Overview of the Paleogene of the Eastern Alps

Walter KURZ[1], Harald FRITZ[1], Werner E. PILLER[1],
Franz NEUBAUER[2], Johann GENSER[2]

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Abstract: The evolution of the tectonic units of the Eastern Alps of Austria during the Paleogene is summarized. Based on the structural, petrological, geochronological, and sedimentological-stratigraphic data, a plate tectonic–geodynamic interpretation of the evolution of the Eastern Alps of Austria during the Paleogene is deduced. The starting-point of this evolution is the consolidation of the Austroalpine nappe complex and subsequent cooling of the Austroalpine unit during the Late Cretaceous. During the Late Cretaceous and the Paleogene subduction of the Penninic oceanic domains and the subsequent collision between the Penninic continental and the Austroalpine units is documented. The latter were in an upper plate position. The orogeny during the Paleogene is primarily controlled by subduction, collision and closure of the Penninic oceanic domains. This is indicated by the deposition of turbiditic sediments both on top of the Austroalpine upper nappe complex (Gosau sediment basins), and on the Penninic-European lower plate (Matrei Zone, Kaserer Group?, Rhenodanubian Flysch). Collision is followed by cessation of flysch sedimentation, and the deposition of Molasse sediments, which started in the Late Eocene. Contemporaneously, subduction and collision is documented by high-pressure metamorphism in the internal zones of the Eastern Alps, which affected the Middle- and Southpenninic units, and by Late Oligocene magmatism along the Periadriatic Lineament, which resulted from the break off of the subducted slab of Penninic oceanic lithosphere. Oblique collision finally resulted in the formation of strike-slip faults parallel to the strike of the orogen, and the initiation of orogen-parallel extrusion at the end of the Oligocene, which continued during the Miocene.

Zusammenfassung: Die Entwicklung der tektonischen Einheiten des österreichischen Anteiles der Ostalpen während des Paläogens wird zusammengefasst dargestellt. Basierend auf strukturellen, petrologischen, geochronologischen und sedimentologisch-stratigraphischen Daten wird eine plattentektonisch-geodynamische Interpretation abgeleitet. Ausgangspunkt ist der konsolidierte ostalpine Deckenstapel in der Späten Kreide, der zu diesem Zeitpunkt bereits abgekühlt ist. Während der Späten Kreide und dem Paläogen ist die Subduktion und anschließende Kollision der penninischen

[1] Institut für Geologie und Paläontologie, Karl-Franzens-Universität Graz, Heinrichstrasse 26, A-8010 Graz, E-mail: walter.kurz@uni-graz.at

[2] Institut für Geologie und Paläontologie, Paris-Lodron-Universität Salzburg, Hellbrunner Strasse 34, A-5020 Salzburg

Einheiten mit der ostalpinen Oberplatte belegt. Die Orogenese während des Paläogens wird primär kontrolliert von der Subduktion der penninischen Einheiten und der Schließung penninischer ozeanischer Bereiche. Dies wird vor allem durch die Ablagerung von turbiditischen, flyschoiden Sedimenten, sowohl auf der ostalpinen Oberplatte (Gosau Becken), als auch auf der penninisch-europäischen Unterplatte (Matreier Zone, Kaserer Gruppe?, Rhenodanubischer Flysch), angezeigt. Nach der Kollision folgt das Aussetzen der Flyschsedimentation und der Beginn der Ablagerung der Molassesedimente, welche im Späten Eozän einsetzt. Gleichzeitig ist die Subduktion und Kollision durch eine Hochdruckmetamorphose, welche die mittel- und südpenninischen Einheiten erfaßt hat, in den Internzonen der Ostalpen belegt, und weiters durch spätoligozänen Magmatismus entlang der Periadriatischen Naht. Dieser resultiert aus dem "break off" der subduzierten penninischen ozeanischen Lithosphäre. Schräge Kollision resultiert schlußendlich in der Bildung von Seitenverschiebungen parallel zum Streichen des Orogens, und in orogenparalleler Extrusion am Ende des Oligozäns, welche im Miozän ihre Fortsetzung findet.

Keywords: Paleogene, Austria, Stratigraphy, Structure, Metamorphism, Tectonics

Contents

1. INTRODUCTION

The Paleogene covers the time span between 65 and 23.8 Ma (Hardenbol et al., 1998). During this time period major changes occured on a global as well as on a regional scale which are to a certain extent also detectable in the development of the Eastern Alps. Although recently a comprehensive volume on the Geology of Austria (Neubauer & Höck, 2000) was published, a modern, general overview on the Paleogene development of the Eastern Alps is still missing. We try to combine within this review article data on sedimentology and stratigraphy, as well as on structure, tectonics and metamorphism. Based on these data we attempt to deduce geodynamic implications including all tectonic units within the Eastern Alps and take also neighbouring areas into account. Since a tremendous number of scientific articles on this field is already published (for a summary see Flügel & Faupl, 1987), we rely primarily on the most recent review articles and text-books, however, we try to include as many papers as possible.

2. OVERVIEW OF THE GEOLOGY OF AUSTRIA

The northernmost part of Austria is made up of the southern sectors of the Bohemian Massif. However, the geological evolution of the Bohemian Massif is related to the Variscan (Tollmann, 1977, 1982, 1986; Brandmeyer et al., 1995, 1999; Neubauer et al., 2000a; Neubauer & Handler, 2000). To the south, wide areas are represented by the sediments of the Alpine Molasse Zone (Fig. 1), and further to the south, by the east-west striking belt of the Rhenodanubian Flysch Zone, and by units that are derived from the Helvetic paleogeographic realm (Fig. 1) (Tollmann, 1977, 1986; Janoschek & Matura, 1980; Oberhauser, 1980; Neubauer & Höck, 2000; Kurz et al., 2000; Kurz & Unzog, 2000). These units have already been highly affected by the Alpine orogenesis.

The Alps are the result of the still ongoing convergence between Africa and Europe. Although they are the mountain chain which has been studied in most detail, many open questions remained. These open questions are related to the overprinting of the southern margin of the Variscan fold belt during the Alpine orogeny, the complex tectonic processes during the evolution of the western part of the Tethyan ocean, the Jurassic strike-slip tectonics connected with the opening of the Penninic oceanic realm and a polyphase, non-coaxial deformation during the Alpine orogeny (Schuster et al., 1999). Models on Alpine geology have been rapidly developed during the last decades, mainly because of detailed paleogeographical, structural, petrological, and geochronological investigations. These, together with deep reflection seismic profiling, provided new insights into the present-day structure and resulted in new models which fundamentally changed ideas on the evolution of the Alpine orogen (e.g., Frisch, 1979, 1980a; Nicolas et al., 1990; Pfiffner, 1992; Oberhauser, 1995; Pfiffner et al., 1997; Neubauer et al., 2000a).

2.1. Tectonic units of the Alps

This review basically follows the tectonic subdivision described by Neubauer et al. (1998) and Neubauer & Höck (2000) and intends to synthesize principal structural data of the Eastern Alps and the distribution of Alpine metamorphism. It also includes some paleogeographic aspects according to the present state of data (Neubauer et al., 1998, 2000; Neubauer & Höck, 2000; Kurz et al., 2000). The Mesozoic time scale calibrations follow Hardenbol et al. (1998).

Aspects of the tectonic evolution of the Eastern Alps were discussed in Tollmann (1977, 1986, 1987), Janoschek & Matura (1980), Oberhauser (1980, 1995), Frank (1987), Thöni & Jagoutz (1993), Froitzheim et al. (1996), Ebner (1997), Faupl (1997), Faupl & Wagreich (2000), and Neubauer et al. (2000a). The evolution of the pre-Alpine basement is reviewed in Von Raumer & Neubauer (1993).

Plate tectonic units involved in the Alpine orogen are the European Plate, which is represented by the Helvetic nappes, the largely oceanic Penninic plate, the Apulian (Adriatic) microplate including the Austroalpine and South Alpine units, and the African plate. Remnants of the Tethyan ocean only occur as fragments in the easternmost part of the Eastern Alps. In the Carpathians these Tethyan elements are called Meliaticum.

In a geographical sense the Alps are divided into the E-W-trending Eastern Alps (Fig. 1), which build up the main part of Austria, and the Central Alps of Switzerland, divided by the Rhine valley south of Lake Constanz and its southward, meridional extension, and the arc of the Western Alps of Italy and France. Eastern, Central and Western Alps are characterized by a fundamentally different geological structure, geological evolution and, in part, a distinct geomorphology. In the central Eastern and the Swiss Central Alps the most prominent mountain peaks are located along the central axis. East of the Tauern Window, the topography gradually changes from high elevations into the Neogene Styrian and Pannonian basin plains of a very low elevation above sea level.

From the footwall to the hanging wall (from N to S, or NW to SE) the Alps as a whole include the following tectonic units (the subdividion of several tectonic units follows e.g., TOLLMANN, 1977; TRÜMPY, 1980, 1997a, b; DEBELMAS et al., 1983; FRANK, 1987; DAL PIAZ, 1992; FAUPL, 1997; NEUBAUER et al., 2000a) (Fig. 1):

1) The stable European continental lithosphere which also carries the Late Eocene to Neogene Molasse basin, the northern peripheral foreland basin (e.g., TRÜMPY, 1980, 1997a, b).
2) The Helvetic units, a thin-skinned fold and thrust belt, that nearly exclusively comprises Late Carboniferous to Eocene cover sequences detached from the European lithosphere (FAUPL & WAGREICH, 2000); in the Western Alps, the External massifs comprise pre-Alpine basement rocks and Helvetic, Late Carboniferous to Cretaceous cover sequences (DEBELMAS et al., 1983).
3) The Valais units which represent the infilling of a mainly Cretaceous rift zone on attenuated continental to likely oceanic crust of Northpenninic paleogeographic origin (PFIFFNER, 1992; OBERHÄNSLI, 1994); the eastward continuation of the Valais paleogeographic realm into the Rhenodanubian Flysch Zone was discussed, for example, by EGGER (1990, 1992), KURZ et al. (1998b), and NEUBAUER et al. (2000a). Remnants of this unit have been observed in the lower units of the Lower Engadine Window (Fig. 1), including the Ramosch ophiolite (KOLLER & HÖCK, 1992).
4) The Brianconnais units which represent a microcontinent rifted off from stable Europe during opening of the Valais trough; therefore, these units are of Middlepenninic paleogeographic origin. Units that are derived from the Brianconnais are predominantely represented in the Central and Western Alps (for a summary see STAMPFLI, 1993, 1996; OBERHÄNSLI, 1994; PFIFFNER et al., 1997, and references therein). However, the eastward continuation of the Brianconnais into the Zentralgneiss Terrane of the Tauern Window was discussed, e.g. by TOLLMANN (1965, 1980, 1986), TRÜMPY (1975, 1983), FRISCH (1977b, 1979, 1980a), OBERHAUSER (1995), and FROITZHEIM et al. (1996). Remnants occur within the Tasna Nappe of the Lower Engadine Window (OBERHAUSER, 1980, 1983; WAIBEL & FRISCH, 1989; KOLLER & HÖCK, 1992).
5) The Ligurian-Piemontais-Southpenninic units with oceanic lithosphere in the Eastern, Central, and Western Alps; the ophiolitic Glockner nappe exposed within the Tauern Window and its correlatives exposed in other windows (e.g., the Lower Engadine Window) along the central axis of the Eastern Alps, and the Ybbsitz ophiolite (DECKER, 1990; SCHNABEL, 1992). Furthermore, overlying flysch sequences as well as the Rhen-

odanubian Flysch Zone with remnants of trench filling sediments without any substrate may represent parts of this zone. The Valais, Brianconnais and Piemontais units are conventionally combined to Penninic units also assigned as North-, Middle- and Southpenninic units, respectively.

6) The Austroalpine unit, which includes: (1) a Jurassic passive continental margin which faced towards the Penninic (Piemontais) oceanic trough (Lower Austroalpine unit); (2) a unit of polymetamorphic continental basement, largely represented in the Middle Austroalpine unit (Fig. 1a); (3) remnants of a Triassic passive continental margin, originally facing towards the S (to the Hallstatt-Meliata basin) (e.g., LEIN, 1987), largely represented in the Tirolic and Bajuvaric nappes of the Northern Calcareous Alps (NCA) (Upper Austroalpine unit) (e.g., TOLLMANN, 1973, 1976a, b; EISBACHER & BRANDNER, 1995, 1996; SCHWEIGL & NEUBAUER, 1997; LINZER et al., 1997; FAUPL & WAGREICH, 2000; MANDL, 2000); (4) the Hallstatt-Meliata units with its remnants of the infilling of a small (oceanic?) trough and an adjacent distal continental margin, largely represented in the Lower Juvavic nappes of the NCA (Upper Austroalpine unit); (5) the Upper Juvavic unit of the NCA that exclusively comprises late Paleozoic to Paleocene cover sequences of a passive continental margin (e.g., MANDL, 2000).
7) And finally, the Southalpine unit juxtaposed along the Periadriatic Fault to the Austroalpine units. The Southalpine unit is another continental unit that is largely similar to the Upper Austroalpine unit. Together with the Austroalpine unit it is considered to represent the northern extension of the stable Adriatic microplate which also includes the Po plain and the adjacent Adriatic sea (e.g., CHANNELL et al., 1979). The Southalpine unit is considered as the southern external retro-arc orogenic wedge within the Alpine orogenic system (e.g., DOGLIONI & FLORES, 1997; SCHMID et al., 1996).

