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Structural, Metamorphic and Geochronological Studies in the Reisseck and Southern Ankogel Groups, the Eastern Alps

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With 4 Plates ("Beilagen 7—10") and 63 Figures

Schlüsselwörter
Ostalpen
Tauernfenster
Zentralgneis
Schieferhülle
Metamorphose
Strukturuntersuchungen
Geochronologie

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Summary

An area of about 250 km^2 extending from the Mölltal on the south east margin of the Tauern Window to the Hochalm in the window centre has been the subject of a structural, metamorphic and isotopic age study, and a map is presented at a scale of 1 : 25,000. The area mapped lies entirely within the Pennine Zone of the Eastern Alps. An intensely deformed "cover" sequence of metasedimentary rocks overlies an older basement complex. Although generally distinguishable on lithological grounds the basement and cover are also distinguished by the following criteria: (1) the basement rocks commonly display evidence of more than one episode of intense folding; the cover shows only one main folding episode although locally a later group of gentle flexures is developed; (2) the basement rocks are virtually free of carbonate while it is a common constituent of the cover rocks; (3) the basement is commonly injected by aplitic veins; these are not found in the cover.

The basement comprises two main lithological units: (a) banded gneisses, amphibolites, paragneisses and schists, here termed the Inner Schieferhülle; (b) the Zentralgneis — a more or less homogeneous and gneissose group comprising granodiorites with subsidiary leucogranites and tonalites; a few members of this group are virtually unfoliated and have igneous textures.

The cover sequence comprises greenschists, amphibolites, calc-schists, pelitic schists, marbles dolomites and quartzites and is known as the Peripheral Schieferhülle. The cover sequence is thought to be largely Mesozoic in age while the basement is pre-Mesozoic. Isotopic evidence (Rb/Sr whole rock) indicates the intrusion of large granodioritic bodies in the Permian and these seem to be among the youngest basement rocks; they cut an earlier sequence of metamorphosed igneous and sedimentary rocks which had by that time already undergone one or more major folding deformations.

The cover sequence may have been deposited, at any rate in part, on top of the basement complex; alternatively it may have been carried onto it in Tertiary times. The main Tertiary deformation affected different units differently. There appears to have been an important horizontal component to the deformation associated with the overriding of the NE-moving East Alpine nappes; superimposed on this was a series of differential vertical movements of the basement which were partially contemporaneous with the horizontal movements but continued after them. The basement complex was sheared into a number of schuppen which were transported over each other in a northeasterly direction. The schuppen were themselves folded both on a scale of kilometres and on smaller scales — early folds were intensified and new folds developed with NW—SE trending axes and SW dipping axial surfaces. Most of the Permian intrusives acquired a foliation at this time although in a few places the Tertiary deformation had little effect and pre-Mesozoic intrusive relations are clearly preserved. The Gössgraben Kern is a domical structure within the thrust slices which brings to the surface a synformal body of metasediments surrounded by orthogneisses.

The cover rocks are nowhere intimately interfolded with the basement and the contact is everywhere a zone of intense shearing; the basement rocks record strains of over 80% in this region by comparison with values of 40 to 70% away from the contact. Within the cover, isoclinal folding overturned to the NE is common on a mesoscopic scale; on a larger scale, however, the Peripheral Schieferhülle appears to have undergone extreme tectonic disordering; few lithological units may be traced any great distance along strike and the sequence has the appearance of a tectonic mélange. Late vertical movements of the Basement produced a series of irregular domes (Kernen) and basins (Mulden).

After the main folding movements were complete the region reached a thermal peak for metamorphic recrystallization and the cover rocks reached the greenschist to amphibolite grades of regional metamorphism; staurolite, chloritoid and kyanite bearing mineral assemblages are locally developed. The temperature and pressure of metamorphism at the cover/basement interface in the area is estimated at 550° C and 5 kb. This pressure was reached at a time of maximum burial under the East Alpine allochthonous thrust sheets which are today eroded in the area of the Tauern Window but are exposed around its margins; ophiolitic extrusions excepted, no igneous activity of Mesozoic or Tertiary age is known within the Tauern Window.

In the late ? Miocene the region underwent rapid uplift and faulting occurred along the line of the present-day Mölltal to delineate the margin of the Tauern Window. K/Ar cooling ages on micas give values close to 20 my.

Zusammenfassung

Es wurde ein 250 km^2 großes Gebiet, welches vom Mölltal, der SE-Begrenzung des Tauernfensters, bis zur Hochalm Spitze in dessen Zentrum reicht, in Hinblick auf Tektonik, Metamorphose und isotopische Altersbestimmungen untersucht. Eine Karte im Maßstab 1 : 25,000 wurde erstellt. Das kartierte Gebiet liegt gänzlich im Pennin der Ostalpen. Eine intensivst durchbewegte „Hülle" von Paragesteinen überlagert ein älteres Grundgebirge. Dieses unterscheidet sich bereits lithologisch von den Hüllserien, ist aber darüber hinaus durch folgende Kriterien charakterisiert:

1. Der Basalkomplex läßt allgemein mehr als eine intensive Faltungsphase erkennen. Die Hüllgesteine zeigen bloß eine Hauptfaltung, nur lokal sind gewisse jüngere Flexuren entwickelt.
2. Die Basalserie ist praktisch frei von Karbonat, während dieses ein verbreiteter Bestandteil der Hüllgesteine ist.
3. Der Basalkomplex ist allgemein aplitisch injiziert, nicht aber die Hüllserien.

Der Basalkomplex umfaßt zwei Lithoeinheiten: a) Bändergneise, Amphibolite, Paragneise und Schiefer, welche hier Innere Schieferhülle genannt werden; b) der Zentralgneis — eine Gruppe mehr oder weniger homogener Gneise, die Granodiorite mit untergeordneten Leukograniten und Tonaliten umfaßt; einige wenige Gesteine dieser Gruppe sind praktisch ungeschiefert und zeigen Erstarrungsgefüge.

Die Hülle wird allgemein als Periphere Schieferhülle bezeichnet. Sie umfaßt Grünschiefer, Amphibolite, Kalkschiefer, phyllitische Schiefer, Marmore, Dolomite und Quarzite. Das Alter der Hülle wird als überwiegend mesozoisch aufgefaßt, das der Basalserie als vormesozoisch. Absolute Altersbestimmungen (Rb/Sr Gesamtgestein) zeigen, daß im Perm große Granodioritkörper intrudiert sind, die anscheinend die jüngsten Gesteine des Basalkomplexes darstellen. Sie durchschlagen eine Folge metamorpher Eruptiv- und Sedimentgesteine, die damals bereits eine oder mehrere Faltungsphasen mitgemacht hatten.

Die Hüllserie wurde, zumindest z. T. auf dem Basalkomplex abgelagert, oder aber während der Gebirgsbildung im Tertiär aufgeschoben. Die tertiäre Hauptbewegungsphase hat die einzelnen Einheiten unterschiedlich beeinflußt. Im Zusammenhang mit der Überschiebung der Ostalpinen Decken gegen NE war anscheinend eine starke Horizontalkomponente bei der Deformation wirksam; diese Deformation war ihrerseits von einer Serie von vertikalen Differenzialbewegungen des Basalkomplexes überlagert, welche z. T. gleichzeitig mit den Horizontal-Bewegungen erfolgten, diese aber überdauerten. Der Basiskomplex wurde in eine Reihe NE-gerichteter Schuppen zerlegt. Diese wurden selbst noch in Falten bis zur Kilometer-Größenordnung gelegt — ältere Falten wurden verstärkt und neue Falten entwickelten sich nach NW—SE-Achsen und SW-fallenden Achsenebenen. Die meisten permischen Intrusiva erhielten hierbei ihre Schieferung; dennoch war an einigen wenigen Stellen der Einfluß der tertiären Durchbewegung so schwach, daß der vormesozoische Intrusivverband deutlich erhalten geblieben ist. Der Gössgraben-Kern — ein den Schuppen eingeschaltetes domartiges Gewölbe — bringt einen von Orthogneis umgebenen synklinalförmigen Paragesteinskörper empor.

Die Hüllgesteine sind nirgends mit dem Basalkomplex innig verfaltet, sie grenzen stets an einer Zone starker scherender Durchbewegung aneinander. In dieser Zone zeigen die Gesteine der Basalserie eine über 80prozentige Verformung, während fern der Grenzfläche nur Werte von 40 bis 70% festzustellen sind. In der Hülle sind NE-gerichtete Isoklinalfalten von Handstück- bis Aufschlußgröße verbreitet; im großen betrachtet, hat die Periphere Schieferhülle aber eine extreme tektonische Durchmischung mitgemacht. Wenige Lithoeinheiten können über größere Entfernungen im Streichen verfolgt werden, und die Gesteinsfolge macht den Eindruck einer tektonischen Mischserie. Junge Vertikalbewegungen des Basalkomplexes haben eine Reihe unregelmäßiger domartiger Kerne und Mulden erzeugt.

Nach Beendigung der Hauptfaltung erlebte das Gebiet ein Wärme-Maximum. Die Regionalmetamorphose erreichte in den Hüllgesteinen die Grünschiefer- und Amphibolit-Fazies. Örtlich sind Staurolith, Chloritoid und Disthen-führende Mineralvergesellschaftungen entwickelt. Die P/T-Bedingungen der Metamorphose werden an der Grenze zwischen Hülle und Basalkomplex auf 550° C und 5 Kb geschätzt. Dieser Druck wurde zur Zeit maximaler Bedeckung durch die Ostalpinen Schubmassen erreicht. Heute sind diese im Bereiche des Tauernfensters wegerodiert, in seiner Umrahmung sind sie aber erhalten. Im Tauernfenster ist, abgesehen von den Ophiolithen, keine magmatische Aktivität während des Mesozoikums und Tertiärs bekannt.

Im jüngeren (?) Miozän wurde das Gebiet rasch gehoben, und es kam zur Bruchbildung entlang des heutigen Mölltales, dem Rande des Tauernfensters. K/Ar-Abkühlungsalter von Glimmern ergeben Werte nahe 20 Millionen Jahren.

I. Introduction

by E. R. Oxburgh, R. A. Cliff and R. J. Norris

History and Purpose of Present Study

The 'Oxford East Alpine Research Project' had its origin in 1961 when one of the present authors (ERO) visited the Eastern Alps in the company of his colleague Dr. R. St. J. Lambert with a view to discovering the possibilities for work by English geologists in the region. Through the kind offices of the then Director of the Geologische Bundesanstalt, Prof. Dr. H. D. Küpper, they were fortunate enough to be invited to join a Geologische Bundesanstalt field excursion in the Hohe Tauern on which they met many of the geologists working in the Tauern. During the following winter after consultation with Professor E. Clar and Professor Ch. Exner of the University of Vienna, the decision was made to begin field mapping in the area described in the series of papers which follows.

The purpose of the study was to make a detailed investigation of part of a young and well exposed orogenic belt by means of a multi-disciplinary approach; it was hoped to combine the traditional methods of field mapping with work on the geochronology, metamorphic petrology and structural analysis, in the conviction that when these approaches are employed together the return from each of them is many times greater than when used on their own.

The area selected was one in which one of the oldest controversies in the Eastern Alps could be examined — the controversy concerning the existence of the Tauern Window. If the tectonic window was real, this region offered an opportunity to study a young metamorphic area which had undergone overthrusting by old basement rocks on a scale documented nowhere else in the world. The first objective was therefore to see whether it was possible to devise any new tests for the existence of the Tauern Window and then, if appropriate, to see how this large scale overthrusting related to the deformational and metamorphic characteristics of the rocks beneath and to their subsequent uplift. In essence the study involved making a detailed time and motion study of the behaviour of a small segment of the earth's crust over a period of several hundred million years. It was hoped not only to examine times and rates of motion of material but heat transfer processes too. This has been attempted, but many gaps remain to be filled.

From a practical point of view the area selected had a number of advantages; about two and a half kilometres of relief allowed a reasonable three-dimensional control of the large-scale structural relationships. Although mountainous, the country was for the most part very strenuous rather than very difficult of access; it proved possible to do the geology of nearly all the ridge crests without serious climbing. In addition there was virtually nothing in the way of modern geological maps of the area.

Oxburgh began mapping in the summer of 1962 and that year completed a reconnaissance of much of the present area. During the following summer he was assisted in the field by Cliff and in 1964 Cliff undertook a detailed study of the petrology and geochronology of granite gneisses in an excellently exposed, but relatively inaccessible high part of the area (Plate 1); Cliff's studies continued until 1968 when he was granted his doctorate. In 1966 Norris began work in an area south and west of that studied by Cliff; he concentrated initially on understanding the structural relationships which are clearer there than elsewhere in the region. Subsequently he extended his work to detailed petrological and fabric studies. He received his doctorate in 1970. Wright became associated with this work first in 1968 and participated in field mapping in 1968 and 1969. Subsequently he did detailed work on air photographs and map preparation.

The authors were fortunately able to spend some time in the field together in the summer of 1969 and were then able to settle most points of contention. The papers which follow represent an attempt to synthesise and reconcile the views of the contributors who have mapped the different parts of the area.

Methods of Work

In all, over a hundred weeks have been spent in the field (R. A. C. 34; R. J. N. 36; E. R. O. 41). Mapping was carried out on 1 : 10,000 enlargements of the standard 1 : 25,000 and 1 : 50,000 maps (Österreichische Karte) of the area, although a few parts were mapped at smaller or larger scales. Extensive use was made of aerial photographs and in much of the high ground, off the ridges, stations at which observations had been made could be located more rapidly and accurately on the air photographs than on maps. The photographs were, however, of relatively little value for the interpretation of the geology, other than the mapping of mega-joints and fracture zones and the accurate delineation of superficial deposits. In general, except in the Göss Graben Kern, structure is not well expressed in the topography which is largely controlled by the interaction of the foliation and the jointing with the subsequent glaciation.

In very steep areas it often proved more useful to record information on detailed photographic panoramas; these were taken with a telephoto lens from distant ridges to minimize distortion effects.

Measurements of structural elements were made with the Clar Gefügekompass. About 1800 rock samples were collected and taken to Oxford for more detailed study and nearly half of these were examined in thin-section.

A wide variety of other laboratory studies were undertaken on restricted numbers of specimens for particular purposes. These included chemical and isotopic analysis for geochronological purposes, x-ray diffraction, x-ray fluorescence and electron microprobe work during study of the metamorphic parageneses. There was also a small amount of wet chemical analysis.

Plagioclase compositions were measured in thin section on the Universal stage, following the method of Turner (1947), or on the fixed stage by the Michel-Levy method, and by comparing the refractive index with Canada Balsam and quartz. This latter method was found to give rapid and useful results for the range of compositions encountered. In addition, some garnets and some plagioclase compositions were determined on the electron micro-probe.

A limited number of modal analyses were made by point counting. (1400—2000 points).

Abbreviations used in the text

ab	— albite	ky	— kyanite
acc	— accessories	liz	— lizardite
act	— actinolite	mg/ank	— magnesite/ankerite
antig	— antigorite	mic	— microcline
bi	— biotite	mt	— magnetite
cc	— calcite	mu	— white mica
chd	— chloritoid	olig	— oligoclase/andesine
chl	— chlorite	phl	— phlogopite
di	— diopside	plag	— plagioclase (general)
dol	— dolomite	qz	— quartz
ep	— epidote/clinozoisite	sph	— sphene
gt	— garnet	st	— staurolite
hb	— hornblende	tc	— talc
ht	— haematite	tr	— tremolite
ksp	— K-feldspar	zo	— zoisite
U-stage	— Universal Stage (5 axis)	b	— bars
R. I.	— refractive index	kb	— kilobars

P — pressure; T — temperature; P_{vap} — vapour pressure; P_f — fluid pressure; ΔV — change in volume; X_{CO_2} — partial molar-volume of CO_2; ε — natural strain; σ_1, σ_2, σ_3 — maximum, intermediate and least principal stresses. RN 13, C 125 etc. — sample numbers in the Oxford collection.

All structural and fabric observations illustrated are plotted on equal area, lower hemisphere, stereographic projections with north at the top.

Outline of the Regional Geology

The Eastern Alps form an east—west chain of mountains which are continuous at their western end with the Central and Western Alps of Switzerland and at their eastern end bifurcate to enclose the Hungarian Plain, the northern bifurcation being made by the Carpathians and that to the south by the Dinarids (Fig. I—1). To the north the Eastern Alps abut against the flank of the Hercynian Bohemian Massif from which they are separated in surface outcrop by thick deposits of mid- to late-Tertiary Molasse shed from the flanks of the rising Alps. The southern margin of the Eastern Alps is taken as the Gailtal line, a pronounced east—west trending topographic lineament which must correspond to a fault of regional importance; the Southern Alps which lie to the south of this line have a stratigraphy and geological history which is distinctly different from that of the Eastern Alps.

Within the Eastern Alps themselves it is possible to recognize two broadly defined east—west trending zones — the Northern Zone and the Axial Zone. The Northern Zone which is bounded to its north by the Molasse is essentially an elongate synformal basement depression within which lies a piled series of allochthonous and parautochthonous thrust sheets. The rocks comprising these thrust sheets range in age from Cambrian to Cretaceous; the Mesozoic rocks are unmetamorphosed but some of the Palaeozoics show low greenschist grade of metamorphism (phyllites of the Grauwackenzone). These thrust sheets are of two origins: some represent the original sedimentary cover of the foreland (Bohemian Massif) which has been imbricated and horizontally displaced northwards by Alpine movements from the south; others (especially the Northern Calcareous Alps) represent the highest structural levels and original cover of the Axial Zone. These cover rocks slid northwards from a zone of net basement elevation to a zone of net basement depression in a series of movements culminating in the Miocene.

