*-like organism plus several other enigmatic forms. The oldest, well preserved, complex, arthropod eyes, including those of *Anomalocaris*, have been reported from this biota (Lee et al., 2011; Paterson et al., 2011).

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*by Christine J. Edgoose*

# The Amadeus Basin, central Australia

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*The Amadeus Basin of central Australia has a depositional history spanning the Neoproterozoic to the Devonian/Carboniferous. It was initiated as part of the Neoproterozoic Centralian Superbasin, which formed in an intracratonic setting related to the break up of Rodinia. Sedimentation continued until the 580–540 Ma Petermann Orogeny, coinciding with the assembly of Gondwana, which resulted in the fragmentation of the superbasin into separate intracratonic basins. The Petermann Orogeny was focused in the Musgrave Province and the southern part of the Amadeus Basin, and involved significant N-directed shortening on large-scale structures that involved both the basement and the overlying Neoproterozoic sedimentary rocks. It significantly transformed the basin architecture, with the development of major basin features that controlled subsequent sedimentation. Deposition of Paleozoic successions was largely concentrated in sub-basins and troughs in the N of the basin, where up to 14 km is preserved. Minor events or uplifts punctuated this depositional history and account for local disconformities and absent sections. The 450–300 Ma Alice Springs Orogeny was a multi-phase, intracratonic event concentrated in the Arunta Region and the northern part of the Amadeus Basin. Like the earlier Petermann Orogeny, the Alice Springs Orogeny involved both basement and basin sedimentary rocks, but with overall S-directed movement. Synorogenic sedimentation accompanied Mid–Late Devonian uplift, with Late Devonian–Carboniferous basin inversion terminating sedimentation, and folding the youngest successions.*

*The Amadeus Basin has known reserves of U, minor historic and recent Au production, and is prospective for base metals, especially Cu, and phosphate. The Ordovician succession supports commercial gas production, and the Neoproterozoic succession is considered prospective for oil and gas.*

## Introduction

The Amadeus Basin is a large (c. 170,000 km$^2$) elongate intracratonic Neoproterozoic–Devonian basin that extends c. 800 km E-W and a maximum of 300 km N-S in central Australia (Figure 1). It overlies Paleo–Mesoproterozoic basement of the Musgrave Province to the S and Arunta Region to the N, and is overlain by the late Paleozoic Pedirka Basin and Mesozoic Eromanga Basin in the SE, and by the Paleozoic stratigraphy of the Canning Basin to the W. The present-day Amadeus Basin is a structural remnant of a broad, shallow basin – it has been significantly tectonically modified during two major intracratonic orogenic events: the 580–540 Ma Petermann Orogeny and the 450–300 Ma Alice Springs Orogeny.

The early history of the Amadeus Basin is as part of the Neoproterozoic Centralian Superbasin (Walter et al., 1995), which also encompassed the Officer, Ngalia, Georgina, and Murraba basins (Shaw et al., 1991; Walter et al., 1995), and probably several smaller basins in Western Australia (Figure 2). Formation of the Centralian Superbasin coincided with NE-SW-directed intracratonic extension across the Rodinia Supercontinent, which eventually led to the break-up between North America and Australia at c. 830 Ma (deVries et al., 2000). Walter and Veevers (2000) described this phase as *Centralian 1*. After the break up of Rodinia, sedimentation continued locally in the Centralian Superbasin until c. 750 Ma. In the *Centralian 2* phase, renewed but localised sedimentation related to the 700–690 Ma Sturtian glaciation took place. *Centralian 3* is associated with the younger Elatina (Marinoan) glaciation. Sedimentation in the Centralian Superbasin was terminated by the 580–540 Ma Petermann Orogeny, which coincided with the final stages of the assembly of Gondwana (de Vries et al., 2000), but continued locally in the now separated basins, for example in the northern part of the Amadeus Basin.

Korsch and Lindsay (1989) and Lindsay and Korsch (1991) recognised three main stages of basin evolution. Stage 1 was a long-lived extensional–thermal relaxation (sag) event (c. 900–590 Ma). In the SW of the basin, Stage 1 sediments are underlain by a rift sequence (bimodal volcanics and associated sedimentary rocks (e.g., Close et al., 2004; Edgoose et al., 2004). The relationship of this rift succession to the initiation and early evolution of the Amadeus Basin is equivocal, as the rift sequence is c. 150 Ma older than the Stage 1 sediments (sag phase). Such a time gap suggests the rift and sag may not be related (Shaw, 1991). However, it is also possible that a significant time break occurs at a disconformity within the sag phase sediments, and that the deposition of sag phase sediments commenced soon after rifting ceased (A. Camacho, per. comm., 2011). Stage 2 (c. 580–450 Ma) comprised an early compressive event (Petermann Orogeny) then extension followed by a thermal relaxation event and resulted in shallow-marine to terrestrial deposition across the basin. Stage 3 (c. 450–300 Ma) was the final phase of basin evolution and was a dominantly compressional phase (Alice Springs Orogeny). Events relating to this stage account for most of the obvious structures (folds, domes and thrusts) evident in the present-day surface geology (Figure 3).

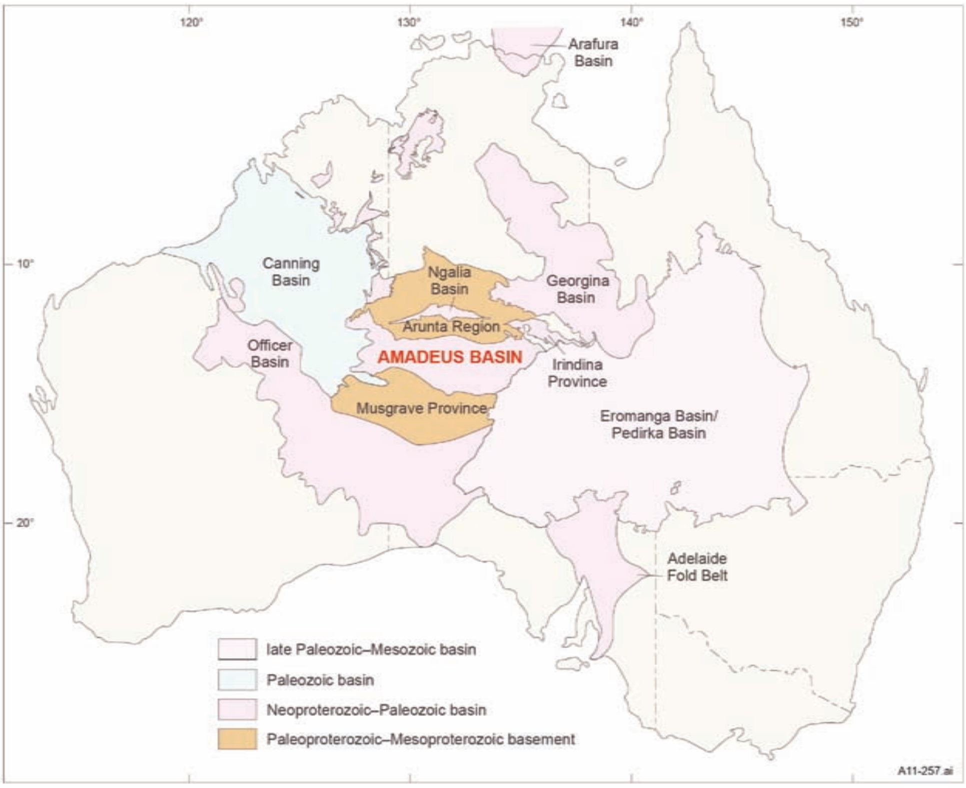


***Figure 1 Location of Amadeus Basin and surrounding tectonic regions.***

Shaw (1991) subdivided the tectonic development of the basin in more detail, and described at least nine tectono-stratigraphic intervals of basin history, separated by regional unconformities, which strongly influenced basin shape and evolution. This tectono-stratigraphic record is interpreted to have largely resulted from far-field forces at plate margins well outside the region. Basement structures also appear to have placed long-term and repeated controls on subsidence patterns. Reactivation of pre-basin Paleoproterozoic and Mesoproterozoic fault zones beneath the Amadeus Basin, during both the Petermann and Alice Springs orogenies, has been a first-order control on basin architecture and the location of basement highs and depocentres (Shaw, 1991; Munroe et al., 2004).

During the early–mid Paleozoic (Stage 2 of Lindsay and Korsch, 1991), Australia was located at low paleolatitudes, and at times of high sea level, transcontinental seaways may have existed (Walley et al., 1991; Lindsay, 1993). During the Cambrian, the seaway axis lay to N of the Amadeus Basin (across the Georgina,Wiso and Daly basins). Most paleogeographic reconstructions of Ordovician Australia (e.g., Nicoll et al., 1988; Walley et al., 1991) show an open E-W seaway (Larapintine Seaway) across the continent. The Ordovician was a period of widespread deposition across Australia, with deeper water continental margin systems established in eastern Australia, and shallow-marine to paralic conditions prevailing in several inland and northern basins. Thick successions accumulated in the Amadeus and Canning basins, and most reconstructions show the two basins linked during this time, effectively operating as a large, single depositional system. However, differences in stratigraphy, depositional style, hydrocarbon systems, and a fairly high degree of faunal endemism have led some workers to doubt this connection (e.g., Veevers, 1976; Haines and Wingate, 2007), and more recent detrital zircon data points to different source areas for the two basins (Haines and Wingate, 2007). The Amadeus Basin shows evidence of having been open to the E and sourced material from eastern Australia. Marine sedimentation ceased at the end of the Ordovician in response to broad, regional uplift. From this time to the Early Devonian, only limited deposition occurred in the basin, in aeolian and lesser fluviatile systems. The Silurian was largely a period of hinterland erosion, followed by fluvio-lacustrine to paralic conditions in parts of the basin in the late Early Devonian. The middle–late Devonian is characterised by an upward-coarsening fluvial system, deposited in response to uplift related to the Alice Springs Orogeny, and sedimentation culminated in the Late Devonian/early Carboniferous as a result of final orogenic processes and basin inversion. Permian glaciation in the area suggests that highlands persisted until this time, and that the area was at relatively high paleolatitudes.

The units of the Amadeus Basin are discussed in more detail below (see also Figure 4).

# Neoproterozoic

## Cryogenian: Supersequence 1

A sequence stratigraphic analysis has identified 4 supersequences in the Neoproterozoic stratigraphy of the Amadeus Basin (Walter et al., 1995). Supersequence 1 forms part of the initial sag phase, equivalent to Stage 1 basin evolution of Lindsay and Korsch (1991) and Interval 1 of Shaw (1991). The sequence comprises a relatively thin, basal clean sand sheet deposited in intertidal and fluviatile environments (Heavitree and Dean quartzites), followed by a thicker succession of marine carbonate rocks, fine siliciclastic sedimentary rocks and evaporites (Bitter Springs Formation and Pinyinna beds).

The characteristics of the Heavitree and Dean quartzites imply an abundant supply of sediment deposited in a high-energy shelf-like environment. Sedimentological and stratigraphical analyses by Lindsay (1993, 1999) show that quartz sandstone sedimentation occurred in a shallow, low-gradient ramp setting. Detrital zircon data for the Heavitree Quartzite suggests a maximum deposition age of 1050–1000 Ma (Zhao et al., 1992), with detrital zircon age data indicating a source dominated by the Arunta Region. Maidment et al., (2007) provided the youngest zircon age at 1.12 Ga, probably sourced from the Musgrave Province.

The Bitter Springs Formation comprises three members consisting of crystalline dolomitic limestone, dolostone and limestone, dololutite, dolarenite, calcarenite, siltstone, gypsiferous siltstone, sandstone, halite and evaporites, with a thin, and probably discontinuous, spilite

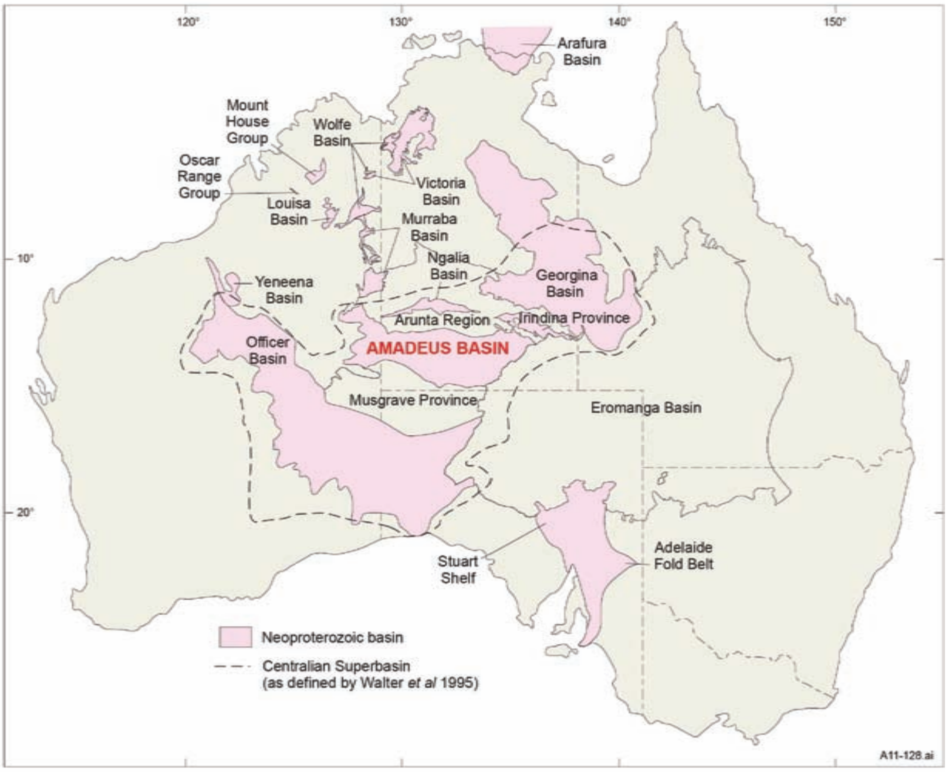


***Figure 2 Centralian Superbasin (after Walter et al., 1995).***

unit within the succession. The Pinyinna beds are the metamorphosed equivalent of the Bitter Springs Formation in the SW of the Basin. The age of the Bitter Springs Formation is interpreted to be c. 820 Ma, from the spilite layers, which have geochemical affinities with dykes dated at c. 800 Ma using Sm-Nd mineral isochrons (Zhao et al., 1994) and at 824 ± 4 Ma by U-Pb dating of baddelyite (Glikson et al., 1996). The upper part of the Bitter Springs Formation contains the Cryogenian stromatolite assemblage *Acaciella australica*, which is older than 802 ± 10 Ma and younger than c. 825 Ma in the Officer Basin in WA (Grey, 2005).

The Areyonga Movement is expressed as an unconformity between the Bitter Springs Formation and the overlying Areyonga Formation in the NE of the basin. However, it is possible that this unconformity is related to halotectonic processes rather than to a discrete tectonic event (Kennedy, 1993).

## Mid–late Cryogenian: Supersequence 2

Supersequence 2 equates with the 700–690 Ma Sturtian glaciation, and comprises the Areyonga and Aralka formations and the equivalent unit in the southern part of the basin, the Inindia beds, and in the western part of the basin, the Boord Formation. The Areyonga Formation is formed predominantly of diamictite (tillite) of variable composition and texture, but includes thin interbeds of sandstone, conglomerate, shale and siltstone, and dolostone. This glaciogenic succession shows marked lithological variation from massive, indurated diamictite/conglomerate to carbonaceous siltstone/shale, feldspathic sandstone, and rarer dolostone in the middle and at the top of the succession. The Aralka Formation is largely siltstone and shale, but includes dolostone and calcarenite that is in part pisolitic and stromatolitic, and a dominantly clastic unit, characterised by pebbly and sandy calcarenite and festoon-bedded sandstone. The dolostone contains the columnar stromatolite *Tungussia inna* Walter (Walter et al., 1995). The Inindia beds comprise a generally thick but variable unit of sandstone with lesser siltstone, chert and jasper, tillite and dolostone that is relatively poorly exposed in the SW and south-central parts of the basin. The Boord Formation contains two glacial units, several disconformities, and significant pre- and post-glacial intervals (Haines et al., 2010).

## Late Cryogenian–early/mid Ediacaran: Supersequence 3

Supersequence 3 is associated with the c. 650–635 Ma Elatina glaciation. It comprises the Olympic Formation, Pioneer and Gaylad sandstones, the Pertatataka Formation and Julie Formation in the central-northern part of the basin. The newly proposed Neoproterozoic stratigraphic correlations of Haines et al., (2010) would suggest that the Inindia beds in the southern part of the basin correlate with both Supersequence 2 and 3, and the overlying Winnall beds, which were formerly considered as Supersequence 3, are now considered to be younger.

The Olympic Formation consists of lenticular units of sandstone, siltstone conglomerate, diamictite, shale and an upper 'cap' dolostone, and is up to 190 m thick. The Pioneer Sandstone (Priess et al., 1978) is a shallow-marine to tidal unit, confined to the central-northern part of the Amadeus Basin. It consists of cross-bedded, medium- to coarse-grained feldspathic and arkosic sandstone, grading up into pink-grey dolostone with red chert nodules. The Pioneer Sandstone is interpreted to be an intertidal periglacial, or glacial outwash facies that correlates with diamictite of the Olympic Formation (Preiss et al., 1978; Walter et al., 1995), although this correlation has been questioned (Lindsay, 1989; Field, 1991). The type section for the Gaylad Sandstone consists of quartz sandstone, lithic sandstone, subarkose, orthoconglomerate, and rare debris-flow diamictite. At its type section, the Pertatataka Formation is c. 350 m thick, and is made up of 320 m of red and green siltstone and shale, with flaser bedding and small-scale cross laminations, overlain by 1.5 m of grey, medium-grained feldspathic silicified sandstone, and 30 m of red siltstone with sandstone laminae (Prichard and Quinlan, 1962). U-Pb SHRIMP dating of detrital zircons by Maidment et al. (2007) gave a youngest concordant age of c. 807 Ma, significantly older than the inferred depositional age of c. 575 Ma.

## Late Neoproterozoic–Cambrian: Supersequence 4

The Petermann Orogeny (580–540 Ma) was a crustal-scale, bivergent intracratonic event localised in the Musgrave Province and in overlying Neoproterozoic stratigraphy of the Amadeus and Officer basins (Scrimgeour and Close, 1999; Close et al., 2004; Edgoose et

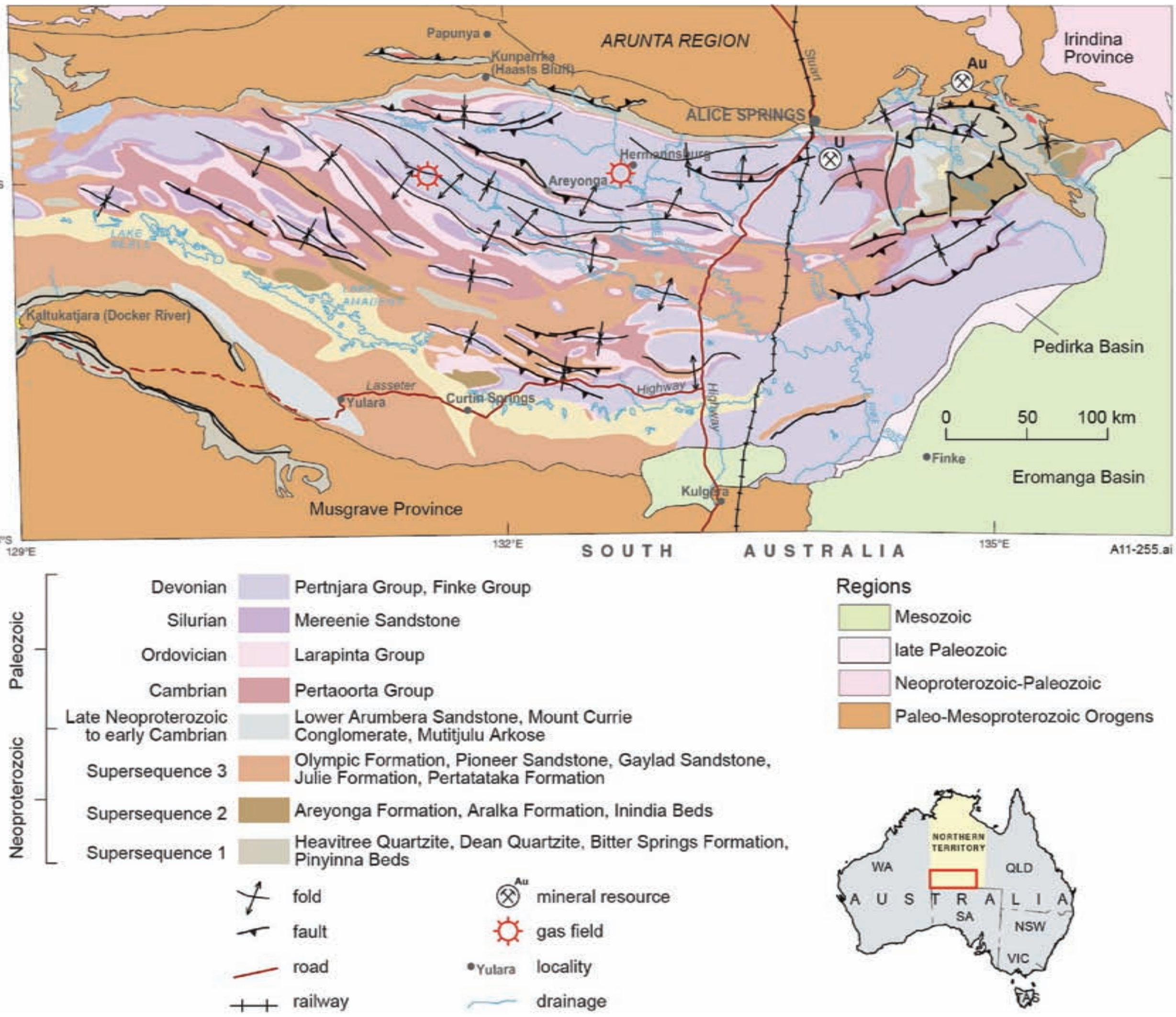


***Figure 3 Generalised geology and major structures of the Amadeus Basin, Northern Territory.***

al., 2004). In the Amadeus Basin, the Petermann Orogeny is traditionally considered to have resulted in only limited foreland basin development, despite N-vergent shortening of c. 100 km and thickening that buried the basal basin succession to depths of 20 km (Flottmann et al., 2004). However, revised stratigraphic correlations now suggest that up to 6 km of sediments in the W and SW of the basin are synchronous with the Petermann Orogeny (Haines et al., 2010).

Within the basal Amadeus Basin succession, progressive deformation during the Petermann Orogeny is recorded in at least three phases of folding, with the earlier phases involving both basement and sedimentary rocks (Scrimgeour et al., 1999; Edgoose et al., 2004). Large detachment zones exhumed interleaved basement rocks and sedimentary rocks of Supersequence 1 from mid crustal levels. North-vergent thrusts repeat the basal Amadeus succession N of a large-scale basement wedge, with significant back thrusting of younger (Supersequence 2 and 3) Neoproterozoic sedimentary rocks to the S accommodating the shortening.

The Amadeus Basin suffered major disruption during the Petermann Orogeny, resulting in fragmentation of the basin architecture, with the development of large sub-basins and troughs N of a central ridge, and a platform in the S (Figure 5). Depositional loci moved northwards and sedimentation was concentrated in these major depocentres. A total of 2,800 m of clastics were shed into the Carmichael Sub-basin and 1,500 m into the Missionary Plains Trough (Lindsay, 1993; Ambrose, 2006). Paleozoic sedimentation was apparently restricted, or for considerable time periods absent, in the southern part of the basin.

The late Neoproterozoic–Cambrian succession in the northern part of the Basin (Pertaoorta Group) is stratigraphically variable from E–W, and comprises many named units that represent facies changes across the geographic area of distribution. The Group is dominated by carbonate rocks in the E, and by siliciclastic successions in the W. The oldest formation, the Arumbera Sandstone, is divided into two depositional successions of upward-coarsening siltstone and sandstone (Lindsay, 1987; Kennard and Lindsay, (1991) – the lower is Neoproterozoic whereas the overlying formations are all early Cambrian in age. The upper Arumbera Sandstone has yielded more than thirty different ichnospecies (Kennard in Kennard and Nicoll, 1986; Shergold, 1986; Shergold et al., 1991). The lower Arumbera Sandstone has yielded specimens of the soft-bodied Ediacarian fauna (metazoan body fossils).

On the southern platform, the newly proposed correlations of Haines et al. (2010) indicate that the Carnegie Formation, Sir Frederick Conglomerate, Ellis Sandstone and Maurice Formation of the far W and SW, and the Winnall beds of the SW and south-central parts of

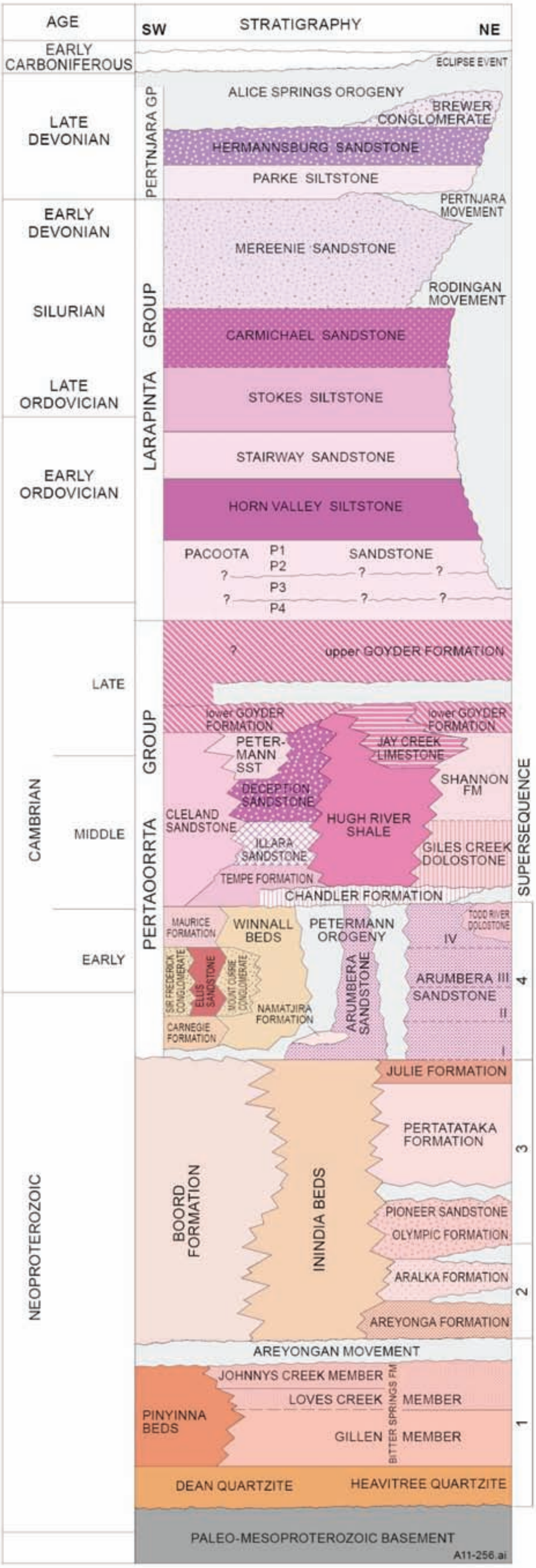


***Figure 4 Stratigraphy of the Amadeus Basin (modified from Marshall et al., 2007; Haines et al., 2010). The supersequence scheme of Walter et al., (1995) is shown on the right.***

the basin are correlatives of the Arumbera Sandstone, rather than correlatives of the late Neoproterozoic Supersequence 3. Coarse clastic sediments of the Mount Currie Conglomerate and Mutitjulu Arkose are correlated with the Sir Frederick Conglomerate and Ellis Sandstone, and were deposited in localised foreland basins in the hinterland of the emergent basal Amadeus succession and Musgrave Province basement during the Petermann Orogeny. These deposits reflect proximal and more distal high-energy fluvial environments of broad shallow channels or sheet floods. In the south-central part of the basin, minor locally developed successions that are interpreted to be laterally equivalent to the Mount Currie Conglomerate represent restricted molasse deposits related to paleohighs created during the Petermann Orogeny.

# Ordovician

The potential for oil and gas in the Ordovician succession was recognised early on, and consequently, it has been the focus of more study than most of the other successions in the basin.