Compared with previous interpretations, the Ybbsitz ophiolite and the subdivision of the Austroalpine into three different tectonic units are discussed by NEUBAUER et al. (1998). Following this discussion, the use of the term Austroalpine unit (s. str.) should be restricted to the pile of nappes of continental origin in the footwall of remnants of the Hallstatt-Meliata trough and transitional continental cover sequences. Consequently, the overlying Upper Juvavic units of the NCA derive from a separate continental unit and are excluded from the Austroalpine units s. str. (NEUBAUER et al., 1998; NEUBAUER & HÖCK, 2000). For further details, see SCHWEIGL & NEUBAUER (1997). However, this is in contradiction to the visible facies patterns, which exhibit all transition from platform to open marine conditions, all oriented in a similar manner, facing towards the S (MANDL, 2000). If the Upper Juvavic units had indeed originated from an opposite southern shelf, separated from the Tyrolean shelf by an ocean (Hallstatt-Meliata realm), the Juvavic units should exhibit facies gradients of opposite orientation (MANDL, 2000). Therefore, the position of the ocean, named "Hallstatt-Meliata-Ocean" (KOZUR, 1991), that bordered the Upper Austroalpine and the Western Carpathians, and the suture in the present Upper Austroalpine nappe stack is still a matter of controversial discussion (e.g., TOLLMANN, 1976a, b; KOZUR, 1991; KOZUR & MOSTLER, 1992; HAAS et al., 1995; SCHWEIGL & NEUBAUER, 1997; MANDL, 2000).

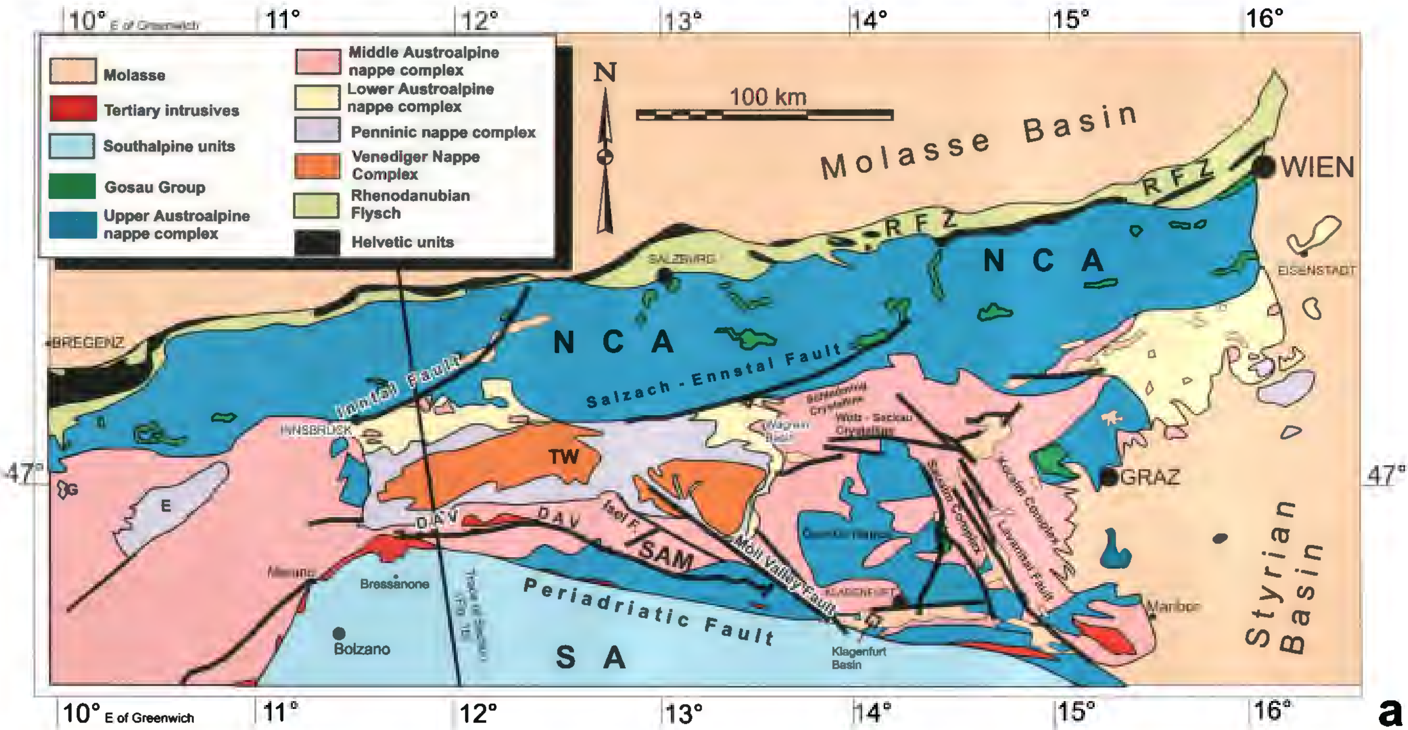
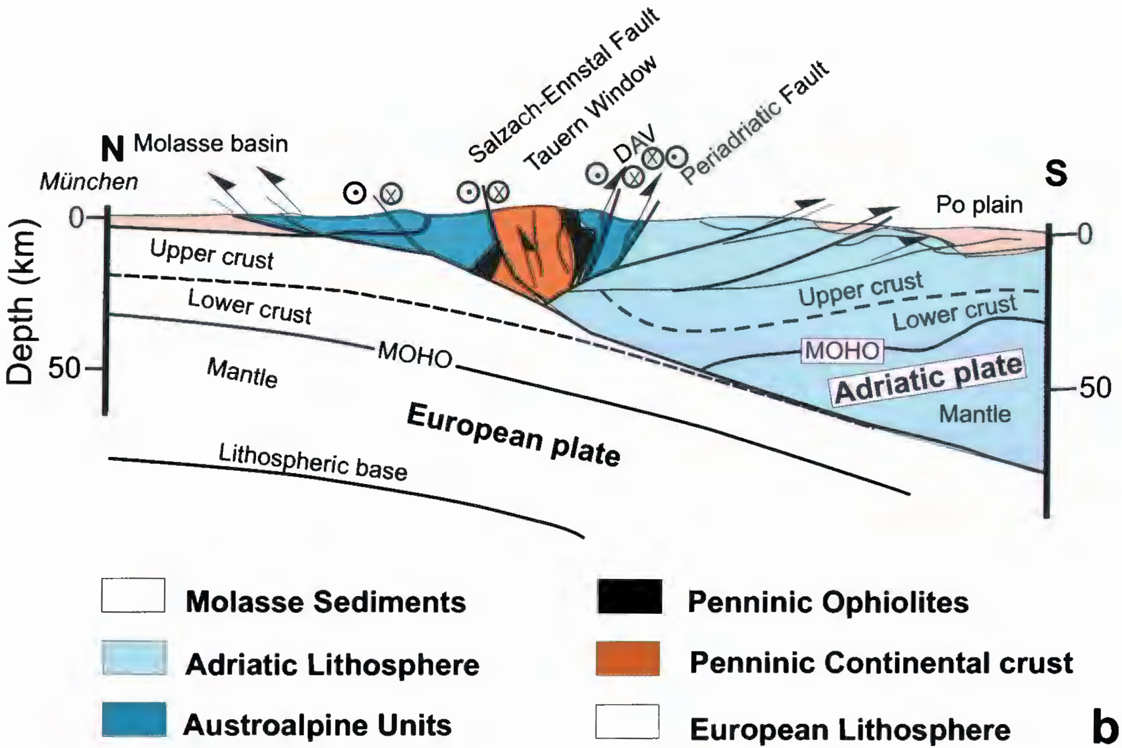


Fig. 1: a – Simplified tectonic map, showing the major tectonic units of the Eastern Alps (following Kurz & Unzog, 1999); DAV: Defereggen-Antholz-Vals Fault; E: Lower Engadine Window; G: Gargellen Window; NCA: Northern Calcareous Alps; RFZ: Rhenodanubian Flysch Zone; SA: Southern Alps; SAM: Southern Limit of Alpine Metamorphism; TW: Tauern Window. b – Transect across the Eastern Alps (modified after Lammerer & Weger, 1998; Neubauer et al., 2000b).


N
S
Molasse basin
München
Salzach-Ennstal Fault
Tauern Window
DAV
Periadriatic Fault
Po plain
Depth (km)
0
50
Upper crust
Lower crust
MOHO
Mantle
European plate
Lithospheric base
Adriatic plate
Molasse Sediments
Adriatic Lithosphere
Austroalpine Units
Penninic Ophiolites
Penninic Continental crust
European Lithosphere
b

3. PALEOGENE GEOLOGICAL EVOLUTION OF THE EASTERN ALPS

From the Late Cretaceous to the Early Neogene the evolution of the Eastern Alps is characterised by two important events:

1. The formation of the extensional Gosau basins at the Early/Late Cretaceous boundary around 90 Ma (e.g., FAUPL et al., 1987), which indicates the termination of nappe stacking in the Austroalpine unit and subsequent extension.
2. The "Tauern metamorphism" (or "Tauern crystallisation") (e.g., SANDER, 1939; FRASL, 1958; FRANK et al., 1987), an event of greenschist to amphibolite facies metamorphism in the Tauern Window (Fig. 1) around 30–25 Ma (e.g., SELVERSTONE, 1993), which affected the Penninic nappe stack (e.g., SELVERSTONE, 1988, 1993; KURZ et al., 1996, 1998a, b).

Referring to the task of this volume, the formation of the Gosau basins started already 35 million years before the Paleogene, the Tauern metamorphic event occurred at the end of the Paleogene. Both events are very well constrained by structural, tectonometamorphic, and sedimentological data. The evolution during the Tauern metamorphic ("Neo-Alpine") event is summarized by e.g., HÖCK & MILLER (1980, 1987), DACHS (1986, 1990), FRANK et al. (1987), GENSER et al. (1996), KURZ et al. (1996), SCHUSTER et al (1999), SELVERSTONE (1993), and KURZ et al. (this volume), and references cited therein. While the Tauern metamorphic event is well constrained in space and time, the pre-Paleogene event affected the Austroalpine units at variable grade and at different times, with continuously decreasing ages from the hanging wall (about 100 Ma) to the footwall (about 70–80 Ma) (e.g., RATSCHBACHER, 1986, 1987; FRANK, 1987; KROHE, 1987; TOLLMANN, 1987; RATSCHBACHER & NEUBAUER, 1989; BEHRMANN, 1990; NEUBAUER & GENSER, 1990; HOINKES et al., 1991, 1999; THÖNI & JAGOUTZ, 1992, 1993; NEUBAUER et al., 1993, 1995, 1998, 2000a; RING, 1994a,b; DALLMEYER et al., 1996; FROITZHEIM et al., 1996, 1997; KURZ et al., 1999). The thermal peak of the Tauern metamorphic event, therefore, was reached about 40–60 Ma after nappe stacking ceased in the Austroalpine unit. Nevertheless, this time gap has been used to formulate some of the classic conductive heating models to explain Barrovian-type metamorphism subsequent to continent-continent collision in Alpine-type orogens (ENGLAND & RICHARDSON, 1977; ENGLAND & THOMPSON, 1984; STÜWE & SANDIFORD, 1995). Actually, a rather big part of this time gap coincides with the Paleogene. This may give rise to the presumption that the Paleogene is a rather "quiet" period concerning the structural and tectonometamorphic evolution of the Eastern Alps. However, especially at the end of this period (in the Oligocene), the structure of the (Eastern) Alps was highly modified. The main events of this structural evolution comprise:

- Closure of the Penninic oceanic domain (e.g., TOLLMANN, 1975; FRISCH, 1976; FRANK et al., 1987; KURZ et al., 1998b).
- Subduction-related eclogite facies metamorphism in the Penninic unit (e.g., DROOP et al., 1990; SELVERSTONE et al., 1994; KURZ et al., 1996, 1998a, this volume).
- Collision between the Penninic continental units and the Austroalpine upper plate, and related nappe stacking within the Penninic unit (e.g., FRISCH, 1974, 1975a, b, 1976; TOLLMANN, 1975; LAMMERER, 1988; GENSER et al., 1996; KURZ et al., 1998b; KURZ et al., this volume).
- Decompression and exhumation of the units within the Tauern Window, and the

Tauern metamorphic event (e.g., DROOP, 1981, 1985; CLIFF et al., 1985; BEHRMANN, 1988, 1990; SELVERSTONE, 1988, 1993; GENSER & NEUBAUER, 1989; AXEN et al., 1995; SELVERSTONE et al., 1995; FÜGENSCHUH et al., 1997; LAMMERER & WEGER, 1998).