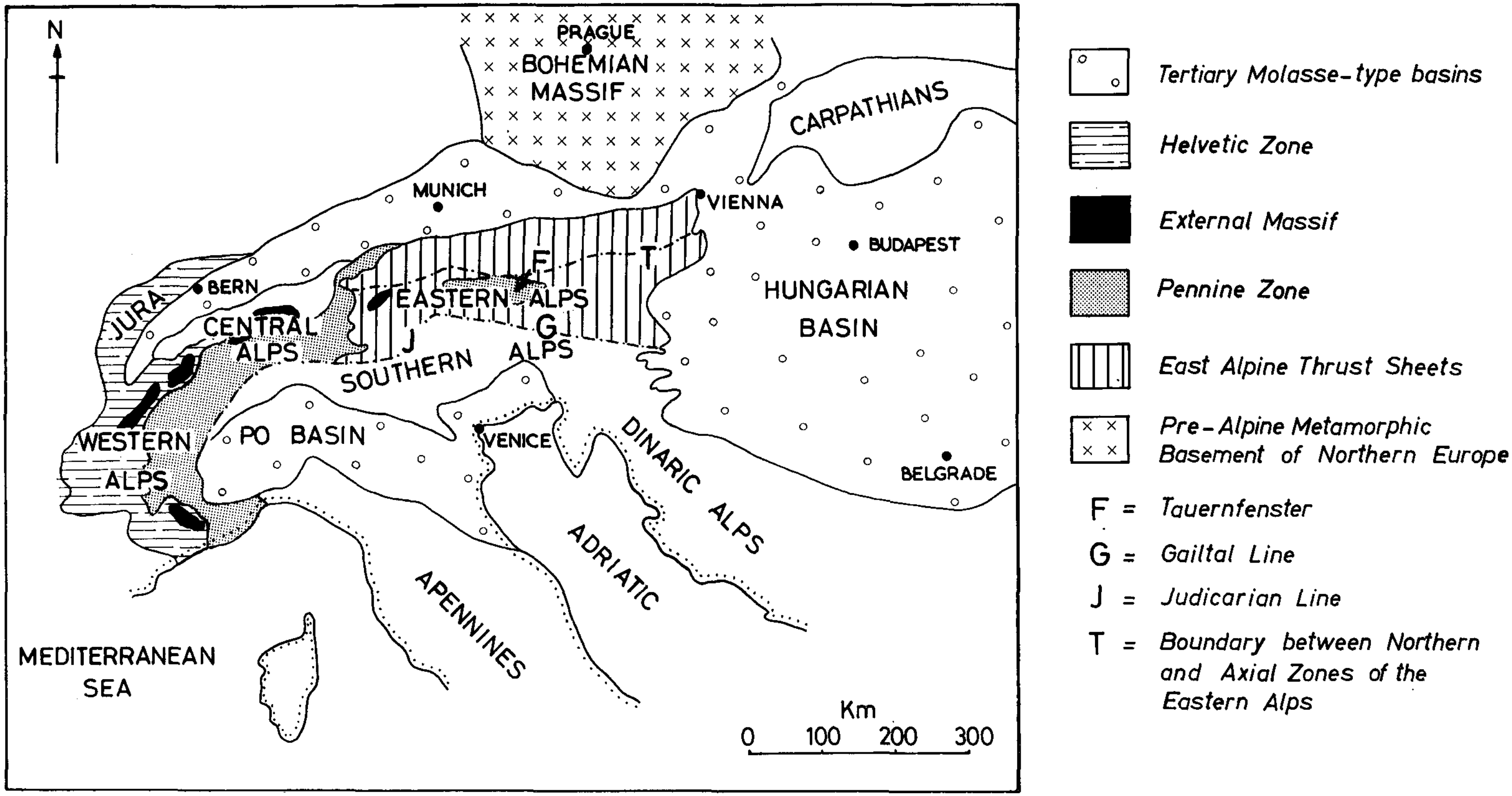

Fig. I - 1 The main features of the Alps; note that ornament has been lost from the Helvetic zone souht of Bern in the Central Alps and that the small black area by the "E" of Eastern Alps should be omitted.

In contrast to the Northern Zone the Axial Zone is an elongate region of antiformal elevation. Here too the structure is characterized by a pile of thrust sheets but in this case the regional upwarping is later than the movement of the sheets within it. There is good stratigraphic and structural evidence that the direction of transport of the sheets was more or less northwards over the rocks beneath. Virtually all rocks of the Axial Zone have undergone regional metamorphism. Differential erosion through the pile of thrust sheets has given rise to tectonic windows. The largest of these is the Tauernfenster (Tauern Window), some 160 *km.* from east to west and 50 *km.* from north to south. In and around the Tauernfenster all the important units of the Axial Zone are exposed.

The lowest unit is the autochthonous or parautochthonous Hercynian basement which together with its metasedimentary cover (largely Mesozoic) comprises the Penninikum or Pennine Zone. These rocks are thought to be laterally continuous with the Pennine Zone of the Central and Western Alps. The internal structure of the Penninikum is complicated by the presence of both Hercynian and Alpine fold structures and extensive Alpine thrusting. The Penninikum is in many places overlain tectonically by the Unterostalpindecke (Lower East Alpine sheet or nappe); this is a unit of complex internal structure and extremely variable thickness; it comprises fossiliferous Mesozoic, and possibly late Palaeozoic, slates, marbles and sands along with thin slivers of their metamorphic basement. The unit is most clearly exposed and best developed in the Radstädter Tauern. It is in turn tectonically overlain by the areally most extensive unit in the Eastern Alps, the Oberostalpindecke (the Upper East Alpine sheet). There is some disagreement as to whether this should properly be regarded as a single tectonic unit and it has been suggested that there is a very important low angle zone of differential movement within it. If this is so, there are profound tectonic and palaeogeographic implications. This has led some authors to split this unit into two parts, designating only the upper part as the Oberostalpindecke and introducing the term Mittelostalpindecke for the lower part. For a more extensive discussion of this problem in English and for the Austrian references see OXBURGH 1968 (a). The Mittelostalpindecke is a sheet comprising schists and gneisses of probably pre-Carboniferous age; these have not undergone extensive Alpine penetrative deformation but have ridden as a relatively inert mass over rocks beneath; they appear to have carried on top of them some Mesozoic sedimentary rocks which now occur as isolated patches, strongly deformed and recrystallized (Central Alpine Mesozoic). The overlying Oberostalpindecke (in the restricted sense) comprises phyllites and other low grade metamorphic rocks overlain unconformably by an unmetamorphosed, weakly folded, Permo-Mesozoic sedimentary sequence (the Drauzug) which has very strong stratigraphic affinities with the Northern Calcareous Alps; this similarity has led many workers to classify the Northern Calcareous Alps as part of the Oberostalpin.

The region with which we are concerned lies entirely within the Penninikum and extends from what is probably one of the deepest levels exposed within the Penninikum to the edge of the window where the Penninikum is in direct contact with the Mittelostalpindecke.

Previous Work in and around the Area Mapped

No attempt will be made here to give a complete or comprehensive account of the early work in the southeast corner of the Tauernfenster. This has already been done by ANGEL and STABER (1952) and EXNER (1957, 1964). We shall here concentrate on the main geological features of the region and mention some of the more modern studies which have a bearing on the problems with which we are concerned.

The main structural features of the southeast part of the Tauernfenster are shown in Fig. I—2. Of special interest is the Mölltal line; this is the topographic lineament which is defined by the valley of the lower Möll and the river Drau between Möllbrücke and Villach. This lineament trends nearly southeast and runs straight for some 70 *km.*; as discussed below, the same line is continued northwards for another 25 *km.* by the Mallnitzer Mulde. The significance of this lineament is not clearly understood and will be considered later; here we note that it appears to affect both the rocks of the Penninikum within the Tauernfenster and the higher tectonic units surrounding the window. Between Obervellach and Möllbrücke the line corresponds to the margin of the window; along this portion of the margin, the Ober(mittel)ostalpindecke lies directly against the Penninikum and the Unterostalpin does not intervene; this could be the result of normal faulting along the Mölltal line. A most interesting discussion of the line has been given by Exner (1962).

The mountains of the Kreuzeck and Polinik Groups, south west of the Mölltal line (i.e. making the edge of the Ober[mittel]ostalpin), comprise a complex of schists and orthogneisses (Oxburgh, 1964) which are quite clearly truncated at the Mölltal line. The details of their contact with the rocks within the window are unfortunately obscured by glacial and alluvial deposits along the Möll valley. The contrast in geology across the Mölltal line is substantiated by a significant difference in the thermal histories of the two sides as shown by their contrasting patterns of K/Ar ages (Oxburgh et al., 1966).

Considering now the rocks within the Tauernfenster we find two main lithological groups: a 'basement' complex of orthogneisses, paragneisses and amphibolites and a 'cover' series of metasedimentary rocks with subsidiary volcanics which are in the greenschist to amphibolite facies of regional metamorphism.

The general distribution of basement and cover rocks (Fig. I—2) has been known for many years and has given rise to numerous tectonic hypotheses for the structure and tectonic evolution of the eastern Tauern. In general the approach was very similar to that which had proved so successful in the Central and Western Alps; orthogneissic and paragneissic masses were regarded as crystalline nappe cores and intervening schist zones (Mulden) were interpreted as "nappe dividers" (e.g. Kober, 1922; Staub, 1924).

There were, however, complications. The age of the orthogneisses was in dispute; while some regarded them as metamorphosed Alpine intrusions (e.g. Angel and Staber, 1952); others regarded the orthogneisses as Variscan intrusions and the paragneisses as the Palaeozoic country rock into which the orthogneisses (as granites) had been intruded (e.g. Exner, 1957, 1964; Frasl, 1957). Yet others believed both Alpine and Variscan intrusions to be present (e.g. Sander, 1912, 1921; Karl, 1959). The cover sequence of low grade metasedimentary rocks was generally regarded as predominantly Mesozoic in age by analogy with fossiliferous rocks from the Pennine Zone of the Central Alps and comparison with the Unterostalpindecke. A number of distinct questions gradually emerged as fundamental to the understanding of the evolution of the central part of the Eastern Alps.

(a) The age of intrusion and age of metamorphism(s) of the orthogneisses (the, so called, Zentralgneis).

(b) The clear distinction between the paragneiss series, closely associated with the Zentralgneis (known as the Inner Schieferhülle, or Altkristallin) and the cover metasedimentary series (the Peripheral Schieferhülle). Unless these were clearly distinguished the significance of a metamorphic 'nappe divider' of sedimentary origin could not be clear because its age, and thus the age of the structures it defined, would be uncertain.

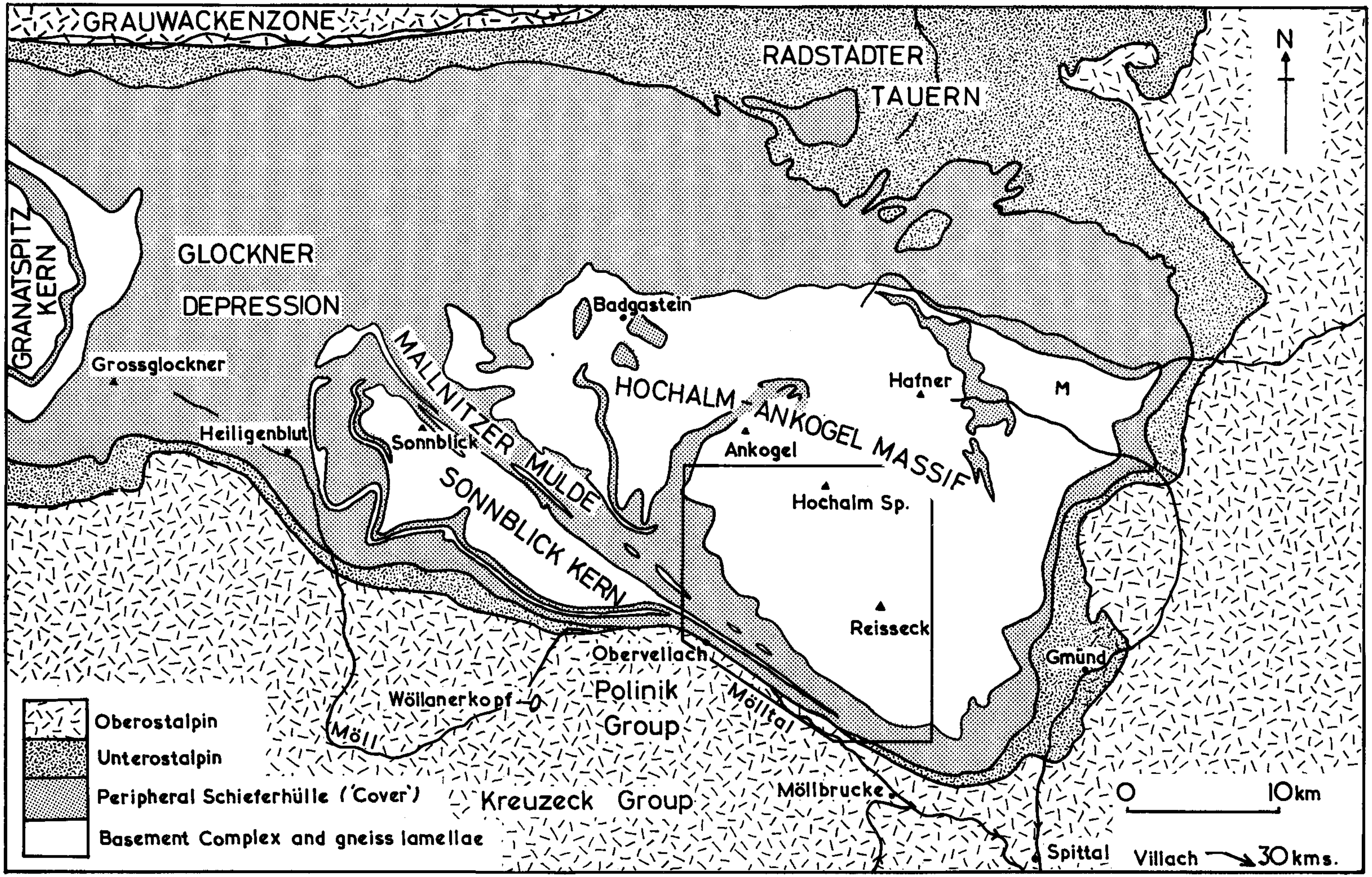

Fig. I - 2 The eastern part of the Tauernfenster showing outline of the area mapped; M = Mureckdecke.

(c) The Mesozoic age of the Peripheral Schieferhülle required substantiation and a more detailed stratigraphic classification was needed.

(d) The form of the metasedimentary zones needed to be accurately known in three-dimensions; only then could their role as nappe dividers be properly assessed.

These, and many other important questions could be answered only by careful and detailed field mapping. In a series of papers over a period of more than thirty years Ch. EXNER (e.g. 1940, 1953, 1954, 1957, 1964) has provided a great deal of information of this kind which has gone far towards solving these problems in the South East Tauern. A lead in this approach had been given by the classic work of CORNELIUS and CLAR (1939) in the Glockner Group of the central Tauern where they made a most detailed and systematic study of the Schieferhülle. This was further elaborated by an important regional analysis of Schieferhülle stratigraphy by FRASL (1957).

Figure I—2 shows that much of the high ground in the Eastern Tauern is made by the Zentralgneis; the Peripheral Schieferhülle generally occupies structural depressions (Mulden) between zones of basement elevation (Kernen) while the Inner Schieferhülle has an irregular distribution around the margins and within the Zentralgneis. The Hochalm—Ankogel Massif is the main Zentralgneis Massif in the Eastern Tauern and may be subdivided into a number of smaller units between which there has been differential vertical movement; in general, however, the Peripheral Schieferhülle dips away from the Hochalm—Ankogel Massif on all sides. On its southwest side the massif is bounded by the Mallnitzer Mulde, a rectilinear elongate synform occupied by Peripheral Schieferhülle which, as noted earlier, continues northwestward the trend of the Mölltal line. Southwest again of the Mallnitzer Mulde is a second, considerably smaller and rather elongate gneiss massif, the Sonnblick Kern. This massif has the form of a northwestwards trending, strongly asymetrical antiform facing northeast. At its southeastern end the gneiss body tapers into an elongate tongue, the Sonnblick Lamella; with vertical sides and flanked by Peripheral Schieferhülle, this body of gneiss extends for some 15 *km*. along the Mölltal parallel to the edge of the Tauernfenster. More remarkable still, EXNER (1962) has shown that several other thin lamellae of gneiss also occur within the Peripheral Schieferhülle which overlies and flanks the Sonnblick Kern; fragments of these lamellae may be found parallel to and on the northeast side of the Sonnblick lamella, apparently pinched back on themselves in a near isoclinal synform which is the southeastward continuation of the Mallnitzer Mulde.

As EXNER has on many occasions pointed out, the structures in this area are largely non-cylindrical; the reasons for this are discussed in a later chapter but it will be clear from Fig. I—2 that it is not possible to draw a single representative cross-section through the region.