The Larapinta Group comprises 5 constituent formations. The Pacoota Sandstone comprises four units of shallow-marine sandstone (Williams et al., 1965). The Horn Valley Siltstone comprises thinly bedded organic rich shale and siltstone, with bedded nodular limestone abundant in some intervals (Shergold, 1986). It has an abundant and well preserved faunal and floral assemblage, which includes trilobites, brachiopods, pelecypods, nautiloids, ostracods, conodonts, chitinozoa, acritarchs, graptolites and gastropods, and simple spores and algae (Shergold, 1986). Conodonts and trilobites indicate a late Early Ordovician (Floian) age (Cooper, 1981; Nicoll and Jones in Kennard and Nicoll, 1986). The Stairway Sandstone has an abundant fossil assemblage which indicates a strong marine influence in shallow subtidal and partly intertidal conditions (Cook, 1972), and includes local phosphorites The assemblage comprises trilobites, brachiopods, sponges, rostroconchs, gastropods, cephalopods, pelcypods, monoplacophorans, vertebrates, chitinozoans and acritarchs, which give an Early Ordovician (Dapingian or Dariwilian) age. U-Pb zircon data (Maidment et al., 2007) gives maximum deposition ages in the range 650–500 Ma. The Stokes Siltstone comprises a lower siltstone and limestone lithofacies, and an upper red and purple sandstone and shale lithofacies (Shergold, 1986). The formation contains relatively few fossils (Nicoll in Kennard and Nicoll, 1986), but includes gastropods, a brachiopod, a bryozoan, rostroconch molluscs and bivalve molluscs, nautiloid cephalopods, trilobites and conodonts (Shergold, 1986). The fauna indicate an Early Ordovician (Dapingian or Dariwilian) age. The Carmichael Sandstone comprises variable interbedded pale brown to red-brown sandstone, siltstone and mudstone, indicating a mixed depositional environment, comprising shallow-marine to possibly hypersaline and fluvial conditions (Owen in Kennard and Nicoll, 1986). The only recorded fossils are rare ichnofossils, including *Cruziana*.

The Rodingan Movement (Cook in Wells et al., 1970; Shaw, 1991; Bradshaw and Evans, 1988) presents as an unconformity between the Carmichael Sandstone and Mereenie Sandstone in the N and NE of the basin. It is interpreted to be a low-angle feature that cuts down to the NE, and probably coincided with the end of marine sedimentation in the basin.

The c. 490–480 Ma Larapinta Event was a high-grade tectono-metamorphic event that affected Amadeus Basin sedimentary

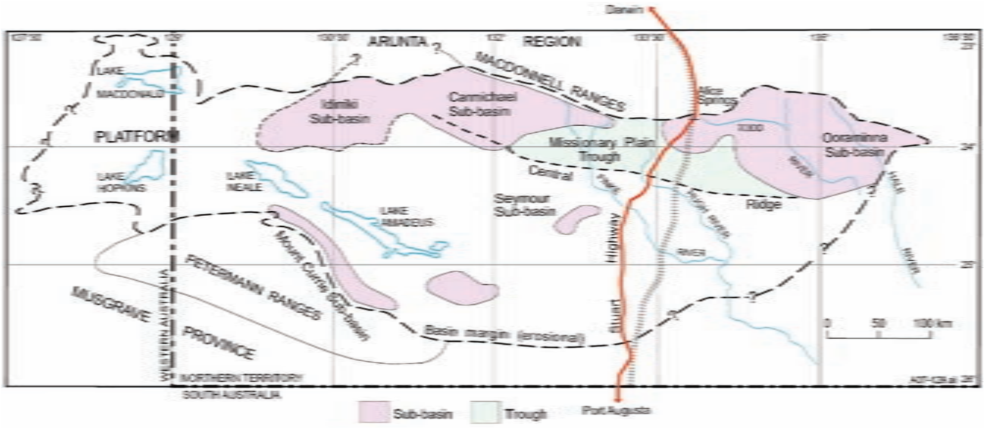


*Figure 5 Architecture of the Amadeus Basin (from Marshall et al., 2007, after Lindsay and Korsch, 1991).*

equivalents deposited in a deep trough (Irindina Province) to the NE of the basin (Mawby et al., 1998, 1999; Maidment et al., 2007). Rapid, deep burial and exhumation in an extensional setting affected this trough during the Larapinta Event, resulting in the exhumation of granulite-facies metasedimentary rocks. The Irindina Province is included in the Arunta Region as a result of its uniquely complex, high-grade tectonothermal history.

## Late Ordovician–Silurian

The Mereenie Sandstone is one of the most widespread units of the basin (Owen in Kennard and Nicoll, 1986). Most of the lower part was deposited in a shallow-marine environment, which transitions to aeolian–fluvial environments for the upper part. Paleocurrent data suggest an E–SE-trending shoreline for the lower marine section, and sediment transport towards the E and S for the upper section (Owen in Kennard and Nicoll, 1986). U-Pb zircon data (Maidment et al., 2007) are dominated by ages in the range 630–490 Ma. There are few older grains, suggesting the Mereenie Sandstone was largely derived from the reworking of underlying units.

The Pertnjara Movement is expressed as an unconformity between the Mereenie Sandstone and Pertnjara Group, in the central-north and NE of the basin. It initiated deposition of the Pertnjara Group, and may have produced gentle folding or thrusting in the NE.

## Devonian–Carboniferous

The Pertnjara Group consists of a series of lacustrine, braided and meandering fluviatile, and alluvial fan deposits, which accumulated on a S-thinning depositional wedge. Variations in the distribution, maturity and facies in the Pertnjara Group reflect successive tectonic pulses of the Alice Springs Orogeny, which were coincident with its deposition (Jones, 1991). Devonian fish fossils have been found and described from several localities (Gilbert-Tomlinson, 1968; Young, 1974, 1985, 1988, 2005), with the fauna giving a Givetian–Frasnian age (Young, 1985). The Finke Group is a correlative of the Pertnjara Group in the southern part of the basin. Sediment source is dominantly from the Musgrave Province to the S, with a later switch to the progressively emergent Arunta Region to the N (Jones, 1973).

The Alice Springs Orogeny (450–300 Ma) was a long-lived, multi-phased, bivergent intracratonic event with its strongest effects in the northern part of the Amadeus Basin and the Arunta Region basement. Some of its earlier phases are recognised as discrete events (e.g., Pertnjara Movement, see above). It resulted in basin inversion and the cessation of Paleozoic deposition in the Amadeus Basin.

The Alice Springs Orogeny involved considerable crustal shortening (50 km – Flottmann et al., 2004; 60–125 km – Haines et al., 2001). The former authors describe two phases of shortening: S-directed overthrusting of a basement wedge; and underthrusting that produced a large-displacement, N-vergent, passive back thrust in the N of the basin. Deformation related to the Alice Springs Orogeny accounts for most of the obvious structures (folds, domes and thrusts) in the present-day surface geology in the Amadeus Basin (Figure 3).

Haines et al. (2001) took a synorogenic sedimentological approach to assessing the distribution and timing of events associated with the Alice Springs Orogeny. The orogeny had a prolonged duration of c. 150 million years (c. 450–300 Ma) with peaks of activity at 450–440 (Rodingan Movement), 390–375 and 340–320 (Eclipse Event) Ma. These episodes were also described by Bradshaw and Evans (1988), who also recognised spatial variation and divided the basin into several provinces with varied structural responses to the orogen as it progressed. The earliest phase of the Alice Springs Orogeny at c. 450–440 Ma is also recorded in the Irindina Province and adjacent Arunta Region basement as extension on major shear zones. In the NE of the basin, the orogeny formed large, S-directed, basement-cored nappes, which are characteristic of Alpine mountain belts and are generally not typical of intracratonic settings (Dunlap and Tessyier, 1995).

Isotopic and geological data indicate re-activation of major Proterozoic shear zones during the Alice Springs Orogeny (e.g., Shaw, 1991; Dunlap and Tessyier, 1995), for example the crustal scale Redbank Thrust Zone on the northern margin. Sedimentological and fission track dating data indicates the removal of a substantial thickness of younger strata from the northern part of the basin late in the Alice Springs Orogeny.

## Halotectonics

The influence of halotectonics on the distribution of sedimentary rocks, structural styles and deformation has been recognised within the Neoproterozoic and early Paleozoic successions in various parts of the Amadeus Basin, but is best shown in the NE. It is generally associated with halite in the Neoproterozoic Bitter Springs Formation and to a lesser extent in the Cambrian Chandler Limestone. Kennedy (1993) reported the effects of diapiric growth on local facies and structure in late Neoproterozoic successions, where salt withdrawal led to normal faulting and syn-sedimentary thickening of adjacent units, as well as local unconformities on units flanking the diapiric structures. Marshall and Dyson (2007) interpreted Neoproterozoic and early Paleozoic sedimentation in the Amadeus Basin to have occurred dominantly in a series of salt nappe complexes and mini-basins, which were formed under the influence of gravity gliding, gravity spreading and salt withdrawal. They attribute many of the fold-thrust nappes previously ascribed to the Alice Springs Orogeny as having most likely formed during salt movement associated with Neoproterozoic sedimentation.

## Petroleum and Mineral Resources

The Amadeus Basin has significant potential for further oil and gas discoveries. For a known producing basin, it is significantly underexplored. Up until 2006, the basin had 33 wells (one well per 5,000 km$^2$) and 7,500 km of seismic data. Renewed exploration for petroleum has occurred in the basin more recently, with seismic acquisition and drilling taking place.

Five petroleum systems are described by Marshall (2004) and Marshall et al. (2007). These are 1 – Heavitree Quartzite/Gillen Member of the Bitter Springs Formation, 2 – Loves Creek Member of the Bitter Springs Formation–Pioneer Sandstone, 3 – Pertatataka Formation–Julie Formation, 4 – Arumbera Sandstone–Chandler Formation, and 5 – Larapinta Group. Currently, the only the latter, encompassing the Mereenie, Palm Valley and West Walker hydrocarbon occurrences (Figure 3), is a commercially productive petroleum system (Marshall, 2004).

The Amadeus Basin has undergone comparatively little exploration for mineral commodities. Mineral production in the Amadeus Basin is confined to minor historical and recent Au production from the Arltunga and Winnecke goldfields (Heavitree Quartzite) in the NE of the basin (Figure 3), and historic working of minor surficial Cu at a few locations in early Paleozoic stratigraphy in the north-central part. The Basin has known reserves of U in Devonian stratigraphy S of Alice Springs (Figure 3; Lally and Bajwah, 2006), and recorded occurrences of base metals, manganese and phosphate (Edgoose, 2012).

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# A review of the geology and uranium, gold and iron ore deposits of the Pine Creek Orogen

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*The Pine Creek Orogen comprises a succession of Paleoproterozoic sedimentary and volcanic rocks, unconformably overlying Neoarchean granitic basement and intruded by Paleoproterozoic mafic rocks and granites. The orogen is subdivided from west to east into the Litchfield Province, Central Domain and Nimbuwah Domain, based on the distinct timing and nature of sedimentation, magmatism and metamorphism. The orogen hosts a wide range of commodities, the most important of which are U and Au. Rifting of Neoarchean basement at 2020 Ma led to deposition of clastic, carbonate, and carbonaceous sedimentary and volcanic rocks in a shallow basin. At 1870 Ma, sedimentation in the Nimbuwah Domain was rapidly followed by burial, I-type granitic magmatism (1867–1860 Ma), compressional tectonism and mid-pressure amphibolite-facies metamorphism (1865–1855 Ma). Major U deposits occur in the Nimbuwah Domain within basal Paleoproterozoic strata, close to tectonised contacts with Neoarchean basement. Metamorphism of the Nimbuwah Domain coincided with sedimentation and volcanism in the Central Domain and Litchfield Province at 1863 Ma. This was followed by extensional high-temperature, low-pressure metamorphism (1855 Ma) and associated felsic and arc-related mafic magmatism (1862–1850 Ma) in the Litchfield Province. At or after this time, greenschist-facies metamorphism and upright folding and shearing occurred at upper crustal levels in the Central Domain, generating structural traps for subsequent Au- and Fe-bearing fluids. Almost all Au occurrences are associated with late to post orogenic, I-type Cullen Supersuite granites (1835–1820 Ma). Shortly thereafter, platform sediments were deposited in braided rivers across the orogen. The strong spatial heterogeneity in the distribution of U and Au suggests that the pre-existing crustal architecture of the orogen was a significant factor controlling their distribution.*

## Introduction

The Pine Creek Orogen is exposed over 47,500 km$^2$ on the northern margin of the North Australian Craton. The orogen comprises Neoarchean (2670–2500 Ma) granitic and gneissic basement that is unconformably overlain by a >4 km thick succession of Paleoproterozoic clastic, carbonate and carbonaceous sedimentary and volcanic rocks. The orogen hosts over 1000 mineral occurrences, with major commodities including Au, U, Pb-Zn-Ag, platinum-group elements, Cu-Co-Ni, Fe ore, Sn-Ta-W and phosphate.

The Pine Creek Orogen is subdivided, from west to east, into the low-pressure amphibolite- to granulite-facies Litchfield Province, greenschist-facies Central Domain, and mid-pressure amphibolite-facies Nimbuwah Domain (Figure 1). Historically, Paleoproterozoic tectonometamorphism in the Pine Creek Orogen was attributed to the 1885–1850 Ma 'Barramundi Orogeny' (Etheridge et al., 1987; Needham et al., 1988; Page and Williams, 1988), but this term has been abandoned because recent studies showed that different parts of the orogen were affected by chronologically distinct events of contrasting character (Carson et al., 2008; Hollis et al., 2009b, 2011; Glass, 2010). Their recognition has had a major influence on understanding the geological context of important mineralising systems.

We present a review of the geological evolution of the Pine Creek Orogen to provide a revised context for understanding the mineralising systems. We discuss three particular commodities – U, Au and Fe ore – focusing on deposits visited during a field trip associated with the 34th International Geological Congress.

## Geology of the Pine Creek Orogen

### Neoarchean

Most of the exposed Archean basement to the North Australian Craton occurs within the Pine Creek Orogen. Neoarchean rocks outcrop in the Central and Nimbuwah domains, and probably underlie much of the Pine Creek Orogen. They comprise c. 2670 Ma, 2640 Ma and 2545–2510 Ma granite and gneiss (Williams and Compston, 1983; Cross et al., 2005; Hollis et al., 2009a). No Archean basement has been identified in the Litchfield Province.

### Paleoproterozoic

#### *c. 2020 Ma sedimentation, magmatism and tectonism*

Rifting of Neoarchean basement at c. 2020 Ma resulted in the

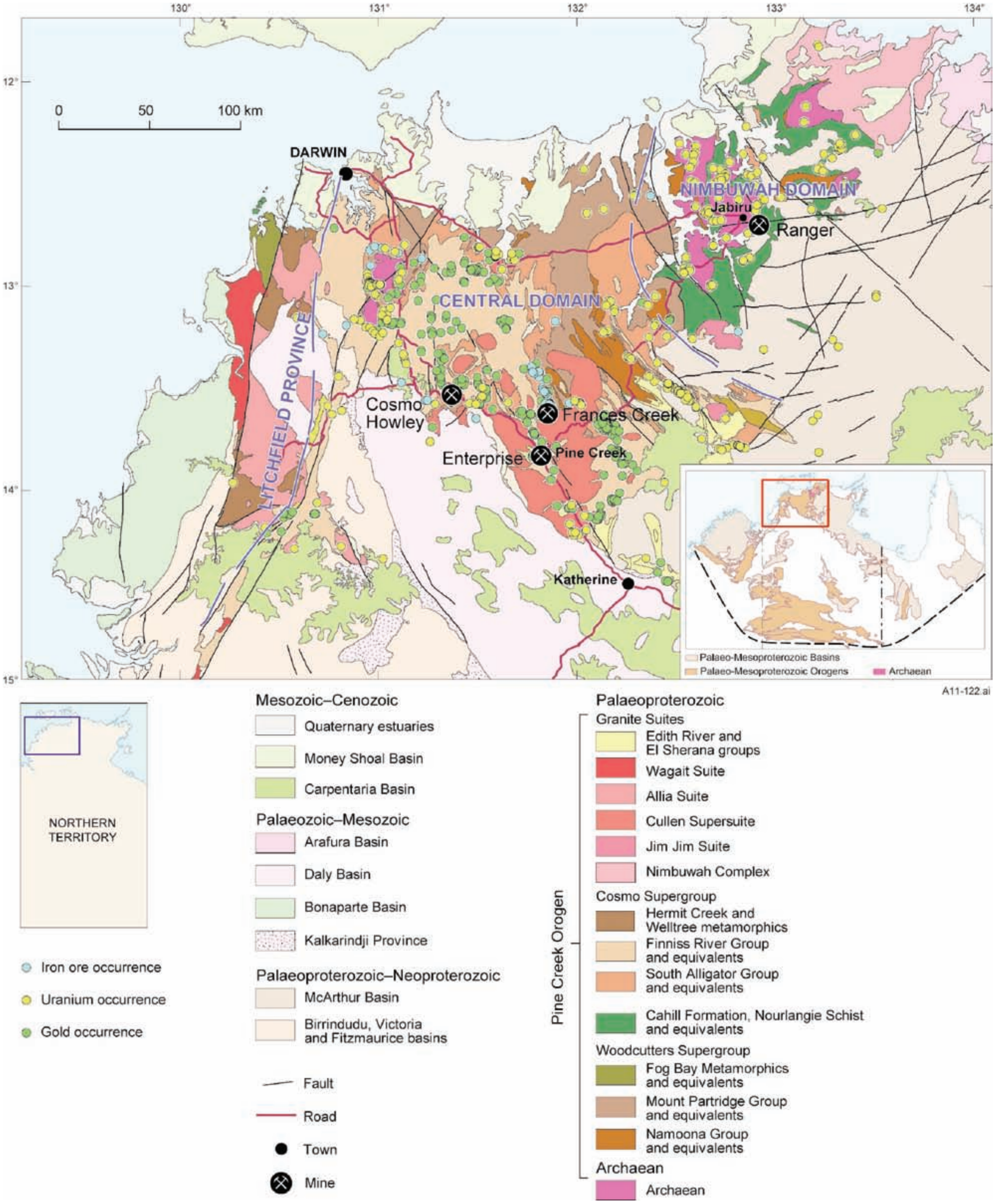


***Figure 1 Generalised geology of the Pine Creek Orogen showing the three tectonostratigraphic domains and some of the major mineral deposits. Inset shows location of the Pine Creek Orogen within the North Australia Craton.***

deposition of clastic, carbonate and carbonaceous sediments and volcanics of the Woodcutters Supergroup across the Central and Nimbuwah domains. These strata comprise conglomerate, sandstone and siltstone (Crater and Beestons formations), stromatolitic dolostone and magnesite (Celia Dolostone), black pyritic and dolomitic shale, tuff, slate, metagreywacke and dolarenite (Masson Formation), and basaltic to andesitic lava and agglomerate (Stag Creek Volcanics; Figure 2; Worden et al., 2008a, b). The presence of reduced pelitic and dolomitic rocks, rare pillow structures and interbedded tuffaceous shale in volcanic units indicate a shallow marine depositional environment. Quartzite and meta-arkose of the Kakadu Group in the Nimbuwah Domain are probable

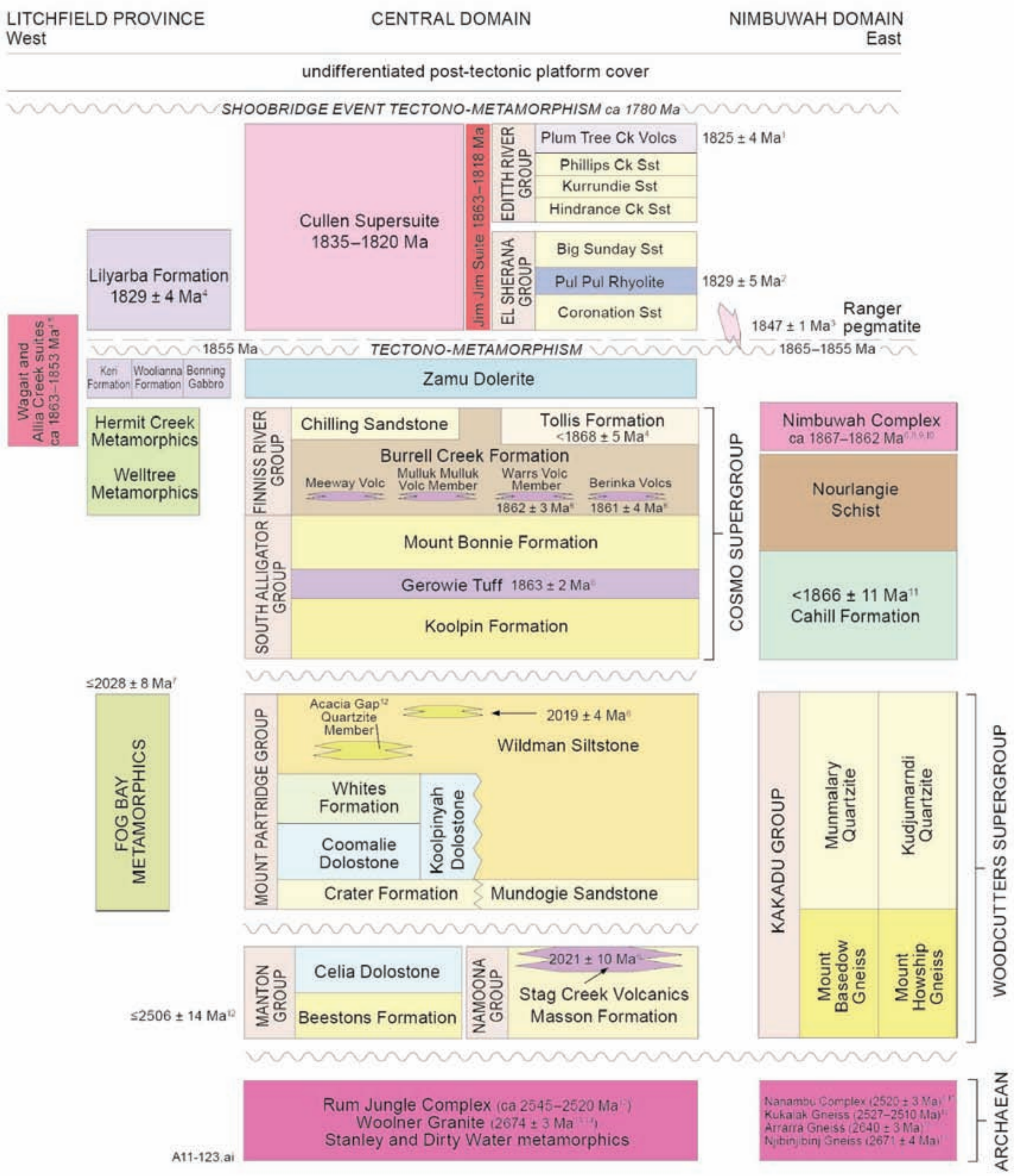


***Figure 2 Simplified stratigraphic column of Pine Creek Orogen modified from (Worden et al., 2008b). [1]Page (1996a), [2]Jagodzinski (1998), [3]Annesley et al. (2002), [4]Worden et al. (2008a), [5]Page et al. (1985), [6]Worden et al. (2008b), [7]Carson et al. (2009), [8]Carson et al. (2010), [9]Hollis et al. (2009b), [10]Page et al. (1980), [11]Hollis et al. (2011), [12]Cross et al. (2005), [13]Glass et al. (2010), [14]McAndrew et al. (1985), [15]Hollis et al. (2009a); Ck = Creek; Sst = Sandstone; Volcs = Volcanics.***

higher metamorphic grade equivalents (Figure 2; Worden et al., 2008a).

## *c. 1870 Ma sedimentation, magmatism and tectonism*

Sedimentation at c. 1870 Ma in an Eastern Trough (Nimbuwah Domain) occurred shortly prior to sedimentation at 1863 Ma in a Central Trough (Litchfield Province/Central Domain); the detritus in the two successions is of distinct ages, indicating different source regions (Hollis et al., 2009b, 2011).

In the Nimbuwah Domain, the Cahill Formation comprises schist (some of it carbonaceous), calc-silicate rock, para-amphibolite, and quartzite, grading upwards into quartzofeldspathic mica schist of the Nourlangie Schist (Figure 2; Needham, 1988). Protoliths to these rocks were metamorphosed at amphibolite facies in the mid crust and cut by W- to NW-vergent thrusts and isoclinal folds at c. 1865–1855 Ma. This occurred during or immediately after the main phase of emplacement of I-type monzogranitic to quartz monzodioritic plutons of the Nimbuwah Complex at 1867–1860 Ma (Worden et al., 2008b; Carson et al., 2010).

In the Central Domain sedimentary and volcanic rocks of the 1863 Ma Cosmo Supergroup were unconformably deposited on the Woodcutters Supergroup coeval with mid-crustal tectonism in the Nimbuwah Domain. The Cosmo Supergroup (Figure 2) comprises iron-rich sedimentary rocks, tuff, carbonate and siliciclastic rocks

(South Alligator Group), overlain by a thick turbidite succession with interbedded felsic volcanic rocks (Finniss River Group; Needham et al., 1988). The turbidites are dominated by c. 1865 Ma detritus (Worden et al., 2008a, b), possibly derived from an actively uplifted Nimbuwah Domain shedding sediment into an evolving Central Trough. The Cosmo Supergroup experienced greenschist-facies metamorphism associated with tight upright N-S folding prior to c. 1835 Ma.

Cosmo Supergroup correlatives are known in the Litchfield Province (Hermit Creek and Welltree metamorphics; Figure 2; Pietsch and Edgoose, 1988; Worden et al., 2008a). These were intruded by 1863–1850 Ma S-type granites of the Allia Creek and Wagait suites. In contrast to the Central and Nimbuwah domains, metamorphism in the Litchfield Province at 1855 Ma is characterised by low-pressure, high-temperature assemblages (Carson et al., 2008); coupled with arc-related mafic magmatism (Woolianna Formation; Figure 2), this style of metamorphism is consistent with a back-arc environment (Glass, 2010).

### *c. 1835–1820 Ma late-orogenic magmatism and tectonism*

In the Central Domain, syn- to late-orogenic granites of the Cullen Supersuite were emplaced at 1835–1820 Ma as a series of plutons termed the Cullen batholith (Figure 1; Worden et al., 2008a). This magmatism was synchronous with the deposition of fluviatile, lacustrine and alluvial fan sedimentary rocks and subaerial volcanic and volcaniclastic rocks unconformably on the South Alligator Group in a fault-bound graben at 1829–1822 Ma (El Sherana and Edith River groups, South Alligator Valley; Figure 2; Needham et al., 1988; Freidmann and Grotzinger, 1994).

### *<1818 Ma sedimentation*

Deposition of braided fluvial sediments of the Katherine River Group at the base of the McArthur Basin started after the 1825 Ma deposition of the Edith River Group and before 1720 Ma granite emplacement (Rawlings and Page, 1999). They were deposited over the Nimbuwah Domain, and at least parts of the Central Domain, and extend far to the southeast. The basal Kombolgie Sub-group comprises conglomerate and quartz sandstone, with mafic volcanic and volcaniclastic intervals in the lower part (Sweet et al., 1999).

## Economic geology

The Pine Creek Orogen is the most fertile part of the North Australian Craton, with over 1,000 mineral occurrences. It contains over 20% of the world's low-cost U resources, has a known resource of about 9 M oz of Au, and produced 3.2 M oz of Au between 1870 and 2007. Considerable resources of Ni-Co-Pb-Cu, Pb-Zn-Ag, Pt-Pd, Sn-Ta-W, Fe ore, magnesite, phosphate and other commodities also exist in this region.

Structurally controlled deposits are predominantly vein type and include Au, base metals, and Sn-bearing veins and Sn-Ta pegmatites. Stratigraphically controlled deposits include stratiform Au and stratabound polymetallic deposits, as well as volcanogenic massive sulfides, U, Fe ore, phosphate and magnesite deposits. The distributions of U, Au and Fe ore mineralisation were largely controlled by large-scale hydrothermal systems, active at distinct stages in the orogenic history. In the following section we provide a brief description of these three mineralisation types (Figure 1).

### *Uranium mineralisation*

Uranium deposits in the Pine Creek Orogen are grouped into four main types: unconformity-related, vein-type, intrusive-related and surficial deposits (Lally and Bajwah, 2006). Major deposits cluster around the East Alligator (Nimbuwah Domain), South Alligator and Rum Jungle U fields (Central Domain, Figure 1), although the only operating mine is Ranger, located in the East Alligator U field, 260 km east of Darwin.

*Ranger* is one of the largest U mines in the world. The original Ranger 1 orebody, mined from 1980–1994, produced 18 Mt of ore grading 0.3% $U_3O_8$. Mining at Ranger 3 commenced in 1997 at the rate of >5,000 t $U_3O_8$ per year. In January 2010, Energy Resources of Australia reported the total resource at Ranger 3 as 127 Mt at 0.09% $U_3O_8$. Recent drilling has outlined an eastern extension, named Ranger 3 Deeps, which is a target of current exploration. The total endowment (past production plus remaining resource) of Ranger 1 and 3 stands at 130,000 t of metal contained in ore grading 0.24–0.37% $U_3O_8$ (MODAT, 2011).