- Contemporaneous thin skinned tectonics related to N-S compression in the Rhenodanubian Flysch and the Helvetic unit (e.g., EGGER, 1989, 1990; FREIMÜLLER et al., 1998; FAUPL & WAGREICH, 2000), and within the Molasse Zone (GENSER et al., 1998).
- Beginning of orogen-parallel extension and "lateral extrusion", and formation of orogen-parallel strike-slip fault systems (Salzach-Ennstal fault, Defereggen-Antholz-Vals fault, Peijo fault, Inntal fault, and the Periadriatic Fault) (KLEINSCHRODT, 1987; GENSER & NEUBAUER, 1989; SCHMID et al., 1989; RATSCHBACHER et al., 1989, 1991; DECKER at al., 1993; DECKER & PERESSON, 1996; PERESSON & DECKER, 1997a, b; LINZER et al., 1997; WANG & NEUBAUER, 1998) (Fig. 1).
- Intrusion of Tonalites and related magmatites along the Periadriatic Fault (KARL, 1959; SCHARBERT, 1975; EXNER, 1976; BORSI et al., 1979; VON BLANCKENBURG & DAVIES, 1995, 1996; DAVIES & VON BLANCKENBURG, 1995; NEMES et al., 1997; VON BLANCKENBURG et al., 1998).

The starting-point of the evolution of the Eastern Alps during the Paleogene is the previous consolidation of the Austroalpine nappe edifice and subsequent cooling in the Late Cretaceous (e.g., HEJL, 1997, 1998; THÖNI, 1999). During the Late Cretaceous and the Paleogene subduction of the Penninic ocean and the subsequent collision between the Penninic and the Austroalpine units is documented (e.g., FRISCH, 1976, 1980c, 1984; FRISCH et al., 1987; FRANK et al., 1987; BEHRMANN, 1990; SELVERSTONE, 1993; KURZ et al., 1998a); the latter is in an upper plate position. This evolution is governed by important plate tectonic processes that are related to the opening of the Middle and Northern Atlantic (e.g., FRISCH, 1977b, 1979, 1980a, 1981; TOLLMANN, 1987b, c; LE PICHON et al., 1988; TRÜMPY, 1988; DEWEY et al., 1989; STAMPFLI, 1993, 1996; NEUBAUER, 1994a; STAMPFLI & MARCHANT, 1997; STAMPFLI et al., 1998), and in particular to the opening of the Gulf of Biscay, which is directly related to the evolution of the North-Penninic domain (FRISCH, 1979, 1980a; SCHÖNENBERG & NEUGEBAUER, 1987; OBERHÄNSLI, 1994; CHANNELL & KOZUR, 1997; STAMPFLI & MARCHANT, 1997). Additionaly, this evolution triggers a counter-clockwise rotation of the Adriatic plate, and the build up of a N-S to NW-SE oriented regional compressional stress field in the Oligocene (LE PICHON et al., 1988; BEHRMANN, 1990; PLATT et al., 1989, CHANNELL et al., 1979). As a direct consequence of the continuous rotation of the Iberian and African plates the regional stress field was re-oriented from a NE-SW (in the Late Oligocene) direction to a NW-SE orientation in the Miocene (LE PICHON et al., 1988; KURZ et al., 1994), which was related to the formation of the Southalpine indenter along the Periadriatic fault (RATSCHBACHER et al., 1991; SPRENGER & HEINISCH, 1992; DECKER et al., 1993; SPRENGER, 1996).

A summary of the evolution of several units during the Paleogene is given in Tab. 1.

3.1. The Paleogene evolution of Austroalpine units and the Gosau Group

In the Late Cretaceous, the major nappe structures within the Austroalpine unit were completed. Only in the Upper Juvavic unit of the NCA, Paleogene shallow water carbonates (Kambühel Limestone) have been documented (FAUPL & WAGREICH, 2000; MANDL, 2000). Subsequent to nappe stacking the Austroalpine nappe assemblage has

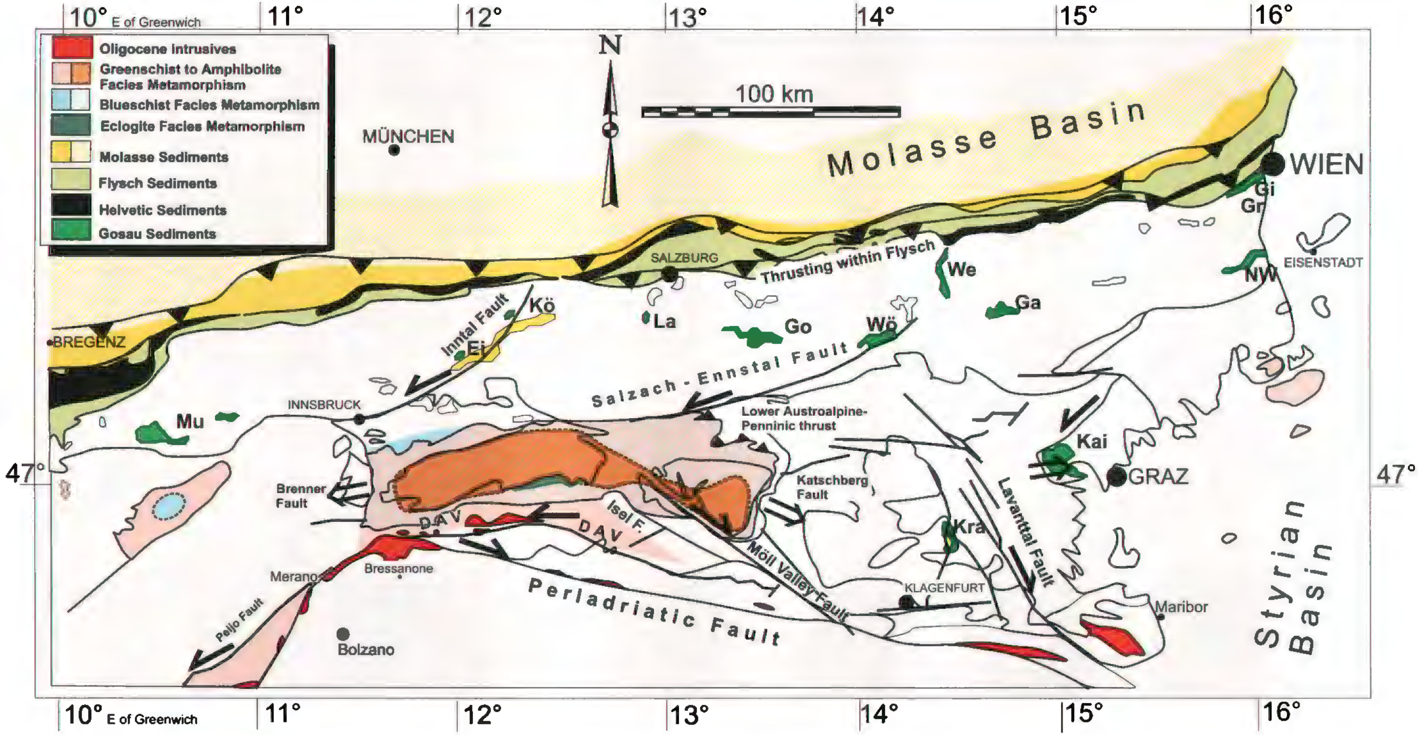


Fig. 2: Simplified map of the Eastern Alps, showing the occurrence and distribution of Paleogene sediments, and tectono-metamorphic and magmatic events during the Paleogene. "Kalkalpine" Gosau Basins with documented Paleogene sequences (after FAUPL & WAGREICH, 1992a; WAGREICH & FAUPL, 1994; WAGREICH, 1995): Mu: Muttekopf; Ei: Kufstein-Eiberg; Kö: Kössen; La: Lattengebirge; Go: Gosau; Wö: Wörschach; We: Weisswasser; Ga: Gams; Gi: Giesshübl; Gr: Grünbach; NW: Neue Welt. "Zentralalpine" Gosau Basins: Kai: Kainach; Kra: Krappfeld.

been affected by crustal extension (Ratschbacher et al., 1989; Neubauer et al., 1995; Froitzheim et al., 1997; Willingshofer, 2000). This triggered the formation of several sedimentary basins, the Gosau basins, on top of the Austroalpine nappe stack (e.g., Faupl et al., 1987). Stretching and basin formation started during the Late Cretaceous (Butt, 1981; Faupl et al., 1987; Faupl & Wagreich, 1992a, b, 2000; Ortner, 1992, 1994a,b; Wagreich, 1993, 1995; Wagreich & Faupl, 1994; Neubauer et al., 1995; Froitzheim et al., 1997; Ebner & Rantitsch, 2000; Wagreich, this volume). The sedimentological evolution is documented up to the Eocene (e.g., Faupl et al., 1987; Wilkens, 1989; Wagreich & Faupl, 1994; Egger et al., 1996). However, structural and sedimentological differences exist between the Gosau basins on top of the NCA ("Kalkalpine Gosau") and on top of the Central Eastern Alps ("Zentralalpine Gosau") (e.g., Neubauer et al., 1992, 1993, 1995; Wagreich & Faupl, 1994). Basically, sedimentation was tectonically controlled.

The sedimentary succession on top of the NCA is subdivided into the lower Gosau Subgroup (Upper Turonian–Campanian), which is characterized by terrestrial to shallow-marine facies associations, and the upper Gosau Subgroup (Santonian-Eocene), which comprises deep-water hemipelagic and turbiditic deposits (Wagreich & Faupl, 1994; Egger et al., 1996; Faupl & Wagreich, 2000; Wagreich, this volume). Rapid subsidence started diachronously from the northwest (Santonian) to the southeast (Maastrichtian) after a short phase of deformation and erosion. The rapid deepening was interpreted to be related to subduction-related tectonic erosion in the footwall of the Austroalpine nappe complex (Wagreich, 1993, 1995). Paleogene sequences are documented in the Gosau Basins of Muttekopf (Danian), Eiberg (Danian), Kössen (Danian), Lattengebirge (Lutetian), Gosau (Ypresian), Wörschach (Ypresian), Weisswasser (Danian), Gams (Ypresian), Giesshübl (Danian-Thanetian), Grünbach (Ypresian), and Neue Welt (Thanetian) (Fig. 2) (Faupl & Wagreich, 1992a; Wagreich & Faupl, 1994; Wagreich, 1995, this volume). The sedimentary sequences of the Paleogene of the upper Gosau Subgroup are characterized by deep water clastics and turbidites. The terrigenous material of the deep-water successions comprises predominantely metamorphic detritus. From the sedimentary record of the distinct basins it can be concluded, that the NCA as a whole had been covered by deep water sediments since the Late Campanian (Wagreich, 1995). In the Paleocene and Eocene, the deep-water deposits already covered most of the NCA. However, in the southeastern parts of the NCA the sedimentation of shallow water carbonates was continued until the Danian (Kambühel Limestone) (Faupl & Wagreich, 2000; Mandl, 2000; Wagreich, this volume). Plankton-rich foraminiferal assemblages in carbonate-rich hemipelagites and radiolarian blooms give evidence for oceanic water masses and speak against sedimentation in small, distinct basins that were separated by shallow-water erosional areas (Wagreich, 1995).

The Gosau deposits of the Central Eastern Alps evolve from alluvial to lacustrine facies, and finally to distal submarine fan-delta sequences (Gräf, 1975; Neubauer et al., 1995; Ebner & Rantitsch, 2000). These deposits are predominantely exposed within the "Kainach Gosau" basin WNW of Graz (Figs. 1, 2), and the "Krappfeld" basin NNE of Klagenfurt. Basically these facies patterns reflect three major steps of evolution:

After the Late Santonian transgression, a carbonate platform developed in front of a terrestrial environment. After a period of lithification of platform sediments, rapid subsidence resulted in a turbiditic facies with local debris flow sediments containing clasts of the earlier platform (Thiedig, 1975; Neumann, 1989). This can be predominantely observed in the "Krappfeld" basin. A last phase is characterized by diminishing limestone clasts, decreasing grain sizes and the transition to an orbitoid-rich assemblage of a submarine fan. Major tectonic pulses, which triggered subsidence occurred during the Late Santonian (start of basin subsidence), the Early Campanian (marine transgression), and the Maastrichtian (Neubauer et al., 1995). Clasts of the Austroalpine crystalline basement and ophiolitic detritus are absent, but fragments of Southalpine provenance have been reported from the Kainach Gosau (Winkler, 1996). These beds are locally overlain by Paleogene sequences (Guttaring Group) with an erosional unconformity (Neubauer, 1992), which is well documented within the "Krappfeld" basin (Rasser, 1994). These successions range from the latest Paleocene to the Middle Eocene. At the base red beds occur above an erosional relief with red clay, quartz gravels, local coal seams and rare horizons of black, and marine detrital limestones. These are overlain by nummulite marls with only minor terrigenous clasts. This sequence is interrupted by coal seams and abundant siliciclastic sediments which indicate enhanced terrigenous input. They are overlain by limestone-marl alternations, followed by nummulite limestones with decreasing contents of siliciclastic material (Wilkens, 1989). In contrast to the underlying Gosau sediments, the sandstones contain ophiolitic material.

A very small occurrence of nummulite-bearing carbonates of Late Eocene age covering the Lower Austroalpine crystalline basement complex in the eastern part of the Eastern Alps, and Eocene marine limestone pebbles in Miocene conglomerates give evidence of a more widespread post-Gosau sedimentary cover of the Eastern Alps. This coincides with the onset of Molasse sedimentation (Faupl & Wagreich, 2000).