Figure I—2 also shows the limits of the present map area in its regional setting· it lies in the eastern half of the Hochalm—Ankogel Massif and extends from close to the mid-point of the Massif southwards to the edge of the Tauernfenster. There has been very little detailed mapping published within this area. On its northwest side the map adjoins the Sonnblick sheet mapped by EXNER (1964) while on its northern side our map overlaps that of ANGEL and STABER (1952). EXNER (1962 and 1964) has published some profiles and a map of the Mölltal, i.e. along the southwestern side of the map area, and in 1954 a sketch map of the area east of it. DEMMER (1968) has also published several profiles across part of the south eastern part of the map; these profiles arise from work done in course of a survey for a hydroelectric scheme.

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II. Field Relations and Petrography of the Main Rock Groups

by R. J. Norris, R. A. Cliff and E. R. Oxburgh

Introduction

Within the southeastern part of the Tauernfenster lies a series of metamorphosed and deformed igneous and sedimentary rocks showing complex relationships to each other. These rocks are subdivided as follows:

1. Peripheral Schieferhülle
2. Basement Complex comprising:
   (a) Zentralgneis (Central Gneiss)
   (b) Inner Schieferhülle.

The term Peripheral Schieferhülle is here used to denote a severely deformed metasedimentary and metavolcanic series which normally

(i) shows only one principal phase of folding deformation, although locally a subsequent warping and flexuring associated with joint development may occur;

(ii) shows a variety of low-temperature, hydrothermal vein-fillings (e.g. albite-chlorite or calcite) but seldom or never contains aplite veins;

(iii) has abundant calcareous units throughout the sequence.

As discussed below, when the Peripheral Schieferhülle is considered in more detail, there are other diagnostic characteristics but these are of relatively less importance. On litho-stratigraphic evidence, the Peripheral Schieferhülle is believed to be largely Mesozoic in age although it could in part be late Palaeozoic.

The Peripheral Schieferhülle surrounds, overlies and occupies depressions within the Basement Complex. Within the complex the Inner Schieferhülle is interfolded with and locally intruded by some of the variously metamorphosed igneous rocks which form the Zentralgneis. Both Inner Schieferhülle and Zentralgneis generally:

(i) show two or more principal phases of folding deformation (note, however, that folding is rarely seen in the Zentralgneis and that parts may show the effects of only one deformation);

(ii) show locally abundant, deformed, or apparently undeformed, aplite veins of one or more generations;

(iii) are typically non-calcareous.

The Inner Schieferhülle is taken to include all metasedimentary rocks, paragneisses, amphibolites and banded gneisses within this complex while the term Zentralgneis is reserved for the quartzofeldspathic rocks (i.e. igneous rocks and orthogneisses, ranging in composition from diorite, through tonalite to granite) which are largely unbanded. No general age distinction is made between the Inner Schieferhülle and the Zentralgneis. The units are in many places intimately interfolded with each other but it is clear that the Zentralgneis represents a series of igneous intrusions of different ages and that some of these post-date at least some metasedimentary rocks of the Inner Schieferhülle, while others could be earlier. Isotopic evidence on the age of some of the younger parts of the Zentralgneis suggest that the Zentralgneis — Inner Schieferhülle complex is Variscan in age, but parts may be older.

These empirical criteria which have been used to distinguish the Basement Complex from the Peripheral Schieferhülle cannot, of their nature, be entirely satisfactory. But the general subdivision of units which they permit is consistent with, and substantiates a subdivision based on lithological criteria. The complexities of variation in metamorphic grade, and in degrees of deformation, and overlapping chemical composition make a subdivision based on lithology alone somewhat hazardous, but when it is substantiated by other criteria such as those employed here, the subdivision carries a much higher degree of confidence.

In so far as the Basement Complex displays aplitic injection not found in the Peripheral Schieferhülle and has also undergone at least one additional episode of deformation, it is clear that if the Peripheral Schieferhülle is autochthonous or parautochthonous it is younger than the Variscan Basement Complex which today it overlies. This is consistent with the stratigraphic evidence for a Mesozoic age for the Peripheral Schieferhülle mentioned above.

The Zentralgneis

Introduction

For the purposes of this paper the Zentralgneis is defined as comprising all the essentially unbanded, quartzo-feldspathic rocks of the Basement Complex. These rocks are thought on account of their texture and/or mineralogical composition to be variously metamorphosed plutonic rocks. The name 'granite' is used in the subsequent description as a general compositional term to describe them, although their mineralogical compositions in fact range from quartz-diorite to alkali-granite. Almost all of the rocks of the Zentralgneis have some kind of metamorphic fabric but the intensity of its development varies greatly; the term '-gneiss', qualified as granite-gneiss and augen-gneiss is used where the metamorphic fabric, either foliation or lineation, is prominent in hand specimen; the majority of augen-gneisses exhibit a blastomylonitic texture in thin section.

The Zentralgneis includes a wide range of lithologies some of which occur as large, more or less homogeneous units which can be readily traced on the ground and can be shown at the scale of the present map. A large part of the Zentralgneis, however, is made up of granite-gneisses which are very inhomogeneous on a scale of tens of metres: different lithologies are intercalated or may grade into one another. It is impossible to show such variations at the scale of the main map and some of the variation, particularly textural variation, is so complicated that it is very difficult to represent it at anything less than natural scale.

Our first aim, therefore, has been to identify and describe those parts of the Zentralgneis that form relatively homogeneous, mappable units, to recognize their field relations to other units and to relate them to the overall plutonic and tectonic evolution of the south-east Tauern. Elsewhere in areas of physical inaccessibility or limited exposure the granitic rocks have simply been mapped as 'undifferentiated granite-gneiss'.

Before discussing the lithologies of the Zentralgneis in detail, certain aspects of interpretation should be mentioned. Although an attempt has been made to describe the lithologies and their field relations, so far as possible, without involving genetic considerations, the latter necessarily influence the approach to field work and the collection of data. Some concepts involved in the interpretation are therefore summarised below; detailed discussion is to be found later in the paper.

While the mineralogical composition of the Zentralgneis is rather variable, by far the most striking variations are in texture. In so far as it is clear that on a somewhat larger scale, deformation has profoundly modified original field relations, these variations in texture are thought largely to result from differing amounts and/or conditions of deformation.

This leads to the concept of tectonic facies within the Zentralgneis i.e. the idea that rocks of similar composition, and related possibly to a single plutonic event, may today exhibit quite different textural characteristics on a mesoscopic and microscopic scale, simply as a result of different deformational histories. Conversely, some rocks which in less deformed facies can readily be distinguished from one another become virtually indistinguishable when strongly deformed. This phenomenon of 'tectonic convergence' is not restricted to the Zentralgneis and can also lead to confusion between rocks that properly belong to the Zentralgneis and parts of the Inner Schieferhülle.

The divisions of the Zentralgneis shown on the map fall into two groups:

1. Relatively undeformed granites: granites of variable composition and grain size, in which a granitic texture is partially preserved and which, at least locally, exhibit relatively undeformed contact relations with one another and with parts of the Inner Schieferhülle. This group includes:

(i) tonalite
(ii) granodiorite
(iii) coarse leucogranite
(iv) fine leucogranite
(v) fine porphyritic granite

In addition, the Zentralgneis and parts of the Inner Schieferhülle are cut by discordant aplite and pegmatite veins which can in part be related to this group.

These granites constitute only a small part of the Zentralgneis.

2. Granite-gneisses: gneissose rocks of variable composition and with a variety of textures among which augen-gneiss is prominent. The contacts of this group with other lithologies are generally tectonic or strongly deformed. Their age relations with the Inner Schieferhülle may only be determined indirectly. The following groups are recognised:

Biotite augen-gneiss	coarse-grained fine-grained undifferentiated
Leucocratic granite-gneiss	augen-bearing fine-grained undifferentiated

Relatively Undeformed Granites

Tonalite and Granodiorite

These two lithologies are characterised by a very distinctive texture defined particularly by the biotites which occur in clots several millimetres across; they differ only in the higher K-feldspar content and somewhat lighter colour of the granodiorite.

In hand specimen the tonalites and granodiorites are of essentially granular appearance and are characterised by the sharp contrast between the clots of biotite and the milky-white feldspar; in the granodiorite the K-feldspar is prominent as megacrysts up to two centimetres long. A slight foliation is usually discernible and in some areas a more intense foliation is developed; locally this is so strong that the tonalite resembles a feldspathic biotite schist. Where the granodiorite is intensely foliated it becomes an augen-gneiss (see below). The essential minerals are quartz, plagioclase, biotite and, in the granodiorite, K-feldspar. Clinozoisite and white mica are important minor constituents and sphene, apatite, zircon, allanite and magnetite occur as accessories. Quartz occurs as aggregates of 0. 5 *mm.* grains showing somewhat undulose extinction. In the poorly foliated tonalites, the quartz occurs as patches of strained grains, not having an equilibrium texture. With increasing deformation, the quartz aggregates become strung out into streaks, and partially incorporated into the plagioclase mosaics.

Oligoclase occurs as large tablets which are packed with clinozoisite and white mica microlites ('gefüllte Plagioklas'); in most samples these tablets have been largely replaced by a fine-grained polyhedral mosaic of inversely zoned, untwinned oligoclase (Table II—1).

Both large and small grains are effectively of the same composition. In some of the small zoned grains no significant difference was found between core and rim, although in others the rims were upto 8% more calcic than the cores. The variation of composition from grain to grain within a single sample is probably real, and is less than that found by some workers (e.g. HUNAHASHI et al., 1968).

The presence of relict oscillatory zoning, simple twinning and igneous habit in the large plagioclase tablets is good evidence that they represent the original igneous plagioclases, somewhat modified in composition (cf. KARL, 1959).

TABLE II—1

Anorthite content of plagioclase in Tonalite

No. and Locality	Microprobe		U-stage Gefüllte grains
	Gefüllte grains	Small grains	
RN 329	26, 28, 30, 29, 28, 27	29	34, 33, 34, 31
Grübelwand			
RN	28, 30, 31, 25, 28, 30, 18	27, 31	31, 28, 28, 29
Villacherhütte	25, 26, 25, 27, 18, 15		
RN 396	29, 29, 29, 30, 30	30, 30, 30, 32	
U. Rieken Graben			
RN 375			29, 31, 30, 35
Villacherhütte			31
RN 388			
Villacherhütte			22, 26, 25, 31, 27, 28
RN 392			
Villacherhütte			34, 28, 30, 31, 31
A 327 (Kaponig Törl — Ebeneck area)	22, 23, 24, 24, 22, 22	22(31), 22(—), 22(26), 19(22), 23(27), 21(29), 22(25), 22(—)	
A 220 (Kaponig Törl — Ebeneck area)	26, 23, 23, 23, 23, 22	22(—), 22(—), 21(—), 21(—), 22(—), 23(—), 22(24), 23(25)	
A 306 (Kaponig Törl — Ebeneck area)	25, 25, 23, 24, 22, 24, 24	18(20), 20(21), 20(—). 21(—)	
A 313 (Kaponig Törl — Ebeneck area)	19, 21, 20, 19	18(—), 19(—), 20(—), 21(—), 19(—)	
A 227 (Kaponig Törl — Ebeneck area)	24, 23, 24, 23, 23, 24, 23, 24	19(—), 24(—), 23(—), 25(—), 21(—), 24(—), 22(24), 22(25), 22(23), 20(—), 20(25)	
A 335 (Kaponig Törl — Ebeneck area)	24, 27, 25	26(—), 26(—), 29(—), 23(28), 26(29), 26(29)	

Note: where separate determinations were made on rims and cores of small grains, the core value is given first, followed by the rim value in brackets ( ).

K-feldspar is monoclinic and, except in very strongly foliated samples, occurs exclusively as megacrysts; occasionally K-feldspar also occurs as antiperthitic intergrowths with oligoclase. Plate 3, a, b shows thin section sketches of tonalite and granodiorite and modal analyses are given in Table II—2.

The outcrop of the tonalite and granodiorite is shown on Fig. II—1. Together they form a semi-continuous outcrop which extends from the Mühldorfergraben in the south, northward to the Schneewinkelspitze and then eastnortheast to the Villacherhütte; from here the outcrop continues northward, beyond the map area. The granodiorite occurs at several localities along the western contact of the tonalite and locally, as in the upper part of the Kaponig Graben, it forms more extensive bodies; it also forms much of the outcrop below the Trippkees. There is a continuous gradation between granodiorite and tonalite and on the map a boundary is drawn arbitrarily at 10% modal K-feldspar. In the area around the Villacherhütte which was not mapped in detail the granodiorite may be more extensive than shown.

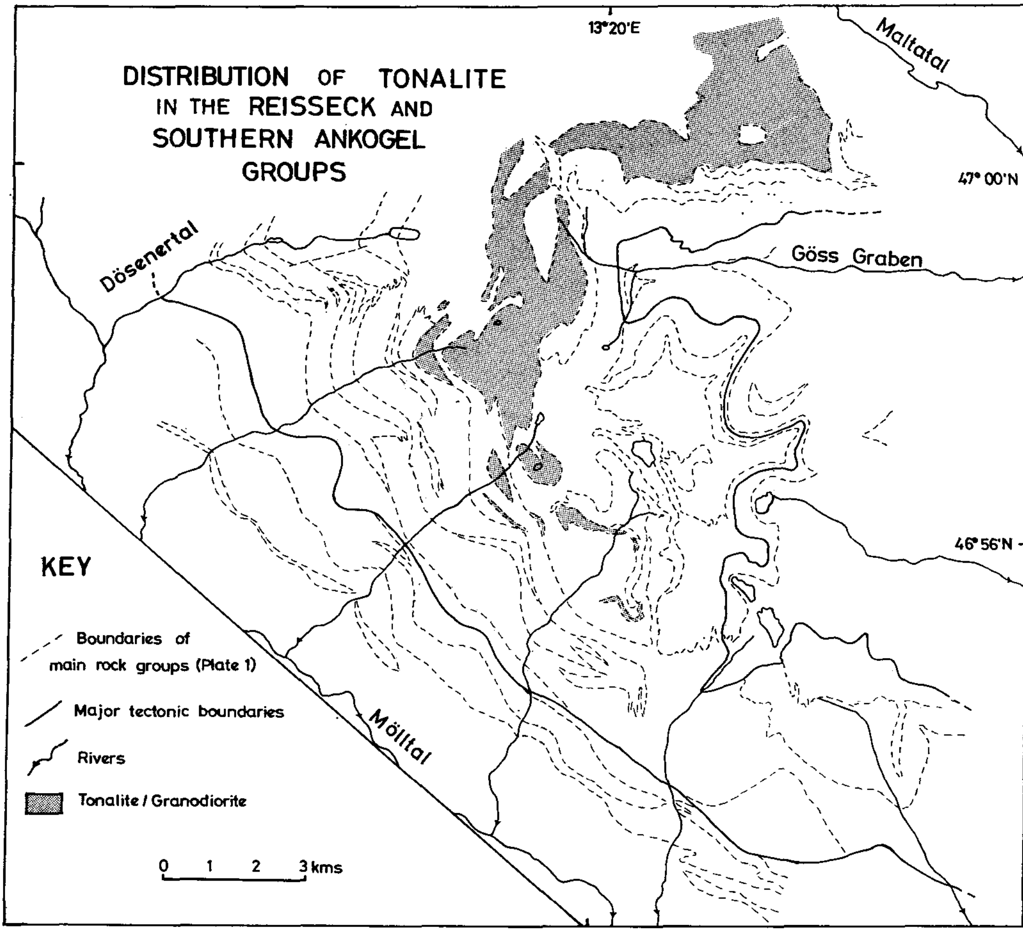


Fig. II - 1

The shape of the tonalite-granodiorite mass is uncertain; it has contacts both with older and younger rocks and at many localities these contacts are probably tectonic (Plate 1). The mass is probably an irregular sheet which is thicker in the north than in the south.

Along the western margin of the outcrop the tonalite/granodiorite is in contact with augen-gneiss as far south as the Zwenberger Ochsnerhütte; locally the contact is sharp (e.g. immediately south of the Ebeneck) but where the granodiorite occurs no sharp boundary can be drawn and, indeed, it is probable that parts of the augen-gneiss represent more strongly deformed granodiorite (see section on augen-gneisses). Some of the augen-gneisses, however, are clearly not related to the granodiorite, but there is no clear evidence whether they are older or younger than the tonalite/granodiorite.

The eastern contact is against mica schists and leucogneisses along most of its length; below the Trippkees a band of very sheared banded gneisses separates the granodiorite from the micaschists and leucogneisses. South of Tristenspitze, however, the tonalite is in contact with the younger fine-grained granite-gneiss. The smaller masses of tonalite in the Rieken- and Mühldorfergrabens are also cut off by the fine-grained granite-gneiss.

Small masses of mica schist, criss-crossed by numerous granite veins and containing diffuse patches of quartzofeldspathic rock are distributed throughout the tonalite outcrop; typically they are only one to two metres thick and several tens of metres long. The contacts with the tonalite are usually quite sharp and they are probably rafts of country rock; in part the lithologies resemble those along the eastern contact, although feldspathic biotite schist similar to that found in the banded gneisses is more abundant.

In addition, the tonalite and granodiorite contain numerous smaller, ellipsoidal xenoliths of biotite schist ranging from a few centimetres to one metre long; they are rather uniform in composition and differ from the rafts of schist described above in having a generally higher biotite content and the absence of leucocratic veins; they show a strong preferred orientation parallel to the local foliation and/or lineation; the long axis may reach ten times the length of the shortest axis. They closely resemble supposed cognate dioritic xenoliths in other tonalite plutons (e.g. Adamello, Bergell).