The mineralisation is broadly stratabound, being hosted in the carbonaceous lower Cahill Formation, close to the contact with the unconformably underlying Neoarchean Nanambu Complex (Figure 1). Amphibolite-facies metamorphism was associated with NNE- to WNW-trending folds, thrusts and shear zones, cross-cut by east-trending pegmatite veins and gently-dipping NNE-trending mafic dykes, historically interpreted as the Oenpelli Dolerite. Uranium was introduced during late extension after emplacement of the dolerite (Hein, 2002). Fluid–rock interaction during mineralisation produced extensive alteration including chloritisation, sericitisation and hematitisation.

The primary U minerals are uraninite and pitchblende, with some coffinite and minor brannerite and curite. Free Au occurs in uraninite. Secondary U minerals include saleeite, sklodowskite, torbanite and kasolite.

Existing mineralisation models for the Alligator Rivers Uranium Field differ in (a) the relative importance assigned to unconformities at the Neoarchean basement/Cahill Formation and Cahill Formation/Kombolgie Subgroup contacts and (b) the inferred source of U-bearing fluids (i.e., basin-derived vs. basement-derived). However, there is increasing consensus that the fluids were basin-sourced and descended from the Kombolgie Subgroup into the Paleoproterozoic basement where mineralisation occurred. Polito et al. (2011) proposed a general model in which the Kombolgie Subgroup was deposited over the Neoarchean and Paleoproterozoic basement and compartmentalised into diagenetic aquifers and aquitards. Diagenetic illite, syn-ore sericite and uraninite formed at 1680 Ma, indicating a link between fluid flow in the sandstones and mineralisation (Figure 3). This is supported by the similar chemical and isotopic nature of fluid inclusions and diagenetic minerals in the Kombolgie Sub-group and in syn-ore quartz veins (Polito et al., 2005). Hydraulic fracturing in basement-rooted structures during basin formation and fluid flow may have provided a mechanism for formation of uraninite-bearing veins and breccias. Their interaction with reduced lithologies in the basement provided a mechanism for U precipitation.

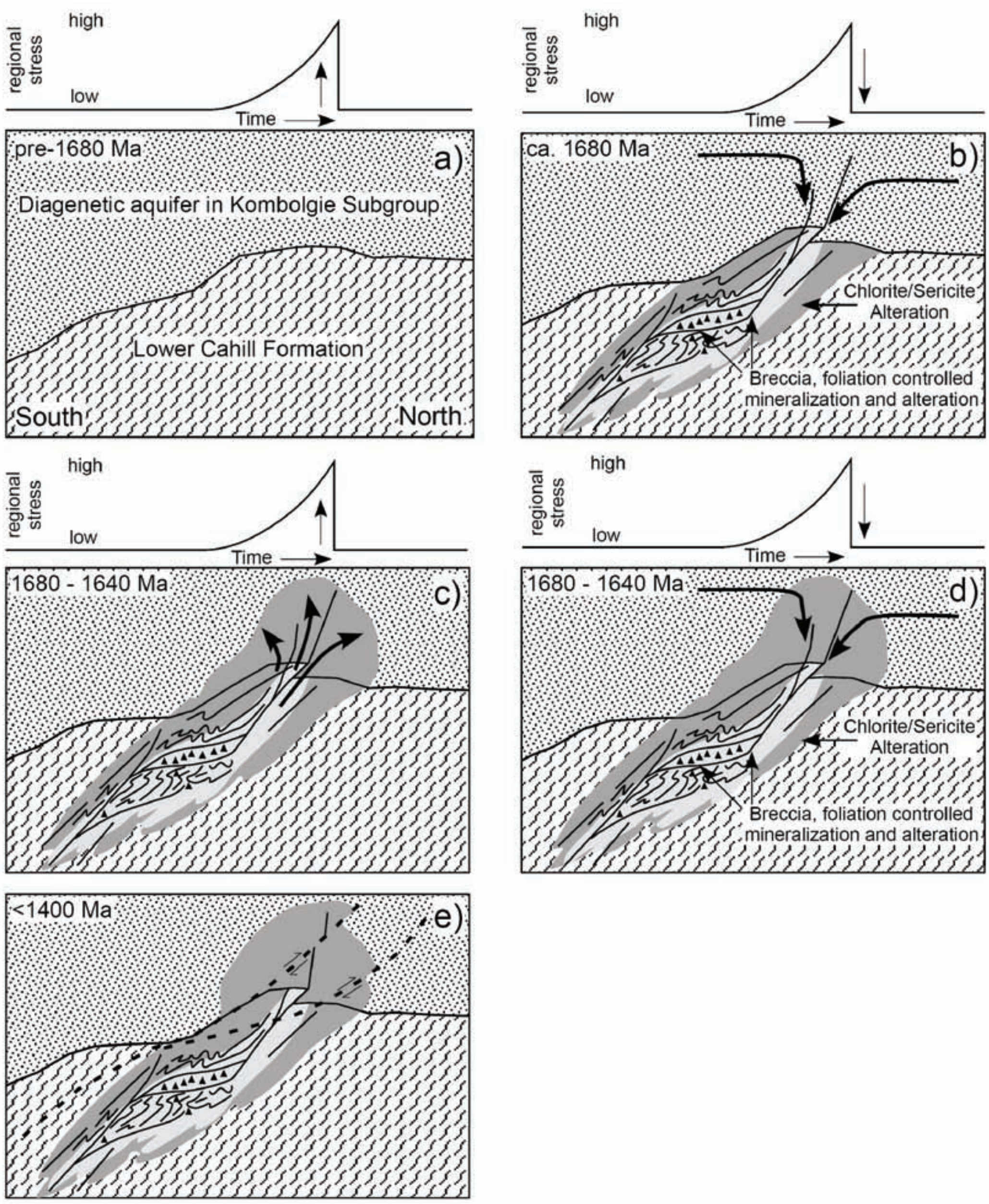


***Figure 3 A schematic representation of the formation of alteration styles surrounding the Jabiluka U deposit (after Polito et al., 2005).***

## *Gold mineralisation*

Almost all Au occurrences in the Pine Creek Orogen are confined to the Central Domain (Figure 1). They are preferentially hosted by the South Alligator Group and the lower parts of the Finniss River Group along anticlines, strike-slip shear zones and thrusts proximal to the Cullen batholith. Two main mineralisation types include gold-quartz veins (e.g., Pine Creek Gold Field) and Au in Fe-rich sedimentary rocks (e.g., Cosmo Howley).

*The Pine Creek Gold Field* is the most productive in the Pine Creek Orogen and comprises a NW-trending belt 6 km long and 1 km wide, about 0.5 km West of Pine Creek, adjacent to the western margin of the Pine Creek Shear Zone. It includes 15 deposits in sheared and contact-metamorphosed turbidites of the Mount Bonnie and Burrell Creek formations. From the 1870s to 2005, the goldfield produced 47.4 t of Au (MODAT, 2011). The majority of production (23.8 t of Au from ore grading 2.95 g/t Au) came from the Enterprise mine that operated from 1985–1995. Mineralisation comprises saddle reefs and, less commonly, discordant quartz veins or fault- and shear-hosted zones. Minor Au is disseminated in the wall rock adjacent to quartz veins. Gold is free milling or is contained in arsenopyrite.

*The Cosmo Howley* mine, 60 km NW of Pine Creek, was another major Au producer in the Pine Creek Orogen. It produced 1.05 t of Au from 1879–1915, and 15.7 t of Au from ore grading 2.04 g/t Au from 1987–1993. The remaining indicated resource for the Cosmo underground operation is stated as 5.30 Mt @ 4.6 g/t Au (MODAT, 2011). Mineralisation is broadly stratabound, being hosted by greenschist-facies banded ironstone and mudstone in the Koolpin Formation (Figure 4) on the Howley Anticline. Mineral assemblages include chlorite and actinolite with minor mica, quartz, garnet, graphite and fine-grained pyrite. Free Au is rare and most Au occurs as sub-

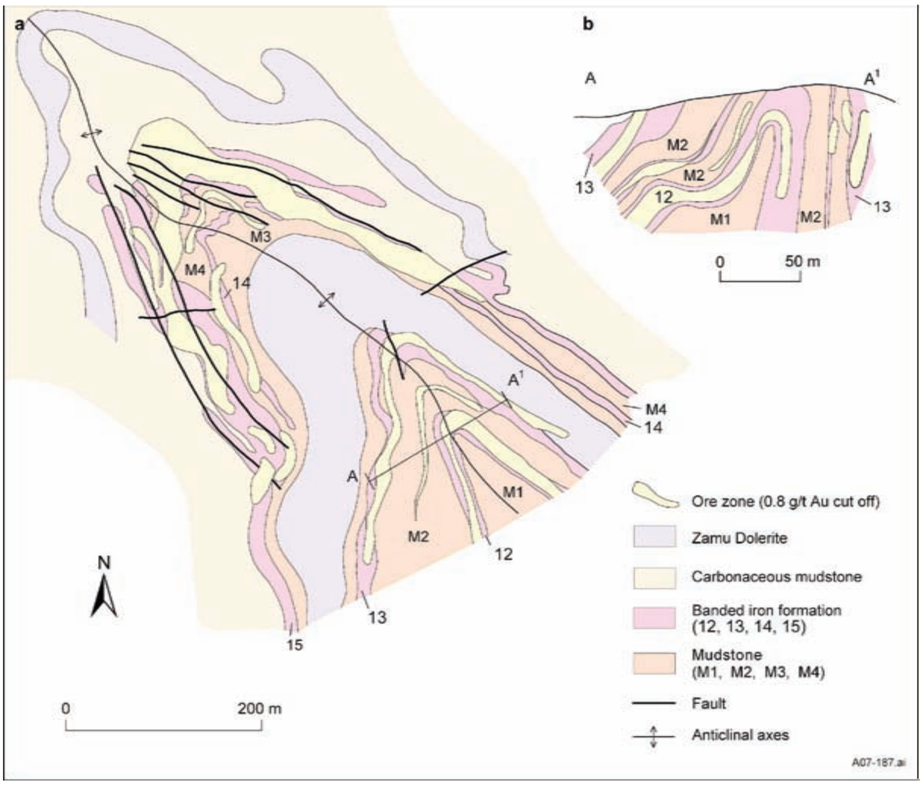


***Figure 4 Geology of Cosmo Howley Au mine. a) geological plan and b) cross-section at Cosmo Howley mine (after Dominion Mining Ltd., 1998).***

of the open pits, suggested that the hematite mineralisation is the result of hydrothermal remobilisation (Bowden, 2000). Fluids derived from the intrusion of the Allamber Springs Granite concentrated Fe oxides from ferruginous banded shale in the Wildman Siltstone into pre-existing favourable structural sites. Later supergene enrichment occurred, possibly leaching phosphorus.

## Tectonic controls on the metallogeny in the Pine Creek Orogen

There are strong indications that distinct tectonic settings in different parts of the Pine Creek Orogen have had an important bearing on the distribution and style of mineralisation in the orogen. Most, if not all, of these systems are related to fluid flow after, or in the late stages of, orogenesis.

Uranium deposits and occurrences are spatially linked with variably tectonised contacts with the Neoarchean basement (Figure 1), but also fall largely within the Nimbuwah Domain. These may have been

microscopic inclusions within arsenopyrite and pyrite. Other sulfides include minor chalcopyrite and pyrrhotite (Ahmad et al., 1999).

Gold deposits in the Pine Creek Orogen occur almost exclusively within 3–5 km of the Cullen Batholith, and are generally thought to be related to granite emplacement at 1835–1820 Ma (e.g., Matthäi et al., 1995). Goldfarb et al. (2001) included these in a distinctive class of deposits they termed 'orogenic Au deposits'. A genetic model incorporating existing observations and data was proposed by Wygralak (1996; Figure 5). However, field observations (Matthäi et al., 1995) and geochronology studies (Sener et al., 2005; Rasmussen et al., 2006) suggest that Au mineralisation at some deposits (e.g., Mount Todd and Goodall) formed 40–100 Myr after granite emplacement. This may be associated with long-lived hydrothermal systems driven by the high-heat producing granites (Rasmussen et al., 2006).

### *Iron mineralisation*

Most known Fe ore occurrences in the Pine Creek Orogen are hosted by the Wildman Siltstone (Central Domain). The largest of these, the Frances Creek Iron Ore Field, was discovered in 1961 and comprises fifty named Fe ore occurrences (Figures 1 and 6). Current indicated and inferred resources (inclusive of reserves) are 8.59 Mt @ 59.2% Fe. The largest resources, at the Helene 5/6/7 deposit, are 5.21 Mt @ 58% Fe. Production from September 2007–June 2010 was 4.16 Mt of ore (Territory Resources Ltd., 2010). The ore consists of massive, fine, micaceous to bladed hematite and contains varying amounts of shale fragments and quartz grains.

Crohn (1968) considered that the deposits formed as a result of supergene enrichment of pyritic shale breccia. A more recent review of mining data, along with petrological studies and field inspections

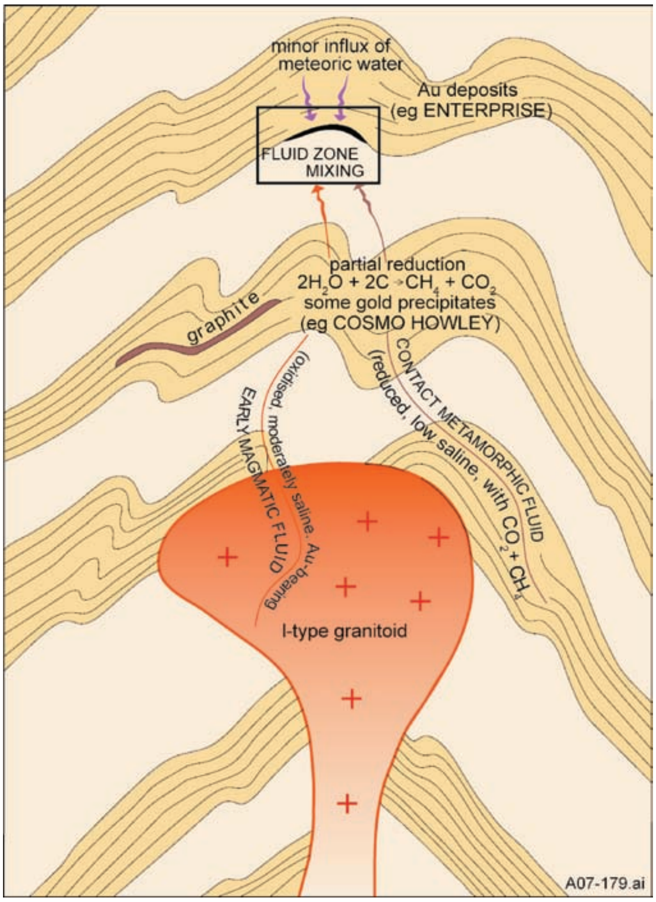


***Figure 5 Model for genesis of Au deposits in the Pine Creek Orogen (after Ahmad et al., 1999).***

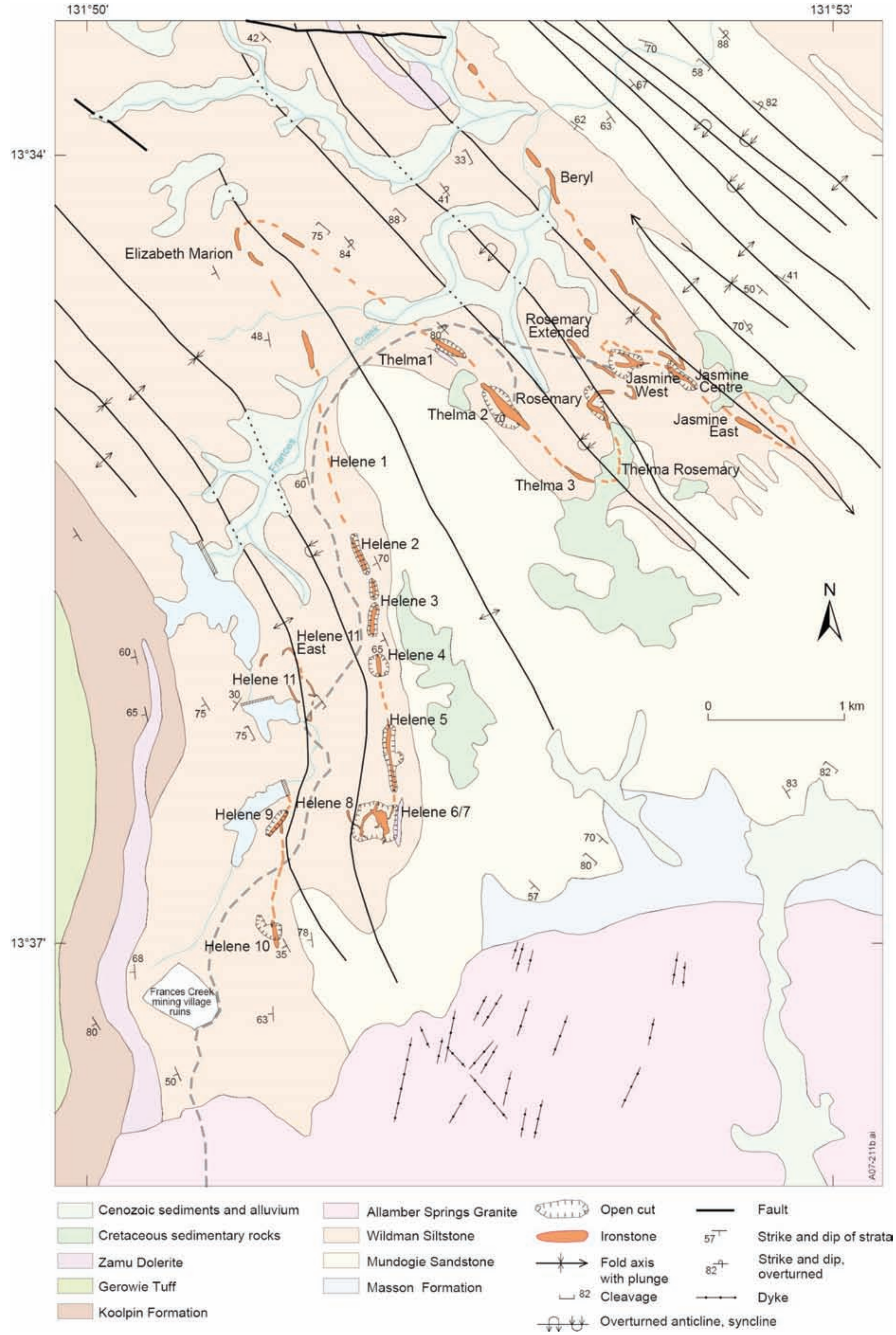


*Figure 6 Geological map of southern part of Frances Creek Iron Ore Field, with significant deposits labelled (after Ferenczi, 2001).*

controlled in part by Paleoproterozoic basement structures associated with amalgamation of the Nimbuwah and Central domains at c. 1863 Ma. In the period 1867–1860 Ma, the Nimbuwah Domain experienced crustal thickening, probably in response to closure of a basin between the two domains, W-vergent thrusting, Nimbuwah Complex magmatism and associated amphibolites facies metamorphism (Hollis et al., 2009b, 2011). Subsequently, the high heat-producing granites (Neoarchean and Paleoproterozoic), coupled with the thermal cap provided by the Kombolgie Subgroup, provided the pre-existing conditions required for generation of large-scale hydrothermal systems. Reactivation of Paleoproterozoic structures may have formed important fluid conduits during subsequent Mesoproterozoic (and later) U mineralisation events. Pre-existing basement structures are known to have a strong control on U mineralisation at Ranger and Nabarlek.

In contrast, the Central Domain is dominated by Au deposits and occurrences, which have a strong spatial, and to some degree temporal, association with late orogenic granites. The Cosmo Supergroup, which hosts Au occurrences, is dominated by turbidites of the Finniss River Group, thought to have been deposited in a foreland basin sourcing detritus from highlands in the Nimbuwah Domain (Hollis et al., 2009b). Greenschist-facies metamorphism and associated folding and shearing during collisional tectonism generated structural trap sites. Subsequent emplacement of 1835–1820 Ma high-level late- to post-orogenic granites resulted in the interaction of gold-bearing magmatic fluids with reduced, low-salinity contact-metamorphic fluids and the precipitation of Au in reduced lithologies in these pre-existing traps. High heat-producing granites probably also generated long-lived hydrothermal systems, accounting for Au mineralisation that significantly postdates granite emplacement. Such long-lived fluid systems are also thought to have been responsible for the Fe mineralisation at Frances Creek.

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# Geology and controls on mineralisation in the Eastern Goldfields region, Yilgarn Craton, Western Australia

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*The Yilgarn Craton in Western Australia contains evidence of the oldest crust on Earth. Greenstone successions developed after c. 3000 Ma show a complex history of juvenile crust generation and crustal reworking. There are at least three periods of greenstone-related magmatism in the Yilgarn Craton. The earliest recognised greenstone development consists of volcanic and sedimentary successions deposited between c. 3000–2900 Ma. A mantle plume at c. 2800 Ma produced large mafic–ultramafic igneous complexes and probably initiated rifting on the eastern side of the craton and incipient rifting in the NW. A second major plume, at c. 2700 Ma, was focussed along the rupture created by the c. 2800 Ma event and may have been associated with the re-accretion of lithospheric blocks created by the earlier event. Komatiites generated by the c. 2700 Ma plume contain world-class Ni deposits, and structures developed subsequent to the peak of plume activity host world-class Au deposits. Recent studies of Ni and Au deposits in the Eastern Goldfields Superterrane have shown how features ranging in scale from the lithosphere to regional structural and stratigraphic controls to local volcanological and sedimentological variations can affect the size and distribution of deposits. This understanding is now being applied in exploration targeting.*

## Introduction

The Paleo–Neoarchean Yilgarn Craton in Western Australia (Figure 1) is a highly mineralised granite-greenstone terrain with world-class deposits of Au and Ni, and significant iron and volcanic-hosted massive sulphide (VHMS) base-metal deposits. Economic Fe deposits are confined to the western part of the craton.

Over the past 15 years, the acquisition of large datasets and major advances in the understanding of the geological evolution of the Yilgarn Craton at all scales have encouraged the application of the holistic mineral systems approach to mineral exploration as a tool for developing targeting criteria, particularly for Ni and Au (McCuaig et al., 2010). In this review, examples from the Kambalda district in the Eastern Goldfields region illustrate how the size, distribution and concentration of Au and Ni deposits are controlled by factors from the craton to the regional scale, down to the deposit cluster and individual deposit scale. While there has been little recent, regional-scale work on Yilgarn VHMS deposits, comparison with similar terrains in Canada suggests that the fundamental controls on Ni mineralisation also influence the distribution and endowment of VHMS mineralisation (Huston et al., 2005).

## Yilgarn Craton

Cassidy et al. (2006) divided the Yilgarn Craton into terranes defined on the basis of distinct sedimentary and magmatic associations, geochemistry and ages of volcanism. The Narryer and South West terranes in the west are dominated by granite and granitic gneiss with minor supracrustal greenstone inliers, whereas the Youanmi Terrane and the Eastern Goldfields Superterrane contain substantial greenstone belts separated by granite and granitic gneiss. Subsequent revision has further subdivided the Eastern Goldfields Superterrane into the Kalgoorlie, Kurnalpi, Burtville and Yamarna terranes (Figure 1; Pawley et al., 2012).

The Ida Fault (Figure 1), which marks the boundary between the western Yilgarn Craton and the Eastern Goldfields Superterrane, is a major structure that extends to the base of the crust (Drummond et al., 2000). Various geophysical techniques, including deep-crustal seismic (Drummond et al., 2000; Goleby et al., 2003), seismic receiver-function analysis (Reading et al., 2007), and magnetotelluric surveys (Dentith et al., 2012), show the Yilgarn crust to be 32–46 km thick, with the shallowest Moho beneath the Youanmi Terrane. The crust is thicker in the SW, and thickest in the eastern part of the Eastern Goldfields Superterrane. Seismic and gravity data suggest that the greenstones are 2–7 km thick (Swager et al., 1997).

Isotopic data, including Sm-Nd (Figure 2; Champion and Cassidy, 2007) and Lu-Hf (Mole et al., 2010; Wyche et al., 2012) data, show that the terrane subdivisions of the Yilgarn Craton reflect regions with distinctive crustal histories. The Narryer Terrane, which contains both the oldest detrital zircons yet found on Earth (back to c. 4400 Ma; Wilde et al., 2001) and the oldest rocks in Australia (back to c.

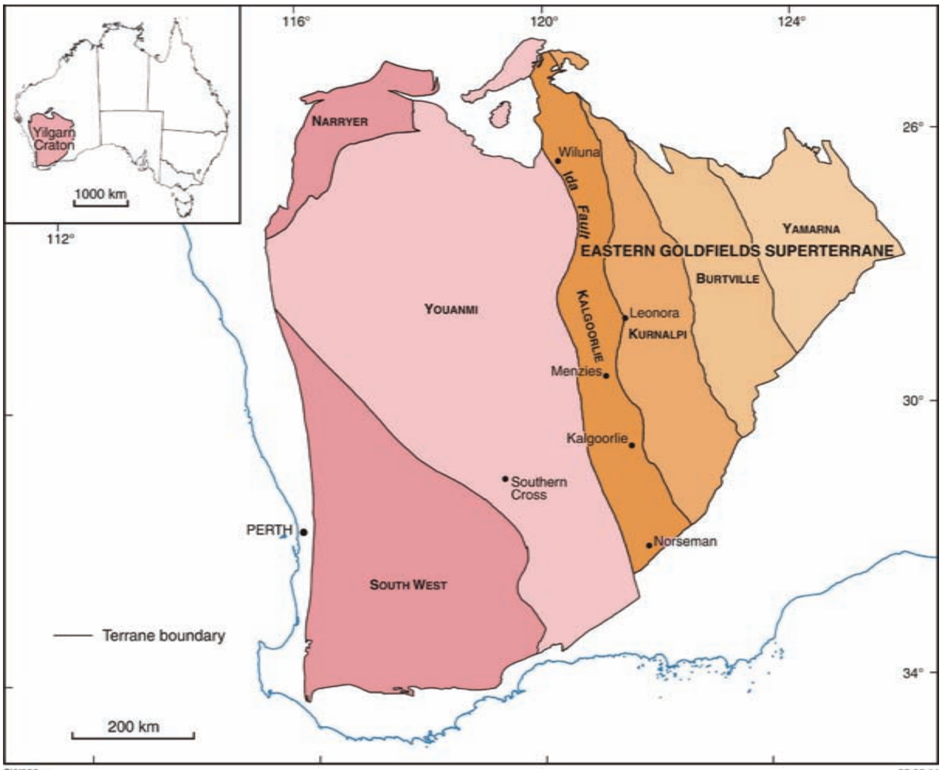


*Figure 1 Subdivision of the Yilgarn Craton (modified from Pawley et al., 2012).*

3730 Ma; Kinny et al., 1988), shows abundant evidence of very old model ages. The Youanmi Terrane has a more mixed history, whereas the Eastern Goldfields Superterrane is distinctly more juvenile than the terranes to the west.

## Yilgarn granite–greenstones

The supracrustal rock record in the Yilgarn Craton dates back to at least c. 3080 Ma in the Youanmi Terrane in the west (Wang et al., 1998; Yeats et al., 1996; Rasmussen et al., 2010; Van Kranendonk et al., 2012) and c. 2960 Ma in the Burtville Terrane in the NE (Pawley et al., 2012). However, greenstone successions across the Yilgarn Craton are dominated by rocks that formed after c. 2820 Ma.

In the central Youanmi Terrane, a cycle of mafic–ultramafic–felsic volcanism between c. 2820–2735 Ma is likely due to a major plume that produced large mafic–ultramafic layered intrusions between c. 2820–2800 Ma (Ivanic et al., 2010), coincident with similar, but less voluminous, magmatism in the eastern part of the craton (Wyche et al., 2012). This event may have resulted in partial break-up of the early Yilgarn Craton with rifting in the east (Czarnota et al., 2010) and incipient rifting marked by younger Nd model ages and the layered intrusions in the Youanmi Terrane (Ivanic et al., 2010). A protracted period of mafic – felsic volcanism and associated sedimentation continued from c. 2800–2735 Ma. Calc-alkaline volcanism was dominant after c. 2760 Ma, and broadly coincided with a period of mafic tonalite–trondhjemite–granodiorite (TTG) and enriched high-field-strength element (HFSE) granite magmatism (Cassidy et al., 2002; Van Kranendonk et al., 2012).

The last recognised regional greenstone-forming event in the Youanmi Terrane was a mafic to felsic volcanic cycle between c. 2740–2725 Ma (Van Kranendonk et al., 2012), which was contemporaneous with high-Ca TTG granite magmatism (Cassidy et al., 2002).

Except for rare greenstones in the South West Terrane (Allibone et al., 1998), after c. 2715 Ma, volcanic activity and greenstone development in the Yilgarn Craton was restricted to the Eastern Goldfields Superterrane. Andesite-dominated calc-alkaline volcanism in the eastern Kurnalpi Terrane (Figures 1 and 3; Barley et al., 2008) and dacitic volcanism in the northern Kalgoorlie Terrane (Rosengren et al., 2005) was prevalent between c. 2715–2705 Ma (Fiorentini et al., 2005; Kositcin et al., 2008). Barley et al. (2008) interpreted the andesite-dominated successions as oceanic intra-arc volcanic centres.