Especially the Gosau basins on top of the NCA have been affected by compressive deformation during the Paleogene, which is related to N- to NE-oriented contraction. This deformation includes generally N- to NE vergent folding, and faulting which partitioned into reverse faults of variable sense of displacement, having been linked by NE-striking sinistral transfer faults (Eisbacher et al., 1990; Eisbacher & Brandner, 1995, 1996). Paleogene thrusts within the NCA have also affected the Gosau deposits (e.g., Faupl & Wagreich, 1992a, 2000). Erosional unconformities within the Gosau Group deposits may be related to these deformational events (e.g., Wagreich, 1993, 1995).

Crustal stretching, extension and the formation of the Gosau basins of the Eastern Alps east of the Tauern Window ("Zentralalpine Gosau") coincides with the exhumation of crystalline basement complexes of the Middle Austroalpine unit (Fig. 1). Exhumation resulted in cooling from initial epidote-amphibolite/upper greenschist facies conditions to temperatures below 300° C at the beginning of the Paleogene. Sphene, zircon and apatite fission track data, for example from the Gleinalm area, indicate cooling to below temperatures of 200–250° C at 65 Ma (Neubauer et al., 1995). The Middle Austroalpine units north of the Mur Valley (Seckau Crystalline) and the northern part of the Koralm Complex (Figs. 1, 2) cooled to temperatures below 200° C already in the Late Cretaceous (Hejl, 1997,1998). Hence, these regions were already near to the surface during the whole Cenozoic. Further westward, in the Schladming Crystalline and in the Gurktal Nappe complex (Fig. 2), a stronger post-Cretaceous denudation can be observed. During the Neogene, the Schladming Crystalline experienced pronounced uplift with respect to the Gurktal Alps (Hejl, 1998). Clastic material of the upper Gosau Subgroup on top of the NCA was derived from source terrains from the south, which was related to continued exhumation and progressive erosional unroofing of Austroalpine basement complexes (FAUPL & WAGREICH, 1992a; WAGREICH, 1995).

3.2. The internal zones of the Eastern Alps

3.2.1. Tectonic units

In the internal zones of the Eastern Alps, the tectono-metamorphic evolution during the Paleogene is generally restricted to the area of the Tauern Window, the Lower Engadine Window, and the Gargellen Window (Figs. 1, 2). A contribution of the tectonometamorphic evolution of the Tauern Window and its surrounding units (Fig. 1) during the Paleogene is given by KURZ et al. (this volume). The Lower Engadine Window has been subdivided into a number of nappes, comprising the central Bündnerschiefer Unit, including the Ramosch ophiolite (WAIBEL & FRISCH, 1989; KOLLER & HÖCK, 1992), the Tasna Zone consisting of a continuous sedimentary sequence deposited on an associated continental basement, and the Arosa Zone including the Idalp ophiolite (KOLLER & HÖCK, 1992). The Gargellen Window comprises the flysch sequences of the Falknis Nappe, the Sulzfluh Unit, and the Arosa Zone representing the uppermost unit (OBERHAUSER, 1980). The Penninic units within the Tauern Window comprise the parautochthonous continental basement and cover sequences that are incorporated in an imbricate nappe stack (e.g., FRISCH, 1975a, 1976, 1977a, 1979; TOLLMANN, 1975; LAMMERER et al., 1981; LAMMERER, 1986, 1988; KURZ, 2000; KURZ et al., 1996, 1998b). These nappes are derived from the Zentralgneiss Terrane. They are overthrusted by the oceanic sequences of the Glockner Nappe, which are supposed to be of Southpenninic paleogeographic origin, and the Matrei Zone, which comprises the remnants of an accretionary wedge in the hanging wall of the subducted oceanic Glockner Nappe (e.g., FRISCH et al., 1987). In the Lower Engadine Window, the Southpenninic unit is represented by the Arosa Zone, and the oceanic sequences of the Idalp ophiolite (DAURER, 1980; HÖCK & KOLLER, 1987, 1989; KOLLER & HÖCK, 1992). The Ybbsitz Zone comprises serpentinites derived from remnants of an oceanic floor, and a sedimentary sequence that reaches up to the Danian (DECKER, 1990; SCHNABEL, 1992). It is supposed to be an equivalent to the Arosa Zone (FAUPL & WAGREICH, 2000).

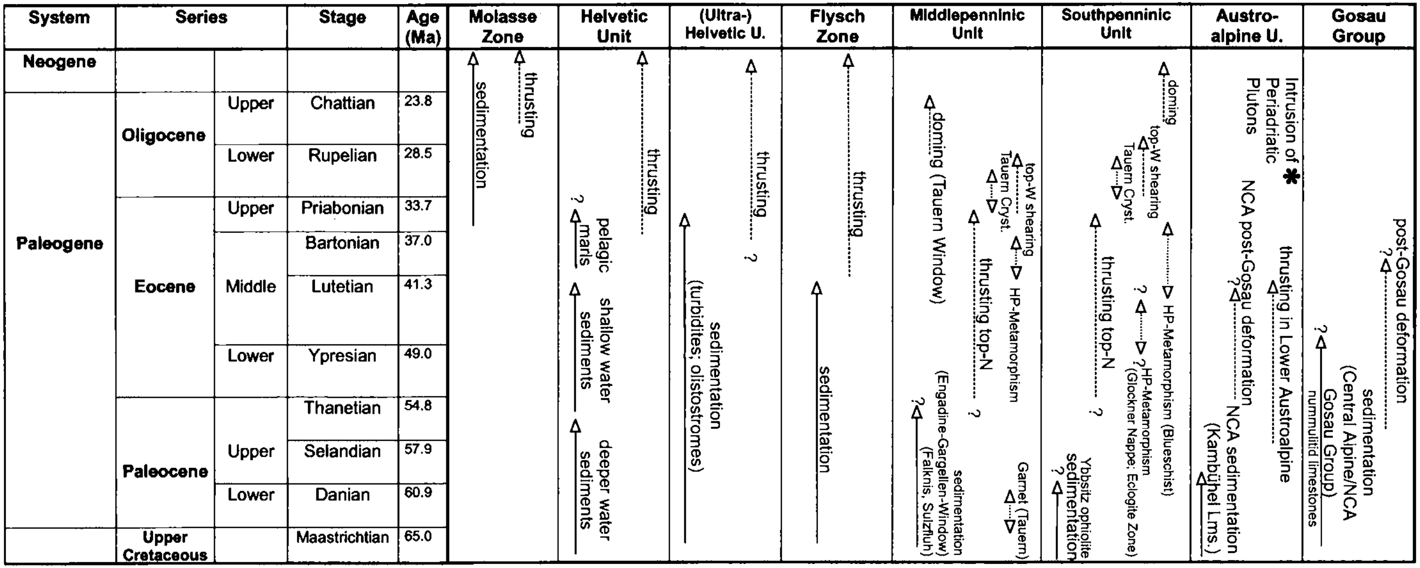
The Lower Austroalpine nappe complex forms the hanging wall in the northeastern and northwestern corner of the Tauern Window units. Remnants of the oceanic crust between the Penninic continental units and the Austroalpine block are preserved in the Glockner Nappe (Southpenninic unit), which is emplaced onto the parautochthonous basement with its Permian to Mesozoic cover and onto the basement-cover units, while the Glockner Nappe itself is overthrusted by the Austroalpine nappe complex. Thus the Glockner Nappe forms a major plate tectonic suture zone within the Eastern Alps. A detailed tectonostratigraphic description of the Penninic units within the Tauern Window is given by FRASL (1958), FRASL & FRANK (1964, 1969), THIELE (1970), EXNER (1973, 1983, 1990), FRISCH (1974, 1975a, b, 1977, 1980a, b), LAMMERER et al. (1981), and KURZ et al. (1996, 1998b, 2000). A detailed description of the evolution of the lithostratigraphic units during the Paleogene is given by KURZ et al. (this volume).

Basically, evidence of the exact stratigraphic ages of several sequences is scarce because of multiple structural and tectonic overprint. E.g., Early Cretaceous pteridophyte spores have been documented by REITZ et al. (1990) in the hanging wall formations of the Matrei Zone in the northeastern part of the Tauern Window. Additionally, lithofacial similarities between the Helvetic unit and the Hochstegen unit of the Venediger Nappe Complex indicate, that the future Zentralgneis Terrane was part of the Helvetic facies realm during the Triassic and Jurassic (e.g., FRISCH, 1975). This gave access to the assumption that the sedimentation within the Kaserer Group could have lasted up to the Eocene (LAMMERER, 1988). However, these assumptions are not supported by stratigraphic data. Although there are strong similarities to the couges rouges assemblages within the Middlepenninic Tasna Nappe of the Lower Engadine Window, this indicates that the Penninic units of the Tauern Window have already been buried by the Austroalpine nappe complex in the Late Cretaceous (OBERHAUSER, 1964, 1980, 1983, 1995; WAIBEL & FRISCH, 1989). Lower Eocene sequences are documented in the wildflysch of the Falknis and Sulzfluh Zone of the Lower Engadin and the Gargellen Window (Northpenninic paleogeographic origin) (e.g., OBERHAUSER, 1983, 1995; WAIBEL & FRISCH, 1989; BOUSQUET et al., 1998).

3.2.2. Metamorphic evolution

The rocks exposed within the Tauern Window experienced a polyphase metamorphic evolution (SELVERSTONE et al., 1984, 1991; SELVERSTONE, 1985, 1988, 1993; SELVERSTONE & SPEAR, 1985; GENSER et al., 1996). Inclusions in garnets document a first stage of metamorphism at ca. 400° C (MILLER, 1977; FRANK et al., 1981, 1987). Eclogite facies metamorphism is only observed clearly within the Eclogite Zone (see KURZ et al., this volume), and in the lowermost sections of the Glockner Nappe (PROYER et al., 1999; STURM et al., 1996). The eclogite facies rocks were buried to a depth of at least 65 km (20 MPa, ± 600° C; HOLLAND, 1979; DACHS, 1986, 1990; FRANK et al., 1987; DROOP et al., 1990; SELVERSTONE et al., 1992; ZIMMERMANN et al., 1994; GETTY & SELVERSTONE, 1994). The entire nappe pile in the Tauern Window was subsequently affected by blueschist facies metamorphism. Within the Eclogite Zone pressures of 7–9 MPa and temperatures of ca. 450° C are estimated by RAITH et al. (1980); 450° C, 10–15 MPa are estimated by HOLLAND (1979b) and ZIMMERMANN et al. (1994), but the P-T data are not well constrained due to the subsequent strong overprint by Barrovian-type metamorphism. Within the other tectonic units peak pressures of up to 10–12 MPa have been evaluated (FRY, 1973; HÖCK, 1974; SELVERSTONE et al., 1984, 1992; SELVERSTONE & SPEAR, 1985; CLIFF et al., 1985; DROOP, 1981, 1985; HOLLAND & RAY, 1985; FRANK et al., 1987; BEHRMANN & RATSCHBACHER, 1989; DACHS, 1990; BEHRMANN, 1990; KRUHL, 1993; SELVERSTONE, 1993; KURZ et al., 1998a, b). Finally, the entire nappe pile was affected by Barrovian-type upper greenschist to lower amphibolite facies metamorphism (e.g., FRANK et al., 1987; SELVERSTONE, 1993). A detailed summary of the tectonometamorphic evolution of these units during the Paleogene is provided by KURZ et al. (this volume).

In the Lower Engadine Window (Figs. 1, 2) the metamorphic assemblages indicate metamorphic conditions no more than lowermost greenschist facies, which increases from the frame to the central part of the window. In the central parts, high pressure phases, such as glaucophane and lawsonite, have been observed (LEIMSER & PURTSCHELLER, 1980).



Tab. 1: Paleogene time scale (after HARDENBOL et al., 1998) and evolution of several geological units of the Eastern Alps.

3.2.3. Geochronological constraints

Especially the timing of high pressure metamorphism and the emplacement of high pressure rocks within the Tauern Window were discussed controversially. Generally, an Early to middle Cretaceous age of eclogite formation was assumed (e.g., BEHRMANN & RATSCHBACHER, 1989; RAITH et al., 1980). Following ERNST (1971, 1975), these high-pressure metamorphic assemblages were suspected to form a "paired metamorphic belt" together with Cretaceous middle pressure metamorphic sequences within Austroalpine units (e.g., FRANK, 1987; FRANK et al., 1987; WALLIS et al., 1993). However, no direct evidence of a Cretaceous age high-pressure metamorphism within the Penninic realm was documented by these authors. In contrast to supposed Cretaceous ages, phengite $^{40}Ar/^{39}Ar$ mineral ages of ca. 36–32 Ma (ZIMMERMANN et al., 1994) from the eclogites are interpreted to represent cooling ages subsequent to Eocene blueschist facies metamorphism (ZIMMERMANN et al., 1994). The possibility of a younger age of high pressure metamorphism was discussed by INGER & CLIFF (1994), too. Garnet Rb-Sr ages of ca. 65 Ma are documented within the Venediger Nappe in the western part of the Tauern Window by CHRISTENSEN et al. (1994). This indicates that the Zentralgneiss Terrane was already subducted at the end of the Late Cretaceous, and that growth of metamorphic garnets was initiated.