TABLE II—2

Modal Analyses of Tonalite

	RN 392	A 330	A 155	A 323	A 325	A 326	A 327	A 335	A 83	A 91	A 109	A 297	A 182	RN 329	RN 388
Plagioclase	47	47	50	49	52	51	47	54	58	60	54	50	63	49	46
K-feldspar	a	3	1	—	4	4	a	5	7	4	5	5	a	4	4
Quartz	31	29	22	31	19	24	13	23	21	22	24	32	21	32	20
Biotite	18	18	20	19	24	19	36	17	14	11	16	12	14	12	19
White Mica	4	1	—	—	a	a	a	—	—	—	—	—	a	—	5
Epidote	1	a	7	a	2	2	a	a	a	3	a	a	1	a	6
Sphene	a	a	a	a	a	a	a	a	a	a	a	a	—	—	a
Apatite	a	a	a	a	a	a	2	a	a	a	1	a	a	a	a
Zircon	—	—	—	a	a	a	a	a	—	—	—	a	—	—	a
Ore	a	—	a	—	a	a	a	a	a	a	—	—	—	a	a
Chlorite	1	—	a	a	—	—	—	—	a	—	—	—	—	—	a

Modal Analyses of Granodiorite (A) and Gneisses derived from it (B)

<table>
<tr><th></th><th colspan="4">A</th><th colspan="4">B</th></tr>
<tr><th></th><th>A 197</th><th>A 275</th><th>A 294</th><th>A 277</th><th>A 161</th><th>A 207</th><th>A 319</th><th>A 395</th></tr>
<tr><td>K-feldspar</td><td>13</td><td>26</td><td>33</td><td>12</td><td>26</td><td>21</td><td>29</td><td>24</td></tr>
<tr><td>Plagioclase</td><td>32</td><td>32</td><td>54</td><td>53</td><td rowspan="2">62</td><td rowspan="2">66</td><td>24</td><td rowspan="2">70</td></tr>
<tr><td>Quartz</td><td>50</td><td>33</td><td>11</td><td>16</td><td>39</td></tr>
<tr><td>Biotite</td><td>6</td><td>10</td><td>2</td><td>17</td><td>11</td><td>11</td><td>8</td><td>5</td></tr>
<tr><td>White Mica</td><td>a</td><td>—</td><td>1</td><td>—</td><td>1</td><td>a</td><td>1</td><td>2</td></tr>
<tr><td>Epidote</td><td>a</td><td>a</td><td>a</td><td>2</td><td>a</td><td>a</td><td>a</td><td>a</td></tr>
<tr><td>Sphene</td><td>—</td><td>a</td><td>a</td><td>a</td><td>—</td><td>—</td><td>—</td><td>a</td></tr>
<tr><td>Apatite</td><td>a</td><td>a</td><td>a</td><td>a</td><td>a</td><td>a</td><td>a</td><td>a</td></tr>
<tr><td>Zircon</td><td>—</td><td>—</td><td>—</td><td>a</td><td>—</td><td>—</td><td>—</td><td>a</td></tr>
<tr><td>Ore</td><td>—</td><td>—</td><td>a</td><td>a</td><td>—</td><td>—</td><td>—</td><td>—</td></tr>
<tr><td>Chlorite</td><td>—</td><td>—</td><td>a</td><td>a</td><td>—</td><td>—</td><td>—</td><td>—</td></tr>
</table>

a: less than 1%

The tonalite and granodiorite are cut by numerous small intrusions of coarse leucogranite and fine-grained porphyritic granite (see below) as well as by numerous, largely undeformed aplites and pegmatites.

Leucogranites

Small masses of extremely leucocratic granite occur in a variety of situations throughout the area of Zentralgneis and Inner Schieferhülle. They vary considerably in texture and grain size; many of the rocks described below as leucogneiss have identical mineralogy and are thought to be deformed equivalents of the leucogranites.

Over much of the map area it is possible to distinguish two main types of leucogranite, although subordinate amounts of intermediate types also occur: (a) coarse leucogranite (b) fine leucogranite.

(a) Coarse leucogranite is characterised by a grain size of 2—5 *mm*. and occurs as small irregular intrusions in the tonalite and granodiorite. There are numerous xenoliths of tonalite and the contacts with the country rock are sharply discordant and appear to be undeformed (Fig. II—2; see page 140). Similar granites occur in the leucogneisses and schists immediately adjacent to the eastern contact of the tonalite in the Hohes Gösskar and between the Zwenberger Seen.

Fig. II - 2 Angular blocks of tonalite intruded by coarse Variscan leucogranite; note the pre-leucogranite aplitic vein in the tonalite and the undeformed character of the block margins; near the Kaponig Törl. Zentralgneis.

Here they contain numerous large xenoliths of mica schist and migmatite. Over much of this part of the area the exposure is not good enough to establish the form of the leucogranite bodies — they may well be isolated lenses as suggested on the map or they may form an interconnecting network within which blocks of leucogneiss, schist and migmatite are "floating".

Generally a very weak foliation is present which crosses the sharp intrusive contacts. Near the contact with the augen-gneisses south of the Ebeneck a stronger foliation is developed and the leucogranite bodies occur in fold cores within the augen-gneisses; these strongly foliated leucogranites are discussed below as coarse leucogneisses.

The coarse leucogranites consist of quartz, K-feldspar and oligoclase with about 2% (occasionally up to 4%) biotite and up to 2% white mica (modal analyses are given in Table II—3). Accessory clinozoisite is ubiquitous and garnet, allanite and zircon are common. An opaque mineral (usually magnetite) is often present and may be more abundant than biotite. The texture is granular; in thin section, however, only the K-feldspar is as coarsely grained as it appears in hand specimen; the quartz occurs as aggregates of irregular 1 *mm.* grains and the oligoclase forms fine-grained polyhedral mosaics.

TABLE II—3

Modal Analyses of Leucogranite

	A 104	A 112	A 188	A 189	A 219	A 324	A 334	A 336	A 307	A 417
Plagioclase	24	25	11	19	28	16	13	15	29	31
K-feldspar	38	34	29	43	37	32	43	41	30	35
Quartz	38	38	59	37	36	49	42	41	34	30
Biotite	a	2	1	1	3	1	1	3	4	1
White Mica	a	2	a	a	a	1	a	a	1	3
Clinozoisite	a	a	a	a	a	1	a	a	a	1
Sphene	—	—	—	—	—	—	—	—	—	—
Apatite	—	a	—	—	—	—	—	—	a	—
Zircon	—	—	a		a	—	a	—	a	—
Opaques	a	a	a	a	—	a	a	a	a	a
Chlorite	—	—	a	a	—	—	—	—	—	—
Garnet	—	a	a	a	a	—	—	—	a	a

a: less than 1%.

(b) Fine-grained leucogranites are associated with the augen-gneisses and the Inner Schieferhülle. Locally they are clearly discordant but they frequently form concordant bodies. They are often foliated and grade into fine-grained leucogneisses; the foliation is generally concordant with that in the country rock but in the discordant bodies foliation parallel to the margins of the intrusion is commonly developed.

In the amphibolites on the north side of the Gössgraben the fine-grained leucogranites form intrusive sheets, from a few centimetres to 2 metres thick; they are strongly discordant to the banding in the amphibolites but have subsequently undergone folding into buckles or boudinage along with the amphibolites (Fig. II—3; see page 142). Similar discordant relationships are also seen in the amphibolites of the Riekener Sonnblick—Reisseck area. The leucogranites at these localities have very little biotite, but garnet and opaque minerals are prominent accessories; they are characterised by patchy variations in the grain size. Similar leucogranites also intrude leucogneisses and the fine-grained granite-gneiss; this suggests that they are probably the youngest plutonic rocks of the map-area; since some of the leucogneisses probably formed by deformation of very similar fine-grained leucogranites, it is possible that there was more than one period of intrusion of fine-grained leucogranite.

Fine-grained leucogranite also occurs as a thin sheet along the contact between the biotite augen-gneiss and banded gneisses around the Kleine Mühldorfer See, and in a similar situation between fine granite-gneiss and biotite augen-gneiss on the east side of the Hohe Leier. It is characteristic of the fine-grained leucogranite that whereas the contacts against the Inner Schieferhülle are sharp, those against the biotite augen-gneiss are often diffuse: at the first locality mentioned above there is a transition zone of several metres in the augen-gneiss which is unusually leucocratic; this contrasts with the upper contact where the same leucogranite body sends off numerous veins into the banded gneiss which is itself relatively unaltered. (Elsewhere the contacts with augen-gneiss may be fairly sharp as on the east side of Hohe Leier.) A similar apparent alteration of biotite augen-gneiss is observed in the upper Hinterregen

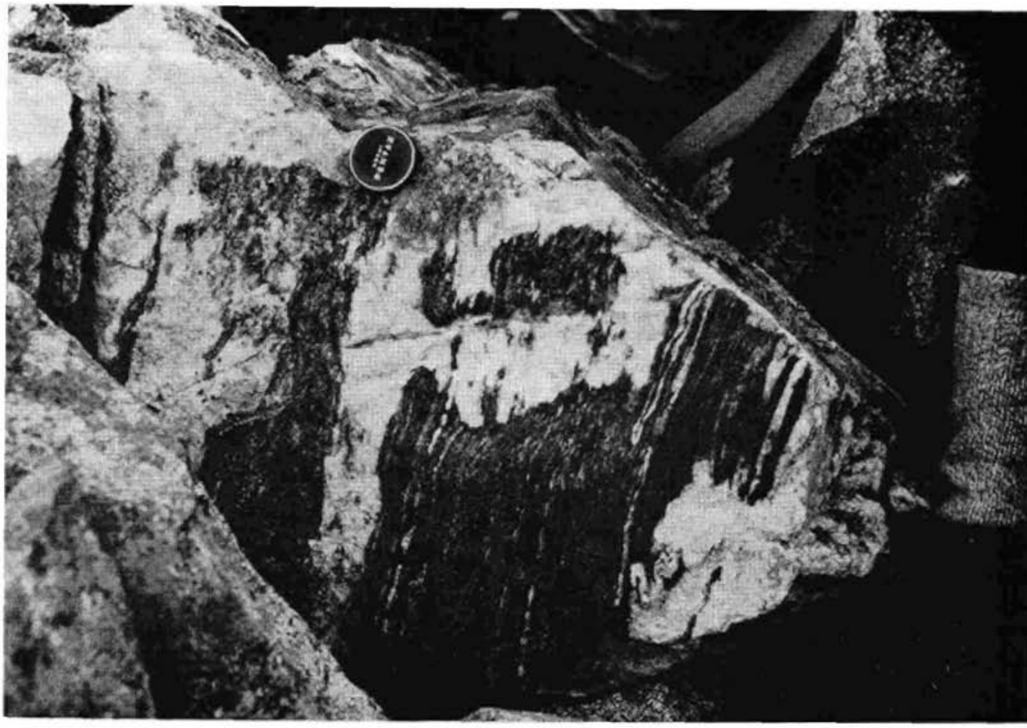


Fig. II - 3 Fragments of amphibolite containing isoclinally folded veinlets, intruded by fine-grained leucogranite; on the Winterleitengrat. Inner Schieferhülle.

Graben where several bodies of leucogranite are surrounded by extensively altered augen-gneiss; these leucocratic parts of the augen-gneiss typically contain garnet and locally show replacement perthite textures.

Late discordant veins

In addition to the relatively large bodies of leucogranite described above, parts of the Zentralgneis and Inner Schieferhülle are cut by numerous late discordant veins. These are petrographically rather variable and include aplites, granites, both biotite and muscovite pegmatites, as well as monomineralic feldspar veins and vein quartz. Some of the granite veins contain abundant magnetite and pyrite. There is good evidence that these veins were emplaced in several episodes. In the Pfaffenberger Seen area one group clearly predates the fine porphyritic granite while others are later; the ore-bearing veins are confined to the first group. Vein quartz forms the latest veins seen. In the gneisses of the Gössgraben and Maltatal there is also evidence for several episodes of vein emplacement; here some early veins have undergone two episodes of folding, whereas others cut these veins and are affected by only the second folding, while yet a third group is undeformed and cuts across all structures, apparently following a major joint pattern. At one locality immediately south of the Seeschartl, thicker leucogranitic bodies also apparently follow the major joint directions. At two localities metamorphosed lamprophyre veins with hornblende and biotite phenocrysts were found.

Fine-grained porphyritic granite

A fine-grained granite with randomly oriented subhedral megacrysts of simply twinned K-feldspar forms numerous small intrusions into the tonalite and granodiorite. These intrusions occur principally in the area between the Zwenberger-Ochsnerhütte and the Ebeneck; the field relations are especially well shown in the slabs on the north-west flank of Tristenspitze.

The individual intrusions seldom exceed 20 metres across but their shape is extremely complicated; nowhere is the porphyritic granite free of angular xenoliths of tonalite (Fig. II—4; see page 143) and there are areas where tonalitic country-rock is criss-crossed by a network of fine porphyritic granite dykes. These dykes cut many of the aplitic and pegmatitic veins in the tonalite but are themselves cut by later veins. Locally the fine porphyritic granite cuts coarse leucogranite.

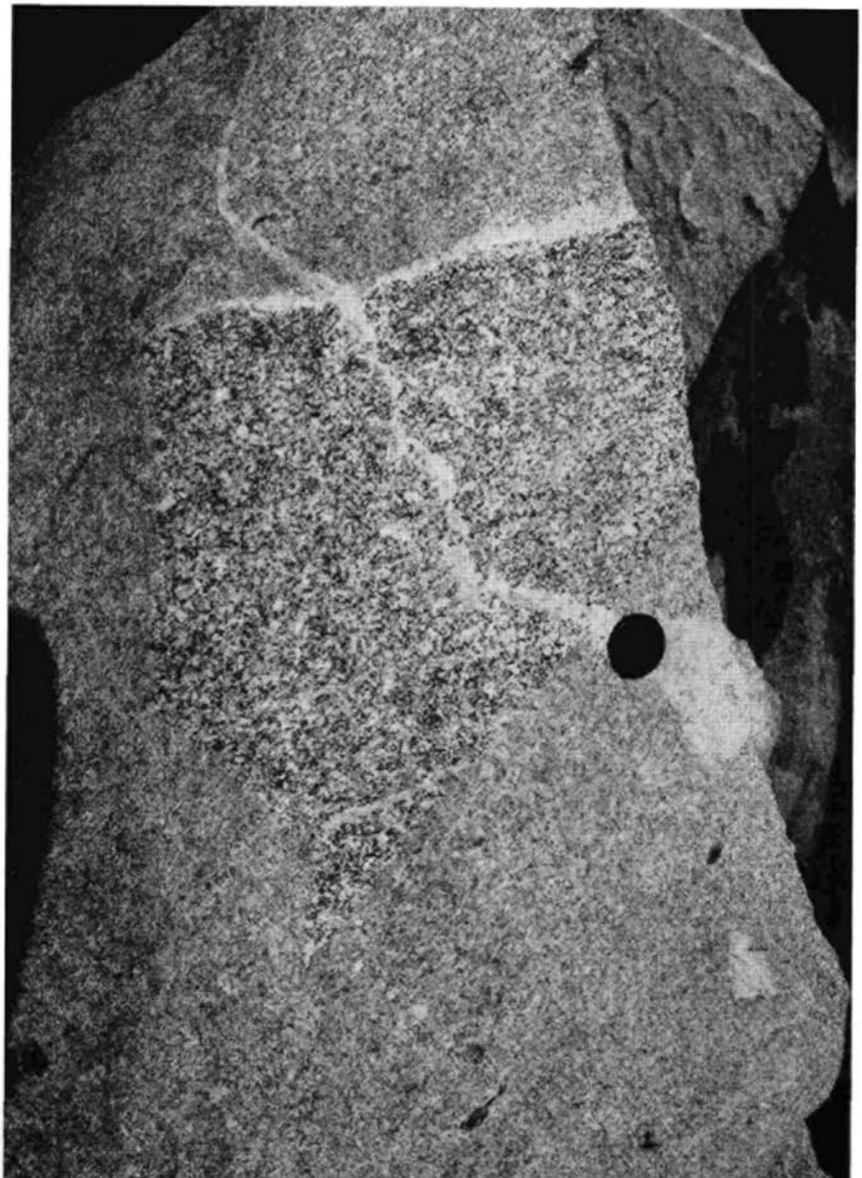


Fig. II - 4 A xenolith of tonalite in fine-grained porphyritic granite. Zentralgneis. The dark coin is 2.5 cm in diameter.

In hand-specimen the random orientation and rather uniform size (1—2 *cm.*) of the subhedral megacrysts is the most striking feature of the rock type. The fine-grained matrix has scattered flakes and small clots of biotite and generally shows at least a weak foliation. Where the foliation is stronger the megacrysts tend to have a distinct preferred orientation and the rock is a fine-grained augen-gneiss (see below). The foliated varieties have a wider distribution and occur outside the area of tonalite outcrop (e.g. south of the Seeschartl).

The petrography of the fine porphyritic granite is illustrated in Plate 3 and by modal analyses in Table II—4.

TABLE II—4

Modal Analyses of fine grained porphyritic Granite (A) and Gneisses derived from it (B)

	A			B			
	A 383	A 92	A 387	A 49	A 62	A 81	A 351
K-feldspar	30	14	22	24	15 (g)	29 (g)	39
Plagioclase	35	43	67	64	45	40	50
Quartz	25	32			30	21	
Biotite	10	10	9	8	9	7	5
White Mica	a	a	1	4	1	3	5
Epidote	a	a	a	a	a	a	a
Sphene	a	a	a	a	a	a	—
Apatite	a	—	a	a	a	a	a
Zircon	—	—	—	—	—	—	—
Ore	—	—	—	—	—	a	—
Chlorite	—	—	—	—	—	—	—
Garnet	—	—	—	—	—	—	—

(g) = groundmass K-feldspar only.