In the Eastern Goldfields Superterrane, a second major plume event (Campbell and Hill, 1988) produced voluminous komatiites which occur as both high-level intrusions and flows (Trofimovs et al., 2004; Fiorentini et al., 2005, 2010). They are preserved in a distinct N–NW-trending belt, 600 x 100 km, between Norseman and Wiluna (Figure 1). The mafic–ultramafic succession also contains tholeiitic and komatiitic basalts (Lesher, 1983; Squire et al., 1998; Said and Kerrich, 2009). It is well constrained between c. 2710–2692 Ma (Kositcin et al., 2008) and partly overlaps in age with the andesite-dominated calc-alkaline volcanism. The Norseman–Wiluna komatiites, which host major Ni deposits, are not only younger than ultramafic rocks in the Youanmi Terrane, but also differ in chemical character. Komatiites of the Youanmi Terrane include Al-depleted, and Al-undepleted and Ti-enriched varieties, whereas those in the Norseman–Wiluna belt are Al-undepleted (Barnes et al., 2007). Abundant SHRIMP geochronological data on greenstones from throughout the Eastern Goldfields Superterrane (Geological Survey of Western Australia, 2011) suggest that thinner and more sparsely distributed komatiite units east of the Norseman–Wiluna belt (e.g., E and SE of Leonora; Figure 3) are mainly the same age as the more voluminous material within the main belt and may represent thin flows or channel deposits which have travelled farther as result of paleotopography.

Between c. 2692–2680 Ma, volcanic centres in the western part of the Kurnalpi Terrane produced bimodal (basalt–rhyolite) volcanic and associated intrusive and sedimentary rocks (Figure 3), coinciding with the main period of high-HFSE granite magmatism (Cassidy et al., 2002). Barley et al. (2008) interpreted these 'Gindalbie' successions as representing an arc-rift environment. Gindalbie-style volcanism, which locally hosts VHMS mineralisation, overlapped in age with, and was succeeded by, TTG volcanism and associated sedimentary rocks and mafic intrusions represented by the Black Flag Group in the Kalgoorlie Terrane (Figure 3). The deposition of the Black Flag Group between c. 2690–2660 Ma coincided with voluminous high-Ca TTG granite magmatism in the Eastern Goldfields Superterrane (Champion and Cassidy, 2007). Krapez and Hand (2008) interpreted the Black Flag Group (their 'Kalgoorlie Sequence') as representing a strike-slip intra-arc basin, whereas Squire et al. (2010) argued that they are the result of volcanism and sedimentation

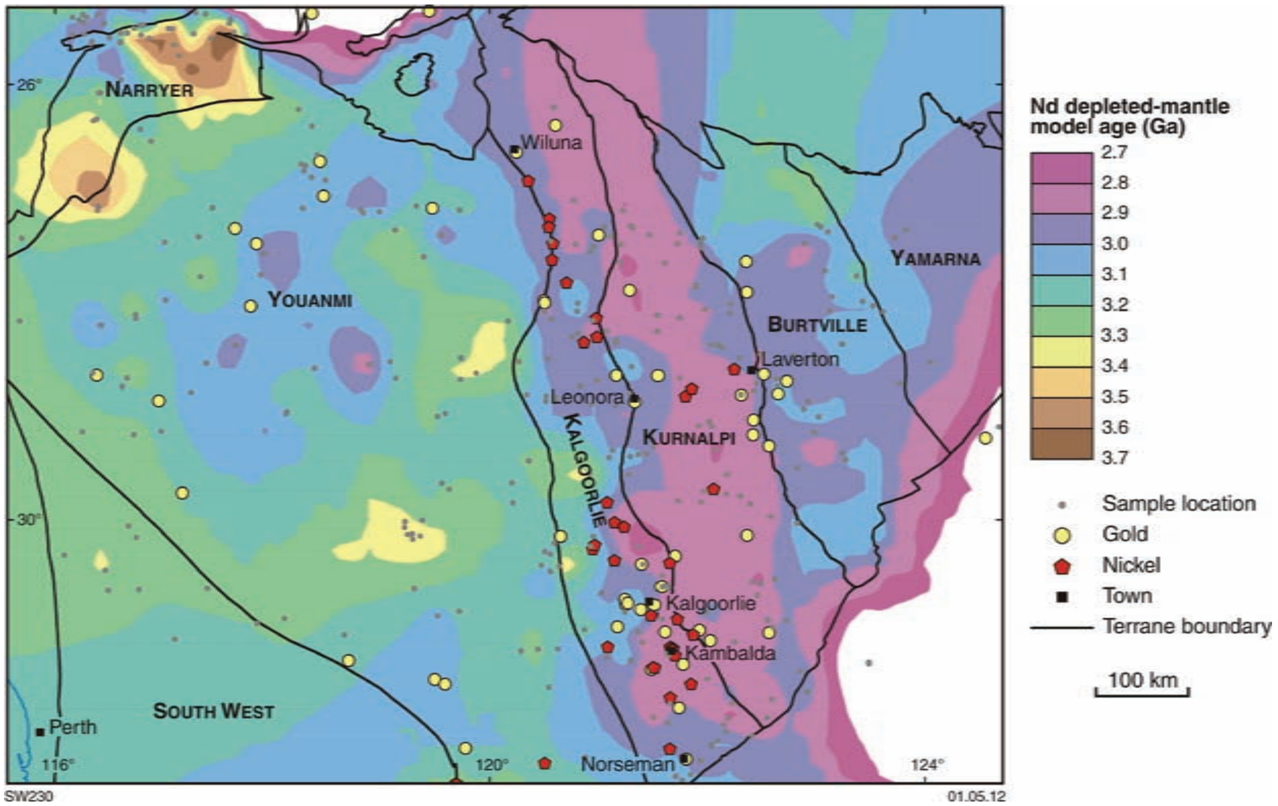


***Figure 2 Nd depleted-mantle model age map for the Yilgarn Craton showing terrane subdivisions and locations of major Ni and Au deposits (modified from Champion and Cassidy, 2007).***

associated with extensional deformation due to the emplacement of large granite batholiths. Felsic volcanic and associated plutonic rocks of this age have also been recorded in a poorly exposed bimodal greenstone succession in the Yamarna Terrane in the far E of the Eastern Goldfields Superterrane (Figure 1; Pawley et al., 2012).

The youngest supracrustal successions in the Yilgarn Craton are the so-called 'late basins', which rest unconformably on all earlier greenstones in the Eastern Goldfields Superterrane. Likely deposited in a very short time (c. 10 Myr) after c. 2665 Ma (Squire et al., 2010), they preserve fluvial and deep-marine facies, which Krapez and Barley (2008) interpreted as having formed in a tectonic-escape corridor after arc closure. The late-basin sediments, which range from turbidites through to coarse, braided-stream sediments (Krapez et al., 2008), contain a range of detrital zircon ages and postdate the cessation of TTG granite magmatism. They contain material derived from both proximal and distal sources during ongoing extension and uplift (Squire et al., 2010).

Finally, the cessation of greenstone deposition was accompanied by craton-wide low-Ca granite magmatism (Cassidy et al., 2002). A distinctive belt of alkaline granites, emplaced at this time, appears to coincide with deeply penetrating crustal structures and is mainly restricted to the Kurnalpi Terrane (Smithies and Champion, 1999).

## Eastern Goldfields: stratigraphy and structure

### Stratigraphy

Poor exposure, deep weathering, lack of detailed geochronology, and structural and metamorphic overprints preclude description of detailed stratigraphy in many of the Yilgarn greenstones. West of the Ida Fault (Figure 1), only the NW part of the Youanmi Terrane has an established stratigraphy (Van Kranendonk et al., 2012). East of the Ida Fault, local stratigraphy has been established in some greenstone belts (Kositcin et al., 2008) but detailed regional stratigraphy has been described only for the southern part of the Kalgoorlie Terrane.

The southern Kalgoorlie Terrane (Figure 4; Woodall, 1965; Gresham and Loftus-Hills, 1981; Swager et al., 1995) comprises a lower mafic–ultramafic succession consisting of the Lunnon Basalt, Kambalda Komatiite (including the Silver Lake and Tripod Hill members), Devon Consols Basalt, Kapai Slate and Paringa Basalt. The mafic–ultramafic succession is unconformably overlain by the Black Flag Group, which comprises extensive volcaniclastic rocks, rhyolitic to dacitic volcanic rocks, intrusive mafic complexes and minor mafic volcanic rocks (Squire et al., 2010). The late-basin sediments are represented in this area by polymictic conglomerate of the Kurrawang Formation, which contains a variety of clasts, including banded-iron formation and granite that indicate a distal provenance, probably in the Youanmi Terrane (Krapez et al., 2008). In less-deformed areas, many primary igneous features and textures are still visible despite locally complete replacement by alteration assemblages.

The abundant pillow lavas and hyaloclastites in basalts, the presence of marine sediments, and quench textures in komatiites and basalts indicate a submarine eruption of the mafic–ultramafic succession (Hill et al., 1995; Squire et al., 1998; Said and Kerrich, 2009). Squire et al. (1998) proposed that the Lunnon Basalt is either distal to a shield volcano, or represents a ponded lava field in an extensional basin distant from the eruptive centre. Similarly, Hill et al. (1995) suggested that the komatiite flows at Kambalda are distal deposits in contrast with the thick, cumulate dunite bodies to the N

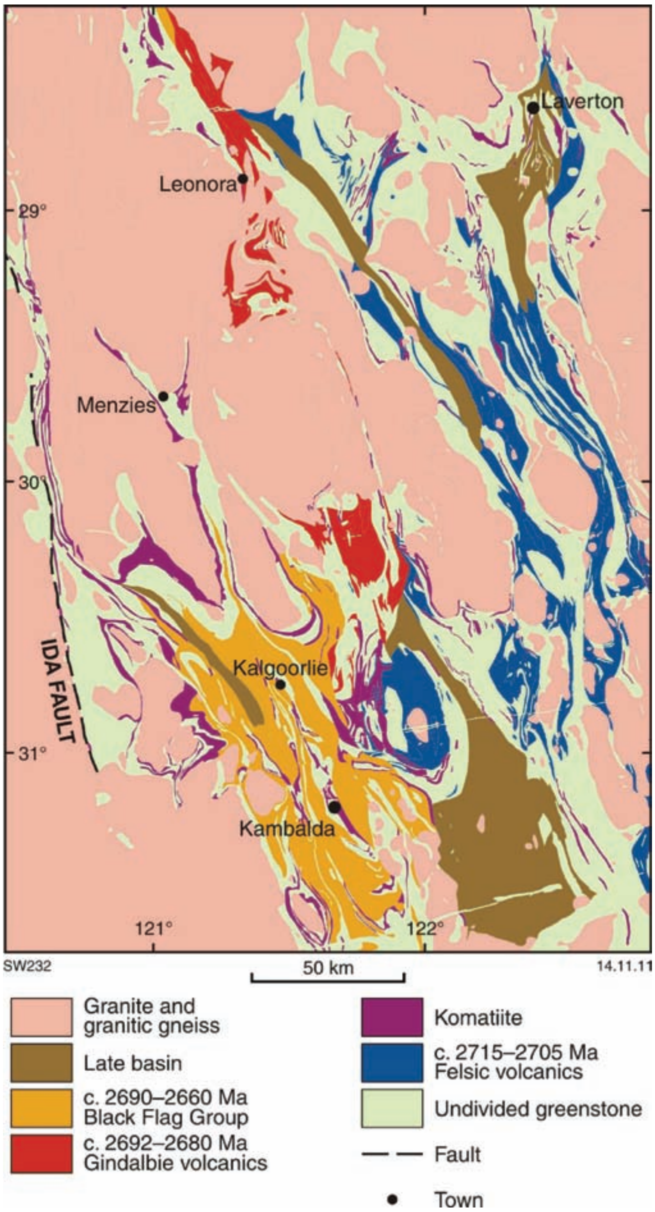


***Figure 3 Distribution of main volcanic facies in the central part of the Eastern Goldfields Superterrane.***

and NW of Kalgoorlie (Figure 3), which are proximal to the eruptive centre.

The nature of basement to the mafic–ultramafic succession in the Kalgoorlie region is unknown. Xenocrystic zircon age data (Compston et al., 1986) and trace-element and isotopic data suggest the mafic–ultramafic succession may have been generated through varying degrees of crustal contamination (Arndt and Jenner, 1986) and mixing of a depleted mantle source with an enriched subcontinental lithospheric mantle (Said and Kerrich, 2009). Model age data based on Lu-Hf analyses on zircons have peaks after 3500 Ma, and mainly after 3100 Ma. This is significantly younger than the earliest model age recognised in the Youanmi Terrane (Wyche et al., 2012).

## Structure and metamorphism

Building on the regional framework established by Swager (1997), Blewett et al. (2010a) produced a six-stage, integrated structural-event framework (Figure 5) for the Eastern Goldfields Superterrane to account for the documented magmatic, depositional, structural and metamorphic history (Czarnota et al., 2010). In this scheme, the period of greenstone deposition between c. 2715–2705 Ma, characterised by calc-alkaline and komatiite magmatism, was a time of dominantly extensional tectonics that marked the initiation of regional-scale granite doming.

The deposition of the Black Flag Group, between c. 2690–2660 Ma, was accompanied by the widespread emplacement of a high-Ca TTG granite suite. Granite doming was probably coeval with local contraction indicated by upright folding and dextral shearing at this time. The peak period of granite doming began during the last depositional phase of the Black Flag Group (Squire et al., 2010). Ongoing doming and extension produced the clastic late-basin sediments. After the cessation of high-Ca magmatism, low-Ca granite magmatism, which appears to be the result of melting of a mid–lower crustal source of TTG/high-Ca composition (Champion and Cassidy, 2007), was accompanied by a major contractional deformation, which produced both upright folds and regional-scale sinistral shearing which may have reactivated earlier structures. Subsequent, relatively minor, brittle contractional and extensional events affected the now rigid Yilgarn Craton (Czarnota et al., 2010).

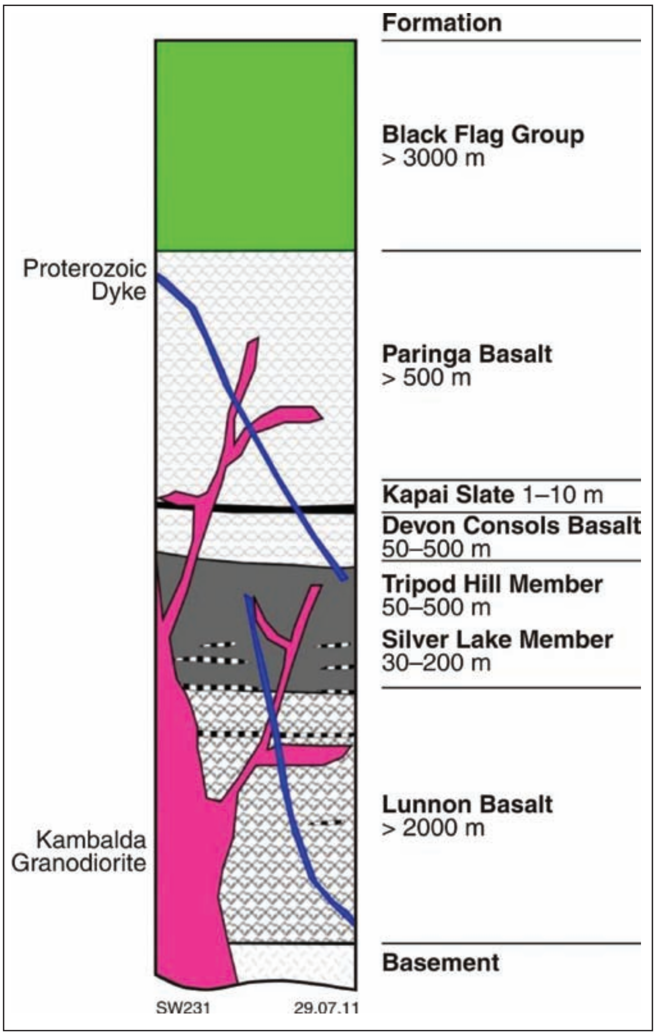


***Figure 4 Kambalda stratigraphy (modified from Beresford et al., 2005).***

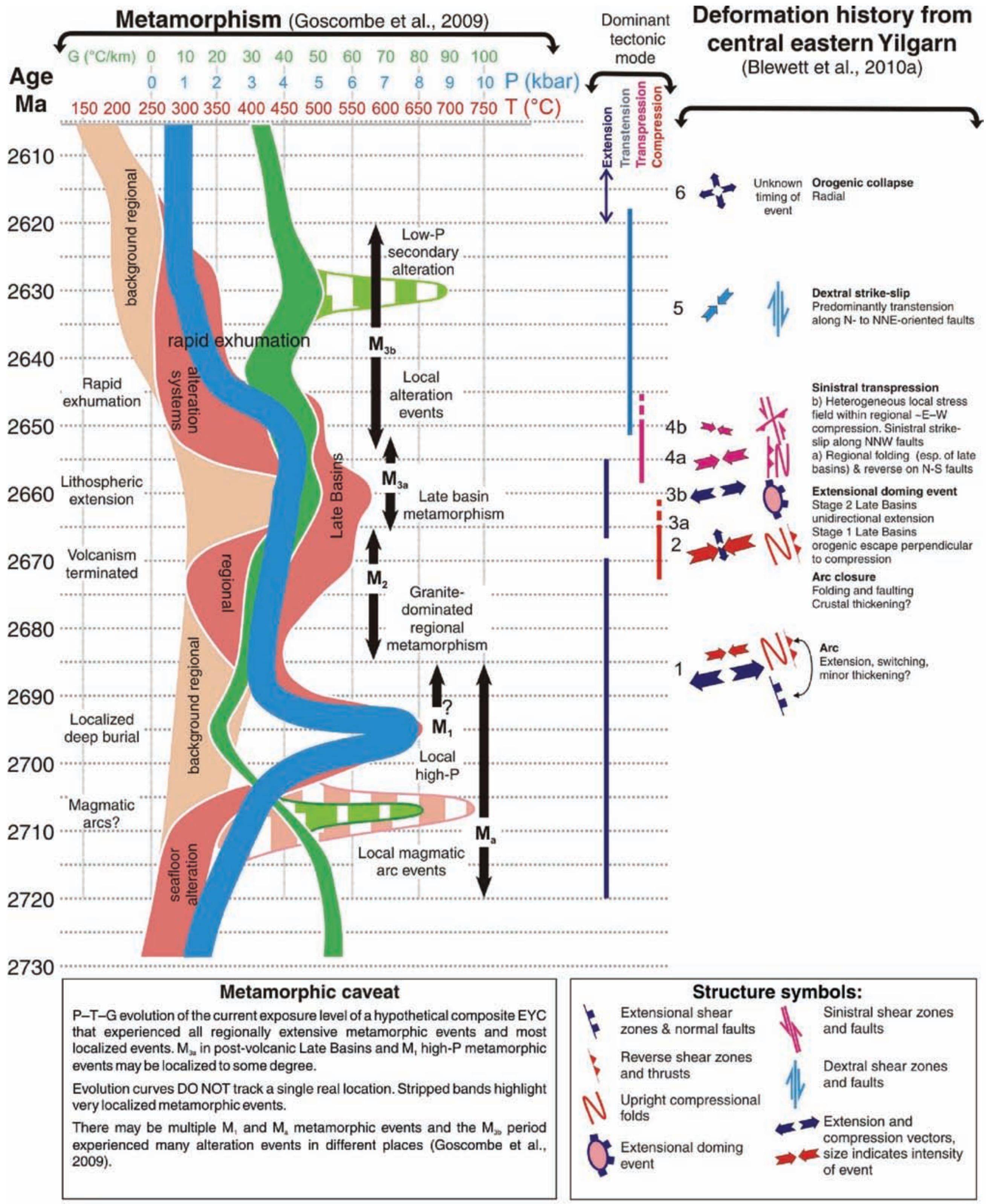


***Figure 5 Metamorphic and structural history of the Eastern Goldfields Superterrane (modified from Czarnota et al., 2010).***

Local evidence of early, low-pressure granulite-facies metamorphism, consistent with a high geothermal gradient (Figure 5; Goscombe et al., 2009), was contemporaneous with the eruption of the Norseman–Wiluna komatiites. Later medium-pressure metamorphism was most likely due to exhumation of deep-seated early structures around granite domes (Goscombe et al., 2009). Subsequent periods of low pressure metamorphism, accompanied by moderate to high geothermal gradients, reflect exhumation during granite doming prior to and during late-basin development. Very low pressures and geothermal gradients mark the end of the period of granite doming and the initiation of widespread exhumation (Goscombe et al., 2009).

## Eastern Goldfields: nickel and gold

The stratigraphy, structure and metamorphic history of the southern Kalgoorlie Terrane remain the most comprehensively studied and documented part of the Yilgarn Craton. Consequently, this region provides the best insight into the multiscale factors that control the distribution of the world-class Ni and Au deposits of the eastern Yilgarn Craton.

### Nickel

The most significant recent advances in understanding komatiite-hosted Ni-sulfide deposits in the Yilgarn Craton (Barnes, 2006) have been the recognition of: (1) different modes of emplacement, sulfur assimilation, and the difference in resultant mineralisation styles; and (2) large-scale architectural control on the location of Ni-sulfide deposit clusters. This new understanding has come from work on a number of major Ni deposits in the Eastern Goldfields such as Mount Keith (e.g., Fiorentini et al., 2007), Black Swan (e.g., Barnes, 2004) and Kambalda (e.g., Gresham and Loftus-Hills, 1981; Lesher, 1983). The Kambalda deposits were the first discovered and have the longest history of both exploitation and research. They provide a useful illustration of how factors at all scales affect the accumulation of major Ni deposits in this geological province.

Nickel-sulfide mineralisation was first discovered at Kambalda (Figure 3) in 1966, c. 70 years after the discovery of Au in the region. Mineralisation is hosted in the Kambalda dome, a doubly plunging anticline cored by granitic intrusions, which post-date the Ni-sulfide mineralisation. Mineralisation is primarily within the volcanic stratigraphy exposed along the flanks of the intrusions and occurs as discontinuous to semi-continuous lenticular bodies termed ‘ore shoots’ (Gresham and Loftus-Hills, 1981). The local stratigraphic, metamorphic and structural history at Kambalda closely reflects the regional history in the Kalgoorlie Terrane.

#### *Geodynamic setting of the Kambalda nickel deposits*

Begg et al. (2010) argued that fundamental lithosphere-scale architecture has a major influence on the distribution of magmatic-hosted Ni deposits and showed that most Ni deposits in the Eastern Goldfields are found along the boundaries of very early developed lithospheric blocks. The Nd model-age map of the Yilgarn Craton (Figure 2) shows the distribution of major Ni and Au deposits in the Eastern Goldfields in relation to the model ages.

The isotopic map can be interpreted as providing a snapshot of the lithospheric architecture at 2650 Ma. This map can be considered as a proxy to image major lithospheric discontinuities, which may have acted as active pathways for large volumes of hot, mantle-derived melt to reach upper crustal levels without undergoing any significant differentiation. In other words, steep colour gradients (interpreted as lithosphere-scale boundaries) in Figure 2 show areas where hot magmas were most likely focussed.

Highly mineralised komatiites are most abundant on the western side of the Eastern Goldfields Superterrane, along the boundary between the isotopically juvenile part of the Kalgoorlie Terrane and the older Youanmi Terrane (Figure 2). This boundary may represent a significant lithospheric discontinuity at 2700 Ma, along which high volumes of hot komatiites were emplaced and interacted with crustally derived sulfur (cf. Bekker et al., 2009) to generate giant Ni-sulfide ore systems.

#### *Nickel-sulfide mineralisation at Kambalda*

The complete mafic–ultramafic stratigraphy of the Kalgoorlie Terrane is exposed at Kambalda, including the tholeiitic Lunnon Basalt (Figure 4). Nickel mineralisation is hosted in the Silver Lake Member of the Kambalda Komatiite. Thin, sulfidic sedimentary units, comprising dominant pale siliceous sediments, dark, carbonaceous slaty sediments, and minor mafic sediments occur throughout the mafic–ultramafic succession, but are most abundant within the Silver Lake Member of the Kambalda Komatiite (Bavinton, 1981; Gresham and Loftus-Hills, 1981). These sediments may have been a source of sulfur for the mineralisation (e.g., Lesher and Campbell, 1993).

Basal contact mineralisation between the ultramafic flows of the Kambalda Komatiite and the underlying Lunnon Basalt occurs within troughs or channels in the top of the footwall basalts (Lesher, 1983). These troughs or channels have been interpreted as primary features formed through thermal-mechanical erosion by flowing ultramafic lavas that cut down into the sediments overlying the pillowed basalts of the Lunnon Basalt (Lesher, 1983; Beresford et al., 2005). Alternatively, troughs could have formed along pre-existing faults with syn-eruptive graben development (Connors et al., 2002), or during subsequent deformation of the greenstone belt (Stone and Archibald, 2004; Stone et al., 2005). A combination of mechanisms is most likely responsible for the current ore surface configuration.

Troughs in the Kambalda dome area vary in size, but are commonly narrow and elongate with lengths up to 2,300 m and widths <300 m (Gresham and Loftus-Hills, 1981). Mineralisation in major troughs is mainly continuous and occurs as small (20–130 m), elliptical orebodies in minor troughs. Stratiform, hanging-wall ore is spatially associated with the basal ore but stratigraphically higher, typically at the contact between first and second flow units and within 100 m of the komatiite–basalt contact (Gresham and Loftus-Hills, 1981). Some secondary orebodies have been produced by the remobilisation of sulfides into areas of dilation and lower tectonic pressure (e.g., fold hinges, fault dilation zones and shear zones) away from the primary accumulation site (Lesher and Keays, 2002).

The Kambalda Ni deposits illustrate how a series of scale-dependent processes, which are reflected in different datasets, have aligned to focus komatiitic magmas and nickel sulfide deposits from the craton to deposit scale.

### Gold

The Eastern Goldfields has produced more than 130 Moz of Au. While there are more than 20 deposits with >1 Moz of contained Au in the region, the Golden Mile at Kalgoorlie is unique in terms of its size and historical production of more than 50 Moz (Western Australia Department of Mines and Petroleum, 2011).

The three recent key advances in the understanding of the distribution of Au mineralisation in the Eastern Goldfields are: 1) the recognition of the influence of large-scale lithospheric architecture on the localisation of mineralisation; 2) the recognition of multiple sources of fluid involved in the deposit genesis, sparking a resurgence of intrusion-related mineralisation models; and 3) the recognition of multiple timings of mineralisation within single Au deposits or deposit clusters.

Other important advances include the recognition that at least some high-temperature deposits have been metamorphosed and the characterisation of large-scale footprints of Au-related hydrothermal alteration that can be potentially be mapped by combinations of spectral and lithogeochemical means.

### *Large-scale lithospheric architecture*

Blewett et al. (2010b) showed that the distribution of large Au deposits in the Eastern Goldfields is controlled by a favourable convergence of factors from the lithospheric to the deposit scale. A variety of approaches to examining deep-crustal structure, including potential field, magnetotelluric, seismic tomography and reflection seismic data in combination with regional isotopic data have allowed interpretation of deep-crustal-penetrating shear zones which link to structures identified in the upper crust (Blewett et al., 2010b). These fundamental structures, which may act as conduits for fluids, reflect the deep structures identified by Begg et al. (2010) as playing a major role in the distribution of nickel deposits.

### *Metal sources*

Recent paragenetic and analytical studies have indicated the involvement of three fluids in Au deposits of the Eastern Goldfields (Walshe et al., 2009): 1) a reduced and acid fluid, interpreted to be derived from the upper crust; 2) an oxidised fluid, interpreted as sourced from oxidised magmas; and 3) a reduced fluid, interpreted as sourced from the lower crust or mantle. Despite debate concerning the timing and genesis of the fluids, there is a mounting body of evidence that differing alteration signatures can be detected in regional datasets, including spectral and lithogeochemical data, when normalised to rock type. Recent work indicates that Au is deposited at gradients in mineralogy and chemistry visible in these datasets and that they provide the potential to map alteration systems and possible sites of deposition (Neumayr et al., 2008).

### *Timing and structural controls on gold mineralisation*

The distribution of gold mineralisation is structurally controlled, and the timing, style and reactivation of structures are major factors in determining the size of deposits. Blewett et al. (2010a) suggested that Au was deposited through most of the deformation history but that the major mineralisation took place after the $D_3$ deformation (Figure 5), with Vielreicher et al. (2010) demonstrating, via a variety of techniques, that the main Golden Mile mineralisation at Kalgoorlie took place at c. 2642 Ma.