Rb-Sr white mica ages of ca. 27 Ma to 29 Ma (REDDY, 1989; REDDY et al., 1993; INGER & CLIFF, 1994) were interpreted as formation ages during the Cenozoic thermal peak of Barrovian-type metamorphism. K-Ar and $^{40}Ar/^{39}Ar$ muscovite and biotite cooling ages of ca. 35 Ma (OXBURGH et al., 1966; LAMBERT, 1970) are supposed to be too high due to excess argon (VON BLANCKENBURG & VILLA, 1988; VON BLANCKENBURG et al., 1989). Within the eastern Tauern Window, amphibole $^{40}Ar/^{39}Ar$ ages of ca. 24 Ma are thought to date the thermal peak of metamorphism (CLIFF et al., 1985). Biotite Rb-Sr ages indicate closure of this isotopic system and cooling through ca. 300° C between 19 and 23 Ma in the central parts of the eastern Tauern Window (REDDY et al., 1993). Apatite fission track ages document cooling through ca. 100° C between 8 and 12 Ma (STAUFENBERG, 1987; HOKE, 1990). Similar ages are reported from the western part of the Tauern Window by GRUNDMANN & MORTEANI (1985), GRUNDMANN (1987), and FÜGENSCHUH et al. (1997). Tension gashes as well as hydrothermal quartz veins are mineralised with quartz, silicates and sulphides. The sulphides formed at temperatures between 365 and 410° C (FEITZINGER & PAAR, 1991). REDEN & GÖTZINGER (1991) assumed 300° C as a minimum temperature of quartz formation within quartz veins. If the vein fluids were in thermal equlibrium with the surrounding rocks this indicates the formation of these veins at around 23 Ma.

The Austroalpine units surrounding the Tauern Window show a partly different chronological evolution in contrast to the Penninic units within the Tauern Window. Whole rock $^{40}Ar/^{39}Ar$ plateau ages from the Lower Austroalpine unit NW of the Tauern Window (Reckner and Hippold Nappe) as well as the Glockner Nappe in the footwall, which all have been affected by high-pressure, blueschist facies metamorphism, record ages around 50 Ma (Reckner Nappe), and 44–37 Ma (Hippold Nappe and Glockner Nappe) (DINGELDEY et al., 1997). These ages are interpreted to date the high-pressure metamorphism. The Zircon fission track ages from the Lower Austroalpine unit (Innsbruck Quartzphyllite) (Fig. 1) range from 67 Ma to 35 Ma (FÜGENSCHUH et al., 1997).

Apatite fission track ages are in the range of 13 ± 2 Ma, similar to the ages within the Tauern Window (e.g., GRUNDMANN & MORTEANI, 1985). This indicates, that cooling was rather slow between 67 and 13 Ma, and that the Penninic and Lower Austroalpine units in this area suffered a common cooling history around 13 Ma. The Middle Austroalpine Ötztal-Stubai basement complex is characterized by Zircon ages ranging from 84 to 42 Ma, and Apatite ages of 62–17 Ma (FÜGENSCHUH et al., 1997, 2000), which preserve evidence of an earlier Cretaceous exhumation and subsequent cooling during the Paleogene.

3.2.4. Structural evolution

The Penninic units within the Tauern Window and the lowermost parts of the Lower Austroalpine unit are characterized by a polyphase deformation history, starting in the Late Cretaceous/Paleogene (e.g., FRISCH, 1976, 1977; KURZ et al., 1996, 1998b). Generally, three main deformational events can be distinguished: (1) Brittle to ductile thrusting with top-to-the N nappe emplacement (e.g., BEHRMANN, 1990; OEHLKE et al., 1993; KURZ et al., 1996); (2) ductile top-to-the WNW shearing approx. contemporaneous to the Tauern metamorphic event; (3) crustal stretching and doming, contemporaneous to exhumation and uplift.

Brittle to semibrittle deformation occurred along distinct thrusts during early stacking within the accretionary wedge between the Penninic lower plate and the Austroalpine upper plate, showing top-to-the N displacement. Basically, this deformational phase is related to the subduction of the Southpenninic oceanic lithosphere, and subsequent accretion of the units derived from the Zentralgneiss Terrane in the footwall of the Austroalpine block.

W- to NW-directed ductile shearing is well documented over the entire Tauern Window starting approximately at the thermal peak of Cenozoic Barrovian-type metamorphism (BICKLE & HAWKESWORTH, 1978; FRANK et al., 1987; LAMMERER, 1988; BEHRMANN, 1990; BEHRMANN & FRISCH, 1990; SELVERSTONE, 1993; KRUHL, 1993; OEHLKE et al., 1993; WALLIS et al., 1993; CHRISTENSEN et al., 1994; KURZ et al., 1996). It affected the entire nappe pile within the Tauern Window. Basically, this deformational phase is related to the collision between the Penninic nappe stack and the main part of the European plate (e.g., GENSER et al., 1996).

The shape of the Tauern Window was highly modified during exhumation and doming after the penetrative deformation events that were related to nappe stacking (NEUBAUER et al., 1999, 2000b). The Tauern Window is characterized by a kinked shape that might be the result of an Alpine indenter in the central part of the window. Indentation in the Late Oligocene (slightly after peak thermal metamorphism) caused clockwise rotation of the eastern part of the Tauern Window and counterclockwise rotation of the western part (KURZ & NEUBAUER, 1996). The rotation is proven by overprinting relationships of cross-cutting NE-trending and younger NNE-trending subvertical mineralized extensional veins in the eastern part of the Tauern Window (KURZ et al., 1994). It resulted in the rotation of several previous structures and kinematic indicators and in the divergence of kinematic data over the entire Tauern Window.

3.3. Paleogene evolution of the Rhenodanubian Flysch Zone and the Ultrahelvetic/Helvetic unit

The *Rhenodanubian Flysch Zone* comprises a Late Cretaceous to Eocene succession of turbidites intercalated with marls, shales and subordinate tuffs (FAUPL, 1996; EGGER et al., 1997; NEUBAUER & HANDLER, 1998). This unit is interpreted to represent the infilling of a deep sea trough located between the Helvetic/Ultrahelvetic slope and the Penninic units. For detailed discussion, the reader is referred to FAUPL (1978), WINKLER et al (1985), WINKLER (1988, 1996), EGGER (1990, 1992, 1995), FAUPL & WAGREICH (1992), PREY (1992), SCHNABEL (1992), OBERHAUSER (1995), and EGGER et al. (1997). The succession started with carbonate-dominated flysch deposits, but passed into turbidites rich in siliciclastic material in the latest Early Cretaceous. The Upper Cretaceous turbidite successions are subdivided by several thin-bedded variegated pelitic intervals, deposited during periods of reduced turbiditic input. In the Campanian, basin-plane deposits in carbonate mud turbidites are documented. The Rhenodanubian Flysch comprises several tectonic units with variable thicknesses. These include from bottom to top the Tristel, Reiselsberg, Seisenburg, Zementmergel, Perneck, and Altlengbach Formations. However, only the uppermost parts of the Altlengbach Formation have been deposited during the Paleogene (Paleocene and Early Eocene) (EGGER, 1989, 1990, 1992). The Altlengbach Formation is subdivided into the Rossgraben, Ahornleiten, Acharting, and Strubach Members (EGGER, 1995). Only the upper section of the Strubach Member reaches up into the Paleogene. The Paleogene deposits of the Rhenodanubian Flysch obtain a thickness of approximately 500 m. They can be subdivided into four lithofacial units (EGGER, 1989, 1990), which can be distinguished by different sub-facies of turbidite facies D (EGGER, 1990); two of them show a hemipelagic facies which indicates very low sedimentation rates in the Middle Paleocene and around the Paleocene/Eocene boundary. Similar intercalations of the same stratigraphic age are described from the Gurnigel Nappe in Switzerland (Gurnigel Flysch, Wägital Flysch, Schlieren Flysch) (WINKLER et al., 1985; EGGER, 1990). The Paleogene deposits are mainly characterized by stable heavy mineral assemblages (FAUPL & WAGREICH, 2000). The change from garnet- to zircon-dominated assemblages has been observed within the Maastrichtian-Upper Paleocene deposits. In the middle part of the Rhenodanubian Flysch Zone thin bentonite layers occur in the Paleocene/Eocene boundary interval (EGGER et al., 1997). Toward the east, thick terrigenous turbidite beds of Late Paleocene-Eocene age were deposited near the present northern margin of the trough (FAUPL & WAGREICH, 2000).

Structurally, the Rhenodanubian Flysch Zone represents a classical thrust-fold belt. The structure is dominated by approximately E-trending, kilometre-scale kink fold anticlines and synclines, blind thrust faults and splay thrusts. Several deformation stages can be distinguished (FREIMÜLLER et al., 1998). These include: (1) overthrusting of the Ultrahelvetic continental margin sequences by the Rhenodanubian Flysch during the Late Eocene, (2) subsequent shortening of this thrust wedge, probably associated with the emplacement onto the southern margin of the Molasse Zone during the Oligocene to Early Neogene, and (3) the disruption of the combined thrust wedge by strike-slip faults during the Neogene.

The *Ultrahelvetic Zone* comprises a stratigraphic sequence from the Lower Jurassic to the Paleogene. In contrast to the Helvetic Zone, the Lower Cretaceous deposits are characterised by a deep-water carbonate facies (FAUPL & WAGREICH, 2000). The deep-water limestones are followed by the "Buntmergelserie", which comprises a variegated marl succession starting in the Albian. In the Paleogene, the stratigraphic succession is characterized by marls, coarse clastic deposits with olistoliths of limestones, granites and related rocks, as well as turbiditic successions (FAUPL, 1978a, b; FRASL, 1980; WIDDER, 1986; FAUPL & SCHNABEL, 1987).

The *Helvetic realm* comprises sedimentary strata which had been deposited on the shelf and upper slope of the European continental plate (FAUPL & WAGREICH, 2000; RASSER & PILLER, this volume). Lower Cretaceous deposits are exposed in the western part of the Eastern Alps, terminating east of Salzburg. They form a continuation of the broad Helvetic Zone of the Swiss Central Alps (e.g., PFIFFNER, 1993). The Gresten Klippen Zone E of Salzburg shows facial assemblages that are different from the Helvetic Zone s.str. (OBERHAUSER, 1980; FAUPL & WAGREICH, 2000). Generally, the Helvetic Zone consists of a sequence of deposits ranging from Late Carboniferous to Paleogene, which has been interrupted several times. The Paleocene and Early Eocene are characterized by the accumulation of nummulite and corallinacean shallow-water sediments (RASSER & PILLER, this volume). Pelagic deposits with globigerinid foraminifera and terrigenous flysch sediments pass finally into the Molasse sediments of the Alpine foreland during the Late Eocene/Oligocene (FAUPL & WAGREICH, 2000).

3.4. Paleogene evolution of the Molasse Basin

The Molasse basin of the Eastern Alps (Fig. 1) is a flexural foreland basin developed in response to the loading of the southern margin of the European plate after the final continent-continent collision (LEMCKE, 1984; BACHMANN et al., 1987; WESSELY, 1987; MALZER et al., 1993; GENSER et al., 1998; ZWEIGEL, 1998; ZWEIGEL et al., 1998; RASSER, 2000; STEININGER & WESSELY, 2000). The overriding Alpine nappe complex comprises, from bottom to top, the outer shelf to slope of the European continental margin, the Flysch nappes, and the Austroalpine complex in an upper plate position. The youngest sediments of these units are of Early Eocene age. In the footwall of the Molasse of the Vienna Basin, Rhenodanubian Flysch sediments reach up to the Middle Eocene (Steinbergflysch; Laaberdecken-Agsbachschichten, Buntmergelserie) (OBERHAUSER 1980, p. 198; WESSELY, 2000). The Molasse basin displays strong lateral changes in shape with a decrease in width from approximately 150 km in the western part of the basin to less than 10 km at the spur of the Bohemian Massif. To the east the basin widens again and changes its strike from E-W to NE-SW. The basement depths at the southern margin of the basin decrease from approximately 3500 m in the west to about 500 m in the east. The ages of the oldest Molasse strata get increasingly younger in the same direction (MALZER et al., 1993; GENSER et al., 1998; RASSER, 2000).