Granite-Gneisses

Introduction

On the map the subdivisions of the gneissose granitic rocks are based on colour index (i.e. biotite content), grain size and the presence or absence of augen. The following groups are distinguished

Biotite augen-gneiss
- coarse-grained
- fine-grained
- undifferentiated

Leucocratic granite-gneiss
- augen bearing
- fine-grained
- undifferentiated and leucogneiss

The two undifferentiated classes include regions which are transitional in character or which have intercalations of the other lithologies or where exposure is limited or where more detailed mapping was impracticable. A rather more detailed subdivision of these lithologies was used in the field but it is not practicable to show this for the whole area on the scale at which the main map is presented.

The biotite augen-gneisses differ from the second group in that they contain more (about 10%) biotite and have very little white mica except as partial sheathes around the augen. Both groups are essentially granodioritic in composition and contain roughly equal quantities of quartz, K-feldspar and plagioclase. The coarse augen-gneisses contain K-feldspar megacrysts up to 5 *cm.* in length and commonly 2—3 *cm.*, while the megacrysts in the fine augen-gneisses rarely exceed 2 *cm.* and mostly range from 1—1.5 *cm.*

Leucogneisses nowhere form mappable units but they occur as common intercalations in other lithologies. Typically they are rather variable in composition but contain largely quartz and feldspar and have very little white mica and no biotite. Some are deformed equivalents of fine leucogranite but others are demonstrably older. Certain coarse leucogneisses are probably related to the coarse leucogranite.

Augen-Gneisses

Four of the six lithologies mentioned above carry augen and although they are of polygenetic origin, it is convenient to discuss some of the characteristics which the augen-gneisses have in common before dealing with the lithological groups separately.

These gneisses comprise a wide variety of rocks of broadly granitic composition in which K-feldspar megacrysts ("augen") are set in a well foliated ground-mass composed of quartz, feldspar and mica. While many of the augen are indeed eye-shaped, euhedral megacrysts are also common. In some places augen-gneisses can be traced with decreasing intensity of foliation into granites with little or no foliation; in contrast to these less foliated rocks, the augen-gneisses rarely preserve recognizable original contact relationships at their margins.

Within the gneisses there are marked variations in the size, shape and distribution of augen, and also in the grain size and composition of the ground-mass, and in the nature of its foliation. These variations are illustrated in Plate 3. The augen range from a few millimetres to ten centimetres in length; in some lithologies they are of rather uniform size but in others there is much variation even within a single hand specimen. The shape varies from euhedral to oval; euhedral crystals make up at least a proportion of the augen in all but the most strongly foliated gneisses. The long axis, or longest diagonal, of the augen is generally parallel to the foliation of the matrix but in a few localities the augen orientation, and a crude banding defined by the augen distribution, defines a foliation oblique to the foliation in the ground-mass. The augen generally contain numerous inclusions of plagioclase oriented parallel to crystallographic directions in their host; these are sometimes subhedral but often show extensive corrosion. Biotite is also a common inclusion. Myrmekite occurs in nearly all the augen-gneisses but is restricted to the margins of the augen where it forms excrescences or semi-continuous streaks and patches.

In the less strongly foliated augen-gneisses the foliation of the matrix is defined principally by the mica orientation; in the coarser varieties this results from the elongation of aggregates of biotite, and in finer varieties from parallel orientation of scattered individual biotite flakes (Plate 3). Both quartz and plagioclase occur as more or less equant aggregates made up of a polyhedral mosaic of grains; plagioclase mosaics are usually appreciably finer than those of quartz (Plate 3, c). In the more strongly foliated varieties the foliation is defined by mica orientation and by parallel orientation of segregations of quartz which are from 0.1 to 0.5 *mm.* wide and up to 10 *mm.* long (Plate 3 c). In these more foliated varieties biotite no longer occurs concentrated into irregular clots.

With progressively more intense development of the foliation there are changes in the micas; in the less foliated varieties, biotite is commonly the main ground-mass mica and white mica is restricted to envelopes around the augen; as the foliation becomes more pronounced white mica becomes important in the ground-mass also (Plate 3 c, d, e); in some extremely foliated examples white mica and a little chlorite occur and biotite is not found. In some cases, however (e.g. the leucocratic augen-gneiss), muscovite is present in all varieties from the least to the most deformed.

The plagioclase occurs mainly as small (0.02—0.5 *mm.*) polyhedral grains, but there are occasional larger gefüllte grains, which are apparently relicts from an earlier fabric. The composition is An_{20-30}. The quartz forms aggregates of strained grains, and smaller recrystallised grains within the feldspar mosaic. In the strongly deformed examples, elongate aggregates of sutured quartz develop, wrapping around the augen. The K-feldspar occurs as megacrysts of varying size, and also as smaller grains in the ground-mass. The megacrysts are sometimes micro-perthitic, with strings and patches of exsolved albite, usually on a fine scale, and sometimes scarcely discernible. Simple twinning, usually on the Carlsbad law, is common. Cross-hatch twinning is abundant in the ground-mass K-feldspar, but in the augen, is generally confined to the margins and to the highly strained zones, and is quite often altogether lacking. The 2 V varies from 65° to 90°, in some cases from the centre to the margin of a single megacryst. Single feldspar megacrysts from 22 augen-gneisses were investigated by means of X-ray diffraction. All examples except one exhibited strong (131) peaks of orthoclase. Two thirds of the samples also yielded subsidiary peaks on either side, which were attributed to microcline (131) and ($1\bar{3}1$). The grains appear to have broken down into parts some of which are triclinic, while others retain their monoclinic symmetry. The triclinic domains appear to be associated with a straining of the grain.

In addition to the inversion to microline, partial recrystallisation has commonly occurred, particularly in zones of high strain. Myrmekite is ubiquitous at the margins of the large K-feldspar megacrysts. Inclusions of tablet-shaped grains of plagioclase are common, these are usually filled with epidote microlites, and are frequently oriented parallel to the margins of the K-feldspar. In many examples they have very narrow rims of clear albite, presumably due to perthite exsolution. Inclusions of biotite, muscovite and allanite are also common; the allanite inclusions frequently lack epidote rims, whereas these are ubiquitous in grains within the ground-mass.

In the more strongly developed augen-gneisses, envelopes of white phengitic mica develop around the augen. They appear to be produced by reaction between K-feldspar and biotite (see Chapter III).

The subhedral shape, and lack of preferred orientation of the K-feldspar megacrysts in the less deformed augen-gneisses, and their demonstrable equivalence in some cases to K-feldspar phenocrysts in undeformed granites suggests that they originated during the igneous crystallisation of the rocks. Other data supporting this conclusion are:

(a) The low triclinity and simple twinning

(b) The oriented inclusions of gefüllte plagioclase. Voll (1960) has suggested that oriented inclusions may arise during metamorphism by preferential preservation of favourably oriented grains. However, the volume of plagioclase inclusions in some of the K-feldspars makes this unlikely in this case. The fact that the inclusions are of gefüllte plagioclase means that the growth of the K-feldspar must have occurred before the breakdown of the calcic plagioclase (Frasl, 1954).

c) The perthitic nature of most of the K-feldspars suggests that they must have formed at sufficiently high temperatures to contain an appreciable amount of albite in solid solution.

Within the ground-mass, however, K-feldspar frequently occurs as narrow films along grain boundaries indicating that a limited amount of intergranular diffusion occurred during the later stages of metamorphism.

Biotite augen-gneiss

Coarse biotite augen-gneiss: this group forms extensive rather homogeneous outcrops in the area of the Mühldorfer Seen and in the Mühldorfergraben to the southwest; it also forms the Grubelwand. Near the Obere Thomanbauer Alm the same rocks form a small outcrop which is the tip of a large mass exposed in the northern 2000 *m.* of the tunnel (Gössstollen) between the Gössgraben and the Riekengraben. In addition to their coarse grain size (ground-mass 1—2 *mm.*) these gneisses are characterised by large, commonly euhedral augen, without the typical white mica envelopes. They are cut by discordant pegmatites and aplites.

Around the Mühldorfer Seen near the contact with fine-grained leucogranite the augen-gneiss becomes much more leucocratic; the K-feldspars show better developed cross-hatch twinning, and the quartz more undulose extinction than usual. In several places the augen-gneiss is traversed by discordant shear zones along which it shows stronger foliation parallel to the margins, smaller augen and increasing abundance of white mica; in extreme cases along the shear-surface itself a white mica, quartz schist is developed.

A further variety of biotite augen-gneiss occurs within and near the margins of the tonalite and granodiorite — a coarse biotite augen-gneiss which, where strongly

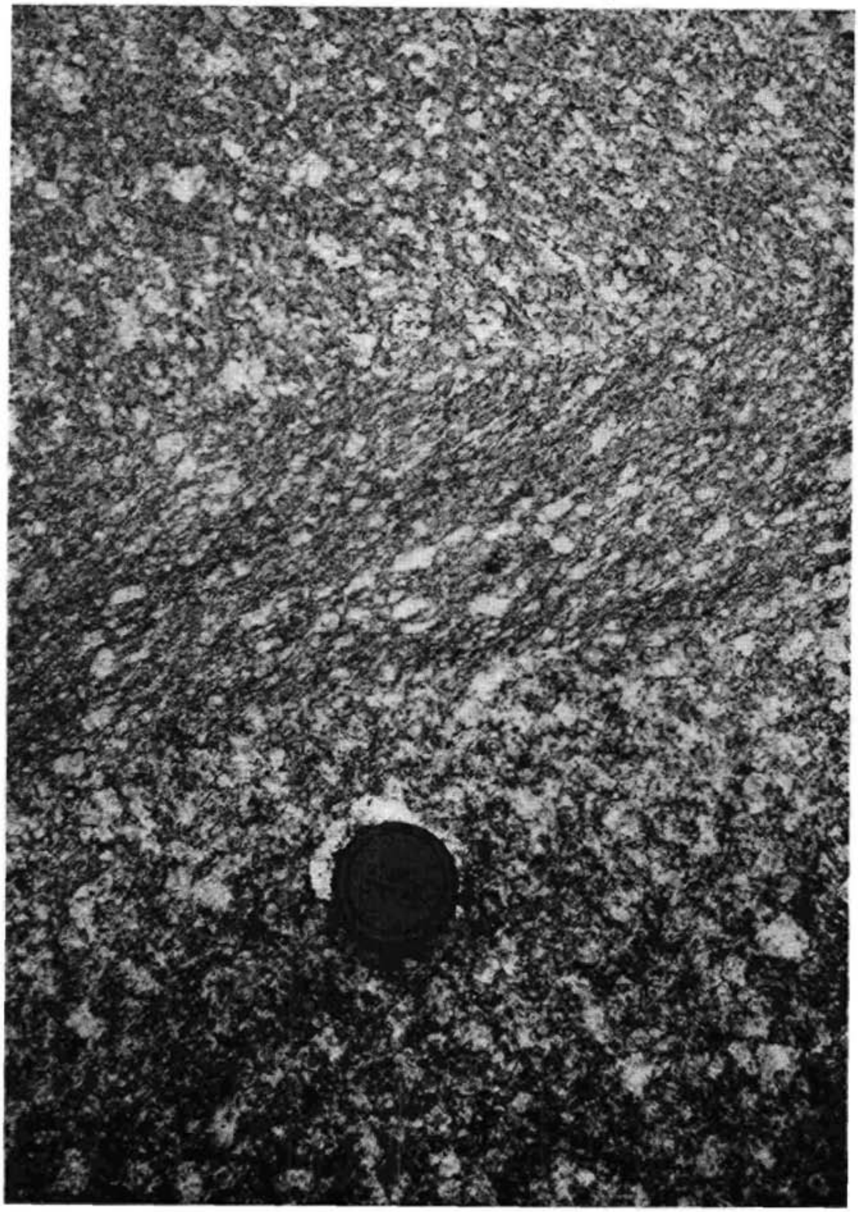


Fig. II · 5 A small-scale demonstration of the development of a gneiss: augen-gneiss developed along a 15 *cm*. wide kink-band in granodiorite; 2600 *m*. on the Schwarzenburgerweg. Zentralgneis.

foliated, is indistinguishable from parts of the biotite augen-gneiss described above but shows a transition through less foliated varieties into the granodiorite. As noted earlier the granodiorite seems to occur principally as a marginal facies along the western limit of tonalite; this transition from granodiorite into augen-gneiss makes the line shown as the contact on the map rather arbitrary. Small zones of augen-gneiss also occur along shear zones within otherwise poorly foliated granodiorite (Fig. II—5; see page 147). In all but the most foliated varieties these gneisses display concentrations of biotite in clots similar to those commonly found in the tonalite-granodiorite lithologies.

Fine biotite augen-gneiss: a further lithology commonly associated with the tonalite and granodiorite, is a fine-grained augen-gneiss with biotite dispersed through the matrix; the augen are of uniform size close to 1 *cm*. in length. This rock can be traced laterally and gradationally into fine porphyritic granite; a good example of such a transition occurs immediately south of the Kleine Gössspitze. Xenoliths of tonalite are abundant and as the foliation in the porphyritic granite becomes more intense so does that in the xenoliths and they become more elongate. Elsewhere (e.g. upper Mühldorfergraben) occasional xenoliths of tonalite occur within the fine biotite augen-gneiss. It is possible, but not certain, that all the fine biotite augen-gneiss is derived from the fine porphyritic granite.

Biotite augen-gneiss: the main mass of undifferentiated biotite augen-gneiss forms an extensive arcuate area of outcrop intercalated with other lithologies and extending from the flanks of the Hochegg in the south, where it is in contact with the Peripheral Schieferhülle, northwestward and then northward across the Schober, Kampleck and Gröneck ridges. On the north side of the Kaponiggraben the field relations are very complex; there are extensive intercalations of other lithologies. The gneisses continue northeastward through the Säuleck and across the Trippkees to the map edge near the Villacherhütte.

The gneisses themselves are rather inhomogeneous and of variable grain size and have numerous intercalations of leucogneiss and banded gneiss. Some of the lithologies are identical to the two varieties of biotite augen-gneiss already described. The coarse parts, however, usually have prominent white mica envelopes around the augen and commonly show streaks of quartz (Plate 3). In addition a fine-grained augen-gneiss with a finer ground-mass and rather uniform (1 *cm*.) Carlsbad-twinned augen may occur as concordant-intercalations; rare discordant contacts and inclusions indicate that this latter rock is younger than the coarser augen-gneiss.

Leucocratic Granite Gneiss

Fine-grained leucocratic granite-gneiss: In the areas shown on the map as undifferentiated leucocratic granite-gneiss, the fine-grained leucocratic granite-gneiss commonly occurs intercalated with a variety of other lithologies. Elsewhere, however, it does form clearly defined homogeneous outcrops, e.g. in the Hohes Gösskar; another body extends from the Schönangersee across to the Gamclnigspitze and down into the Riekengraben; rocks of this latter body also crop out in the Gössstollen hydroelectric tunnel.

In these outcrops the rock is a massive fine-grained, muscovite-biotite granite-gneiss; it is typically very homogeneous but the grain size varies from 0.1 to 2.0 *mm*. and locally small K-feldspar augen are dispersed through the rock. Xenoliths are generally rare but in the Hohes Gösskar they occur near the contacts with coarse biotite augen-gneiss and fine-grained biotite augen-gneiss or porphyritic granite. Below the Törlspitze also there is a raft of tonalite in the fine-grained leucocratic granite-gneiss. These xenoliths suggest that the fine-grained leucocratic granite-gneiss is one of the youngest lithologies of the Zentralgneis. This is confirmed by the contacts of the body as a whole which are characteristically discordant. There is, however, considerable local variation in the nature of the contacts, which seems to be controlled at least in part by the lithology of the country rock. Examples are again taken from the Gösskar; in the north where contacts are with tonalite, granodiorite and biotite augen-gneiss the contact is sharp and clearly discordant; along its eastern margin the intrusive displays a coarser, more leucocratic facies which has sharp contacts against the granodiorite and the biotite augen-gneiss. A subsidiary dyke of leucocratic gneiss extends northeastward into the granodiorite for 100 *m.*; beyond this it is obscured by moraine; southward the contact becomes difficult to define in the field. Along the western contact, there is a gradual transition from homogeneous leucocratic granite-gneiss to leucogneiss with variable proportions of mica schist. A similar contact is seen near the Zwenberger Seen. This complete gradation suggests that the leucogneiss schist mixture may represent a contact facies of the leucocratic granite-gneiss. The contact relations are also well exposed some distance to the south where a sharp, apparently concordant, contact against the banded gneisses was mapped at an elevation of 2100 to 2300 *m.*, north of the Riekener Sonnblick. Above 2300 *m.* and extending across to Riekenkopf and down into the Riekengraben, leucocratic augen-gneiss overlies the fine-grained leucocratic granite-gneiss. No sharp contact is seen; the two lithologies are intercalated on scales from 1 *cm.* to several meters. In places the two rocks are distinguished only by the presence or absence of augen; but generally leucocratic augen-gneiss has a somewhat coarser matrix. Similar relations between these two lithologies are found in the Radlgraben and in the region between the Ritteralm and Tröska Alm. At all of these localities there are small fragments of the leucocratic augen-gneiss in the fine-grained leucocratic granite-gneiss and it seems likely that the relation is an intrusive one, the augen-gneiss being the older. In the Riekengraben the fine-grained leucocratic granite-gneiss is again in contact with banded gneisses and has a sharp, locally strongly discordant contact.