Although the mineralisation event at Kalgoorlie was quite late with respect to the overall structural history of the region, various stages of the structural evolution were responsible for the creation of favourable sites for Au deposition. Weinberg et al. (2004) recognised that deviations on the Boulder–Lefroy shear zone, a major structure common to a number of large deposits in the Kalgoorlie district, provided a first-order focus for the concentration of mineralising fluids in the Kalgoorlie region. These deviations or jogs, which are spaced c.30 km apart, were developed during the sinistral transpression stage ($D_{4b}$; Figure 5) of Blewett et al. (2010a), probably at about the same time as the main mineralising event. Going back in time, the extensional event that is associated with granite doming and the development of the late basins ($D_3$; Figure 5) played a major role in the creation of sites favourable for Au deposition (Hall, 2007) through the development of suitable fluid pathways (Blewett et al., 2010a).

Going farther back in time, a significant recent advance in understanding Au mineral systems has been the recognition of the role that the very early structural architecture of greenstone belts plays on the clustering of Au deposits. Studies at St Ives near Kambalda (Figures 2 and 3) have shown that a structural architecture, established at the time of mafic–ultramafic volcanism at c. 2700 Ma, has been continually reactivated through c. 70 Myr, controlling the subsequent greenstone depositional events, all subsequent responses of the crust to deformation and the location of Au deposits (Miller et al., 2010).

## Summary

The Yilgarn Craton preserves evidence of the oldest crust on Earth, back to c. 4400 Ma. The earliest recognisable volcano-sedimentary greenstones were deposited after c. 3000 Ma. Isotopic data show that there have been several episodes of crust generation and recycling, the earliest of which are not recognised in the Eastern Goldfields Superterrane. Two major episodes of plume-related magmatism, at c. 2800 Ma and c. 2700 Ma, had major consequences for the development and evolution of the craton. The c. 2800 Ma event, the scale of which has only recently been recognised, produced huge mafic–ultramafic igneous complexes in the central part of the craton and was associated with rifting and break-up of the preserved eastern part. The c. 2700 Ma plume, which is responsible for the creation of the world-class nickel deposits of the Eastern Goldfields Superterrane, was focussed between the older cratonic blocks created at the time of the c. 2800 Ma break-up. The arc-like volcano-sedimentary successions in the Kalgoorlie and Kurnalpi terranes of the Eastern Goldfields Superterrane probably formed as result of the re-assembly of the older crustal blocks (Czarnota et al., 2010). Deeply penetrating, extensional structures which developed at this time allowed large-scale fluid fluxes and provided loci for the deposition of Au deposits (Blewett et al., 2010a).

Taken together, the advances in understanding craton-scale control on deposit clusters, and emplacement controls on deposit style and mineralisation potential can be used as scale-dependent proxies for exploration (McCuaig et al., 2010). The isotopic datasets can be used as proxies for identifying regions with high potential magma and fluid flux, whereas stratigraphic components such as felsic volcanic and associated sediments, mappable komatiite thickness and presence of assimilated volcanogenic sulfur may mark the position of rifts within these belts, and therefore regions with the highest potential for Ni sulfide- and VHMS- ore concentrations. Furthermore, these fundamental flaws in the crust appear to control the subsequent response of the crust to deformation and have again focussed fluids during late-cratonic Au mineralisation. An understanding of the local structural architecture and chemically receptive host rocks for hydrothermal mineralisation becomes important at the scale of clusters of deposits and individual deposits.

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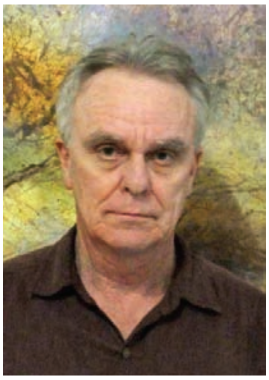

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# Early Earth evolution: evidence from the 3.5–1.8 Ga geological history of the Pilbara region of Western Australia

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*The Pilbara region of Western Australia is one of only two areas on Earth – the other being the Kaapvaal Craton of southern Africa – that contain well preserved, near-continuous geological records of crustal evolution from the Paleoarchean into the late Paleoproterozoic. The Pilbara is famous for hosting fossil evidence of early life (stromatolites and microfossils), and for containing a record of the early Archean atmosphere. The geological record extends from granite–greenstone terranes and overlying clastic basins of the 3.53–2.83 Ga Pilbara Craton, across a major unconformity, to a series of 2.78–1.79 Ga volcanic and sedimentary successions. Between 3.53–3.23 Ga, a succession of mantle plume events formed a thick volcanic plateau on older continental crust, remnants of which include enclaves of c. 3.6 Ga granitic gneiss and abundant 3.8–3.6 Ga inherited and detrital zircons. During each of the plume events, the volcanic plateau was intruded by crustally-derived granitic rocks, leading to vertical deformation by partial convective overturn. By 3.23 Ga, these processes had established thick continental crust that was then rifted into three microplates separated by c. 3.2 Ga basins of oceanic crust. Subsequent plate tectonic processes to 2.90 Ga included subduction, terrane accretion, and orogeny. From 2.78–2.63 Ga the northern Pilbara Craton was affected by minor rifting, followed by deposition of thick basaltic formations separated by felsic volcanic and sedimentary rocks (Fortescue Basin). Rifting in the southern Pilbara resulted in progressively deepening marginal basin sedimentation, including thick units of banded iron formation (Hamersley Basin: 2.63–2.45 Ga). At c. 2.45 Ga, sedimentation in the southern Pilbara changed to a mixed assemblage of clastic and carbonate sedimentary rocks of the Turee Creek Basin, including one unit of glacial diamictites. Deposition of the unconformably overlying 2.21–1.79 Ga Wyloo Group in the Ashburton Basin followed the Ophthalmian Orogeny, and all of these rocks were deformed by the Panhandle (c. 2 Ga) and Capricorn (c. 1.78 Ga) orogenies.*

## Introduction

The Pilbara region of Western Australia is one of only two areas in the world – the other being the Kaapvaal Craton of southern Africa – that provides extensive exposures of well preserved Paleoarchean–late Paleoproterozoic crust from which extensive information about the evolution of crustal processes, the biosphere and atmosphere has been obtained. The Pilbara contains three major Archean–Paleoproterozoic tectonic divisions: (1), the Pilbara Craton, composed of early crust (3.80–3.53 Ga), granite-greenstone terranes (3.53–3.07 Ga), volcanosedimentary basins (3.05–2.93 Ga), and post-orogenic granites (2.89–2.83 Ga); (2), the Fortescue, Hamersley, and Turee Creek basins (2.78–2.42 Ga), composed of a thick succession of interbedded clastic and chemical sedimentary rocks and volcanic rocks; and (3), the Ashburton Basin (2.21–1.79 Ga), composed of the volcano-sedimentary Wyloo Group.

From 1990 until 2007, the first two of these divisions were combined as the 'Pilbara Craton' (Trendall, 1990) on the grounds that there had been a continuum of depositional and igneous events from Paleo- to Mesoarchean granite-greenstones (prior to 1990 referred to as the 'Pilbara Block' in the N Pilbara) to the Proterozoic Turee Creek Group. Accordingly, it was considered that 'tectonic stability' (defining establishment of a craton) had not been attained until 2.4 Ga. However, a major geological mapping project, conducted jointly by the Geological Survey of Western Australia and Geoscience Australia between 1995 and 2002 (Huston et al., 2002a), provided a much improved geological understanding of the craton (Van Kranendonk et al., 2002, 2007a; Huston et al., 2002b). Van Kranendonk et al. (2006) revised the lithostratigraphy and tectonic units of the Paleoarchean and Mesoarchean rocks, and used the name 'Pilbara Craton' to encompass only units older than the Fortescue Group (the lowermost stratigraphic group of the Hamersley Basin). Hickman et al. (2006) explained that the change had been made because the last major deformation event to affect the granite-greenstones of the craton occurred at 2.90 Ga, after which there had been a c.130 Myr period of crustal stability (apart from the intrusion of post-orogenic granites). In 2006, the Fortescue, Hamersley, and

Turee Creek groups of the Mount Bruce Supergroup were still included within the 'Hamersley Basin', but Tyler and Hocking (2008) ascribed the succession to the Fortescue, Hamersley, and Turee Creek Basins (see also Hickman et al., 2010). The status of the Ashburton Basin has not changed since it was described by Thorne and Seymour (1991).

# Pilbara Craton

Regional gravity and magnetic data indicate that the Pilbara Craton underlies 250,000 $km^2$ of the Pilbara region (Hickman, 2004), but apart from a 60,000 $km^2$ area in the northern Pilbara, it is largely concealed by unconformably overlying Neoarchean–Paleoproterozoic rocks of the Fortescue, Hamersley and Turee Creek basins (Figure 1). The tectonic subdivision of the northern Pilbara Craton was established by Van Kranendonk et al. (2006), and revised by Hickman et al. (2010).

## Overview of tectonic evolution

The oldest part of the Pilbara Craton is 3.80–3.53 Ga crust, which has been identified by geochronology in rare outcrops of gneissic granite and gabbroic anorthosite. Geochemical and geochronological evidence shows that the early crust was present and widely exposed throughout the evolution of the craton.

Between 3.53–3.23 Ga, mantle plume activity resulted in the deposition of at least eight successive volcanic cycles on the early crust. The resulting volcanic plateau is now exceptionally well preserved as the 15–20 km-thick Pilbara Supergroup. Large volumes of granitic magma were intruded during the same period, and thick

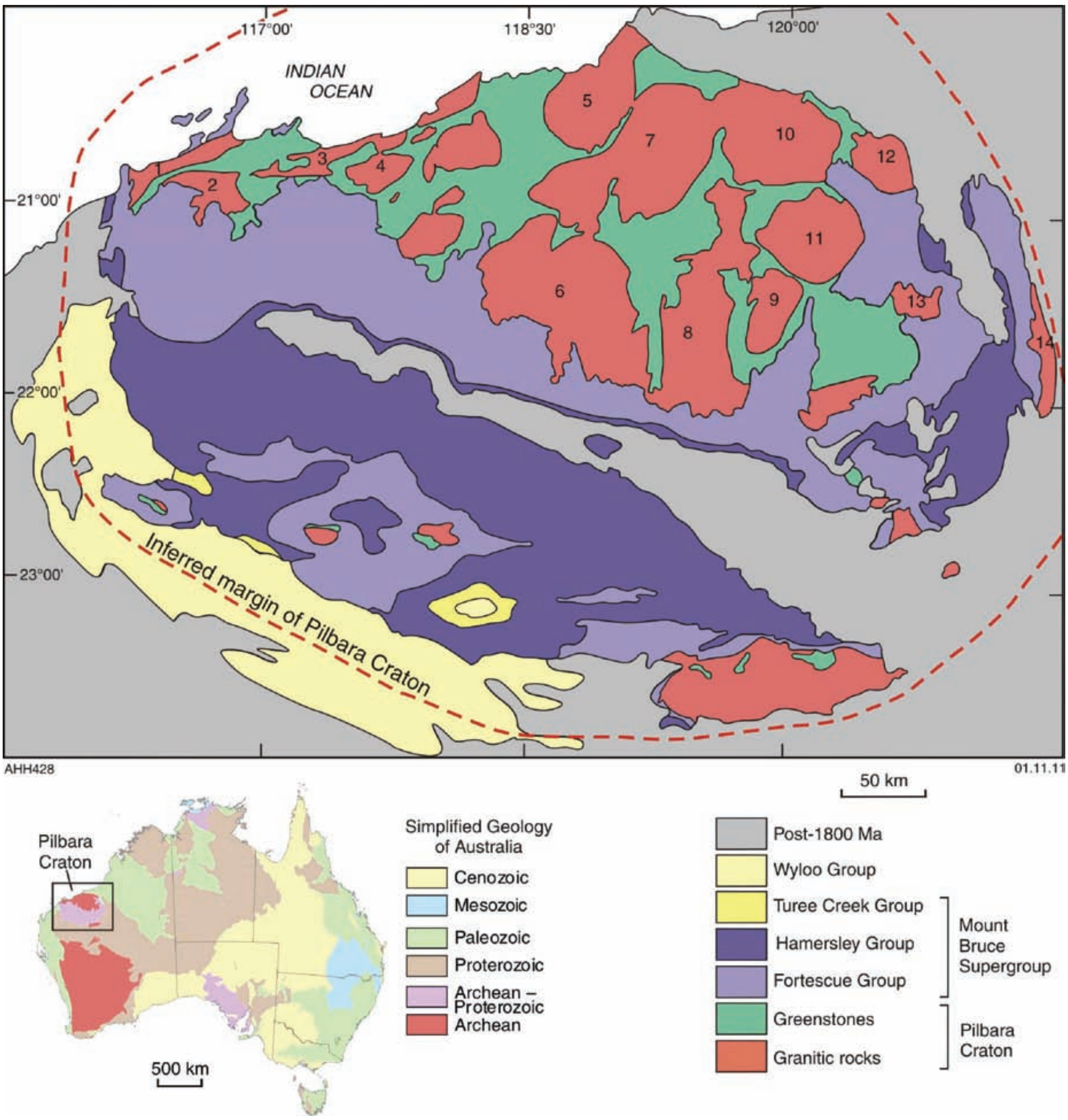


*Figure 1 Simplified geology of the Pilbara region of Western Australia, showing the interpreted margin of the Pilbara Craton beneath younger rocks. Granitic complexes: 1, Dampier; 2, Cherratta; 3, Harding; 4, Caines Well; 5, Pippingarra; 6; Yule; 7, Carlindi; 8, Shaw;9, Corunna Downs; 10, Muccan; 11, Mount Edgar; 12, Warrawagine; 13, Yilgalong; 14; Gregory (after Thorne and Trendall, 2001).*

continental crust had been established by 3.23 Ga. A major rifting event between 3.23–3.16 Ga split this crust into three continental microplates (Karratha, East Pilbara and Kurrana terranes; Figure 2) separated by two NE-trending basins of oceanic crust. Early in the rifting stage, clastic deposition on the edges of the microplates formed passive margin successions, including that of the Soanesville Basin, which is the best preserved representative and unconformably overlies the northwestern side of the East Pilbara Terrane. Later in the rifting, thick successions of pillow basalt and komatiitic basalt were erupted, and the underlying crust was intruded by ultramafic and mafic dykes and sills of the 3.18 Ga Dalton Suite.

Between 3.16–3.07 Ga, plate convergence caused part of the oceanic crust of the northwest basin to be obducted across the Karratha Terrane, establishing the Regal Terrane. Further SE, a subduction zone and intra-oceanic arc (Sholl Terrane; Figure 2) existed between 3.13–3.11 Ga. At 3.07 Ga, the NW-SE convergence responsible for the formation of the Sholl and Regal terranes culminated in the accretion of the Karratha, Sholl and Regal terranes to form the West Pilbara Superterrane. At the same time, collision of this superterrane with the East Pilbara Terrane resulted in major deformation and widespread granitic intrusion of the Prinsep Orogeny.

Crustal relaxation, extension, and moderate subsidence followed the Prinsep Orogeny and led to the development of the 3.05–2.93 Ga De Grey Superbasin across most of the northern Pilbara Craton. The De Grey Superbasin unconformably overlies the East Pilbara Terrane and the West Pilbara Superterrane. In the W Pilbara it is composed of three unconformity-bound basins: the basal 3.05–3.02 Ga Gorge Creek Basin (banded iron formation and clastic sediments); the 3.01–2.99 Ga Whim Creek Basin (volcanics); and the 2.97–2.94 Ga Mallina Basin (sandstone, wacke, and shale). The Mallina Basin records a history of alternating extension and compression during which there were periods of sedimentation, volcanism, and both felsic and mafic magmatic intrusion. In the SE Pilbara, the 2.98–2.93 Ga Mosquito Creek Basin is lithologically similar to the Mallina Basin but does not contain contemporaneous igneous rocks. Deposition in both basins was terminated by major deformation (North Pilbara and Mosquito Creek orogenies) between 2.94–2.90 Ga. Except for post-orogenic intrusion of highly fractionated

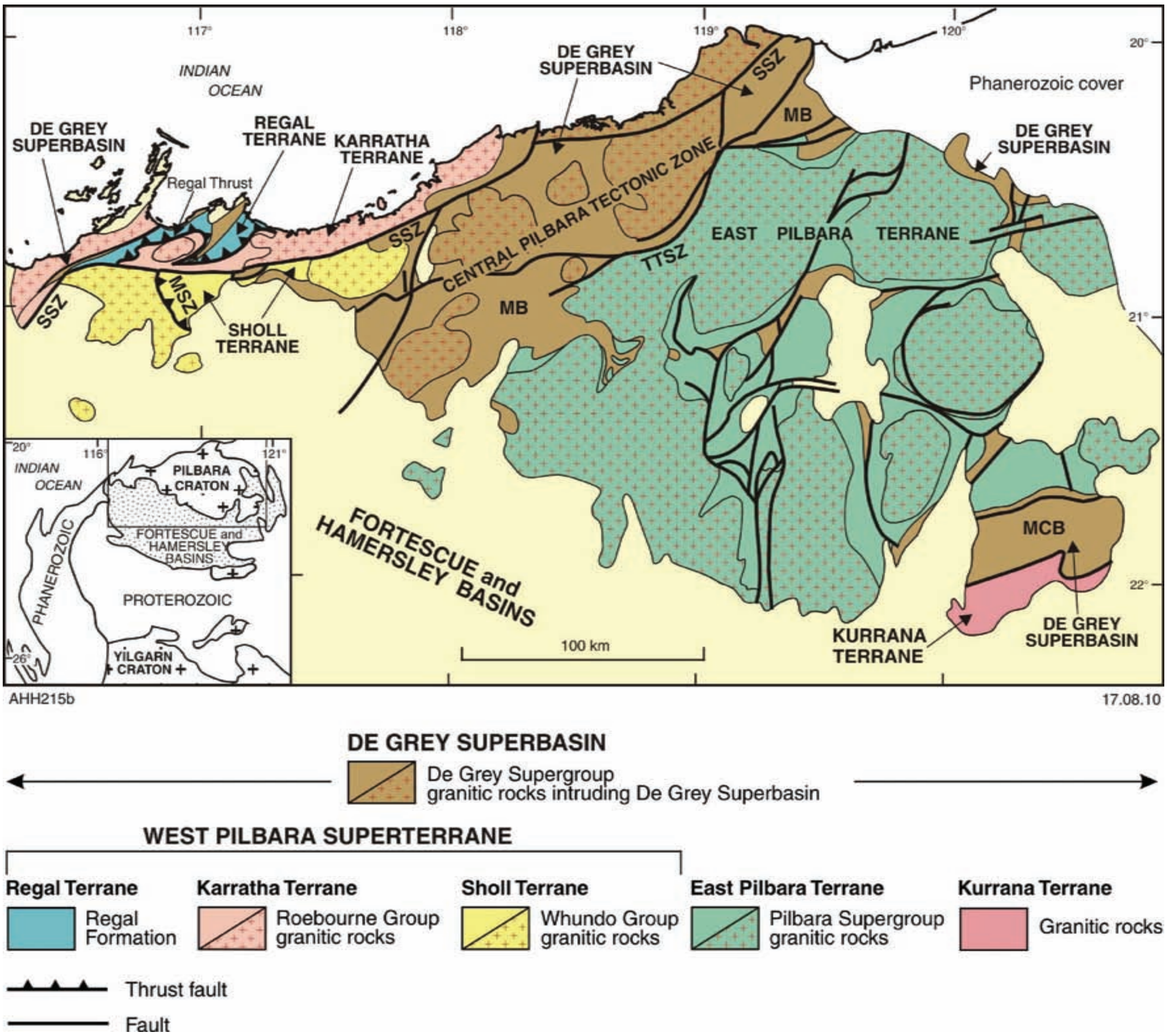


***Figure 2** Simplified geology of the northern Pilbara Craton, showing terranes and the De Grey Superbasin. MB, Mallina Basin; MCB, Mosquito Creek Basin; SSZ, Sholl Shear Zone; MSZ, Maitland Shear Zone; TTSZ, Tabba Tabba Shear Zone.*

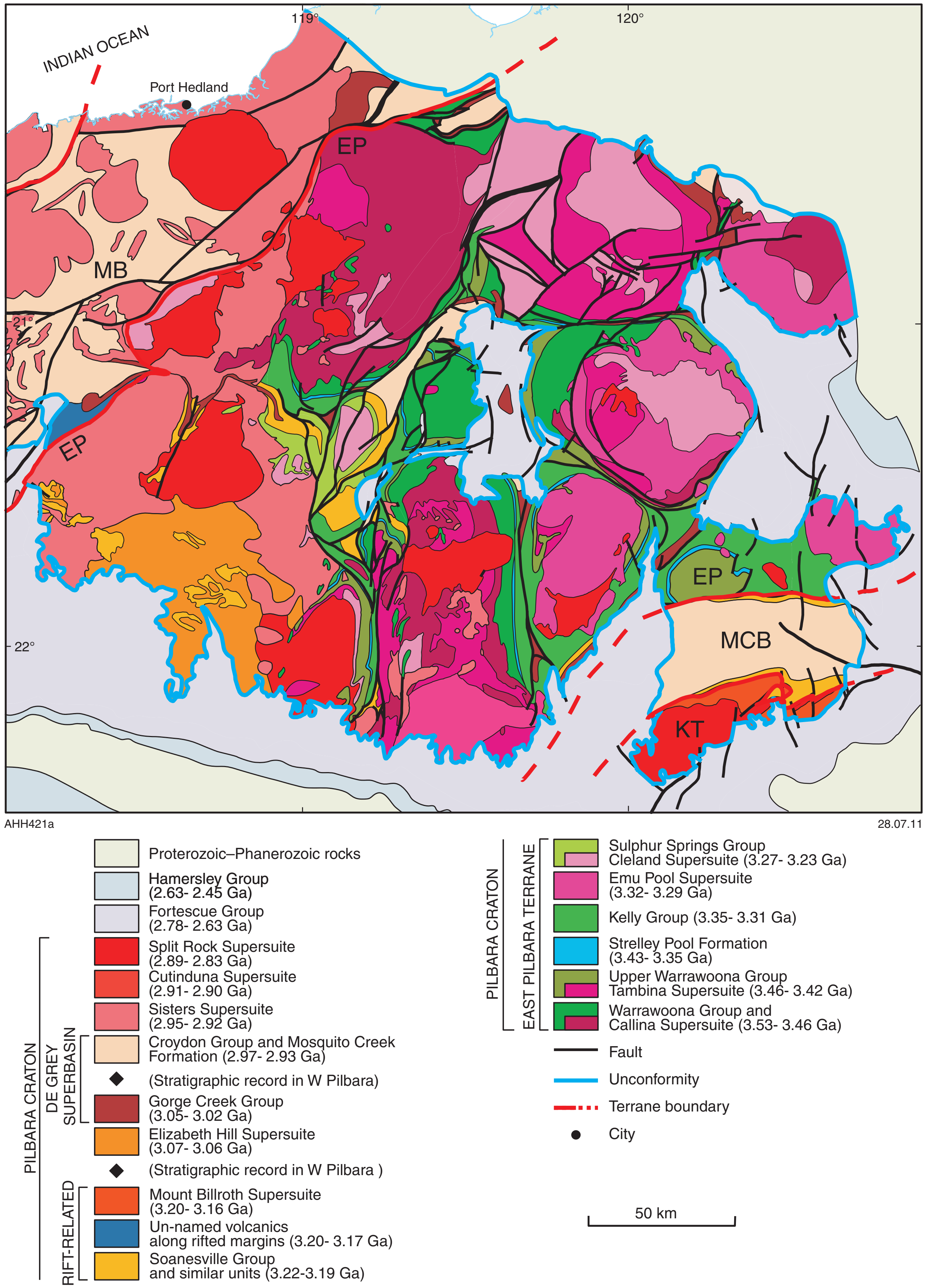


*Figure 3 Geology of the East Pilbara Terrane and adjacent tectonic units, showing all greenstone belts except those W of 118°45'E.*

granites between 2.89–2.83 Ga (Split Rock Supersuite) the craton remained stable until c. 2.78 Ga.

## Early crust (3.80–3.53 Ga)

The Warrawagine Granitic Complex (Figure 1) includes enclaves of 3.66–3.58 Ga biotite tonalite gneiss within younger granodiorite and monzogranite. Farther W, in the Shaw Granitic Complex, xenoliths of 3.58 Ga gabbroic anorthosite (McNaughton et al., 1988) occur within 3.43 Ga granitic rocks. Many Paleoarchean and Mesoarchean siliciclastic formations contain abundant 3.8–3.6 Ga detrital zircons, indicating erosion of crust up to 300 Myr older than the East Pilbara Terrane (Hickman et al., 2010). The pre-3.53 Ga crust or mantle sources for rocks of the East Pilbara Terrane is also supported by Nd-isotopic data (Jahn et al., 1981; Gruau et al., 1987; Bickle et al., 1989; Van Kranendonk et al., 2007a, b; Tessalina et al., 2010). Champion and Smithies (2007) provided evidence that most pre-3.3 Ga granites of the Pilbara Craton were sourced through infracrustal melting of material that was older than 3.5 Ga.

## East Pilbara Terrane (3.53–3.23 Ga)

The 3.53–3.23 Ga East Pilbara Terrane (Figures 2 and 3) provides the world's most complete record of Paleoarchean crustal evolution. Stratigraphy, structure, geochronology, and geochemistry collectively testify that the evolution of this terrane was dominated by volcanism, magmatic intrusion, and deformation during repeated episodes of heating and melting of underlying older crust (including felsic crust) and mantle over 300 Myr (Van Kranendonk et al., 2002, 2007a, b; Smithies et al., 2005b).

The 3.53–3.23 Ga Pilbara Supergroup of the East Pilbara Terrane is predominantly volcanic (Figure 4) and 15–20 km thick. Thick partial sections of this succession are recognized in almost all greenstone belts of the terrane, except in the NW where younger greenstones are preserved. The three component groups (Warrawoona, Kelly, and Sulphur Springs) of the Pilbara Supergroup are separated by two major erosional unconformities (Figure 4; Buick et al., 1995, 2002; Van Kranendonk et al., 2002). The time gap between the Warrawoona and Kelly groups was c. 75 Myr (3.427–3.350 Ga), and the gap between the Kelly and Sulphur Springs groups was c. 60 Myr (3.315–3.255 Ga). In both cases, the long pause in volcanic activity was preceded by deformation and metamorphism, and accompanied by subaerial erosion and deposition of clastic sediments. Shallow-water sediments between the Warrawoona and Kelly groups are preserved as the Strelley Pool Formation (up to 1 km thick; Van Kranendonk, 2010a), whereas siliciclastic rocks of the Leilira Formation (up to 3.9 km thick; Van Kranendonk and Morant, 1998) occur at the base of the Sulphur Springs Group.

Where best preserved, the Pilbara Supergroup is composed of eight ultramafic-mafic-felsic volcanic cycles (Figure 4). Geochronology on the felsic formations of successive cycles, and on contemporaneous granitic intrusions, some of which are subvolcanic (Hickman, 2012), indicates that most of the cycles spanned no more than 10–15 Myr; these cycles are interpreted to have resulted from successive mantle plume events (e.g., Arndt et al., 2001; Van Kranendonk et al., 2007a, b; Smithies et al., 2005b).

A characteristic feature of the terrane is the regional outcrop pattern of granitic domes separated by arcuate belts of volcanosedimentary rocks (greenstones) visible on geological maps (Figures 1, 2 and 3) and satellite imagery that has been variously described as "dome-and-syncline", "dome-and-basin", or "dome-and-keel" structure. Whereas some previous workers have interpreted this pattern as either the result of cross-folding (e.g., Blewett, 2002) or core complex formation (Zegers et al., 1996; Kloppenburg et al., 2001), neither of these are consistent with the full set of geological and geophysical features of the region (Hickman and Van Kranendonk, 2004). The most widely accepted interpretation is that the dome-and-keel structure arose from punctuated episodes of doming by partial convective overturn of the thick, dense greenstone succession into a partially molten granitic middle crust from 3.46–2.94 Ga (e.g., Hickman, 1975, 1984; Collins et al., 1998; Hickman and Van Kranendonk, 2004; Van Kranendonk et al., 2004, 2007a, b).

Geochemical evidence supporting growth of the East Pilbara Terrane as a thick volcanic plateau has been presented and reviewed in several papers (Smithies et al., 2003, 2005b, 2007a, b; Van Kranendonk et al., 2006, 2007a, b). Detailed mapping, geochronology, and geochemical traverses have established that the 15–20 km thickness of the Pilbara Supergroup is autochthonous and not the result of tectonic duplication by folding or thrusting as proposed by some workers (e.g., Bickle et al., 1980, 1985; Boulter et al., 1987; van Haaften and White, 1998). Smithies et al. (2009) noted that c. 3.52 Ga basalts and andesites, well exposed near the base of the Pilbara Supergroup, are enriched in K, LILE (large-ion lithophile elements), Th, and LREE relative to typical Archean basalt. They concluded that these 3.52 Ga rocks were derived from a mantle source enriched in felsic crustal components by previous episodes of crustal recycling. They also suggested that these 3.52 Ga enriched volcanic rocks, or chemically similar crustal material of similar age, were most likely the source for Paleoarchean tonalite-trondhjemite-granodiorite (TTG) intrusions in the East Pilbara Terrane.