The Molasse sequence started in the Late Eocene with the subsidence of the European plate, with a deepening towards the south (LEMCKE, 1984; BACHMANN et al., 1987; MALZER et al., 1993; WAGNER, 1996, 1998; GENSER et al., 1998; POLESNY, 1998; RASSER et al., 1999; STEININGER, 1999; RASSER, 2000; STEININGER & WESSELY, 2000). The Late Eocene sediments show a paralic facies and nearshore sands (WAGNER, 1998). The clastic input was derived from the Bohemian Massif in the N. These sands interfinger and continue to the S with corallinacean limestones ("Lithothamnium Limestone"; WAGNER, 1998; RASSER et al., 1999), passing into slope sediments. Similar sediments have been observed, squeezed in between the allochthonous parts of the Molasse, within the nappes of the Rhenodanubian Flysch Zone and the Northern Calcareous Alps. Sediments of the southern part transgressively overly the nappe complex of the Northern Calcareous Alps (Wagner, 1998; Steininger & Wessely, 2000). Massive conglomerates and contemporaneous littoral sediments are known from deep wells (Steininger & Wessely, 2000). The beginning of the Paratethys in the Molasse Basin is marked in the Oligocene. This phase is characterized by a remarkable facies change, with mainly fine-layered to laminated sequence with nannochalks and diatomites (Wagner, 1998; Krhovsky et al., this volume; Roegl et al., this volume), and by a massive bloom of a few nannoplankton and planktic foraminiferal taxa and an endemic mollusc and ostracode fauna, which thrived in an euxinic-sapropelic environment in the Early Oligocene (Steininger, 1999; Steininger & Wessely, 2000; Krhovsky et al., this volume). This entire sedimentary cycle is confined to the deeper parts of the basin, overlain by younger sediments and only known from deep wells (Wagner, 1996). In the Early Oligocene, the basin strongly subsided to greater water depths and nearly reached the maximum amounts of tectonic subsidence (Genser et al., 1998; Schlunegger & Jordan, 1997). Sediments range from terrestrial to shallow marine clastics and limestones to marls of a deeper shelf. From the Late Eocene to the Early Oligocene the basin subsided quickly to water depths of several hundred meters, indicated by the deposition of dark shales to marls (Genser et al., 1998). They are overlain by the Rupelian (32–28 Ma) sequence of light chalky marls, overlain by banded marls and then by silty shales to marls (Krhovsky et al., this volume). These pass into sandstones and conglomerates towards the south, which are derived from the rising Alpine mountain chain. Along the northern rim of the present basin and along the spur of the Bohemian Massif, terrestrial deposits prevailed. In the Rupelian and Chattian (Tab. 1), the basin configuration in the northern part remained essentially the same, still characterized by terrestrial to shallow marine conditions along its northern margin, and an open marine flora and fauna (Steininger, 1999; Steininger & Wessely, 2000). In its southern part thick turbiditic fans were shed into the basin. At the same time the southernmost part of the basin was overridden by the advancing nappe complex, incorporating also Molasse sediments (Malzer et al., 1993). With the early Neogene the orogenic wedge reached essentially its present position and the basin axis shifted to the north. The basin shows northward transgression; the spur of the Bohemian Massif shows a first major phase of subsidence. The Neogene evolution is characterized by continuous shallowing of the basin, uplift and subsequent erosion.

From subsidence analysis a major flexuring event can be reconstructed from the Late Eocene to the Early Miocene (Aquitanian). This phase is characterized by a later onset of subsidence from north to south, increasing rates of subsidence with time, and a strong N-S asymmetry in subsidence rates with higher rates in the south, closer to the orogenic front. This pattern was related to a flexure caused by the collision of the European and Adriatic plates, and the onset of thrusting. The northward advance of the nappes and the increase of internal deformation account for the increasing rates of subsidence with time (Zweigel, 1998).

Intra-montainous Paleogene Molasse sediments can be observed in the Lower Inn Valley along the Inntal Fault (Figs. 1, 2) E of Innsbruck (Häring and Reith im Winkel) (Ortner, 1996; Ortner & Sachsenhofer, 1996; Ortner et al., 1999; Steininger & Wessely, 2000; Ortner & Stingl, this volume) and indicate a transgression from the Molasse into the NCA (Faupl & Wagreich, 2000). These sediments were deposited in a "piggy back" basin with a sequence of fluviatile/limnic/paralic to marine Oligocene sediments, grading into limnic/fluviatile deposits (Steininger & Wessely, 2000). Sedimentation in the Lower Inn Valley area began in the Priabonian on the previously folded and deeply eroded rocks of the Northern Calcareous Alps. In the area between Reith im Winkel and Rattenberg sedimentation began in the Rupelian (Ortner & Sachsenhofer, 1996; Ortner et al., 1999). The transgression is strongly overprinted by tectonic subsidence of the basin due to transtension along the sinistral Inntal Fault (Figs. 1, 2). The falling sea-level in the Late Rupelian coincides with the start of fluviatile coarse clastic sedimentation in a paleo – Inn valley. Upper Eocene and Oligocene deposits in the Lower Inn Valley area show vitrinite reflectance values, that require a thick Lower Miocene sedimentary cover (Ortner & Sachsenhofer, 1996). Temperature estimates from stable isotopes are in accordance with a 1500 m thick sedimentary sequence, leading to maximum temperatures of about 90° C during the Early Miocene. Sedimentary thicknesses seem to be controlled by rates of erosion in the Tauern Window area and adjacent regions. Strong periods of erosion from Rupelian to the Early Miocene were interrupted by slower rates in the intervening period, when several generations of ancient land surfaces evolved (Ortner & Sachsenhofer, 1996).

3.5. Intrusion of tonalites and related magmatism along the Periadriatic Fault

Oligocene magmatic intrusions are widespread along the Periadriatic Fault (Figs. 1, 2). In Austria, these intrusions comprise (from east to west) the granites and tonalite gneisses of the Karawanken mountains, of Eisenkappel, and the granites and granitic dykes of Nötsch (Karl, 1959; Scharbert, 1975; Exner, 1976). Their geochemistry indicates a mantle origin and a strong crustal contamination (von Blanckenburg & Davies, 1995). Most of the Late Oligocene intrusions took place during a very short time interval between 33 and 29 Ma (e.g., Scharbert, 1975; Gratzer, 1984; von Blanckenburg & Davies, 1995). Geobarometric calculations ranging from 7–3 MPa indicate intrusion depths of 20–10 km (Gratzer, 1984). Andesitic volcanism took place probably around 33 Ma. Its feeder dikes are known with certitude. Among nine dated mafic dykes, three younger ages of between 24 and 27 Ma have been reported (Deutsch, 1984). Remnants of the volcanic rocks have also been observed as pebbles within the Alpine Molasse (Frisch et al., 1998).

The tonalites show features of synmagmatic deformation producing a very strong (locally mylonitic) fabric found at the base of the plutons and subsequently led to the large-scale folds that shortened this contact. Final intrusion, back folding and initial stages of back thrusting along the Periadriatic Lineament are contemporaneous and related to ongoing N-S shortening (Nemes et al., 1997; Neubauer et al., 1998; Kurz & Unzog, 2000).

3.6. Beginning of orogen-parallel extension and lateral extrusion

The concept of orogen-parallel extension and lateral extrusion (Ratschbacher et al., 1991) is in agreement with both differences in crustal thickness and the actual topographic pattern of the Eastern Alps. The eastward decrease in elevation along the strike of the Eastern Alps and the Styrian Basin (Fig. 1), which graduates into the Pannonian Basin to the E goes conform with crustal thinning. Hence, lateral extrusion plays an important role in shaping the actual surface of the Eastern Alps (FRISCH et al., 1998, 2000). Lateral extrusion mainly occurred in Early and Middle Miocene time (approx. 23–13 Ma) and followed Eocene-Oligocene collision and nappe stacking. In contrast to the Western Alps, strike-slip and normal faults dominate the Late Oligocene-Miocene tectonic style of the Eastern Alps (RATSCHBACHER et al., 1991; DECKER et al., 1993; KURZ & NEUBAUER, 1996; LINZER et al., 1997; PERESSON & DECKER, 1997a, b; WANG & NEUBAUER, 1998). Contraction along sub-horizontal faults terminated in the Late Oligocene – Early Miocene in the foreland, like the Molasse Basin (ROYDEN et al., 1982; ROYDEN, 1988, 1993; RATSCHBACHER et al., 1991; DECKER et al., 1993). Thrusting-related N-S compressional structures encompass large-scale folds with E-W trending axes and E-W oriented thrust branch and cut-off lines, with south-dipping ramps in the Rhenodanubian Flysch (DECKER et al., 1993). These compressional structures result from N-S shortening during the Late Eocene to Early Miocene (DECKER et al., 1993; LINZER et al., 1997). Strike-slip faults and extensional structures depicting N-S compression and E-W extension formed during eastward lateral extrusion (GENSER & NEUBAUER, 1989; RATSCHBACHER et al., 1991; DECKER et al., 1993; KURZ et al., 1994; KURZ & NEUBAUER, 1996; FRISCH et al., 1998). Deformation resulted in the eastward lateral motion of the Eastern Alps, that started during the Miocene (DECKER et al., 1993; NEMES et al., 1995, 1997; DECKER & PERESSON, 1996; PERESSON & DECKER, 1997a, b; LINZER et al., 1997; WANG & NEUBAUER, 1998).

This evolution is also documented by the formation of several sedimentary basins during the Miocene (KUHLEMANN et al., 2001), that are related to strike-slip faults, like the Vienna Basin (ROYDEN et al., 1982; DECKER et al., 1993; PERESSON & DECKER, 1997a, b) (Fig. 1), the Klagenfurt basin (e.g., POLINSKI, 1991; POLINSKI & EISBACHER, 1992; NEMES et al., 1997) (Fig. 1), and some minor basins along the Salzach-Ennstal-Fault (e.g., the Wagrain basin) (e.g., EXNER, 1979; WANG & NEUBAUER, 1998). Intra-Alpine Paleogene basins have been formed along the Inn valley. These are related to a segment of the Inntal Fault (Embach Fault) (Figs. 1, 2) (EISBACHER et al., 1990; EISBACHER & BRANDNER, 1995, 1996; DECKER et al., 1993). Along this fault Paleogene sinistral displacement of approx. 20 km has been oberved (EISBACHER & BRANDNER, 1996). Oligocene displacement is also constrained by fission track data (FUEGENSCHUH et al., 1997). In the southern part of the Eastern Alps, Paleogene contraction produced NW-SE- trending fold-thrust structures and related sinistral cross faults. Miocene deformation produced NW- and SE-directed thrusts and NW-striking dextral cross faults (POLINSKI & EISBACHER, 1992; SPRENGER & HEINISCH, 1992; SPRENGER, 1996).

The southern limit of Alpine Cretaceous metamorphism (SAM) (HOINKES et al., 1999) represents an approximately east-trending system of faults with a multiphase deformation history, that juxtapose Austroalpine units with a strong, amphibolite and eclogite grade metamorphic imprint to the north against very-low grade Austroalpine units in the south. The SAM consists of (from the west to the east) the Peio-, Passeier-Jaufen-, Defereggen-Antholz-Vals- (DAV) (Fig. 1), Isel-, Zwischenbergen-Wöllatratten-, Ragga-Teuchl-, Siflitz-, and Viktring fault zones. The western sector of the SAM represents a zone of important, mostly Oligocene sinistral strike-slip displacement with an oblique slip component (KLEINSCHRODT, 1987; GENSER & NEUBAUER, 1989). The eastern sector (east of the Isel fault) represents a zone of Late Cretaceous sinistral strike-slip shear zones with a subordinate component of normal displacement.

The development and the geometrical arrangement of strike-slip and normal faults in the central part of the Eastern Alps is highly constrained by the lithospheric structure of the Eastern Alps. In particular, the abnormal crustal thickness in the area of the Tauern Window (NEUBAUER & GENSER, 1990; NEUBAUER et al., 1999, 2000a), which is related to the accretion of Penninic continental units, governed the initiation of confining strike-slip faults to the N and the S of the Tauern Window, which was still hidden beneath the Austroalpine Nappe stack during the Paleogene (FRISCH et al., 1998) (Fig. 2). In this area crustal thicknesses of 50–60 km are documented (ARIC et al., 1987; MEURERS et al., 1987; NEUBAUER et al., 1999, 2000b). Especially the formation of both confining E-W trending strike-slip faults (Salzach-Ennstal Fault; DAV) (KLEINSCHRODT, 1987; GENSER & NEUBAUER, 1989; RATSCHBACHER et al., 1991; NEMES et al., 1995; WANG & NEUBAUER, 1998) and N-S trending low-angle normal faults at the eastern and western margins of the Tauern Window (SELVERSTONE, 1988; BEHRMANN, 1988; GENSER & NEUBAUER, 1989) are constrained by the northern and southern limits of the former Zentralgneiss Terrane. According to the published age data and the documented structural evolution, ductile faulting along several faults started already during the thermal peak of metamorphism in the eastern part of the Tauern Window, between 27 and 29 Ma (KURZ & NEUBAUER, 1996), within a N-S oriented compressional field. Following REDDY et al. (1993) the deformation style changed from shortening to extension around 19 Ma.

Actually, the main part of lateral extrusion is documented during the Miocene and comprises both the exhumation of the Tauern Window (e.g., NEUBAUER et al., 1999, 2000b), of the Rechnitz Window (e.g., RATSCHBACHER et al., 1990; DUNKL & DEMENY, 1997; DUNKL et al., 1998), and contemporaneous crustal stretching and formation of the Styrian and Pannonian basins (ROYDEN et al., 1982; ROYDEN, 1988, 1993; ROYDEN & BURCHFIELD, 1989; BERGERAT, 1989; NEUBAUER & GENSER, 1990; LILLIE et al., 1994; EBNER & SACHSENHOFER, 1995; PERESSON & DECKER, 1997a, b; SACHSENHOFER et al., 1997; NEMCOK et al., 1998).