The contact with the tonalite south of Tristenspitze is concordant with the general trend of the foliation and the two rocks are interfolded on a small scale; deformation has here left no trace of the original contact relations. Other important outcrops of the fine-grained leucocratic granite-gneiss occur between the Obere Mooshütte and Schoberspitze; they also extend from the Radlsee around to the Ritteralm. In the latter area the leucocratic granite-gneiss contact cuts obliquely across a weak banding in the biotite augen-gneiss.

Leucocratic granite-gneiss with augen: As noted above this lithology is intimately associated with the fine-grained leucocratic granite-gneiss. Together they form coherent bodies of granite-gneiss within the Inner Schieferhülle and over large areas it was not possible to show them separately at the scale of the map.

A relatively homogeneous sheet underlies the Riekener Sonnblick/Reisseck amphibolites. The Tröska Alm is also in fairly uniform leucocratic augen-gneiss, but on the ridge dividing it from the Ritteralm it is intercalated with fine-grained leucocratic granite-gneiss as noted above. The Ritteralm itself is made up largely of the same rock but intercalations of other granite-gneisses, leucogneisses and banded gneiss are more abundant.

Shear-Zones

As described in chapter III, shear zones are quite widespread in the Zentralgneis, and mineralogical changes seem to have occurred within them. Three typical examples are described.

The first is at the base of the body of leucocratic granite-gneiss west of the Mühldorferseen, along the line of the proposed thrust at the margin of the Gössgraben Kern. The gneiss becomes richer in white mica towards the contact, and an 0.5 *m.* thick zone of a micaceous rock with nodular feldspar occurs at the base. In thin section (RN 440) the feldspar is seen to be largely chess-board albite occurring as large (5 *mm.*) tablets. A second generation of plagioclase (An_{20-30}), occurring as smaller (1 *mm.*) untwinned grains, with irregular, rounded margins, appears to be replacing the albite.

The large albite grains are somewhat fractured and deformed, and are partially recrystallised to the later plagioclase. The rock contains practically no K-feldspar. About 20—30% of quartz occurs as small (0.5 *mm.*) grains with the smaller plagioclases, and white mica and biotite make up 10—15%. The white mica has a very pale green pleochroism, suggesting a phengitic composition. Thus K-feldspar seems to have been replaced by chess-board albite and phengite, albite later recrystallising and being replaced by oligoclase.

The second example is at the base of the sheet of biotite augen-gneiss south of the Rosskopf (RN 473). The augen-gneiss grades into about 1 metre of white mica phyllonite at the contact.

The third example (RN 153) occurs at the margin of a small body of leucogranite in the upper Hinterregengraben. The leucogranite grades over about 5 *cm.* into a 10 *cm.* thick, white mica+quartz phyllonite. The white mica forms fairly large (5 *mm.*) strain-free flakes, having a good preferred orientation, and appears to replace K-feldspar, plagioclase and biotite until only white mica and quartz remain.

These latter two examples are similar to the phyllonites at the margins of the Sonnblick lamella.

Discussion

From the descriptions of the petrography and field relations which have preceded, a number of provisional conclusions may be drawn. The Zentralgneis is composed in large part of granodioritic gneissose rocks within which a number of varieties may be distinguished; in general the distinctions are reflected in texture and fabric rather than in composition. A number of the gneissose rocks may locally be traced laterally into relatively weakly deformed facies which retain many of their original intrusive textural features and in which original contact relations with other rock groups are also commonly preserved. Why deformation should have affected these rocks so inhomogeneously is not clear; the problem is discussed in more detail in a later chapter. It is, however, in part related to the distribution of the large metasedimentary units into which the igneous rocks were intruded.

On the basis of relationships established in the less deformed facies, particularly the existence of discordant intrusive contacts and the presence of xenoliths of one lithology in another, it has been possible to establish their relative ages; these age relationships within the less deformed rocks may, subject to certain limitations, be extended to those more deformed lithologies into which the less deformed show gradational transitions. Difficulties arise, however, because of the frequent occurrence of 'tectonic convergence': many of the highly gneissose lithological groups are polygenetic and although virtually identical today, have been derived from various lithologies which were originally distinct — this is probably true of the leucogneisses, some very sheared augen-gneisses and, in the Inner Schieferhülle, the banded gneisses (see below). Furthermore, it appears that although some of the very deformed groups may have been derived from lithologically similar parent rock types, these parents were of differing age: the best example is the coarse biotite augen-gneiss which is, in part, the deformed equivalent of the granodiorite, while elsewhere its fabric relationships show it to be older. Clearly an age relationship established for part of one of these strongly deformed gneisses cannot necessarily be extended to the whole. The proposed correlation for the more deformed lithologies is given in Table II—5.

TABLE II—5

Summary of igneous rock types and their deformational derivatives

Undeformed		**Deformed**
Tonalite	→	"Biotite schist"
Granodiorite	→	Coarse-grained biotite augen-gneiss
Fine porphyritic granite	→	Fine-grained biotite augen-gneiss
Coarse leucogranite	→	"various leucogneisses"
Fine leucogranite	→	

The age relations of the various parts of the Zentralgneis deduced in this way are shown in Table II—6. It is shown in Chapter V that some parts of this intrusive sequence including the youngest parts, are Variscan. Other parts could be significantly older but there is no evidence. The suite of granitic lithologies of the Zentralgneis could, however, easily have been part of a single, granodioritic-tonalitic, polyphase Variscan intrusive series; such series characterize certain orogenic belts today, e.g. the Sierra Nevada batholith of the western cordillera of North America (Bateman et al., 1963).

TABLE II—6

Age relations in the Zentralgneis

	Decreasing age →			
Part of Biotite augen-gneiss	Tonalite	Aplites	Coarse Leucogranites	Fine grained porphyritic granite
				Aplites
? Part of Leuco-gneiss	Granodiorite		Fine-grained Leucocratic Granite-gneiss	Fine Leucogranites

Note: these relationships would also hold for the deformed equivalents of these rocks.

The Inner Schieferhülle

Introduction

The Inner Schieferhülle represents, at any rate in part, the country rocks into which the Zentralgneis was intruded. Evidence is presented in later chapters that these rocks had already undergone an amphibolite facies regional metamorphism and at least one phase of folding before the intrusion of much of the Zentralgneis. In the central Tauern Frasl (1958) recognizes two main pre-mesozoic metasedimentary groups, the Habach series and the Altkristallin. The rocks described here as Inner Schieferhülle have more in common with Frasl's Altkristallin than the Habach Series, which he describes as having undergone little folding and only mild metamorphism at the time of intrusion of the Zentralgneis. Rocks laterally continuous with the Inner Schieferhülle have been termed Central Schieferhülle by Exner (1964) in the area covered by his Gastein map.

In the area of the present map, the main outcrop of the Inner Schieferhülle forms the "Reisseck Mulde", a nearly flat-lying, wedge-shaped body of schists, banded gneisses and amphibolites overlying the gneisses of the "Gössgraben Kern" which outcrop to the east and underlying a large body of Zentralgneis which outcrops to the west. This latter body of gneiss is termed the Hochalm Kern by Exner. Similar lithologies to those of the Reisseck Mulde also occur closely intercalated with each other in a narrow band along the northeast flank of the Mölltal where in most places they lie between the Peripheral Schieferhülle and the main part of the Basement Complex. These rocks have generally undergone more intense deformation than those of the Reisseck Mulde; they and the associated very sheared parts of the Zentralgneis have been termed Randgneis by EXNER (1954). We here consider the banded gneiss, schist and amphibolitic parts of the Randgneis along with the Inner Schieferhülle, which they closely resemble except in intensity of deformation.

Banded Gneisses

Banded Gneisses make up a large portion of the Inner Schieferhülle. In the present work, any gneiss having a banded or striped appearance in the field is termed a banded gneiss, and a non-genetic classification based on the nature of banding will be used. It is to be noted that these rocks are quite distinct from migmatites, as discussed later. Two major types occur:

(1) Finely banded gneisses: these are characterised by leucocratic (plagioclase + quartz) and melanocratic (plagioclase + quartz + mica + epidote) layers, interbanded on a 0.5—10 *cm.* scale. The origin of the banding is not known, but as it occasionally exhibits two phases of folding, there is presumably a considerable deformational element in its production. The banded gneisses within the Randgneis unit (Fig. II—6; see page 152) are more finely banded, and frequently develop small (1—2 *mm.*) porphyroblasts of plagioclase, giving the rock a 'knobbly' appearance. However, these are also included as finely banded gneisses.

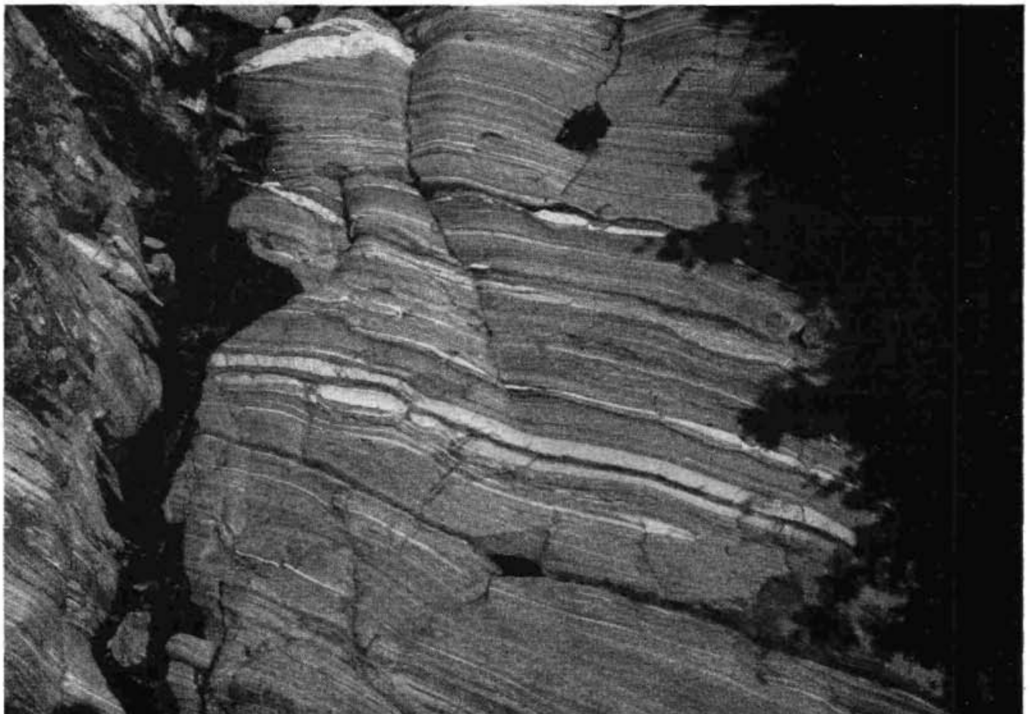


Fig. II - 6 Fine banded gneiss; note both boudinage and tight isoclinal folding, low centre right; Randgneis in the Riecken Graben. Field of view 1.3 *m.* high. Inner Schieferhülle.

(2) Grey banded gneisses: these are grey biotite gneisses interbanded on a 10—100 *cm.* scale. Some layers contain dispersed augen, usually small (0.5—1 *cm.*) but up to a maximum of 2 *cm.*, and always strongly deformed as eyes within the foliation. Other layers consist of a very fine-grained biotite-plagioclase-quartz gneiss totally lacking augen; locally, as in the Hohes Gösskar and around the Villacherhütte the development of grey banded gneiss through deformation of intercalated granite-gneisses with cross-cutting leucocratic veins, can be demonstrated. Much of the grey banded gneiss, however, appears unrelated in space to any of the main masses of Zentralgneis and, in any case, exhibits relict structures pre-dating much of the latter. Consequently these rocks seem best considered as a separate lithology, and of uncertain, probably polygenetic, origin. The differences between them and the fine banded gneisses are: (1) the layers are thicker and (2) the layers are much less strongly contrasted in colour and composition. Thin lenses of amphibolite are occasionally found within this lithology. At a number of localities, the two types of banded gneiss just described, together with thin units of amphibolite and leucogneiss are interlayered on such a scale that it is not practicable to distinguish them on a map. In these cases the lithology is shown, simply, as banded gneiss.

Petrographically the banded gneisses are fairly monotonous; the following assemblages appear to be in equilibrium:

(1) qz+olig+bi+ep±mu±ore+acc

(2) qz+olig+ksp+bi±mu±ep+ore+acc

The quartz and plagioclase generally occur as a fine-grained (0.1—0.4 *mm.*) polygonal mosaic, with 120° triple junctions. The plagioclase tends to exhibit weak zoning and a little twinning. If it is zoned, the inner zone is generally around An_{20} and the outer zone An_{25-30}. Occasionally, larger grains of plagioclase or quartz may occur, and are partially recrystallised to smaller grains. K-feldspar is abundant only in the augen-bearing portions of the grey-banded gneiss. Here it forms large (10 *mm.*) crystals, rather deformed and largely recrystallised to a polygonal mosaic. Weak cross-hatch twinning is common. In other gneisses with assemblage (2), the K-feldspar forms less than 10%, and occurs as intergranular growths, and as patches and films parallel to cleavage and composition planes in the plagioclase. Myrmekitic intergrowth of quartz and plagioclase occurs in the K-feldspar rich rocks. White mica is most abundant in the K-feldspar bearing rocks but is generally less than 10%. Biotite usually displays a greenish pleochroism (X = light yellow, Y = olive green to green-brown, Z = dark green/brown), and may vary from 1% in the lightlayers to 30—40% in some of the dark layers.

Epidote commonly occurs as rounded, moderately to highly birefringent grains similar to, or smaller than, the plagioclase in size. Frequently they contain optically disoriented, brown pleochroic cores of allanite, suggesting two stages of epidote growth.

In the banded gneisses of the Randgneis, the plagioclase tends to form large (5 *mm.*) porphyroblastic grains, generally untwinned, and containing pools and stringlets of quartz (e.g. RN 421). Similar porphyroblastic plagioclase also occurs in the amphibolites of the Randgneis, and seems restricted to this unit.

The principal ore minerals are magnetite and haematite, the latter occasionally containing remanent sulphide cores.

Veined Gneisses

Any one of a number of lithologies may be cut by distinct veins of aplite, microgranite and pegmatite. Most lithologies contain concordant and discordant leucocratic veins, but these are not usually sufficiently abundant to warrant treatment as a separate unit. However, in the Maltatal above Pflüglhof, in the floor of the middle-upper part of the Gössgraben, and at a number of other rather more restricted localities, extensively veined rocks are abundant. In the Maltatal and Gössgraben, the rock is essentially a grey banded gneiss, with numerous irregular layers of amphibolite and biotite schist, which have been strongly folded by two phases of folding and intensively veined by three phases of aplites and pegmatites (Fig. II—7; see page 154). No detailed study has been made of this group but one set of veins appears to predate the early folds, and some veins apparently postdate the folding altogether. At most of the other localities, the veining is fairly clearly associated with the intrusion of leucocratic granite-gneiss and leucogranite. The term migmatite is reserved for lithologies in which distinct veins are not seen, and the leucocratic material forms a greater part of the rock as a whole.

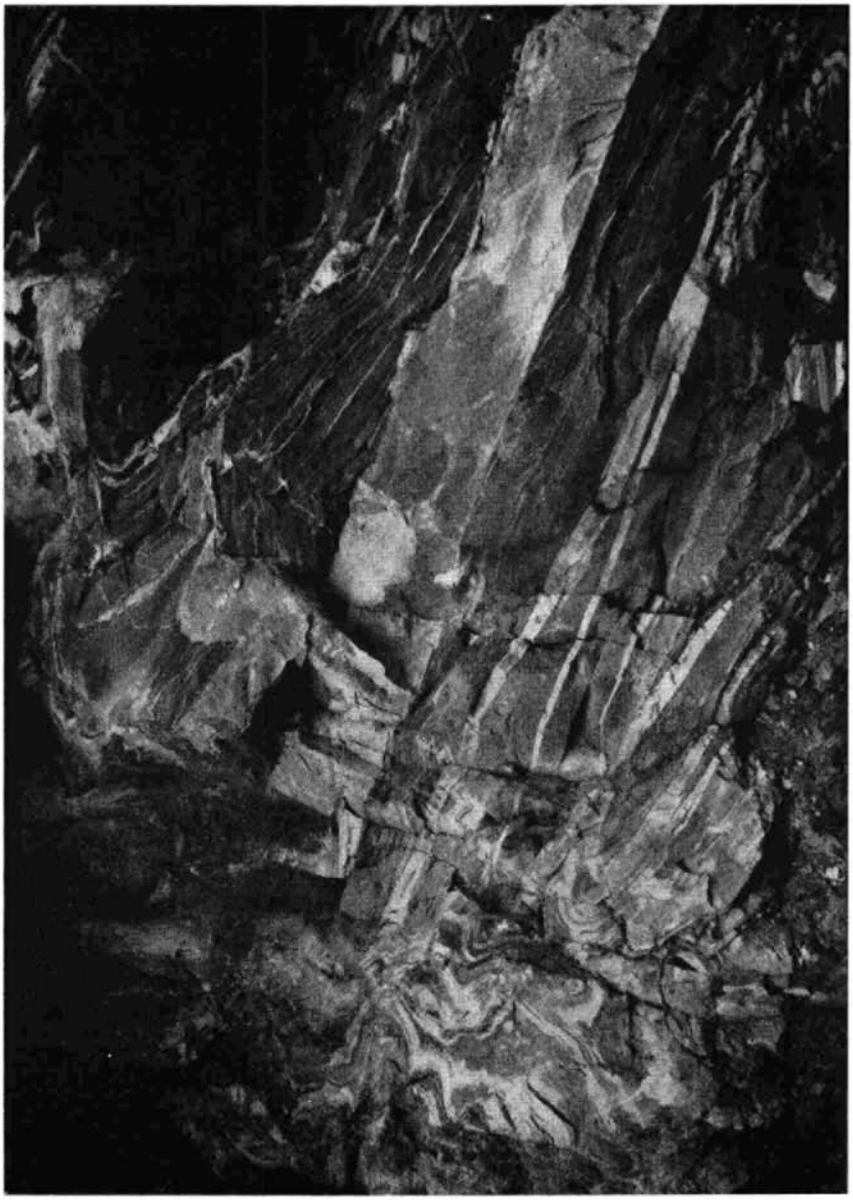


Fig. II - 7 Veined gneisses of the Göss Graben; three generations of cross-cutting veins occur in a biotite quartzo-feldspathic paragneiss host. Exposure about 4 m high. Inner Schieferhülle.