Previous workers have interpreted a variety of tectonic settings for parts of the East Pilbara Terrane, including mid-ocean ridge (Ueno et al., 2001; Komiya et al., 2002; Kato and Nakamura, 2003), oceanic island arc (Komiya et al., 2002) and convergent margins involving continental magmatic arcs above subduction zones (e.g., Bickle et al., 1983, 1993; Barley et al., 1984). However, all of these settings are invalidated by the current data. For example, not only is the thickness of the succession too great for it to represent oceanic crust, but the presence of thick, repeated felsic volcanic and shallow-water sedimentary formations all preclude a mid-ocean ridge origin. Oceanic island arc settings for the felsic volcanic formations are not supported by the geochemical evidence (e.g. Smithies et al., 2007a, b) and geochronological evidence for considerably older (c. 300 Myr. older) crust underlying the Pilbara Supergroup. Interpretations that the felsic volcanics or granitic rocks of the terrane can be explained by convergent margin settings are not supported by regional stratigraphy and structure (Hickman, 2012), or by geochemistry which shows the derivation of contemporaneous felsic volcanic and granitic rocks from the fractionation of tholeiitic parental magmas and from melting across a range of depths within the crust (Van Kranendonk et al., 2007a; Champion and Smithies, 2007; Smithies et al., 2007a, b).

## Karratha and Kurrana Terranes (? 3.53–3.16 Ga)

From 3.23 Ga onwards, the cyclic volcanism of the Pilbara Supergroup ceased, and the crustal evolution of the craton became dominated by Phanerozoic-style plate tectonic processes (Hickman, 2004; Van Kranendonk et al., 2006, 2007a, b, 2010). Crustal extension

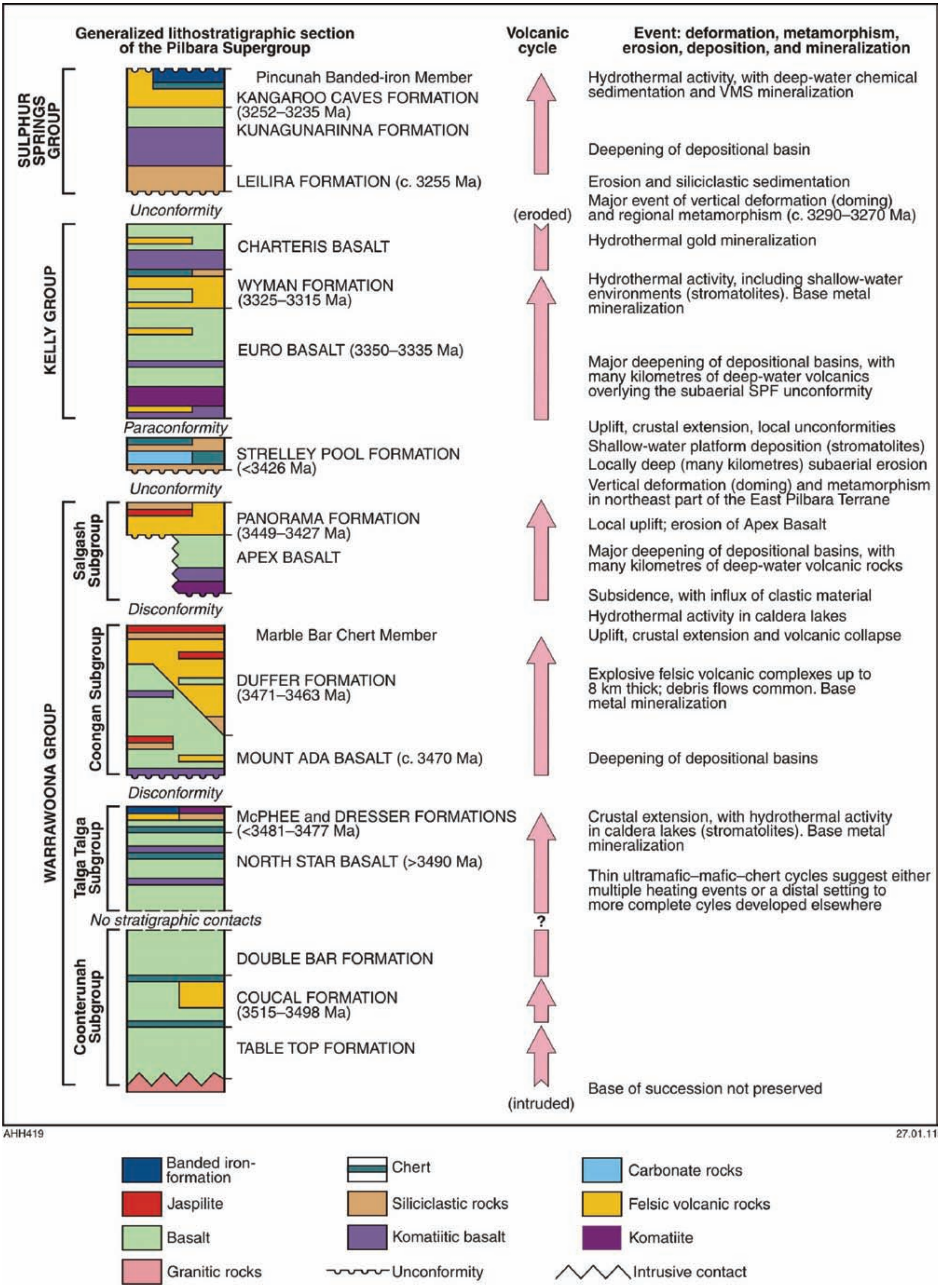


***Figure 4 Generalised stratigraphy of the predominantly volcanic Pilbara Supergroup, which is represented in almost all greenstone belts of the East Pilbara Terrane (after Hickman, 2011).***

established two NE-trending zones of rifting and crustal thinning across the East Pilbara Terrane. With continued spreading, these zones developed into rift basins containing c. 3.2 Ga sedimentary and volcanic passive margin successions and oceanic crust. The progenitor to the East Pilbara Terrane became divided into three separate segments of continental crust: a NW microplate, the Karratha Terrane; a remaining central segment of the East Pilbara Terrane; and a SE microplate, the Kurrana Terrane. The Karratha and Kurrana terranes are interpreted to be fragments of the East Pilbara Terrane based on common early histories back to at least 3.45 Ga.

The main components of the Karratha Terrane are the undated Ruth Well Formation and the intrusive 3.27–3.26 Ga Karratha Granodiorite (Smith et al., 1998). The Ruth Well Formation is 2 km thick and consists of metamorphosed ultramafic and mafic volcanic rocks (peridotite to tholeiitic basalt) with a few thin chert intercalations. The overlying Nikol River Formation includes felsic volcaniclastic rocks dated at 3.27–3.25 Ga. Nd $T_{DM}$ model ages of 3.48–3.43 Ga for the Karratha Granodiorite indicate that magma generation involved older crust or enriched lithospheric mantle (Sun and Hickman, 1998).

In the SE Pilbara, the present exposure of the Kurrana Terrane represents only a small fraction of its total extent, most of which is concealed by the Fortescue and Hamersley basins. Rift-related 3.20–3.16 Ga granitic rocks of the Mount Billroth Supersuite are widespread in this terrane and intrude undated, strongly deformed supracrustal rocks. Evidence of Paleoarchean crust is provided by xenocrystic zircons ranging in age between 3.58–3.46 Ga, and by Nd $T_{DM}$ model ages between 3.45–3.41 Ga (Van Kranendonk et al., 2007b). Mafic and ultramafic sills intruded along the northern boundary of the terrane have not been dated, but are likely to be 3.18 Ga intrusions of the Dalton Suite.

## Soanesville Basin, and similar units (3.20–3.165 Ga)

Subsidence along the rifted margins of the three separating continental microplates was accompanied by the deposition of shallow-water clastic successions and thick volcanic piles, in passive margin settings. The best preserved succession is that of the Soanesville Group on the NW margin of the East Pilbara Terrane (Figure 3; Van Kranendonk et al., 2006, 2010; Hickman et al., 2010). Similar successions are the Budjan Creek Formation on the SE side of the East Pilbara Terrane, the Coondamar Formation on the NW margin of the Kurrana Terrane, and the Nickol River Formation on the Karratha Terrane (Hickman et al., 2010).

With increasing crustal extension, the rifted margins were intruded by dykes and sills of dolerite, gabbro, and ultramafic rocks of the Dalton Suite (3.18 Ga), and thick pillow basalt was deposited above the clastic sediments (Van Kranendonk et al., 2010). The Dalton Suite also intruded the greenstone belts of the East Pilbara Terrane. Between 3.19 and 3.16 Ga the rifted margins were locally intruded by granitic intrusions of the Mount Billroth Supersuite (Van Kranendonk et al., 2006).

## Regal Terrane

From 3.20 Ga onwards, NW-SE extension between the East Pilbara and Karratha Terranes opened up a basin floored by oceanic crust (Hickman, 2004). Evidence for this basin is mainly geochronological and geochemical (Ohta et al., 1996; Sun and Hickman, 1998; Smithies et al., 2005a, 2007a), because most of it is interpreted to have been destroyed by subduction from 3.13–3.07 Ga. However, the c. 3.2 Ga basaltic Regal Formation is interpreted to be a remnant of this Mesoarchean oceanic crust, forming the Regal Terrane (Figures 2 and 5). The Regal Formation is a 2–3 km-thick sequence of metamorphosed pillow basalt, local basal komatiitic peridotite, and rare chert units that has a lithological composition and geochemistry consistent with oceanic crust (Ohta et al., 1996; Sun and Hickman 1998). The metabasalt has flat REE patterns, εNd of c. +3.5 (close to the depleted mantle value (3.2) at 3.20 Ga), and no geochemical evidence of crustal contamination (Smithies et al., 2007a). The Regal Terrane overlies the Karratha Terrane and the Nickol River Formation above a major zone of horizontal thrusting, the Regal Thrust (Hickman, 2001, 2004; Hickman et al., 2010). The present interpretation is that the Regal Formation was obducted onto the Karratha Terrane at some time between 3.16–3.07 Ga (Hickman, 2004; Hickman et al., 2010).

## Sholl Terrane

The fault-bounded Sholl Terrane (Figures 2 and 5) is composed of volcanic rocks of the 3.13–3.11 Ga Whundo Group, and granitic and mafic rocks of the contemporaneous Railway Supersuite. The Whundo Group is 10 km thick and consists of a lower volcanic package of calc-alkaline and boninite-like rocks, a middle package of tholeiitic rocks with minor boninite-like rocks and rhyolite, and an upper package of calc-alkaline rocks, including adakite, Mg-rich basalt, Nb-enriched basalt and rhyolite (Smithies et al., 2005a). Sedimentary rocks, which make up less than 1% of the group, include chert, banded iron formation (BIF), and quartzite. The basal contact of the group is the Maitland Shear Zone, which was originally a low-angle thrust above the Railway Supersuite, and the group is unconformably overlain by the 3.05–3.02 Ga Gorge Creek Group of the De Grey Supergroup (see below).

Nd-isotopic compositions indicate that during its deposition, the Whundo Group was not underlain by crust older than 3.25 Ga (Sun and Hickman, 1998). This evidence, combined with the geochemical features of the group and the fact that it is fault-bounded against crustal remnants with distinct histories, prompted Smithies et al. (2005a) to suggest an intra-oceanic arc origin for the Whundo Group at 3.13 Ga. Previous interpretations as a back arc setting (Krapez and Eisenlohr, 1998; Smith et al., 1998; Smith, 2003) are not supported, due to the lack of evidence for felsic basement, by the presence of boninites, and by low Th/La, La/Nb, and Ce/Yb ratios that are more consistent an intra-oceanic arc setting (Smithies et al., 2005a).

## West Pilbara Superterrane and Prinsep Orogeny

At c. 3.1 Ga, the Sholl Terrane was isolated from the Karratha and Regal terranes, and all three of these terranes developed separate histories from the East Pilbara Terrane. However, at 3.07 Ga the Sholl and East Pilbara terranes were each intruded by tonalite and granodiorite of the Elizabeth Hills Supersuite (Van Kranendonk et al., 2006). This intrusive event, present in both terranes, was accompanied by recumbent folding, thrusting, and metamorphism along the Regal Thrust, major sinistral strike-slip movement on the Sholl Shear Zone, thrusting of the Whundo Group across the Railway Supersuite, and recumbent folding and thrusting on the NW side of the East Pilbara Terrane (Hickman, 2001, 2004; Hickman et al., 2010; Van Kranendonk et al., 2010). This tectonomagmatic event, referred to as the Prinsep Orogeny, marks the accretion of the Karratha, Regal and Sholl terranes to form the West Pilbara Superterrane, and collision of this with the East Pilbara Terrane (Van Kranendonk et al., 2007a, 2010).

## De Grey Superbasin

The De Grey Superbasin unconformably overlies the East Pilbara Terrane and the West Pilbara Superterrane. It is composed of four basins: the Gorge Creek Basin (3.05–3.02 Ga), Whim Creek Basin (3.01–2.99 Ga), Mallina Basin (3.2–2.94 Ga), and the Mosquito Creek Basin (2.99–2.90 Ga).

The regionally extensive Gorge Creek Basin is composed of the Gorge Creek Group, which in most areas consists of basal conglomerate and sandstone overlain by a 1,000 m-thick unit of BIF, chert, and black shale (Cleaverville Formation). Deposition was initially in shallow-water, and included evaporite and fluviatile deposits (Sugitani et al., 1998). Deposition of the group followed widespread erosion after the 3.07 Ga Prinsep Orogeny, and was probably developed in response to post-orogenic crustal relaxation and subsidence.

The 3.01–2.99 Ga Whim Creek Basin is located immediately to the SE of the Sholl Shear Zone and contains the Whim Creek Group of volcanic, intrusive, and volcaniclastic rocks. Reactivation of this shear zone by N–S convergence between 3.01–3.00 Ga resulted in transpressional, tight to isoclinal folding of the Gorge Creek Group and metamorphism of adjacent parts of the Whundo Group, prior to deposition of the Whim Creek Group. The depositional setting of the Whim Creek Group was interpreted as a pull-apart basin by Barley (1987), but Pike and Cas (2002) suggested it formed in an ensialic back-arc basin. Smith (2003) considered that the TTG of the contemporaneous Maitland River Supersuite (3.00–2.98 Ga)

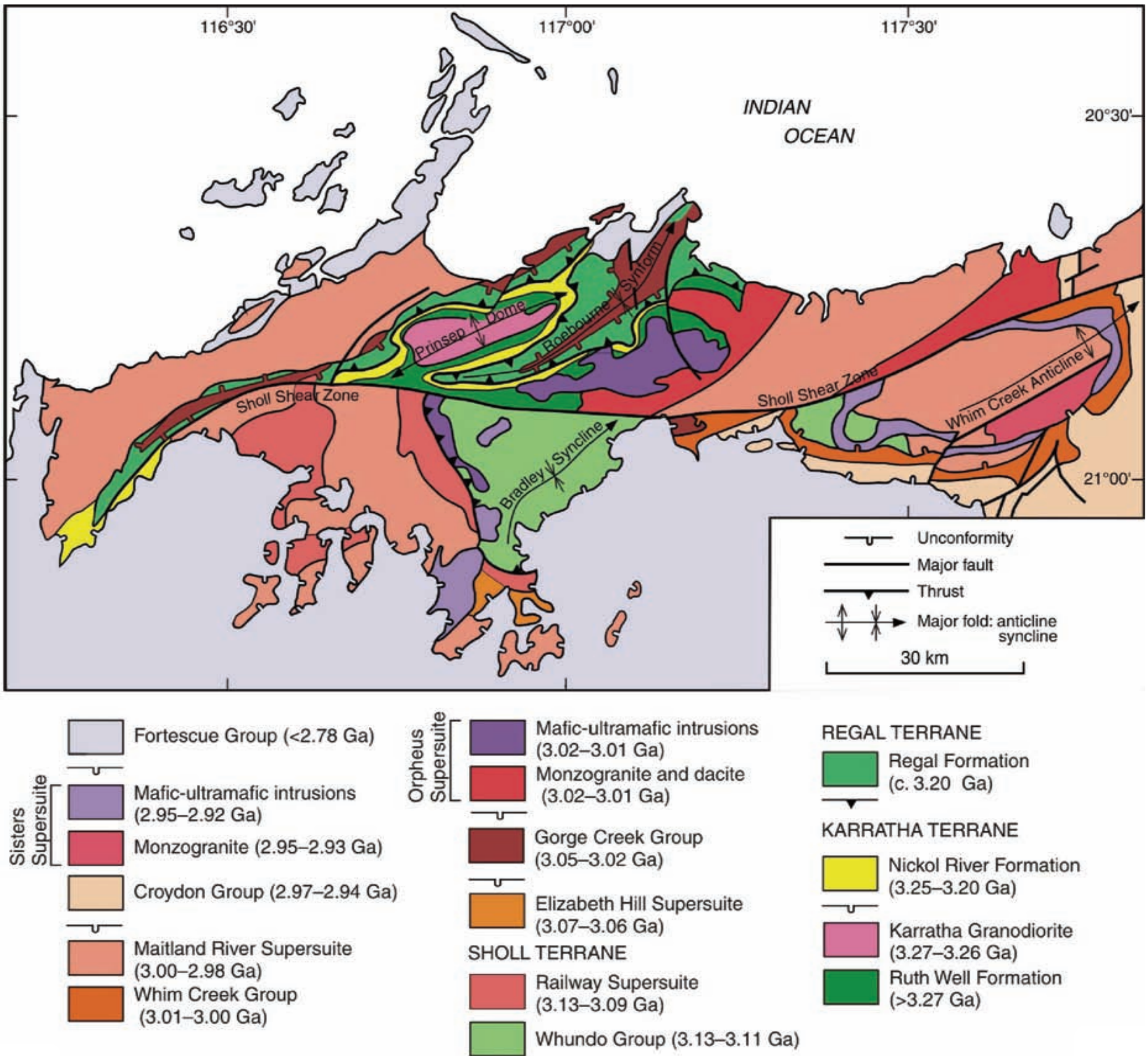


***Figure 5 Simplified geology of the NW Pilbara Craton, showing stratigraphy and major structures. The three terranes (Karratha, Regal, and Sholl) are separated by major faults, and each has an entirely different stratigraphy and tectonic history.***

represents the roots of a continental arc, and a subduction origin is supported by the geochemistry of basalt in the group, in particular its enrichment in Th and LREE, which is consistent with an enriched mantle source. However, such mantle enrichment is also present in magmas that later intruded the Mallina Basin, and these have not been related to subduction (Smithies et al. 2004). A within-plate extensional setting was interpreted by Van Kranendonk et al. (2007a).

The 250 x 100 km Mallina Basin (2.97–2.94 Ga) is filled by metasedimentary rocks of the Croydon Group and overlies the interpreted 3.07 Ga collision zone between the East Pilbara Terrane and the West Pilbara Superterrane. This group comprises the Bookingarra Formation (volcaniclastic and mafic-felsic volcanic rocks) and the Constantine Sandstone and Mallina Formation in the main part of the basin. The Mallina Formation is a 2–4 km thick succession of conglomerate, sandstone, and shale deposited in submarine fans (Eriksson, 1982). The stratigraphy and structural history of the basin records periods of extension and rift-related deposition of clastic sediments, intrusion of ultramafic-mafic sills, and minor basaltic magmatism, alternating with folding, strike-slip faulting, and thrusting during NW–SE compression. Alkali granites, monzogranites, and high-Mg diorites (sanukitoids) of the Sisters Supersuite intruded the basin between 2.95–2.94 Ga. (Smithies and Champion, 2000).

The Mosquito Creek Basin (Figure 2) originated at c. 3.2 Ga as an ENE-WSW-trending rift basin between the East Pilbara and Kurrana terranes. Juvenile basaltic crust is interpreted to have formed in the centre of the basin, but it was later overlain and concealed by clastic sediments of the Mosquito Creek Formation. Thickness estimates for the complexely folded Mosquito Creek Formation range between 1 and 5 km, and the maximum age of deposition of the formation is approximately 2.98 Ga (Nelson, 2004). Nijman et al. (2010) used sedimentological and structural observations to interpret the basin as a late-stage intramontane basin occupying a synclinal depression between domes of the East Pilbara Terrane. However, the northern and southern margins of the basin are in contact with 3.20–3.16 Ga intrusive units (Mount Billroth Supersuite and Dalton Suite) that in the NW Pilbara Craton are associated with rifting. Secondly, most detrital zircons in the Mosquito Creek Formation have crystallization ages inconsistent with derivation from the East Pilbara Terrane, suggesting a more distant source, perhaps by longitudinal currents within a large rift basin (Bagas et al., 2004).

### North Pilbara and Mosquito Creek orogenies

Two separate events of deformation and metamorphism affected the NW and SE parts of the Pilbara Craton between 2.95–2.90 Ga. In the NW Pilbara, large-scale NE-trending folds and faults associated with the North Pilbara Orogeny deformed the West Pilbara Superterrane between 2.95 – 2.92 Ga (Hickman, 2001). In the central Pilbara, the Mallina Basin was affected by regional transpression at this time, resulting in upright folding and strike-slip faulting (Krapez and Eisenlohr, 1998). In the central part of the East Pilbara Terrane, complex folding and shear deformation of the Lalla Rookh–Western Shaw structural corridor were formed under NW-SE compression at 2.93 Ga (Van Kranendonk and Collins, 1998; Van Kranendonk, 2008).

In the SE Pilbara, fold-thrust style deformation of the Mosquito Creek Formation at c.2.90 Ga (Huston et al., 2002b) during the Mosquito Creek Orogeny was the final result of N–S closure of the Mosquito Creek Basin as the Kurrana Terrane was accreted to the East Pilbara Terrane (Van Kranendonk et al., 2007a). Intrusion of monzogranite and lesser amounts of other granitic rocks accompanied both orogenies and, as with the deformation, was older in the NW (Sisters Supersuite, 2.95–2.92 Ga) than in the SE (Cutinduna Supersuite, 2.91–2.90 Ga).

### Split Rock Supersuite

The Split Rock Supersuite consists of highly fractionated, Sn–Ta–Li bearing, post-orogenic monzogranites that were emplaced in a SE–NW-trending linear belt across the Kurrana and East Pilbara terranes between 2.89–2.83 Ga (Van Kranendonk et al., 2007a). Nd model age data indicate derivation of the Split Rock Supersuite from partial melting of much older granitic crust, commonly with model ages of between 3.7–3.4 Ga (Bickle et al., 1989; Smithies et al., 2003).

## Fortescue, Hamersley and Turee Creek basins (2.78–2.42 Ga)

The Neoarchean–Paleoproterozoic successions of the Fortescue, Hamersley, Turee Creek basins unconformably overlie the Pilbara Craton and record a history that commenced with 2.78 Ga crustal extension and volcanic plateau volcanism (Fortescue Group), through passive margin settings (2.63–2.45 Ga Hamersley Group), to basin deposition (<2.45 to >2.21 Ga, Turee Creek Group), in advance of the Ophthalmian Orogeny (>2.21 Ga; Figure 6).

### Fortescue Basin

The Fortescue Basin is entirely composed of the Fortescue Group, a 6 km-thick, predominantly volcanic succession that unconformably overlies the Pilbara Craton across 250,000 $km^2$. In the N Pilbara, the Fortescue Group was deposited in four distinct stages spanning 150 Myr: (1) 2.78–2.77 Ga crustal extension, with local rifting and extrusion of basalt through a swarm of N–NE-trending dolerite dykes; (2) 2.77–2.75 Ga folding and faulting accompanied by rapid deposition of fluvial conglomerate and sandstone and lacustrine shale within shallow rift basins; (3) 2.75–2.71 Ga eruption of plume-related ultramafic–mafic–felsic volcanic cycles (Arndt et al., 2001) separated by volcaniclastic rocks and stromatolitic limestone of the 2.73–2.72 Ga Tumbiana Formation; (4) 2.71–2.63 Ga deposition of the Jeerinah Formation involving a northerly marine transgression with a basal near-shore facies of sandstone and stromatolitic limestone overlain by an off-shore facies of mudstone, carbonate rocks, carbonaceous pyritic shale, and chert (Thorne and Trendall, 2001). In some areas, clastic units up to several hundred metres thick, and possibly also basaltic units up to 2 km thick, underlie the Mt Roe Basalt of stage 1, and have been assigned to the Bellary Formation and an un-named formation (Van Kranendonk, 2010b). However, in the absence of geochronology, these local basal units could vary considerably in age.

In the SE Pilbara, stages 1 and 2 are not represented due to non-deposition across an elevated area of the underlying Pilbara Craton known as the Yule–Sylvania High (Thorne and Trendall, 2001). In the SW, stages 2 to 4 are in most cases represented by deeper-water facies than in the northern part of the basin. This southerly deepening

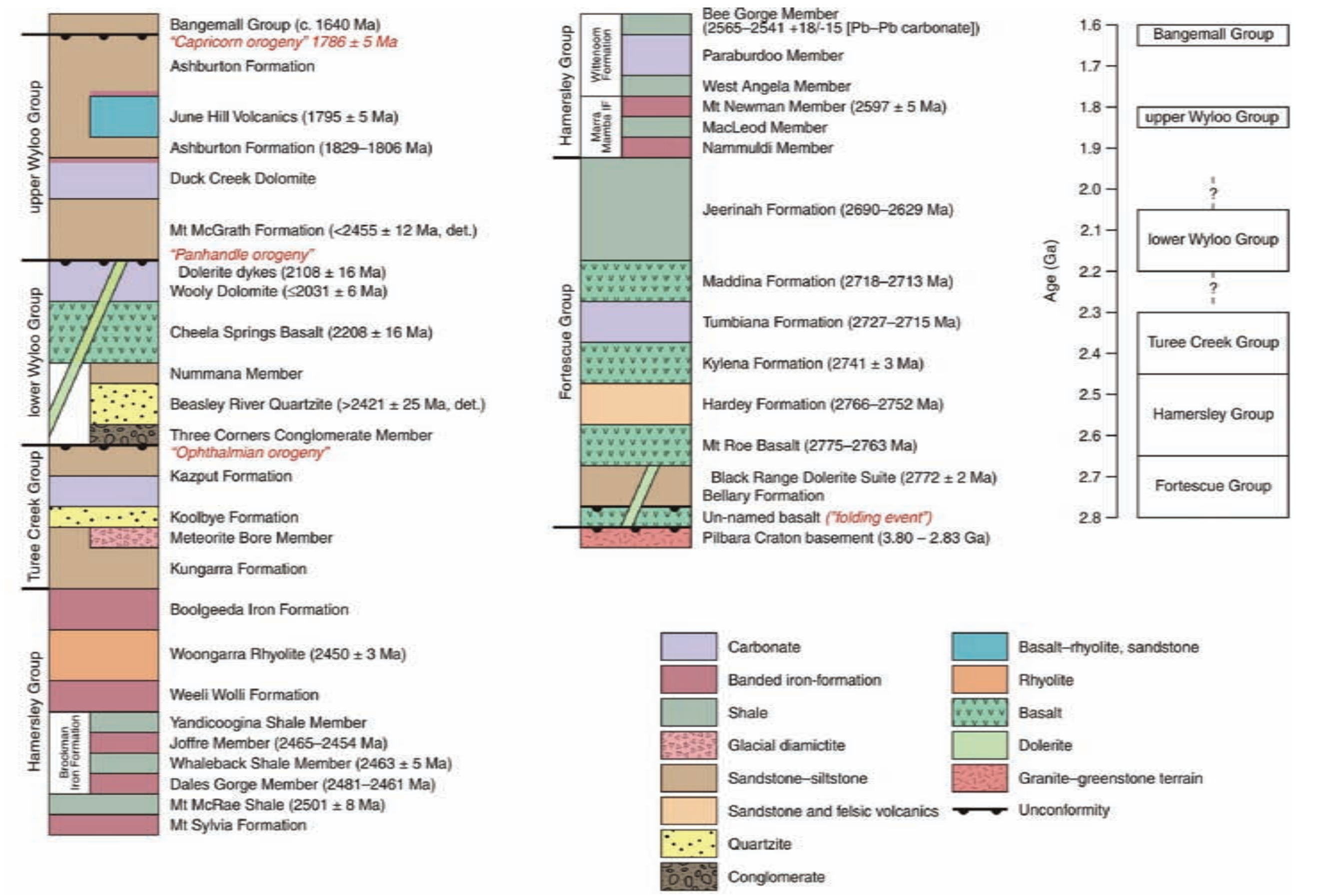


***Figure 6 Stratigraphic column of the Neoarchean–Paleoproterozoic succession of the Pilbara region, summarising lithology and geochronological data.***

of the Fortescue Basin was accompanied by a marked thickening of the succession in the S, consistent with a major rift axis to the SSW. In the S Pilbara, the Tumbiana Formation of stage 3 is represented by the Pyradie Formation, a 1 km-thick marine unit of komatiite, komatiitic basalt, hyaloclastite, argillite, and chert. In the S Pilbara, the Jeerinah Formation is much thicker than in the N, and consists of 'deeper' shelf deposits, including basalt, as well as being extensively intruded by dolerite sills.