4. GEODYNAMIC EVOLUTION AND PALEOGEOGRAPHIC SITUATION

Based on the structural, petrological, geochronological, and sedimentological-stratigraphical data summarized in the previous chapters, a plate tectonic-geodynamic interpretation of the evolution of the Eastern Alps during the Paleogene is given below. This interpretation basically follows the recent descriptions by FROITZHEIM et al. (1996), KURZ et al. (1998b), FAUPL & WAGREICH (2000), and NEUBAUER et al. (2000a). For recent Meso- and Cenozoic paleogeographic models of the Alps the reader is referred to, e.g., TRÜMPY (1975, 1983, 1988, 1997a), FRISCH (1977b, 1979, 1980), CHANNELL et al. (1979), TOLLMANN (1980, 1987a, b), ROYDEN et al. (1982), FRANK (1987), ROYDEN (1988), ROYDEN & BALDI (1988), DEWEY et al. (1989), STAMPFLI (1993, 1996), OBERHAUSER (1995), FROITZHEIM et al. (1996), SCHMID et al. (1996, 1997), STAMPFLI & MARCHANT (1997), STAMPFLI et al. (1998), SCHMID & KISSLING (2000).

A palinspastic reconstruction of the Eastern Alps during the Oligocene, based on FRISCH et al. (1998, 2000), prior to the onset of orogen-parallel lateral extrusion is shown

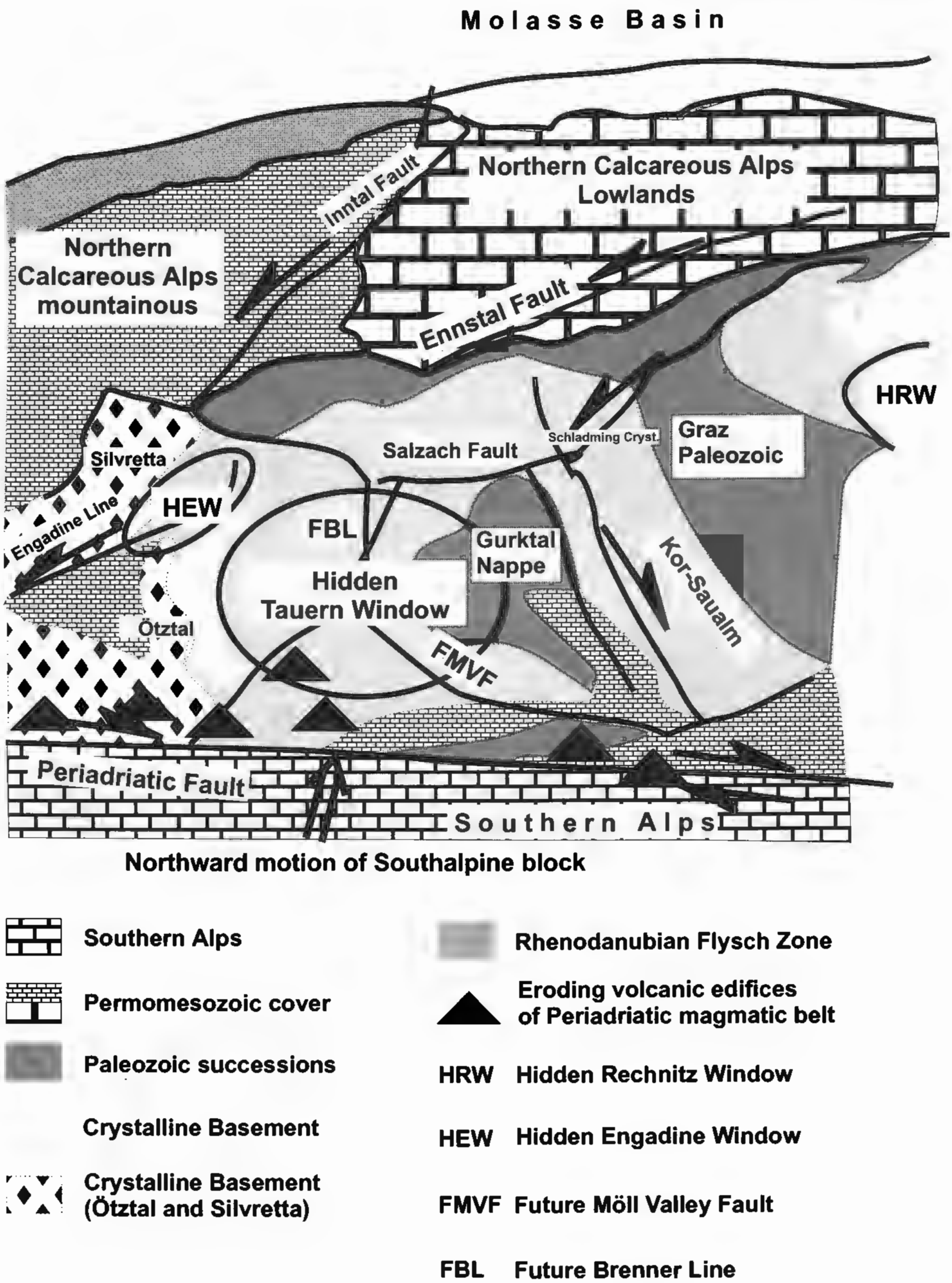


Fig. 3: Palinspastic reconstruction of the Eastern Alps for the pre-extrusion situation in Oligocene time around 30 Ma (following Frisch et al., 1998).

in Fig. 3. This reconstruction shows, that the Penninic units within the Tauern Window are buried in the footwall of the Austroalpine nappe complex. This indicates, that the Brenner Fault, the Katschberg and the Möll Valley Fault (Figs. 2, 3) had not been activated yet. Tectonic activity is only proven for the Inntal Fault, parts of the Salzach Fault, the Lower Engadine Line, and the Periadriatic Fault. The western part of the Northern Calcareous Alps had already been elevated above sea level, and forms a topographic mountain chain. This paleogeographic arrangement of the lithotectonic units results from subsequent distinct tectonic processes during the Late Cretaceous, the Paleocene and Eocene:

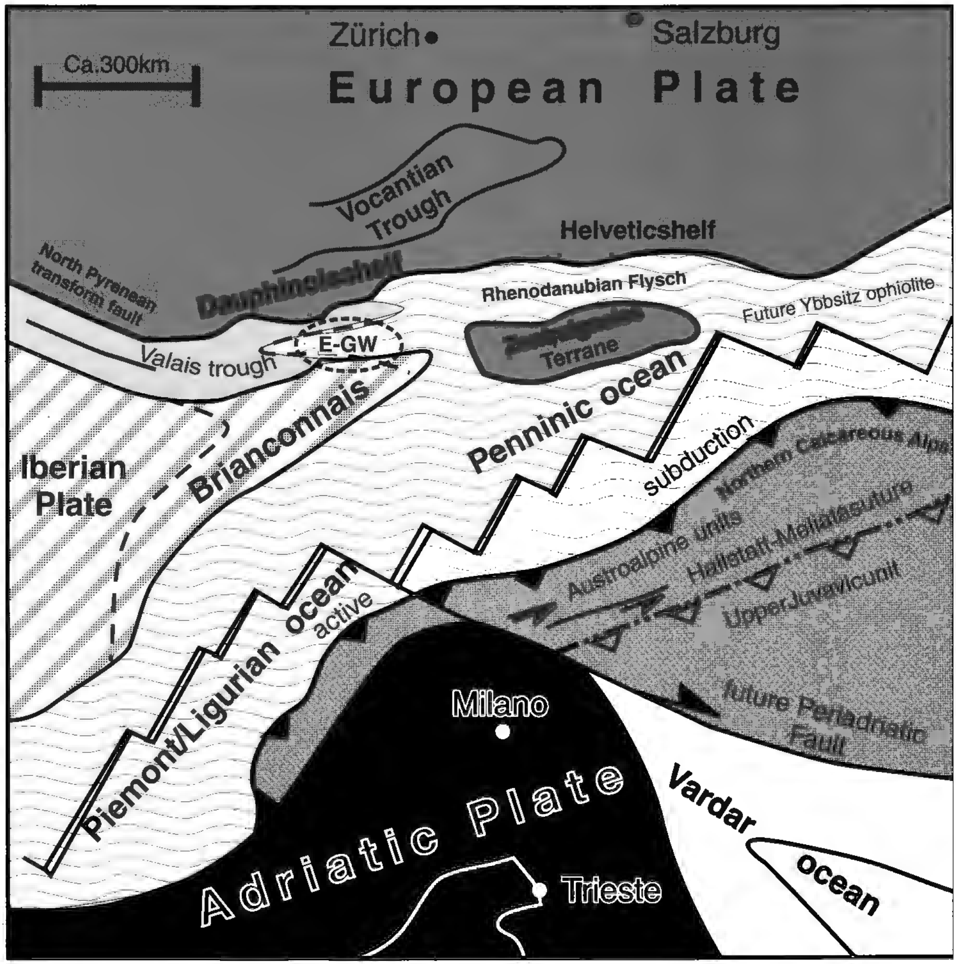


Fig. 4: Simplified paleogeographic map of the Alps representing roughly the situation during the Late Cretaceous/Early Paleogene (based on CHANNELL & KOZUR, 1997; STAMPFLI & MARCHANT, 1997; NEUBAUER et al., 2000a). The arrows indicate inferred directions of plate movement. E-GW: Lower Engadine and Gargellen Window.

During the Late Cretaceous and the Paleogene subduction of the Penninic ocean and the subsequent collision between the Penninic and the Austroalpine units is documented. Within the Penninic units a change from mainly oceanic to terrigenous sedimentation is obvious. This sedimentary evolution indicates the closure of the Penninic oceanic basin (Figs. 4 to 6). The main amount of clastic material is exposed within the Matrei Zone, which forms an accretionary wedge in the hanging wall of the subducting oceanic lithosphere of the Penninic oceanic basin (Frisch et al., 1987) (Fig. 6). Within the future Venediger Nappe Complex (Zentralgneiss Terrane) the Cretaceous was dominated by terrigenous sediments deposited within a pelagic environment (Kiessling, 1992) (Kaserer Group). Eocene sequences are only documented from the Lower Engadine and the Gargellen Window. This gives rise to the assumption that the closure of the Penninic oceanic basin occurred during the Late Cretaceous in the eastern and central parts of the Eastern Alps (Tauern Window), and reached up to the Paleocene in the western parts. Subduction comprises subsequent top-to-the N emplacement of the Glockner Nappe and final collision between the Austroalpine and Penninic continental units in the Eocene, including the accretion and subduction of the Middlepenninic unit west of the Tauern Window (exposed in the Tasna Nappe).

Indirect evidence of of this process may be given by the formation of the Gosau basins (Fig. 6) (e.g., Faupl et al., 1987). The formation of these basins could have been

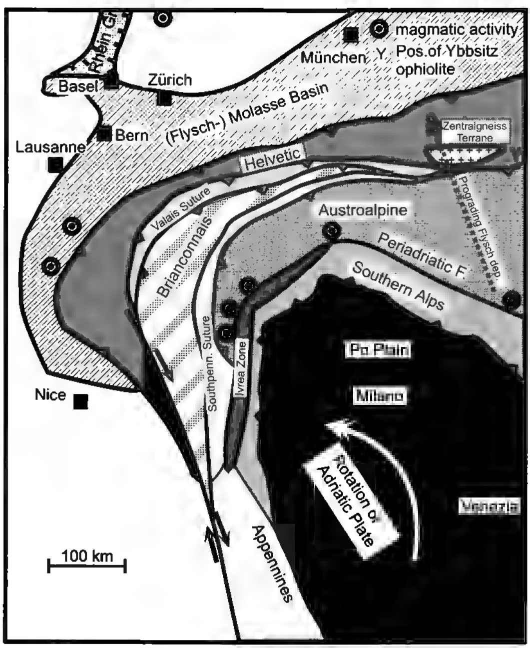


Fig. 5: Simplified paleogeographic map of the Alps representing roughly the situation during the Early Oligocene (based on Stampfli & Marchant, 1997).