Amphibolite Group

The amphibolite group as shown on the map is a heterogeneous collection of rock types, among which amphibolites make up about 90%. The remaining 10% may include leucogneiss, banded gneisses, biotite schists and garnet-mica schists. However, these all occur in quantities too small to be represented individually on the map.

The amphibolites themselves consist of a variety of types, the principal ones being:

(1) Fine-grained banded amphibolite: this is predominantly a homogeneous black, fine-grained amphibolite consisting of about 50% hornblende, together with plagioclase, and subsidiary epidote and biotite. These form an even-grained texture, with a grainsize of around 1 *mm.* or less. These are inter-layered with thin (0.5—5 *cm.*) bands richer in feldspar or epidote.

This lithology typically occurs in the Hohes Gösskar, on the south side of the Gössgraben, and above the Stapniksee.

(2) Coarse feldspar amphibolites: these consist of short, stubby prisms of hornblende, between 3 and 10 *mm.* long, in a plagioclase matrix, with subsidiary biotite and epidote. The ratio of hornblende to plagioclase is rather variable. In some examples, about equal proportions of the two form a coarse, even texture, while in others, small grains of hornblende are associated with plagioclase in a fine-grained matrix in which much larger (0.5—1 *cm.*) prisms of hornblende occur. This type forms most of the amphibolite outcrop between the Riekentörl and the Stapniksee, but appears to die out to the north. They are also found to the south of the Riedbock, and in the upper Hinterregengraben. They are sometimes interbanded with the fine-grained amphibolites, but are more usually associated with the types described below.

(3) Coarse massive biotite amphibolites: these may contain up to 90% hornblende, with subsidiary plagioclase, although there is a gradation into type 2. Large porphyroblastic flakes of biotite up to 1.5 *cm.* long are characteristic of this type of amphibolite, and the hornblende grains themselves may reach similar dimensions, although generally they are a little smaller. Within other lithologies such as the grey banded gneiss, they occur as pods, or more commonly as concordant lenses and boudinaged layers within the feldspar amphibolite. At a number of localities, for instance above the Stapniksee, angular fragments of massive biotite amphibolite occur embedded in feldspar-rich amphibolite. Interpretation of this relationship is difficult. It is unlikely to have been produced by disruption of the biotite amphibolite within the feldspar amphibolite during deformation, as the fragments are angular and include occasional mica-schists, a lithology which is not found locally. These rocks are rather restricted in distribution as they are known only from single outcrops and cannot be traced to neighbouring exposures. This could represent an intrusive relationship between the two rocks, although there is no other evidence for the intrusive character of the feldspar amphibolites. A further possibility is that if the amphibolite group is a sequence of metavolcanics, the fragmental rock is some form of metamorphosed volcanic breccia.

(4) Garnet amphibolites: these are rather rare throughout the northwestern part of the area, although somewhat more abundant in the Randgneis. They consist of small, 2 *mm.* diameter, garnets set in a fine-grained amphibole and plagioclase groundmass. Garnets are occasionally found in the feldspar amphibolites, but are not common.

(5) Actinolite amphibolites: these are composed of almost 100% actinolite, which has an equant rather than acicular habit, producing a pale green, granular appearance in hand specimen. They are not common, but occur in two localities — between the Sonnblick and the Hochegg, and on the Rosskopf. Here they take the form of thin (50 *cm.*) bands and lenses in the feldspar and biotite amphibolites. At both these localities, they are associated with the next lithology.

(6) Serpentinite: three small, oval bodies, about 10 *m.* by 5 *m.*, occur in the amphibolite group of the upper Hinterregengraben. One is on the south side of the Sonnblick, one in the valley near to the Ochsen Alm, and one on the east face of the Rosskopf. The banding in the gneiss and amphibolite appears to wrap around the body near the Alm, and a talc envelope is developed at its margin. Although a little carbonate is associated with the serpentinites, this does not reach the proportions found in the Peripheral Schieferhülle.

Petrography of the amphibolites: assemblages recorded are:

(1) antig+tc+dol+mt+ht
(2) hb+olig+bi±ep±qz±ore+acc
(3) hb+bi+ep±qz±ore+acc
(4) hb+olig+acc
(5) hb+olig+bi+ksp+ep+ore+acc
(6) hb+olig+bi+gt±ep±ore+acc
(7) act+acc
(8) hb+di+plag+ep±ore+acc
(9) hb+olig+bi+chl±gt±ep±ore+acc
(10) olig+phl+chl+mu±hb±gt±ore+acc

In the absence of evidence to the contrary, assemblages 2—7 are considered to be equilibrium assemblages. Textural evidence strongly suggests that assemblages 8—10 are in disequilibrium.

Assemblage (1) is that of the serpentinite bodies in the upper Hinterregengraben described above. Although the assemblage is similar to that found in the serpentinites of the Peripheral Schieferhülle the textural relationships and proportions are quite different. Antigorite makes up practically 90% of the rock. The carbonate occurs as small (0.2—0.4 *mm.*) rounded grains within the serpentine, while the talc occurs as a few large (0.5—2 *mm.*) porphyroblasts, generally kinked and fractured. The textures do not suggest that either of these minerals are replacing the serpentine.

Assemblages (2), (3) and (4) are those typical of the bulk of the amphibolites of the Reisseck Mulde. Of these, the coarse feldspar amphibolites are the most abundant. The amphibole in these is a blue-green hornblende (X = pale yellow, Y = green, Z = blue green), usually with a deeper pleochroism than those in the Peripheral Schieferhülle. Hornblende occurs both as small (0.3 *mm.*) grains within the ground-mass, and larger megacrysts (up to 5 *mm.* long) usually in the form of short stubby prisms. In many rocks (e.g. RN 189, A 346) the latter appear to have an inner zone full of minute opaque inclusions, and an outer clear zone (this has also been described in metamorphosed pre-Alpine dykes in the Sonnblick Kern by Exner, 1964). The larger grains also frequently exhibit fractures filled by the quartz, plagioclase and epidote of the ground-mass.

In addition, the large grains show the development of low-angle sub-boundaries, along which new grains have nucleated, until eventually the larger grain is replaced by an aggregate of smaller grains. If the grain was disrupted during deformation, it may now be represented by an elongate aggregate or string of small grains within the foliation.

The plagioclase within the feldspar amphibolites also occurs in two generations. The earlier grains are large (2—4 *mm.*) grains, frequently exhibiting polysynthetic twinning, and full of microlites of epidote or clinozoisite, suggesting a compositional adjustment from a higher An content (gefüllte Plagioklas). These have partially or completely recrystallised to a fine-grained (ca. 0.03 *mm.*) polygonal mosaic. In many rocks, however, there are no large grains and it is difficult to discover whether they ever existed.

The composition of the "gefüllte" grains appears to range from An_{18} to An_{25}. The smaller grains are usually zoned, with up to three zones being present. The intermediate zone usually has the highest refractive index, but either the centre zone or outside zone may be missing. In many grains, the centre zone has sharp, irregular boundaries and exhibits twinning not reflected by the outer zones, suggesting that it might be a remnant of an earlier grain. The compositions of these smaller grains within the feldspar amphibolites, ranges from An_{23} to An_{40}, with a concentration between An_{25-30}. Fig. II—8 is a histogram of feldspar compositions from all types

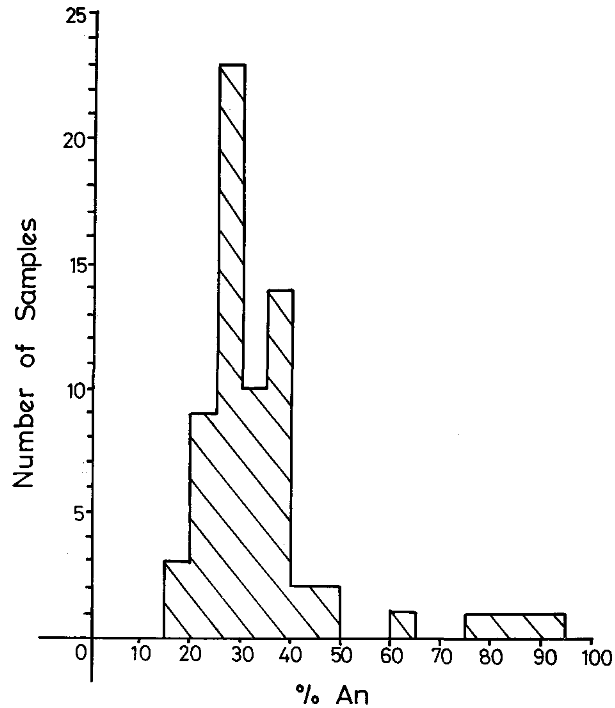


Fig. II - 8 Histogram of anorthite contents of plagioclase from the Inner Schieferhülle.

of amphibolite within the Reisseck Mulde. Modal analyses of a few typical feldspar amphibolites are presented in Table II—7.

TABLE II—7

Modes of feldspar Amphibolites

	RN 38	RN 65	RN 458	RN 339 (schist in shear zone)
Hornblende	20	31	37	1
Plagioclase	47	40	38	31
Quartz	15	1	—	—
Biotite	16	12.5	9.5	63
Epidote	1.5	14	14	acc
Ore	} 0.5	—	0.2	1.5
Acc (sphene, apatite)		1.3	2.5	0.5
Chlorite	—	—	—	3

The banded amphibolites are similar in mineralogy but more fine-grained than the feldspar amphibolites, with no large hornblendes or biotites. The banding is defined by variable proportions of hornblende and plagioclase. In the coarse, massive amphibolites, the hornblendes, which usually have rather pale pleochroism (Z = pale blue-green) suggesting an approach towards actinolite in composition, may reach up to 1 *cm.* in length, and make up over 80% of the rock. They are frequently fractured and partially recrystallised to smaller grains, especially around the margins. Quartz and plagioclase have crystallised interstitially, but biotite frequently forms large (0.5—1 *mm.*) poikiloblastic flakes. The pleochroic scheme of the biotite varies, but the two commonest are: X = pale yellow, Y = olive green, Z = green-brown and X = pale yellow, Y = Z = yellow-brown. Green pleochroic biotite is quite abundant, in contrast to the Peripheral Schieferhülle.

Epidote occurs in two forms: as small, weakly birefringent microlites (0.1—0.2 *mm.*) within the plagioclase, and as highly birefringent (2nd order green) rounded grains 0.5—1 *mm.* in diameter. These latter grains frequently exhibit a very pale yellow pleochroism. This, together with the birefringence, suggests that they are rich in ferric iron.

Sphene and apatite are practically ubiquitous as accessory minerals, and the former may occasionally reach 1 or 2%. Rutile also commonly occurs, and in some rocks, is rimmed by sphene. The principal ore minerals are magnetite and haematite. Occasionally, haematite may have a yellow reflective core suggesting alteration of a sulphide.

Assemblage (8) occurs in two rocks, both from near to the Obere Zwenbergersee.

One such rock is a thin (2 *cm.*) vein of diopside-hornblende-plagioclase-epidote bearing rock enclosed in a feldspar amphibolite. The diopside forms very large (1—2 *cm.*) rectangular grains, somewhat fractured and granulated around the margins. It is slightly altered to a very pale green amphibole at its margins and along fractures but is otherwise quite fresh. At the sides of the vein, however, the pyroxene forms smaller grains rather altered to pale green actinolite and a dusty material, which under high power, appears to consist of tiny needles of clinozoisite. Amphibole grains occur within the vein, and many of these appear to have rather patchy cores of brownish amphibole. These were analysed using the electron microprobe. The results are given in Table II—8.

TABLE II—8

Microprobe analyses of Amphiboles

	RN 249		RN 248
	Brown core (average of 3) wt%	pale green rim (average of 3)	(average of 5)
Al_2O_3	9.6	3.3	7.5
FeO	11.8	10.3	9.5
MgO	10.6	16.3	18.2
TiO_2	0.63	0.17	0.3
CaO	12.6	13.2	12.7
At. proportions based on Ca = 2			
Ca	2.00	2.00	2.00
Al	1.63	0.54	1.31
Mg	2.59	3.47	4.00
Fe	1.48	1.21	1.16
Ti	0.07	0.02	0.04

The rims are actinolite in composition, containing only 0.54 atoms per formula unit of Al (DEER, HOWIE and ZUSSMAN, 1963). The cores are of hornblendic composition, and also contain more Fe, Ti and considerably less Mg than the rims. The plagioclase in the vein is in the form of a polygonal mosaic (0.5 *mm.* diam.) with 120° triple junctions. Nearly all the grains are, however, strongly zoned, the inner zone being in general of higher An content than the outer zones. The centres of most grains contain abundant needles of clinozoisite, suggesting an adjustment from a high An content. The composition was determined by three different methods; the results are set out in Table II—9 for comparison.

TABLE II—9

Michel-Levy method	**U-Stage** (after TURNER 1947)	**Electron microprobe**
An 47	An 38 to An 45	An 30 to An 46
(Maximum extinction angle)	(real variation)	(range of 21 grains or zones)

The probe results confirm that the cores tend to have a higher An content; the average for the core is An_{42} whlie that for the outer zone is An_{37}.

The surrounding feldspar amphibolite (RN 248) contains aggregates of small grains of pale blue-green hornblende (see Table II—8), suggesting recrystallisation from larger grains, set in a fine-grained (0.1 *mm.*) mosaic of plagioclase and epidote. The plagioclase grains are full of tiny needles of clinozoisite, which crowd every feldspar/feldspar boundary. These also, were investigated with the microprobe. The compositions show a considerable variation, but cluster into two groups, An_{58-65} and An_{76-94}, with no compositions in between. The plagioclase is clearly adjusting from a higher An content by the production of epidote, the discontinuity in the An contents may reflect a compositional gap in the plagioclase series (e.g. NISSEN, 1968).

A second diopside-bearing rock occurs as a 20—60 *cm.* thick band of coarse feldspar amphibolite within fine-grained amphibolite. Pale blue-green hornblende is in two forms: large (3—6 *mm.*) rectangular prisms, and small (0.1—0.3 *mm.*), though frequently euhedral, grains. The larger grains are in the process of recrystallising to the smaller grains. The diopside occurs as remnants and small grains within the amphiboles and is clearly being replaced by the latter. As in the other example, the plagioclase forms a mosaic (0.2—0.5 *mm.*) absolutely full of minute clinozoisite needles. The composition is An_{35-57} (by the maximum extinction method). A similar diopside-bearing rock was recorded from a raft of amphibolite in the tonalite.

Garnet-bearing amphibolites occur occasionally within the Reisseck Mulde and commonly in the Randgneis. The garnets occasionally exhibit two phases of growth; those in the Randgneis are frequently fractured and partially replaced by chlorite, biotite or epidote, and atoll forms are common.

Chlorite is often present as an accessory mineral; it has green pleochroism, anomalous interference colours and is clearly a late stage alteration of the biotite. In contrast to the Peripheral Schieferhülle, it does not occur as part of the normal assemblage.

Garnet Mica Schists

At many localities these are associated with, and grade into, the migmatites discussed below. Xenoliths of this lithology are found within the tonalite to the north of the Villacher Hütte. These are fairly fine-grained quartz-mica schists studded with small (2 *mm.*) garnets, which are usually abundant. Below the main tonalite outcrop at this locality and also further south, similar garnet schists and mica schists occur, here containing lenses and veins of quartzo-feldspathic material concordant to the foliation, and developing garnets up to 5 *mm.* diameter. Apart from thin layers within the migmatites, the other main occurrence is in the Randgneis. The petrography of these rocks is discussed below. Locally the mica schists contain thin, banded, locally graphitic layers of quartzite.

Schistose Gneisses

These are an extremely fissile series of mesocratic rocks which are transitional in their properties between fine-banded gneisses and mica schists. The main area of outcrop of these rocks is in the northwestern corner of the map area where they occur on Sicker Kopf and south of it. The rocks are medium-grained, quartzofeldspathic, micaceous, locally finely banded, platy gneisses. At many horizons they carry garnets which may show peripheral retrogression to chlorite. With increasing mica content they grade into mica schists; with decreasing mica content and the development of compositional banding they grade into finely banded paragneisses; they may carry either muscovite or biotite or both. The petrography of these rocks is discussed below.