## Hamersley Basin

The Hamersley Basin, which contains the BIF-dominated Hamersley Group (Trendall and Blockley, 1970), overlies most of the southern part of the Pilbara Craton, and its lower stratigraphy is also preserved close to the E and W margins of the craton (Figure 1). In the S and W, the basin stratigraphy represents continuation of the shelf subsidence in stage 4 of the Fortescue Basin, but in the E, deposition was in shallow water and there is no evidence that the central and upper parts of the Hamersley Group were ever deposited here, or across most of the N Pilbara. In the S Pilbara, the Hamersley Group is c. 2.5 km thick, lies conformably on rocks of the Fortescue Group, and includes four BIF-dominated formations (Marra Mamba, Brockman, and Boolgeeda Iron Formations, and the Weeli Wolli Formation) individually up to 360 m thick. In most cases, the BIF-dominated units are separated by shale, but the carbonate-dominated Wittenoom Formation separates the Marra Mamba and Brockman iron formations. Near the top of the group, the Woongarra Rhyolite has been interpreted by Trendall (1995) as an intrusive, sill-like unit, although it also has obvious extrusive portions.

## Turee Creek Basin

Conformably overlying the Hamersley Basin, close to the southern margin of the Pilbara Craton (Figure 1), the Turee Creek Basin consists largely of greyish-green siltstone, fine-grained greywacke, and fine-grained sandstone, but also contains thin–thick carbonate units and glacially-deposited conglomerates and diamictites (Trendall, 1981). Trendall (1979) identified a lower Kungarra Formation, including the Meteorite Bore Member of glacial diamictites that is up to 270 m thick at the type locality where it consists of diamictite, fine-grained sandstone and shale (Martin, 1999). Detrital zircon dating indicates a maximum depositional age of c. 2.42 Ga (Takehara et al., 2010). The Turee Creek Basin has been interpreted to reflect an accretionary, or collisional, foreland setting in front of the northward-advancing Ophthalmian Fold Belt (Martin and Morris, 2010).

## Ashburton Basin

The Wyloo Group (Figure 6) of the Ashburton Basin consists of low-grade sedimentary and volcanic rocks with a thickness of about 12 km (Thorne and Seymour, 1991). The group has been divided into unconformity-bound upper and lower parts, the lower part resting unconformably on the Hamersley Basin, and the upper part unconformably overlain by younger Proterozoic basins.

## Lower Wyloo Group

The lower part of the Wyloo Group consists of the basal Beasley River Quartzite (up to 360 m thick), which is conformably overlain by the Cheela Springs Basalt and the Wooly Dolomite (Figure 6). This succession lies with marked unconformity on rocks of the Turee Creek and Hamersley basins. The basal Three Corner Conglomerate Member of the Beasley River Quartzite contains abundant clasts of BIF, including pebbles of enriched ore derived from erosion of the Hamersley Group. The 2.21 Ga Cheela Springs Basalt is up to 2 km thick and consists of basalt derived from a subduction-modified source (Martin and Morris, 2010). The overlying 2.03 Ga Wooly Dolomite is a shallow-marine carbonate succession with locally abundant stromatolites. All of the rocks in the lower Wyloo Group are cut by dolerite dykes emplaced at 2.01 Ga.

## Upper Wyloo Group

Deposition of the upper Wyloo Group at 1.83–1.79 Ga included early passive rift deposits, pre- to syn-collisional volcanics (arc?), and younger syn-collisional deposits culminating with the Capricorn Orogeny. The basal Mount McGrath Formation unconformably overlies folded lower Wyloo Group and older rocks, and is composed of upward-fining deltaic and shallow-marine cycles composed of channelized conglomerate and siltstone (Thorne and Seymour, 1991). The conformably overlying Duck Creek Dolomite is up to 1 km thick and consists of repeated upward-shallowing sequences formed by transgressive and regressive sedimentation on a barred carbonate shoreline (Grey and Thorne, 1985). The formation includes slope, barrier-bar, lagoon, intertidal and supratidal facies, together with major developments of subtidal stromatolitic bioherms. The Duck Creek Dolomite is conformably overlain by relatively deep-water sedimentary and volcanic rocks of the >5 km-thick, 1.83–1.795 Ga, Ashburton Formation.

# Archean–Paleoproterozoic biosphere

Fossil occurrences in the Pilbara region are widespread and preserved in a variety of depositional environments (e.g., Buick and Dunlop, 1990; Brasier et al., 2002; Van Kranendonk, 2007; Hickman, 2012). Within the Pilbara Supergroup, stromatolites, microbial mats, and microfossils have been identified within thin sedimentary units that separate volcanic cycles. The most common habitat was hydrothermal, such as around hot springs or adjacent to hydrothermal vents in settings associated with the waning stages of volcanic activity; such environments were provided by the 3.48 Ga Dresser Formation (Walter et al., 1980; Buick and Dunlop, 1990; Philippot et al., 2007; Van Kranendonk et al., 2008) and the 3.24 Ga Kangaroo Caves Formation (Duck et al., 2007). However, in the case of the Strelley Pool Formation, the growth of stromatolites was prolific in a shallow-water carbonate shelf setting (Hofmann et al., 1999; Allwood et al., 2006). Evidence of Paleoarchean life may also lie in some of the BIF units of the succession because oxidising bacteria or photosynthesising microbiota could have caused BIF deposition through oxidation of dissolved $Fe^{2+}$ (e.g., Trendall and Blockley, 1970, 2004). The oldest known BIF in the Pilbara Craton is present within the c. 3.52 Ga Coucal Formation.

The Early Mesoarchean successions have so far yielded relatively little evidence of early life. In the passive margin deposits of the E Pilbara, Duck et al. (2007) identified structures resembling microbial mats near the base of the Soanesville Group. In the W Pilbara, Kiyokawa et al. (2006) reported various structures suggestive of microbial mats and filamentous organic remains from 3.19 Ga oceanic crust at the top of the Regal Formation. In the shallow-water sedimentary rocks at the base of the Late Mesoarchean De Grey Superbasin, microstructures like threads, films, hollow spheres and spindles are interpreted as remnants of microfossils and microbial mats (Sugitani et al. (2009) and within the overlying Cleaverville Formation of the W Pilbara, carbonaceous spheroids occur in black chert (Ueno et al., 2006).

The Neoarchean Fortescue Basin contains widespread stromatolites in lacustrine carbonate units of the Kylena Formation and Tumbiana Formation (Sakurai et al., 2005; Awramik and Buchheim, 2009). Microbial structures within Tumbiana Formation stromatolites were described by Walter (1983) and Lepot et al. (2008). Stromatolites also occur in shallow-water chert at the base of the Jeerinah Formation (Packer and Walter, 1986). Water depths in most formations of the Hamersley Basin were too great for stromatolites, but the shallow-water Carawine Dolomite of the E Pilbara is an exception and is abundantly stromatolitic. The Wooly and Duck Creek Dolomites of the Wyloo Group also contain abundant stromatolites.

# Archean–Paleoproterozoic atmosphere

It is generally accepted that there was a major increase in the $O_2$ content of Earth's atmosphere between 2.4–2.0 Ga (Farquhar et al., 2000), commonly referred to as the 'Great Oxidation Event' (GOE; Holland, 2002). However, some workers have argued either that the atmosphere was oxygenated as early as 4.0 Ga (Ohmoto et al., 2006), or that oxygen increased more gradually or fluctuated from the Paleoarchean to the Paleoproterozoic (e.g., Anbar et al., 2007).

Many workers have used sulfur isotopes to argue that atmospheric oxygen levels were very low prior to the GOE (e.g., Farquhar et al., 2000; Pavlov and Kasting, 2002), based on observations that Archean sedimentary rocks exhibit mass independent fractionation (MIF-S), whereas sedimentary rocks deposited after the GEO do not. Because MIF-S can originate through ultraviolet radiation of volcanic $SO_2$ in the absence of ozone or oxygen, they argued that the MIF-S data provide evidence that Earth's atmosphere changed from anoxic to oxic during the GOE. However, Ohmoto et al. (2005) determined $\delta^{33}S$ and $\delta^{34}S$ values on bulk-rock S from the 2.93 Ga Mosquito Creek Formation and reported an absence of MIF-S in all but one of the 40 samples – inconsistent with an Archean anoxic atmosphere. They also obtained the same result when they analysed shale from the 2.76 Ga Hardey Formation. Similarly, Kato et al. (2009) reported hematite pre-dating veins of non-oxidised 2.76 Ga pyrite in drill core through the 3.46–3.45 Ga Apex Basalt, and concluded that the hematite must have been formed by oxygenated groundwater prior to 2.76 Ga. Kaufman et al. (2007) analysed S isotopes in organic-rich shale and carbonate in the 2.50 Ga Mt McRae Shale and found evidence for oxygenation of the seawater. Anbar et al. (2007) found enrichment of Mo and Rh in the Mt McRae Shale, indicating oxidation of Archean crustal sulfide minerals.

Kirschvink and Kopp (2008) suggested that major glaciations would have contributed to the atmospheric rise in oxygen because the melting of glacial ice in seawater releases $O_2$ from the photochemically produced $H_2O_2$ in the ice. The oldest evidence of glaciation in the Pilbara is provided by the <2.45 but >2.21 Ga Meteorite Bore Member of the Turee Creek Group.

# Archean–Paleoproterozoic mineralisation

## Pilbara Craton

The recognition that the Pilbara Craton is composed of geologically distinct terranes and basins, formed in different tectonic environments, has explained the large variety of mineralisation styles across the craton. Huston et al. (2002b) reviewed mineralisation of the craton within the context of its crustal evolution. Mineralisation in the East Pilbara Terrane occurred during a sequence of magmatic pulses related to mantle plumes, and extended over almost 300 Myr. Deposit types associated with these events include: synvolcanic Cu-Zn-Pb-barite volcanic-hosted massive sulfide (VHMS) deposits; hydrothermal barite in the form of both quartz-barite veins and sediment-replacement bedded chert-barite deposits; polymetallic and base metal deposits in porphyritic felsic stocks; porphyry Cu-Mo mineralisation; and mesothermal Au deposits in shear zones around granitic domes. Several of the Paleoarchean ore deposits are the oldest of their type in the world: hydrothermal barite in the Dresser Formation (3.48 Ga); volcanic-hosted massive sulfides (Cu-Pb-Zn, with 20% barite) in the Duffer Formation (3.465 Ga); polymetallic (Cu-Pb-Zn-Au-Ag) mineralisation in a porphyritic felsic stock (3.45 Ga); porphyry Cu-Mo (3.31 Ga); epigenetic lode Au deposits as old as 3.40 Ga (Huston et al., 2002b).

The c. 3.2 Ga rift-related passive margin successions contain little known mineralisation other than Au in the Nickel River Formation. The Au has been mined from shear zones within the Regal Thrust Zone, and it is uncertain if the source of the Au is detrital or hydrothermal. Rift-related ultramafic-mafic intrusions of the Dalton Suite (3.18 Ga) contain Ni-Cu mineralisation, but grades are generally <1% Ni. The Regal Terrane (oceanic crust) is apparently unmineralised, but the Whundo Group of the Sholl Terrane (3.13–3.11 Ga volcanic arc succession) locally contains economic VHMS Cu-Zn mineralisation.

The De Grey Superbasin contains the most economically important mineralisation in the Pilbara Craton. Iron ore from various deposits within the Gorge Creek Basin (3.05–3.02 Ga) has been mined and exported for almost 50 years. The Croydon Group contains c. .2.95 Ga VHMS and sediment-hosted Pb-Zn and Cu deposits at several localities. These deposits were formed during the early stages of alternating extension and compression in the basin, as were Ni-Cu and V-Ti-magnetite deposits in ultramafic-mafic layered intrusions intruded into the basin between 2.95–2.92 Ga. Orogenic vein- and shear-hosted Au deposits were formed during closure of the Mallina Basin at about 2.92 Ga (North Pilbara Orogeny). In the SE Pilbara Craton, the Mosquito Creek Orogeny was accompanied by similar orogenic Au deposits between 2.93–2.90 Ga.

The final period of mineralisation in the Pilbara Craton occurred between 2.89–2.83 Ga with pegmatite-hosted Sn-Ta deposits around the margins of highly fractionated post-orogenic granites of the Split Rock Supersuite. Shear-hosted Au mineralisation occurred at 2.89 Ga in the Mount York area (NW part of the East Pilbara Terrane).

## Fortescue, Hamersley, and Turee Creek basins

The basal unconformity of Fortescue Basin is locally overlain by conglomerate-hosted Au mineralisation in areas where Neoarchean paleodrainage systems were eroding lode Au deposits in the underlying Pilbara Craton. An important lode Au deposit, Paulsens, has recently been developed in the lower Fortescue Group of the Wyloo Dome (SW Pilbara). Underground mining has reached a depth of 400 m, and current production is approximately 75,000 oz Au per annum. In the NE Pilbara, hydrothermal quartz veins have locally been mined for Pb, Ag, fluorite, Cu and V. BIFs of the Hamersley Basin, notably the Brockman and Marra Mamba iron formations, contain some of the world's largest deposits of Fe ore, owing to enrichment that has locally increased grades to approximately 60% Fe (Blockley, 1990).

## Ashburton Basin

Since the review of mineralisation in the Ashburton Basin by Thorne and Seymour (1991), significant Au deposits have been mined from quartz veins in the Ashburton Formation. Sener et al. (2005) dated the Mount Olympus deposit at 1.74 Ga, and interpreted it to represent orogenic Au mineralisation. Syngenetic stratiform Au-Ag mineralisation in the Ashburton Formation has been interpreted to represent submarine hot-spring deposits (Davy et al., 1991). Other mineralisation in the Ashburton Basin includes Cu, Pb, Zn and Ag along fault zones.

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*by Ian M. Tyler, Roger M. Hocking and Peter W. Haines*

# Geological evolution of the Kimberley region of Western Australia

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*The history of the Kimberley region in the far north of Western Australia began in the Paleoproterozoic with rifting along the North Australian Craton margin at 1910–1880 Ma, followed by plate collision as part of a series of 1870–1790 Ma events that formed the Diamantina Craton within the supercontinent Nuna. Collision involved the accretion of an intra-oceanic arc to a continent that included the Kimberley Craton before final collision and suturing with the North Australian Craton. The c. 1835 Ma Speewah Basin formed as a retro-arc foreland basin to the W. The post-orogenic, c. 1800 Ma shallow-marine to fluvial Kimberley Basin and its equivalents had a provenance to the N and extended across both the Lamboo and Hooper provinces. Subsequent late Paleoproterozoic and Mesoproterozoic basins formed broadly similar depositional settings during break-up and reassembly into the supercontinent Rodinia. The intracratonic Yampi Orogeny generated large-scale folding and thrusting and sinistral strike-slip faulting between 1400–1000 Ma.*

*The Neoproterozoic Centralian Superbasin formed as a broad intracratonic sag basin throughout central Australia between c. 830 Ma and the earliest Cambrian, including a series of basins across the Kimberley. Glacigene rocks are present with the most widespread being equivalent to the c. 610 Ma Elatina ("Marinoan") glaciation. Folding, thrusting and strike-slip faulting during the c. 560 Ma King Leopold Orogeny caused a widespread unconformity at the base of the Ord and Bonaparte basins marked by the c. 508 Ma Kalkarindji Continental Flood Basalt Province. In the Early Ordovician, thermal subsidence initiated the Canning Basin. Paleozoic sedimentary rocks, including Devonian reef complexes, were deposited on the Lennard Shelf and in the Fitzroy Trough. In the Halls Creek Orogen, Devonian sedimentary rocks were deposited in sub-basins of the Ord Basin during the c. 450–300 Ma Alice Springs Orogeny. A widespread glacigene succession followed in the Canning Basin, but by the early Triassic deposition was restricted and the remainder of the Mesozoic succession forms a veneer over much of the basin.*

## Introduction

The geological history of the Kimberley region in the far N of Western Australia (Figures 1 and 2) is complex and spans almost 2 billion years of earth history. A period of Paleoproterozoic plate collision during the assembly of the supercontinent Nuna produced crystalline basement rocks forming the Hooper and Lamboo provinces. Collision and suturing was followed through the rest of the Proterozoic and the Phanerozoic by repeated phases of sedimentary basin formation, orogenic deformation, and associated fault reactivation at upper crustal levels, first within Nuna, then Rodinia and finally Gondwana. Deposition essentially ceased in the Late Paleozoic, and in the Neogene the region began to bow downwards as Australia met the Indian plate.

## Paleoproterozoic plate collision (1910–1805 Ma)

The oldest exposed rocks in the Kimberley region are igneous and low- to high-grade meta-igneous and metasedimentary rocks that occur within the 1910–1790 Ma Lamboo and Hooper provinces (Figures 1 and 2). The Halls Creek Orogen (Figure 2) is a suture formed by plate collision between the southeastern margin of a continent including the Kimberley Craton (which underlies the Speewah and Kimberley basins) and the NW margin of the North Australian Craton, and the Halls Creek Orogeny was one of a series of linked collisional events that formed the Diamantina Craton as part of the supercontinent Nuna (Cawood and Korsch, 2008). Collision was complex, involving the accretion of the Tickalara arc to the Kimberley Craton during the 1870 –1850 Ma Hooper Orogeny (Figures 3 and 4), before suturing during the 1835–1805 Ma Halls Creek Orogeny (Blake et al., 2000; Griffin et al., 2000; Sheppard et al., 1999, 2001; Tyler, 2005).

### The Lamboo and Hooper Provinces

The Lamboo Province in the Halls Creek Orogen can be divided into three zones; the Western zone, the Central zone and the Eastern zone (Figure 2). The Hooper Province in the west Kimberley is an

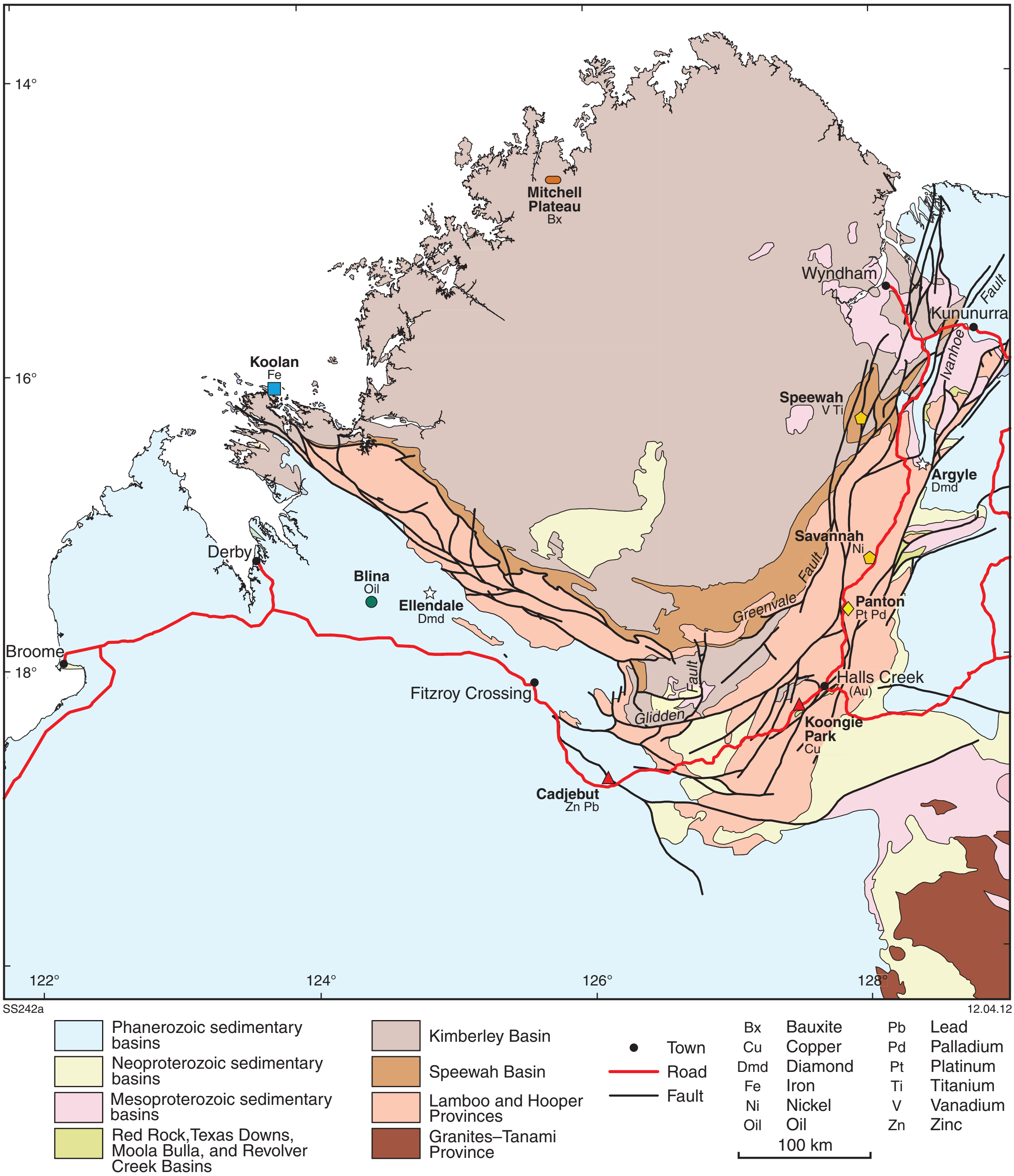


***Figure 1 Tectonic units in the Kimberley region (after Tyler and Hocking, 2001).***

extension of the Western zone (Tyler et al., 1995; Griffin et al., 2000). Stratigraphic units cannot be correlated across the zone boundaries, which is consistent with the zones forming Paleoproterozoic tectonostratigraphic terranes (Tyler et al., 1995).

## *Hooper Province and the Western zone of the Lamboo Province*

The oldest rocks in the Hooper Province and the Western zone of the Lamboo Province (Figure 2) are low- to high-grade turbiditic metasedimentary rocks of the c. 1870 Ma Marboo Formation representing rifting marginal to the Kimberley Craton following accretion of earlier Paleoproterozoic exotic terranes (Figure 3; Griffin et al., 1993, 2000; Tyler et al., 1999). Partial melting of intermediate – felsic, calc-alkaline rock within these source terranes formed the felsic volcanic rocks of the unconformably overlying the Whitewater Volcanics, which were deformed and metamorphosed, and extensively intruded by potassic, I-type granitic and sub-volcanic rocks, as well as gabbroic rocks, and layered mafic–ultramafic intrusions of the Paperbark Supersuite during the 1865–1850 Ma

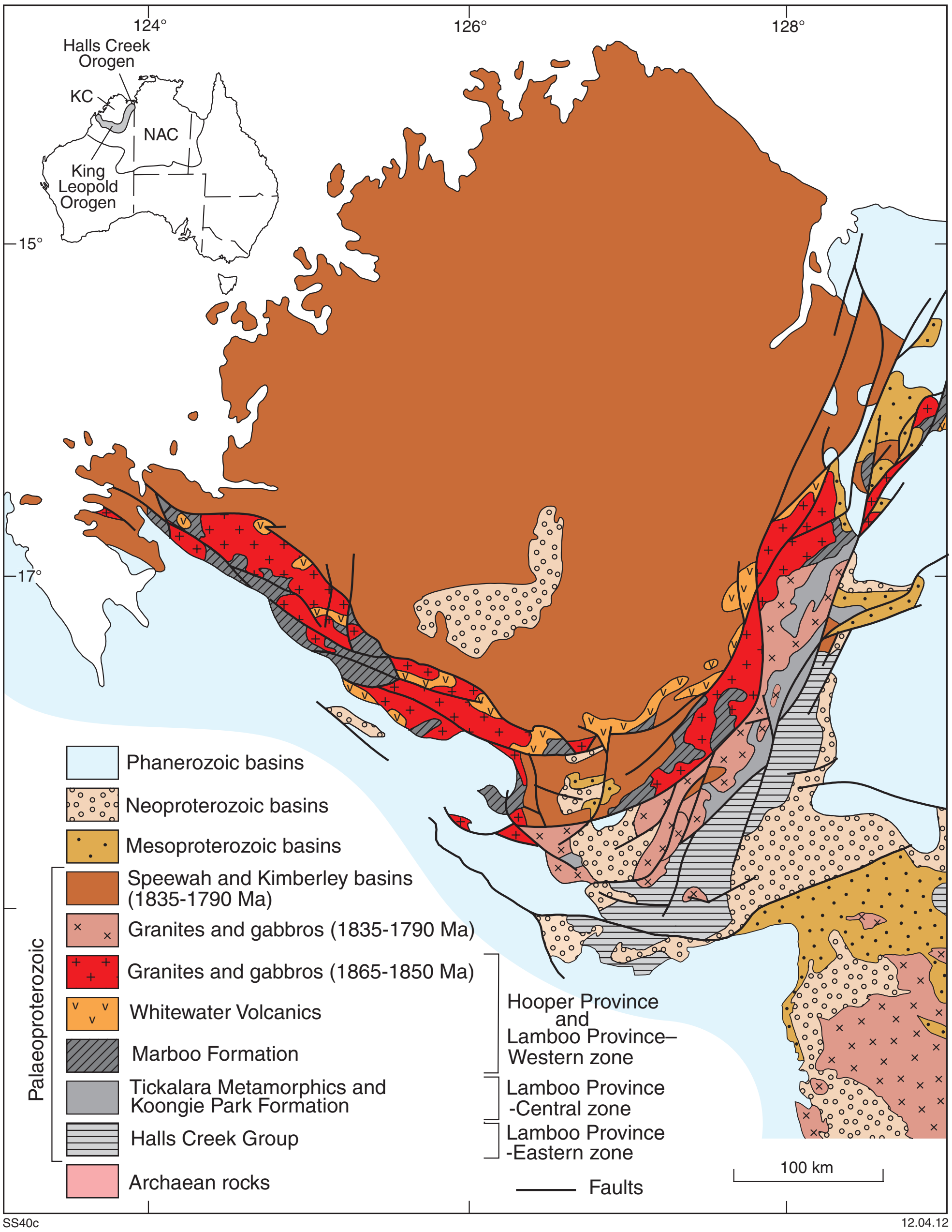


***Figure 2** Paleoproterozoic geology of the Kimberley region. The inset shows the locations of the North Australian Craton (NAC) and the Kimberley Craton (KC) (from Griffin et al., 2000).*

Hooper Orogeny (Griffin et al., 2000; Page and Hoatson, 2000; Page et al., 2001).

## *Central zone of the Lamboo Province*

The Central zone (Figure 2) is dominated by medium – high-grade turbiditic metasedimentary and mafic volcanic and volcaniclastic rocks of the Tickalara Metamorphics, interpreted as an oceanic island arc developed at c. 1865 Ma (Figure 3; Sheppard et al., 1999; Griffin et al., 2000). These were intruded by tonalitic sheets, and deformed and metamorphosed between c. 1865–1856 Ma and at 1850–1845 Ma (Bodorkos et al., 1999, 2000a; Blake et al., 2000; Page et al., 2001). In the southern part of the Central zone, sedimentary rocks and mafic and felsic volcanic rocks of the Koongie Park Formation were deposited at 1845–1840 Ma during rifting of the arc (Page et al., 1994; Tyler et al., 2005). Layered mafic–ultramafic bodies were intruded into the Central zone at c. 1856, c. 1845 and 1830 Ma (Page and Hoatson, 2000). Large volumes of granite and gabbro of the Sally Downs Supersuite intruded the Central zone during the Halls Creek Orogeny at 1835–1805 Ma (Tyler and Page, 1996; Page et al., 2001).

## *Eastern zone of the Lamboo Province*

The Eastern zone (Figure 2) c. 1910 Ma mafic and felsic volcanic rocks of the Ding Dong Downs Volcanics and associated granitic rocks are unconformably overlain by low-grade metasedimentary and metavolcanic rocks of the Halls Creek Group (Tyler et al., 1998). At the base of the Halls Creek Group the quartz sandstone of the Saunders Creek Formation contains exclusively Archean detrital zircon populations (3600 Ma – 2512 Ma; Blake et al., 1999; Tyler et al., 2005). Overlying mafic volcanic rocks of the Biscay Formation in the lower part of the Halls Creek Group were erupted at c. 1880 Ma on a passive continental margin along the western edge of the North Australian Craton (Figure 3; Sheppard et al., 1999).

The overlying turbiditic metasedimentary rocks of the Olympio Formation record a transition from a passive to an active margin and can be divided into upper and lower units separated by alkaline volcanism at c. 1857 Ma and c. 1848 Ma (Blake et al., 1999, 2000). The turbiditic rocks form part of two submarine-fan systems in an evolving foreland basin, with sediment derived from predominantly continental, granitic sources to the NW, the oldest at c. 1873 Ma and the youngest at c. 1847 Ma (Hancock, 1991; Tyler, 2005; Tyler et al., 2005). The Halls Creek Group was deformed and metamorphosed during the 1835–1805 Ma Halls Creek Orogeny, and were stitched to the central zone by the 1820–1810 Ma granites of the Sally Downs Supersuite.