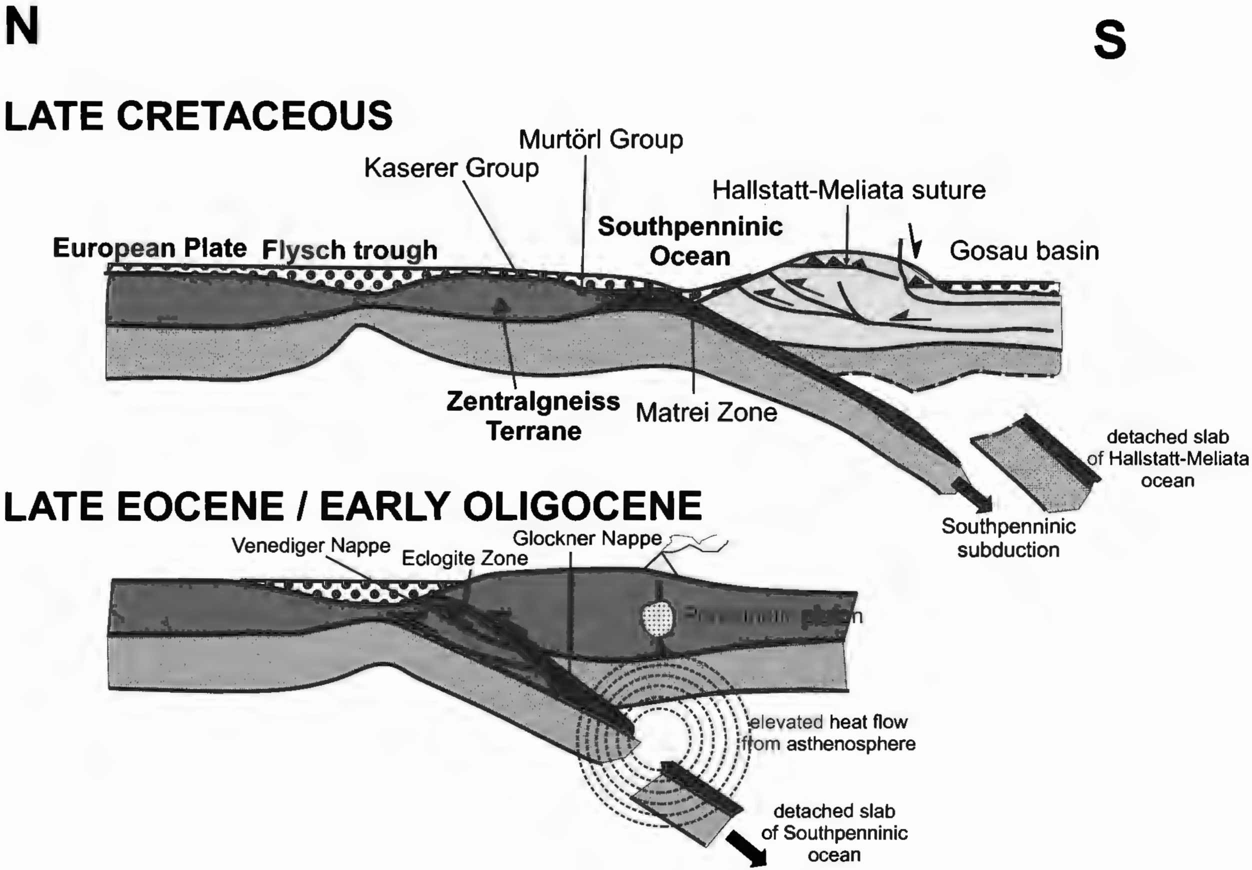


Fig. 6: Plate tectonic evolution of the Eastern Alps from the Late Cretaceous (top) to the Eocene/Oligocene (bottom) (based on VON BLANCKENBURG & DAVIES, 1995; FROITZHEIM et al., 1996; SCHMID et al., 1996; FAUPL & WAGREICH, 2000; NEUBAUER et al., 2000a; STAMPFLI & MARCHANT, 1997; KURZ et al., 1998b).

triggered either by footwall plate accretion of either the Middlepenninic unit (Genser et al., 1996; Kurz et al., 1999), by the stacking of the lowermost Austroalpine units at the southern margin of the Penninic oceanic basin (Neubauer, 1994b), or the roll-back of the subducting Southpenninic oceanic lithosphere (Froitzheim et al., 1997). These processes were compensated by extension within the hanging wall plate (Ratschbacher et al., 1989; Neubauer et al., 1995; Froitzheim et al., 1997; Willingshofer, 2000).

Subduction and accretion of the Lower Austroalpine is documented by blueschist facies metamorphism within this unit (Fig. 2) (Enzenberg, 1967; Enzenberg-Praehauser, 1976; Dingeldey & Koller, 1994; Dingeldey et al., 1997). However, blueschist facies metamorphism is documented between approx. 50 and 40 Ma (Dingeldey et al., 1997), which would require a time gap of several million years from subduction to subsequent high-pressure metamorphism and exhumation. Additionally, the Lower Austroalpine Innsbruck Quartzphyllite has already been exhumed and was cooled below temperatures in the range of 200° C from 67 Ma to 35 Ma (Fügenschuh et al., 1997). Therefore, these data may be better interpreted to record the deformation and associated metamorphic imprint during the final emplacement of the Austroalpine nappe complex onto the Penninic nappe complex (Fig. 2), and its subsequent exhumation.

Sedimentation of the Gosau deposits ceased in the Eocene, which may coincide with the termination of the subduction of the Middlepenninic unit. While the sedimentation terminated on top of the Austroalpine upper plate, deposition of the Molasse sediments was initiated in the Eocene, which indicates the beginning of the formation of a topographic mountain chain (e.g., Frisch et al., 1998). Subsidence of the Molasse basin is related to the flexure of the European lower plate, due to the load of the Austroalpine upper plate (Genser et al., 1998; Zweigel, 1998; Rasser, 2000). The deposition of turbiditic and chaotic sediments in the Ultrahelvetic Unit during the Paleocene and Eocene indicates that the collisional front had already approached the southern parts of the European foreland. This resulted in the creation of a peripheral bulge, which acted as a source area (Faupl & Wagreich, 2000).

Actually, the future Middlepenninic (Brianconnais) unit, and the Zentralgneiss Terrane (Fig. 6) were part of the stable European continent from the Permian to the Jurassic. Lithofacial similarities between the Helvetic unit and the Hochstegen unit of the Venediger Nappe (Zentralgneiss Terrane) indicate, that these units were part of the Helvetic paleogeographic facies realm during the Triassic and Jurassic (e.g., Frisch, 1975). Subsidence of the Rhenodanubian Flysch basin started in the early Late Cretaceous (Winkler, 1988; Egger, 1989, 1990) (Figs. 5, 6). Since this time, a northern and a southern basin could be distinguished, separated by a continental block (either a Middlepenninic block or the Zentralgneiss Terrane). The southern oceanic basin (Southpenninic), which is partly preserved within the Glockner Nappe of the Tauern Window, was already to be consumed, as indicated by the deposition of flysch sediments within the Matrei Zone (Fig. 6). During this subduction process, the flysch deposition prograded continuously northwards from the Matrei Zone to the Rhenodanubian Flysch basin (Figs. 4–6). Based on these assumptions, the subdivision into several Penninic paleogeographic realms is only valid during the Late Cretaceous and the Paleogene, as long as the Southpenninic oceanic basin was not subducted completely. East of the area of the actual Tauern Window, both the Southpenninic basin and the Rhenodanubian Flysch basin merged into a single basin (Egger, 1989, 1990, 1992) because of the termination of the Zentralgneiss Terrane to the east (Figs. 4, 5). This is, for example, documented in the Ybbsitz ophiolite, which forms the base of turbiditic deposits (e.g., DECKER, 1990) (Figs. 4, 5). This is similar to the evolution in the Arosa Zone in eastern Switzerland (e.g., RING et al., 1988a, b, 1989, 1990; FAUPL & WAGREICH, 1992a, 2000), and hence, the Matrei Zone.

After the accretion of the Zentralgneiss Terrane the subduction was blocked, and the zone of active subduction was transferred to the north and resulted in the consumption of the Rhenodanubian Flysch basin, which is supposed to be in a Northpenninic paleogeographic position (FAUPL & WAGREICH, 2000). This is indicated by the cessation of sediment deposition in the Rhenodanubian Flysch during the Eocene (WINKLER, 1988; EGGER, 1989, 1990, 1992; ELIAS et al., 1990), the formation of folds and thrusts within the Flysch Zone, and subsequently within the Molasse Zone. This final phase of collision is recorded by backthrusting along the Periadriatic Fault, too (RATSCHBACHER et al., 1990; POLINSKI & EISBACHER, 1992; SPRENGER & HEINISCH, 1992; SPRENGER, 1996; NEMES et al., 1997).

The structural and metamorphic evolution of the Penninic units within the Tauern Window documents the burial history and the subsequent exhumation (FRANK et al., 1987; BEHRMANN, 1990; SELVERSTONE, 1985, 1993; KURZ et al., 1996; KURZ & NEUBAUER, 1996). The detailed tectonometamorphic evolution of these units during the Paleogene will be described by KURZ et al. (this volume). Subduction is best evidenced by eclogite facies metamorphism within the Eclogite Zone and the Glockner Nappe, and blueschist facies metamorphism which affected the entire Penninic nappe pile of the Tauern Window. During the subduction process a main part of the oceanic lithosphere (the future Glockner Nappe) and the outermost continental slope (the future Eclogite Zone) are subducted to depths of 60–70 km proven by eclogite facies metamorphism. These high-pressure units are emplaced to the N onto the Zentralgneiss Terrane along a S-dipping thrust in the course of the accretion of this unit, and the collision with the Austroalpine block. Sm-Nd garnet ages of ca. 42 Ma from the Eclogite Zone are cited by DROOP et al. (1990) and INGER & CLIFF (1994). Sm-Nd dating of garnet from the Upper Penninic Cima Lunga Nappe (Lepontine Dome, Swiss Alps) yielded consistent mineral ages of ca. 40 Ma (BECKER, 1993). 52±18 Ma are reported from the Zermatt-Saas ophiolite (BOWTELL et al., 1994; AMATO et al., 1999; RUBATTO et al., 1998), approx. 60 Ma are reported from the Monviso ophiolite (CLIFF et al., 1998), which are in a similar tectonic position as the Glockner Nappe. In the Western Alps, ultra-high pressure metamorphism is dated at about 38 Ma in the Dora Maira Massif (HENRY et al., 1993), which is in a similar tectonic position as the Zentralgneiss Terrane. As the future Eclogite Zone was positioned south of the Zentralgneiss block, this implies that the Eclogite Zone was already subducted, too. At which time the crustal level of eclogite facies conditions was reached is poorly constrained.

After the accretion of the main part of the Zentralgneiss Terrane subduction of Penninic units was blocked. Indirect evidence of accretion and collision may be given from the structural evolution of the Gosau basins in the hanging wall Austroalpine plate, which are affected by N-S shortening at ca. 65 Ma (EISBACHER & BRANDNER, 1996). This permitted heating of the accreted continental crust until conditions of amphibolite to greenschist facies metamorphism were reached ("Tauern Crystallisation") at approx. 30 Ma. Blocking of subduction triggered the initiation of the subduction of the North-Penninic oceanic basin further to the N. During this phase the Penninic units of the Tauern Window are penetratively deformed due to transpression and top-to-the W simple shear (Kurz et al., 1996). Later top-to-the W shear is contemporaneous to decompression along the exhumation path and furthermore documents a change of the kinematic boundary conditions. This phase is interpreted to record the emplacement of the Penninic nappe stack onto the European foreland. It developed at or slightly prior to upper greenschist to amphibolite facies metamorphic overprint (ca. 550° C, 6–7 MPa in the central southern part of the Tauern Window) (e.g., Dachs, 1990). The change from top-to-the N to top-to-the W emplacement documents a counterclockwise kinematic path within the Penninic units of the Tauern Window. This path is in accordance with the counterclockwise rotational path of the Adriatic plate during its northward movement during the Late Cretaceous and the Cenozoic (Channell et al., 1979; Platt et al., 1989; Behrmann, 1990).

The doming and exhumation history of the central part of the Eastern Alps is characterized by deformation partitioning and shear localization along the margins of the Tauern Window. Strike slip faults form the northern, the southern, and the south-eastern margins of the Tauern Window, low-angle normal faults form the eastern and western margins (Figs. 1, 2). The central part of the Tauern Window is dominated by folding. This phase is extensively discussed by Selverstone (1988), Behrmann (1988), Genser & Neubauer (1989), Kurz et al. (1996), and Kurz & Neubauer (1996). Doming is driven by the northward indentation of the Southalpine block and dextral transpression along the Periadriatic Fault (Lammerer, 1988) (Figs. 2, 3). The succession of deformation events is interpreted to result from oblique collision of a continental Penninic plate in a footwall position and the Austroalpine upper continental plate after consumption of the oceanic Penninic domain (Glockner Nappe) (Fig. 5). Convergence and collision occurred within a NNE-SSW directed contractional regime. Oblique convergence results in a number of effects that include partitioning of deformation into orthogonal shortening and displacement parallel to the leading edge of plate margins. Collision resulted in frontal accretion and nappe stacking (top-to-the N) within the lower plate and subsequent, orogen-parallel simple shear deformation (top-to-the W) within a transpressional regime, and subsequent doming and the activity of confining E-W trending strike-slip faults, documenting orogen-parallel escape during continued oblique N-S convergence (e.g., Peresson & Decker, 1997a, b; Neubauer et al., 1999, 2000a).

Contemporaneously, both basaltic and granitoid magmatism occur between 42 and 25 Ma (von Blanckenburg & Davies, 1995; Frisch et al., 1998), following the closure of oceanic basins by subduction and continental collision (von Blanckenburg & Davies, 1995, 1996; Schmid et al., 1996, 1997; Froitzheim et al., 1996). Actually, all Alpine magmatic rocks intruded almost synchronously along a strike-slip fault, the Periadriatic Fault (Figs. 2, 3). The occurrence of these magmatites may be interpreted in terms of breakoff of a subducted slab of oceanic lithosphere (e.g., the Southpenninic) (von Blanckenburg & Davies, 1995) (Figs. 5, 6). Once initiated, rapid lateral migration of slab breakoff results in a linear trace of magmatism in locally thermal weakened crust. Breakoff will lead to heating of the overriding lithospheric mantle by upwelling asthenosphere, melting of its enriched layers, and thus to magmatic activity. Geochronological ages from Penninic high-pressure units (see above) indicate that subduction took place at approx. 55–40 Ma, followed by exhumation and uplift at 40–35 Ma (von Blanckenburg & Davies, 1995). The short time interval between this uplift and the onset of magmatism may indicate that both processes have been induced by the breakoff.

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