## Hooper Orogeny

The Hooper Orogeny was recognised first in the Hooper Province (Tyler and Griffin, 1993; Griffin et al., 1993), and took place between c. 1870 Ma and c. 1850 Ma. Rocks of the Marboo Formation are affected by two early phases of deformation. The first was between

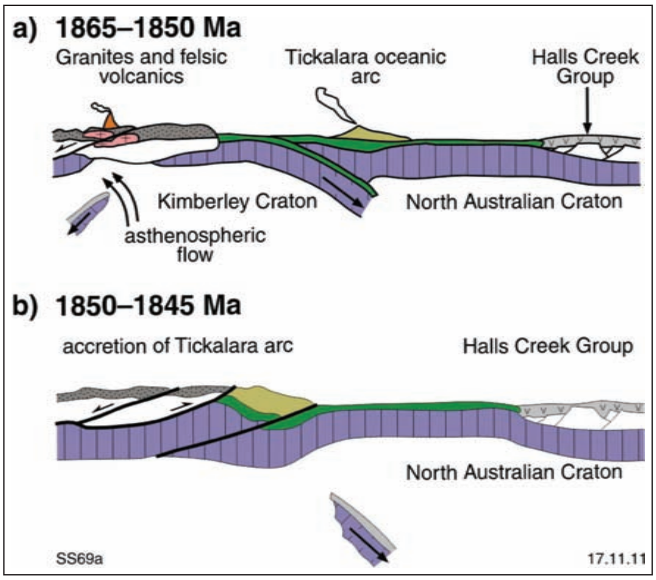


*Figure 3 Plate tectonic setting of the Kimberley during the Hooper Orogeny (from Griffin et al., 2000).*

c. 1870 Ma, the age of the youngest detrital zircons in the Marboo Formation, and c. 1865 Ma, the age of the intrusion of the oldest Paperbark Supersuite intrusions (Tyler et al., 1995, 1999; Griffin et al., 2000). The second deformation and accompanying metamorphism was coeval with the emplacement of the Paperbark Supersuite between 1865 and 1850 Ma, and deformed the c. 1855 Ma Whitewater Volcanics (Griffin et al., 1993; Tyler et al., 1995, 1999).

The first deformation in the Central zone post-dates the c. 1863 Ma age of the Rose Bore Granite, while a minimum age is provided by the c. 1850 Ma granitic rocks of the Dougalls Suite, which post-dates the first deformation but pre-dates the second (Blake et al., 2000; Tyler, 2005). These age constraints are similar to those for the second deformation in the Western zone, suggesting both deformations are linked to the accretion of the Tickalara arc (Thorne et al., 1999; Blake et al., 2000).

### Halls Creek Orogeny

Deformation and metamorphism during the Halls Creek Orogeny affected the entire Lamboo Province (Tyler et al., 1995; Blake et al., 2000). In the Central zone the first deformation occurred synchronously with the intrusion of the c. 1835 Ma Mabel Downs Tonalite (Bodorkos et al., 2000b; Page et al., 2001). The second deformation refolds earlier structures and is cut by c. 1810 Ma granite intrusions (Page et al., 2001).

The Sally Downs Supersuite was intruded between c. 1835 Ma and c. 1805 Ma and its granitic rocks range from syn-collisional, crustally derived adakites through to post-collisional, potassic types (Sheppard et al., 2001).

### Speewah Basin

During the Halls Creek Orogeny, siliciclastic sedimentary rocks of the 1.5 km thick Speewah Group were being deposited in the c. 1835 Ma Speewah Basin, unconformably overlying the northern and western margins of the Western zone of the Hooper and Lamboo provinces (Figure 1). The Speewah Group thins dramatically to the W, and is overlapped by the Kimberley Group in the SE. The presence of c. 1834 Ma felsic volcanic rocks within the Valentine Siltstone, near the base of the succession, suggest the basin developed at the same time as the intrusion of granitic and gabbroic rocks into Lamboo Province at c. 1835 Ma (Tyler et al., 1995; Blake et al., 2000; Tyler, 2000; Sheppard et al., 2012). Paleocurrent data indicate a provenance from the NE, from the core of the uplifted Lamboo Province with deposition taking place in a retro-arc foreland basin behind the active eastern margin of the Kimberley Craton (Gellatly et al., 1970; Sheppard et al., 2012).

## Post-collisional Kimberley Basin and its equivalents (c. 1800 Ma)

The Kimberley Group comprises a 3 km-thick succession of siliciclastic sedimentary rocks and mafic volcanic rocks deposited in the Kimberley Basin (Figure 1), which disconformably overlies the Speewah Group, but pre-date the c. 1797 Ma Hart Dolerite (Sheppard et al., 2012). Paleocurrent data indicate a provenance from the N, beyond the present Australian plate, with deposition taking place in a broad, semi-enclosed, shallow marine basin (Gellatly et al., 1970).

The Moola Bulla, Red Rock, Texas Downs and Revolver Creek basins are probable equivalents to the Kimberley Group that outcrop across the Halls Creek Orogen (Figure 1; Thorne et al., 1999; Blake et al., 1999).

### Hart Dolerite and Wotjulum Porphyry

The Hart Dolerite is a large igneous province that consists of a network of connected sills and dykes of massive dolerite and less extensive granophyres, with a combined thickness of up to 3 km, which are mainly intruded into the Speewah Group and lower Kimberley Group around their deformed SE and SW margins (Sheppard et al., 2012). The intrusions underlie an area of c. 160,000 $km^2$ and have an estimated volume of 250,000 $km^3$ (Griffin et al., 1993). The granophyres have been dated at c. 1797 Ma. The Hart Dolerite records part of a post-collisional magmatic event related to plate re-organisation, and was sourced from subduction-modified mantle beneath the Kimberley Craton (Sheppard et al., 2012).

The c. 1740 Ma Wotjulum Porphyry intrudes as a sill into the upper Kimberley Group in the Yampi Peninsula (Tyler and Griffin, 1993; Sheppard et al., 2012).

## Intracratonic Proterozoic Basins and orogenies (1800–508 Ma)

Collision and suturing was followed by two broad cycles of sedimentary basin formation, followed by intracratonic orogenic deformation and associated fault reactivation. The first spanned the late Paleoproterozoic and Mesoproterozoic as Nuna broke up and Proterozoic Australia was reassembled into Rodinia; the second reflects Neoproterozoic breakup of Rodinia followed by the assembly of Gondwana (Tyler, 2005).

### Paleoproterozoic – Mesoproterozoic basins

The ages of the Bastion, Crowhurst and Osmond basins are uncertain but they all overlie the Kimberley Group or its equivalents

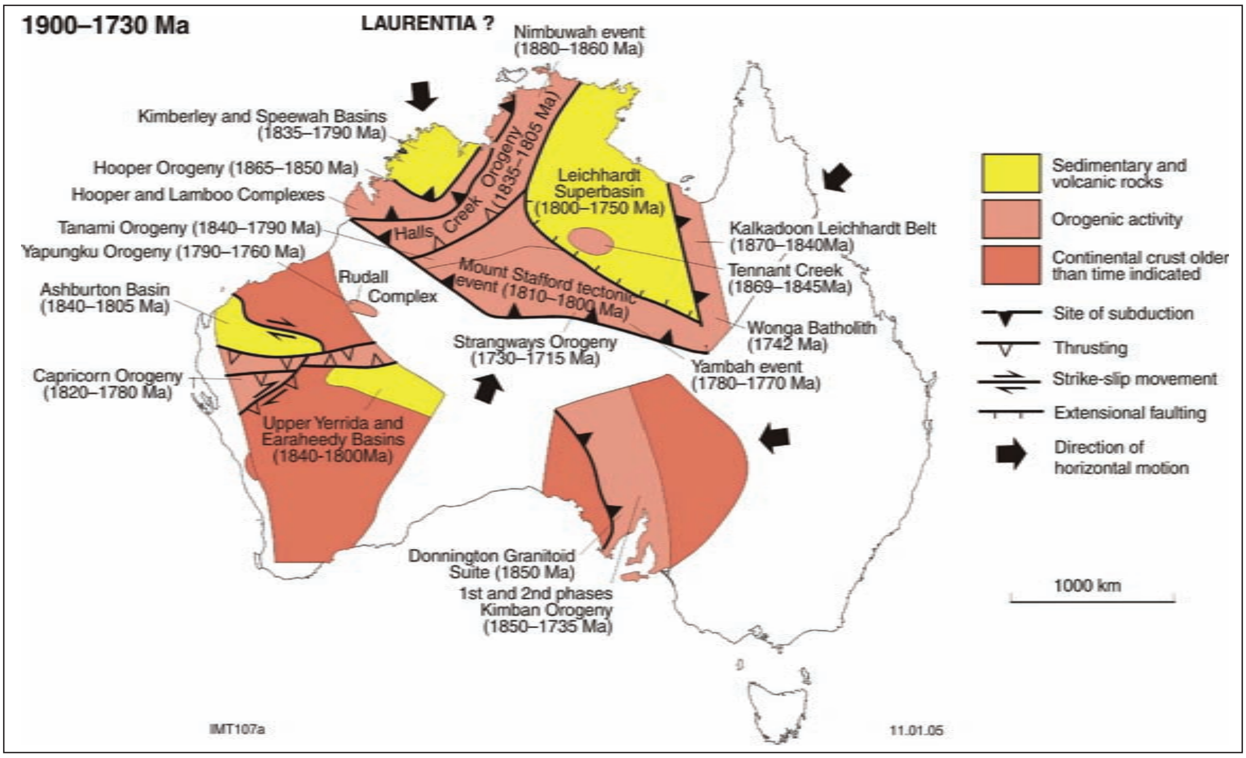


***Figure 4 Model for the assembly of the Diamantina Craton, as proposed by Tyler (2005).***

(Figure 1) and are regarded as younger than c. 1797 Ma. Based on limited isotopic dating and correlation with other units, Thorne et al. (1999) suggested deposition took place in the late Paleoproterozoic – early Mesoproterozoic. Siliciclastic sedimentary rocks of the Bastion Group, deposited in a similar environment to the Kimberley Group, unconformably overlie the Kimberley Basin in the northeastern Halls Creek Orogen, and are overlain by the c. 508 Ma Cambrian Antrim Plateau Volcanics (Plumb and Gemuts, 1976; Thorne et al., 1999). The Crowhurst Group, which includes stromatolitic dolomite, occupies a similar setting overlying the Kimberley Basin at the southern end of the Halls Creek Orogen, and the siliciclastic Mount Parker Formation and the overlying stromatolitic Bungle Bungle Dolomite were deposited in the Osmond Basin on the eastern side of the Halls Creek Orogen (Tyler, 2000).

The siliciclastic rocks of the 4.4 km-thick Carr Boyd Group unconformably overlie both crystalline rocks of the Lamboo Province and equivalents of the Kimberley Group (Thorne et al., 1999) in the northeastern Halls Creek Orogen (Figure 1). They are in turn unconformably overlain by Neoproterozoic glacial deposits of the Wolfe Creek Basin. Deposition of the Carr Boyd Group probably took place in a deltaic to shallow-marine setting at c. 1200 Ma, the age of intrusion of the Argyle lamproite diatreme into wet sediments (Thorne and Tyler, 1996; Jaques et al., 1986; Pidgeon et al., 1989).

The Glidden Group and Wade Creek Sandstone have been correlated with the Carr Boyd Group (Tyler, 2000; Blake et al., 2000).

## Yampi Orogeny

The Mesoproterozoic Yampi Orogeny took place between c. 1400–1000 Ma, based on K-Ar dating (Shaw et al., 1992) and Ar-Ar dating (Bodorkos and Reddy, 2004) and produced large-scale, NE-facing folds and thrusts in the Speewah and Kimberley Basin successions on the Yampi Peninsula (Yampi Fold Belt) (Tyler and Griffin, 1990, 1993; Griffin et al., 1993). Thrusts can be followed into NW-trending, SW-dipping ductile shear zones in the Hooper Province. Deformation was accompanied by a low- to medium-grade metamorphic event with the development of large porphyroblasts of garnet, andalusite and staurolite (Griffin et al., 1993; Bell and Mares, 1999). Large-scale strike-slip faulting took place in the Halls Creek Orogen during the Yampi Orogeny, and established a pattern of N-NE-trending synthetic sinistral faults, and E-NE-trending antithetic dextral faults (White and Muir, 1989; Tyler et al., 1995; Thorne and Tyler, 1996), consistent with the thrusting in the Yampi Fold Belt.

## Neoproterozoic Basins

The Paleoproterozoic and Mesoproterozoic rocks are overlain unconformably by Neoproterozoic sedimentary rocks of the Wolfe Creek Basin, the upper Victoria River Basin, and the Louisa Basin in the Halls Creek Orogen, and by the Mount House Group and Oscar Range Group in the King Leopold Orogen (Figure 1). These linked basins are part of the Centralian Superbasin (Grey and Blake, 1999; Tyler, 2005), interpreted by Walter et al. (1995) as a broad intracratonic sag basin that extended throughout much of central Australia between c. 830 Ma and the earliest Cambrian. Stromatolitic dolomite in the Ruby Plains group can be correlated with Supersequence 1 (Grey and Blake, 1999). Several glacigene intervals are present, with the most widespread being equivalent to the c. 610 Ma Elatina ("Marinoan") glaciation of Supersequence 3 (Grey and Corkeron, 1998). Glacial straie are preserved across the Kimberley region.

## King Leopold Orogeny

The c. 560 Ma King Leopold Orogeny produced extensive, well exposed, W-NW- trending fold and thrust structures in the King Leopold Ranges (Precipice Fold Belt), along the SW margin of the

Kimberley Basin, together with reactivation of shear zones in the Hooper Province (Tyler and Griffin, 1990, 1993; Shaw et al., 1992; Griffin et al., 1993). South-directed thrusting is linked to sinistral strike-slip faulting in the Halls Creek Orogen and deformation affected Neoproterozoic glacigene rocks. Deformation occurred about the same time as the Paterson Orogeny at the NE edge of the Pilbara Craton, and the Petermann Orogeny in central Australia (Tyler, 2005).

# Phanerozoic Basins

A thick Paleozoic and Mesozoic succession is present in the Canning Basin, in the SW Kimberley, and in the Ord and Southern Bonaparte basins, in the eastern and northern Halls Creek Orogen (Figure 1). The King Leopold Orogeny gave rise to a widespread unconformity prior to extrusion of c. 508 Ma basaltic rocks of the Antrim Plateau Volcanics and deposition of middle – upper Cambrian sedimentary rocks in the Ord and Southern Bonaparte basins (Mory and Beere, 1988). The volcanic rocks are part of the extensive Kalkarindji Continental Flood Basalt Province that originally extended across 300,000 km$^2$ of northern Australia (Hanley and Wingate, 2000; Glass and Phillips, 2006). In the Halls Creek Orogen, sinistral strike-slip faulting and associated folding and thrusting related to the c. 450 Ma – 300 Ma Alice Springs Orogeny in central Australia (Mory and Beere, 1988; Thorne and Tyler, 1996) resulted in the development of a series of Devonian sub-basins (Ord Basin) adjoining basin-bounding faults.

Cambrian rocks are absent SW of the Halls Creek Orogen. There, Ordovician and younger rocks of the Canning Basin (Hocking et al., 2008) unconformably overlie or are faulted against Proterozoic rocks (Haines, 2009; Griffin et al., 1993). Ordovician rocks are exposed in a thin sliver along the northern margin of the basin on the Lennard Shelf, beneath the exhumed Devonian reef complexes and associated conglomerates (Playford et al., 2009, and references therein), reflecting a much thicker Ordovician and Silurian succession in the subsurface of the central and southern Canning Basin.

## Devonian reef complexes

Middle and Upper Devonian (Givetian, Frasnian, and Famennian) reef complexes form a belt of rugged limestone ranges, some 350 km long and as much as 50 km wide, on the Lennard Shelf, along the northern margin of the Canning Basin (Playford et al., 2009). The reef complexes are spectacularly exposed as a series of fringing reefs, atolls, and banks, which grew along the mountainous mainland shore of the Kimberley, and around rugged islands. Three main facies are recognised in the reef complexes: platform, marginal slope, and basin (Figures 5–7). The reefs are almost undeformed, with only local tilting due to normal fault movement.

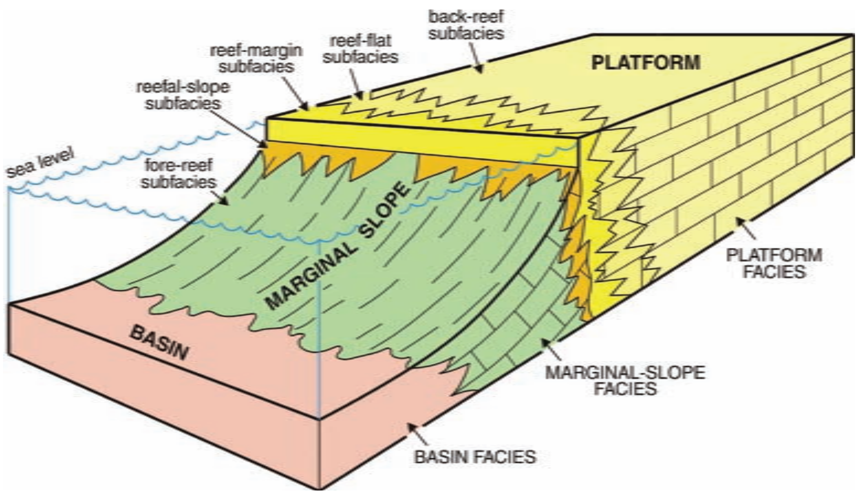


*Figure 5 Reef complexes and facies subdivisions (from Playford et al., 2009, Figure 16).*

Conglomerates interfinger with and cut through the reef complexes at prominent drainage notches related to the scarps of active faults in adjoining Proterozoic basement (Figure 6), and have been intersected (but not studied in detail) by some petroleum exploration drillholes basinward of the outcrop belt. In outcrop, conglomerate bodies interfinger with cyclic platform carbonates. They span many metre-scale cycles, indicating deposition of each body extended over a few million years rather than tens or hundreds of thousands of years, and tectonic rather than eustatic control (Hocking and Playford, 2002). The Famennian phase of conglomerate deposition is attributed to a major episode of the Alice Springs Orogeny of central Australia.

The outcropping complexes preserve continued highstand deposits, with corresponding lowstand deposits preserved basinwards (to the S) in the subsurface. In outcrop, reefal development shows a broad transgressive-regressive cycle upon which several events are superimposed. In the Givetian and Frasnian, reefs showed an overall backstepping pattern, in which vertical reef growth was punctuated by abrupt backsteps, related to a sharp but small sea level fall followed by rapid transgression and drowning of the reef platform. These events appear to be irregular, and thus possibly of tectonic rather than eustatic origin. In the latest Frasnian, at the peak of transgression and just prior to the Frasnian / Famennian global extinction event, there was rapid progradation and horizontal reef growth. Regression in the Famennian led to the reefs prograding in shallower water conditions, although subsidence still outstripped the rate of regression. Many Famennian reef builders migrated upwards from deeper water areas to take the place of newly extinct Frasnian builders such as stromatoporoids.

## Carboniferous–Permian glaciations

In the Late Carboniferous and Early Permian, continental-scale glaciation by N-moving ice sheets possibly several kilometres thick planed off the tops of the limestone reefs. Beneath the ice sheets, subglacial water led to extensive karstification, forming cave systems, karst corridors, solution dolines, and subglacial channels (Playford et al., 2009). Some of these channels are preserved today as the major gorges cutting through the reefs: most notably Billy Munro Gorge, Windjana Gorge (Figure 7) and Geikie Gorge. Clays of probable Permian age can be seen locally in some gorges and caves.

Permian deposition was thickest and most continuous in the Fitzroy Trough, S of the Lennard Shelf. There, deposition continued sporadically until the Late Permian in siliciclastic-dominated, mixed deltaic to shelfal settings (Mory, 2010).

## Mesozoic transgressions

A veneer of Mesozoic sedimentary rocks up to c. 200 m thick covers much of the central and southern Canning Basin, thickening to 500–

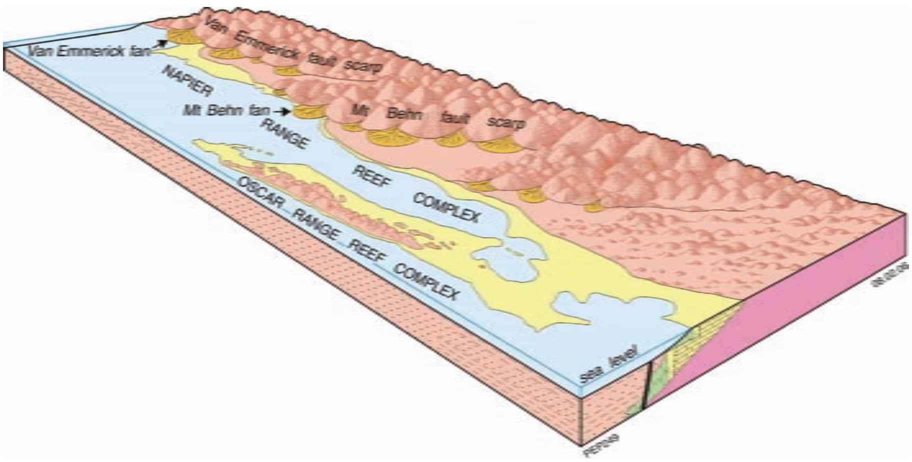


***Figure 6 Oblique block diagram (based on actual geography) of Oscar and Napier Range reef complexes and associated conglomerates at Late Frasnian time (from Playford et al., 2009, Figure 180).***

600 m near the coast. The Triassic succession, characterised by a transgressive marine shale overlain by a prograding deltaic to fluvial succession, is an onshore extension of a much thicker succession in the Roebuck Basin, and is largely restricted to the Fitzroy Trough and outer Lennard Shelf. The Triassic succession is common to the Canning, Roebuck, Northern Carnarvon (where it hosts the giant gas fields of the NW Shelf) and Perth basins, reflecting the onset of Gondwana breakup along Australia's NW margin (Longley et al., 2004).

Along the seaward coast of the Dampier Peninsula, the Lower Cretaceous Broome Sandstone is notable for widespread dinosaur trackways patchily preserved on coastal platforms. The Broome Sandstone marks the start of a post-breakup transgression that extended over much of Australia's interior. In the Canning Basin, deposition was primarily of a fluvial to nearshore, sandstone-dominated veneer. West and S of the Canning Basin (Northern Carnarvon and Gunbarrel basins), and over much of northern South Australia and western New South Wales, the transgression culminated in the Albian with deposition of a radiolarian-rich siltstone.

## Evolving landscapes

The Kimberley region has an ancient landscape that has been evolving following the Neoproterozoic glaciations c. 600 Ma, Devonian tectonism in the Alice Springs Orogeny, Permian glaciation c. 280 Ma, and tilting of the Australian plate in the Neogene. At least two uplifted planation surfaces are present. The High Kimberley surface formed around 280–200 Ma and is preserved as remnants on the main Kimberley Plateau. Between c. 200–100 Ma, uplift and erosion established the Low Kimberley surface surrounding the main plateau. Either or both surfaces may be related to rifting and Gondwana breakup offshore (Longley et al., 2004) to the N and then NW.

Australia has drifted N towards India in the last 24 Myr, with the Australian plate bowing beneath the Indian plate. Its southern margin has lifted up, forming the various barrier plains of the Murray River

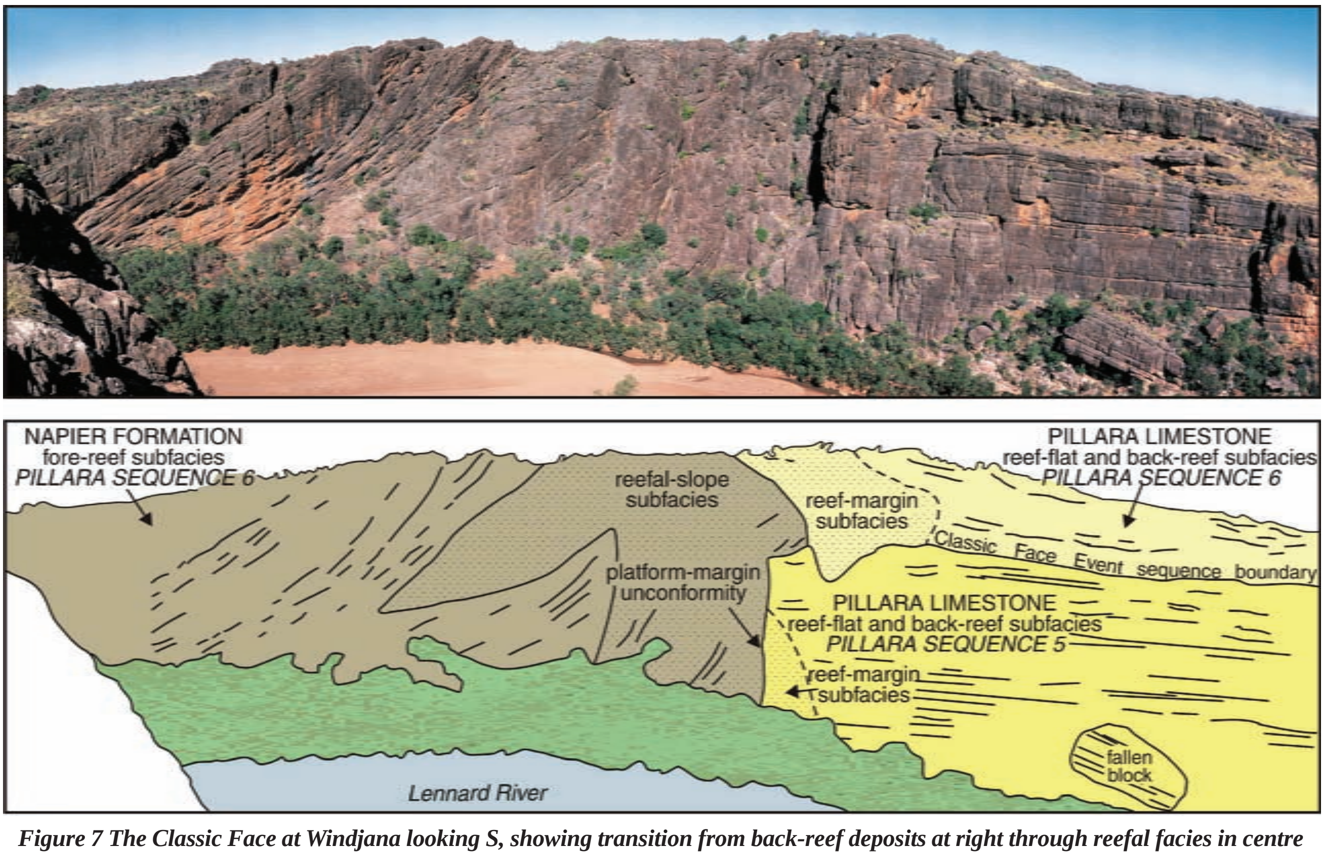


***Figure 7 The Classic Face at Windjana looking S, showing transition from back-reef deposits at right through reefal facies in centre to fore-reef and marginal slope facies at left (after Playford et al., 2009, Figure 486).***

in South Australia and the elevated Nullarbor Plain of the Eucla Basin, and the northern continental margin has bowed downwards, implying the northern Kimberley coastline may have been progressively drowning since the Miocene, rather than simply during the Quaternary. Drowning has been accentuated by sea level rise after the last ice age c. 17 ka.

# Mineral deposits and petroleum fields

The Kimberley region (Figure 1) was the site of Western Australia's first Au rush in 1886 following the discovery of payable Au in the vicinity of Halls Creek in 1885. The small size of the deposits, the remoteness of the area and the lack of fuel and water meant most of the prospectors left in 1892 on news of new discoveries near Kalgoorlie. In 1951 hematitic Fe ore mines were developed by BHP on Cockatoo and Koolan islands in Yampi Sound, a decade or more before the discovery and development of hematitic Fe ore in the Pilbara region. Exploration for diamonds began in the late 1960's and saw the discovery of two major diamond provinces, Ellendale in the west Kimberley and Argyle in the east Kimberley. The Argyle diamond pipe (AK1), a world class deposit, was discovered in 1979, with mining commencing in 1985 (Jaques et al., 1986).

The area around Halls Creek remains highly prospective for Au and a number of small-scale alluvial and lode-gold mines have been developed. The Savannah Ni Mine, N of Halls Creek, together with the sub-economic Panton Pt deposit, highlights the considerable potential for the discovery of further mafic–ultramafic intrusion-related mineralisation. Zinc, Cu and Pb occurs in volcanic-hosted massive sulfide deposits at Koongie Park, SW of Halls Creek, and Mississippi Valley-type Zn-Pb deposits have been mined from the Devonian limestones of the west Kimberley at Pillara and Cadjebut. Bauxite deposits occur on the Mitchell Plateau, and represent deep weathering of underlying basalts in the Kimberley Basin (Smurthwaite, 1990).

A small producing oilfield occurs at Blina in the Canning Basin, and huge reserves of gas are known in the offshore Browse Basin.

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