Migmatite

The use of this term is hazardous, because of the variety of meanings it has carried. Sederholm (1923) used it to designate metasediments invaded by granitic material producing a 'mixed rock' with both metamorphic and igneous components. Since then it has been used to include rocks undergoing partial melting (Barth, 1962) and commonly to signify that partial melting has occurred. The term is used here non-genetically to refer to a 'mixed rock' in the sense of Sederholm. This does not mean that the 'igneous component' has necessarily been introduced by magmatic intrusion, but merely that it has a magmatic appearance.

The main belt of migmatites crosses the head of the Gössgraben, runs across the Zwenbergerkar, and is found on the Gamolnigspitze—Kampleck ridge, and less abundantly on the Schoberspitze.

In all these localities, the rock consists of approximately 50% leucocratic granite gneiss and 50% schistose material, although locally, it tends to grade towards garnet-mica schist on the one hand and granite-gneiss on the other. The granitic component does not form distinct veins or lenses (cf. the veined gneisses) but forms an interconnected matrix in which wisps of schist occur. However, within the granitic part, one finds rather patchy variation in composition, such as biotite-rich patches, occasionally with garnets, suggesting destruction of an original schist fragment. On passing towards zones richer in schist, the granite occurs as more distinct veins and patches.

The origin of these rocks, whether by granite intrusion, granitisation or partial melting, is not clear. They seem to be everywhere associated with the leucogratic granite-gneiss. In some cases, (e.g. on the Zwenbergerkar), they are clearly the marginal zone of the granite and pass gradationally into it; possibly volatile-rich magmatic solutions permeated the schist, finding penetration easy along the foliation; it is not clear that this could cause partial melting.

Petrography of Mica schists and Migmatites and schistose gneisses: The following have been recorded:

(1) qz+olig+bi+ep+ore+acc
(2) qz+olig+bi+mu+chl+gt±ep+ore+acc
(3) olig+bi+mu+chl+gt±ep+ore+acc
(4) qz+olig+bi+mu+ksp+ep+gt+ore+acc
(5) qz+olig+bi+mu+chl+ep+gt+ky+ore±cc+acc
(6) qz+olig+bi+mu+chl+gt+ky+st+ore+acc

Of these only (1) is clearly an equilibrium assemblage. The rest exhibit signs of textural disequilibrium. Assemblage (1) occurs in thin biotite schist layers in the banded gneisses, and in schist xenoliths in the tonalite of the Zentralgneis.

Assemblage (5) occurs in two specimens from the Villacher Hütte area. In one (RN 390), leucocratic lenses occur in mica schist. The quartz forms a well recrystallised polygonal mosaic; the plagioclase is also recrystallised but rather sericitised (An_{25-30}). The garnets are large (2—7 *mm.* diam.) and fragmentary. Anomalously birefringent chlorite, and biotite occur along the cracks, together with haematite, which sometimes completely rims the grain. The garnets are also honeycombed with plagioclase. The reactions, garnet $\longrightarrow$ plagioclase + biotite + haematite, and biotite $\longrightarrow$ chlorite, appear to have taken place.

The kyanite occurs as rather ragged blades altering to white mica. This would suggest some mobility of K^+ and water. In the second specimen, the rather sparse prisms of kyanite are apparently unaltered, although the garnet is again very fractured, the cracks being filled with plagioclase, biotite and chlorite. In this rock chlorite occurs both as highly birefringent (1st order white) grains with the micas, and anomalously birefringent flakes replacing the biotite. The rock also contains abundant (10%) patches of small equant grains of a light-brown carbonate material (? ankerite).

The migmatites around the Zwenberger Seen and on the Schoberspitze contain a large amount of garnet-mica schist, somewhat invaded by leucocratic granite. At an exposure near to the Obere Zwenbergersee, three specimens of a garnet-mica schist showing progressive amounts of invasion were collected.

The first comes from a small body of mica schist containing practically no granitic material. Biotite (20%) and muscovite (60%) are the principal constituents. The biotite is a deep red-brown variety. The muscovite consists of large flakes which are beginning to break down into mats of tiny flakes with a large number of small prisms of zoisite as inclusions. The plagioclase (An_{28}) also contains a large number of zoisite needles and flakes of white mica. Garnet makes up about 2% and forms rounded grains about 1—1.5 *mm.* in diameter usually surrounded by a rim of biotite.

The second sample is from a similar lithology making a thin (20 *cm.*) layer enclosed in the granite material while the third consists of wisps of the schist within the leucocratic granite. The large (1—2 *mm.*) garnets occur within the granite and have the appearance of xenocrysts. The changes are described in Table II—10.

Thus, with increasing introduction of granitic material, quartz and plagioclase increase in quantity, K-feldspar appears, the large flakes of muscovite break down, and new garnets are produced at the expense of biotite. These changes are seen to occur in all the samples of migmatite examined, from the Schoberspitze to the Kaponigtörl. In some examples (e.g. RN 313) patches of a very fine chlorite and clinozoisite mesh appear to be pseudomorphing some earlier porphyroblastic mineral.

Garnets in three of these rocks (RN 245, 311, 313) were studied using the microprobe (Fig. II—9 a, b, c, d). The first sample is from the Obere Zwenbergersee area, while the other two are from the Schoberspitze. The four large garnets scanned show identical patterns of zoning which are, however, very unlike the majority of published examples. The most prominent zoning is that of manganese, but this element increases in concentration towards the margins, the opposite of the normal situation. One of the garnets has an inner zone richer in inclusions and this shows a sharper change in concentration — others are more continuously zoned. Ca tends to be higher in the centre, as are Fe and Mg, while Ti is lower. Because accurate counts could not be made, only estimates of the composition are presented. Both the zoning pattern and the composition differ from the garnets in the Peripheral Schieferhülle. As these garnets are frequently altered and deformed, and occur as xenocrysts in the granitic component of the migmatites, their formation is Pre-Alpine. In some examples (e.g. RN 345) they contain an internal planar fabric at an angle to that outside.

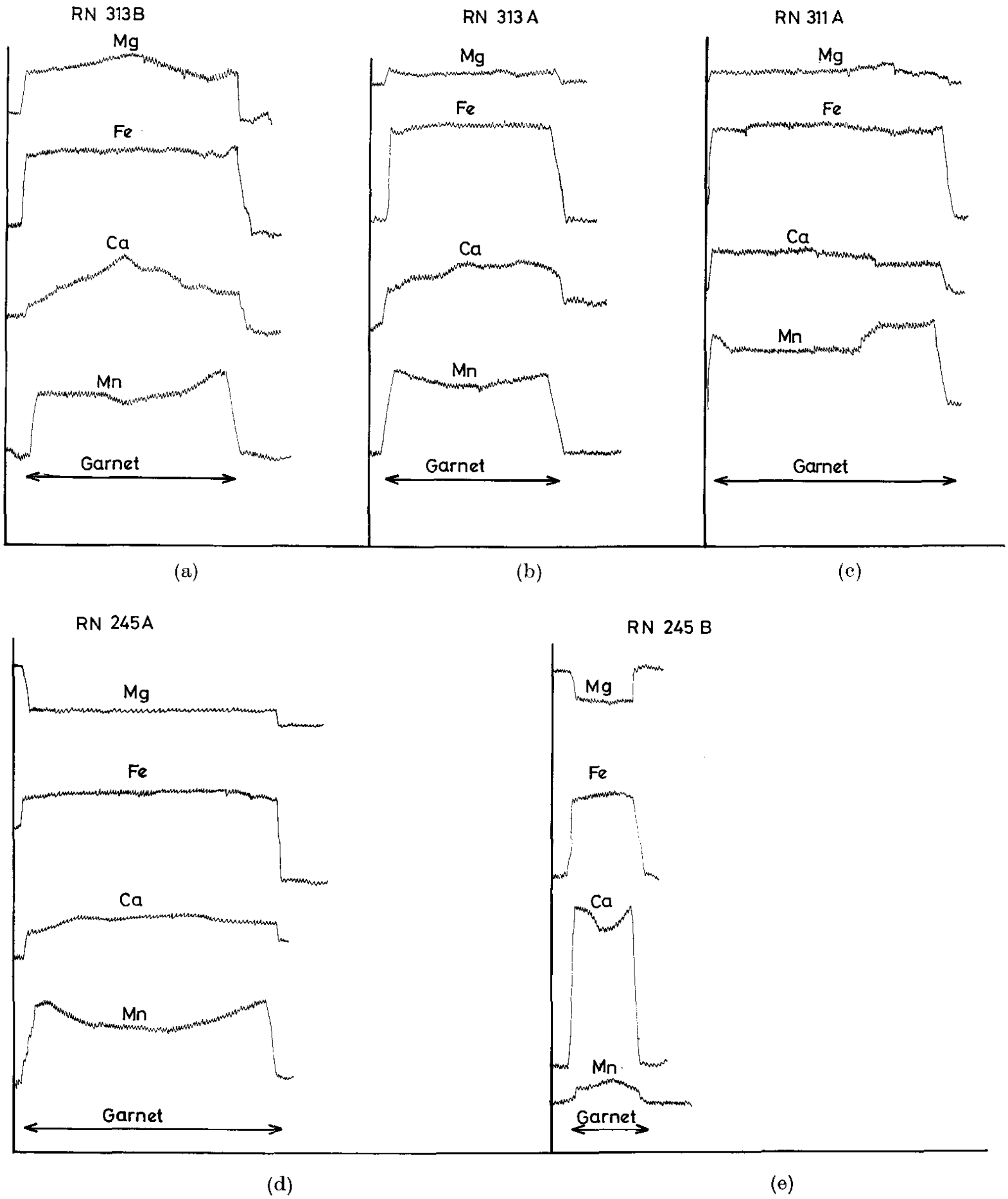


Fig. II - 9 Electron-microprobe scans across garnets from the migmatites of the Inner Schieferhülle.

The fact that the mats of mica delineate the $F_A{}^1$ folds, suggests that at least the minerals they are replacing are pre-Alpine also.

The garnets replacing the biotite are mostly too small to study by scanning. One larger grain of this generation was scanned, however (Fig. II—9 e), and it exhibits a somewhat different zonary pattern and composition. Mn is higher in the centre,

TABLE II—10

Progressive invasion of Migmatites, Oberer Zwenbergersee

'Normal' garnet-mica schist (RN 245)	Schist Lens in Granite (RN 246)	Wispy Schist Patches in Granite (RN 247)
Large (1—2 *mm.*) gt in matrix of 1—2 *mm.* long flakes of bi+mu. A small amount of plag An 25—30	2 generations of gt i) large (1—2 *mm.*) grains as in RN 245 ii) tiny (0.01 *mm.*) grains rimming and replacing bi.	2 generations of gt i) large (1—2 *mm.*) as before; appear as xenocrysts in granite ii) small gts replacing bi;
	Mu flakes broken down partially or completely to fine-grained mesh of mu+chl.	Mu completely replaced by fine-grained mu+chl mesh. Elongate needles of zo within these meshes, especially where meshes are partially replaced by and invaded by plag+qz.
	30% plag (An 28) irregularly shaped grains partially recrystallized to fine (0.01 *mm.*) mosaic. Abundant inclusions of mu+ep; qz appears as 1 *mm.* grains.	30% plag (An 27); lobate boundaries with qz.
		Qz 40—45% large (6 *mm.*) unstrained grains ksp appears as irregular patches around mu+chl meshes.

but its total concentration is much lower. Ca is lower in the centre, but its total concentration is much greater. Thus the later generation does seem to differ chemically from the earlier one.

It is in one of these samples (RN 313) that assemblage (6) occurs. The kyanite forms rare, narrow blades, relatively unaltered, whereas the staurolite, also very sparse, occurs as small grains rather altered to biotite.

The age of the minerals in the migmatites is difficult to ascertain. The older garnets have been discussed already. The kyanite and staurolite could be either Alpine or pre-Alpine. Kyanite also occurs in schists incorporated in the tonalite west of the Kaponigtörl, where it has been altered, first to albite and subsequently to white mica; again there is no indication of the age of these reactions.

The age of the biotite to garnet reaction is also open to speculation. The fact that it is seen only in the migmatites suggests that it is in some way connected with the introduction of the granitic material.

In many of the schists, two generations of mica occur; the first is sharply folded and mimetically recrystallised in the $F_A{}^1$ crenulation cleavage, while the second tends to grow parallel to the axial plane. In one example (RN 31), the first generation of biotites have a pale yellow to yellow green pleochroism, are folded into $F_A{}^1$ folds and form an intricate myrmekite-like intergrowth with quartz, whereas the second generation is pale yellow to deep reddish brown in colour and is in the form of elongate flakes parallel to the axial plane.

The Peripheral Schieferhülle

Introduction

The Peripheral Schieferhülle may be distinguished in the field from the Inner Schieferhülle by the criteria listed at the beginning of this chapter. It is found in the lower parts of the steep-sided valleys on the northeast flank of the Mölltal. The strike is generally parallel to the Mölltal, and the dip steep over much of the outcrop, so that good sections may be observed in the bottoms of these valleys. Most of the observable sequence lies between the Sonnblick lamella and the main outcrop of the Basement Complex, thus forming a southerly continuation of the Mallnitzer Mulde, although here there is no symmetrical arrangement of lithologies suggestive of a simple synform.

The sequence may, however, be followed more or less directly into the Peripheral Schieferhülle of the Sonnblick group, with which very close lithological comparisons can be made: there is thus little doubt that the rocks here termed Peripheral Schieferhülle are the same as those described by EXNER (1964), and as those of FRASL (1958) — Seidlwinkl Trias and Bündnerschiefer — in the type localities of the Glockner depression, with which they are in continuation around the Sonnblick Kern.

The Zentralgneis of the Sonnblick Lamella

The Sonnblick lamella is a thin tongue of Zentralgneis extending down the NE flank of the Mölltal. Northwards, it can be traced into the Knappenhauswalze (EXNER, 1962, 1964) which is thought to be part of the Sonnblick Kern. In the lower Kaponiggraben, the lamella is approximately 300 *m.* thick; in the Riekengraben it is only 200 *m.*, and by the Mühldorfergraben, the thickness is no more than 20 *m.* This is the furthest south that it can be traced. The river bed of the Riekengraben immediately above the railway viaduct exposes a good section through the lamella. At the downstream contact there are one or two metres of phyllonite, grading over about five metres of very sheared gneiss into recognisable augen-gneiss. The phyllonite consists largely of phengite and quartz with remnants of K-feldspar and plagioclase. The latter minerals increase in amount through the zone of sheared gneisses until, in the augen-gneiss, each is approximately equal in proportion to the quartz. The augen-gneiss itself, however, contains very little biotite and is exceptionally rich in phengite. The K-feldspar augen may reach four to five centimetres although the average size is somewhat less. They frequently retain sharp corners, although usually having well developed tails. In thin-section they are seen to be microcline perthites, with well developed cross-hatch twinning and abundant fracturing and recrystallisation and are usually partially replaced by chessboard albite. Small (1—10 *cm.*) biotite-rich inclusions are fairly abundant, as in the Zentralgneis. Two types of larger (1—3 *m.*) inclusion occur. The first is a fine-grained quartzo-feldspathic rock occurring as discontinuous patches in the augen-gneiss, while the other is a melanocratic rock containing hornblende, epidote, biotite and chlorite as mafic minerals, with 1—2 *mm.* porphyroblasts of albite. This latter occurs as irregularly shaped inclusions in the gneiss; in thin-section and hand-specimen, it closely resembles basic dykes recorded by EXNER (1964) as intruding the gneiss of the Sonnblick Kern. At the upstream contact, the augen-gneiss again passes through two or three metres of highly sheared gneiss into a metre-thick quartz-phengite phyllonite; here, however, a small amount of banded gneiss also occurs in the contact zone. In the Kaponiggraben, for instance in the exposures below the Laskitzer viaduct, the gneiss has a similar appearance. In the Mühldorfergraben, all the gneiss is very sheared, and resembles the highly sheared zone at the margins, but is still clearly recognisable as augen-gneiss.

There can be little doubt that the Sonnblick lamella is indeed a thin tongue of Zentralgneis extending 20 *km.* down the northeast flank of the Möll, surrounded on all sides by Mesozoic Schieferhülle; it is almost certainly not a metamorphosed arkose in the Schieferhülle sequence as suggested by FRASL (1958) for gneiss lamellae in the Glockner depression. A little below the lower contact of the Sonnblick lamella in the Riekengraben, EXNER records another such tongue of gneiss, the Rote Wand—Modereck lamella. This is represented by a couple of metres of very sheared quartzo-feldspathic rock, however, and is more problematical. No other lamellae are found in the Riekengraben but further north the present authors have mapped another between the Sonnblick lamella and the main outcrop of the Basement Complex (see also EXNER, 1964); here again the rock is very sheared and gneissose, with occasional augen, and appears to be in the form of discontinuous lenses rather than a continuous layer.

The argument for these latter being thin tongues of Zentralgneis is reasonable but not indisputable, but there is no other reasonable interpretation of the Sonnblick lamella and its existence adds credibility to the existence of others.

Mineral assemblages found in rocks thought to belong to gneiss lamellae are:

(1) qz+mic+mu±ab±bi±ep±cc+ore+acc
(2) qz+mic+mu±olig±bi±ep±cc+ore+acc
(3) qz+mu+cc±bi+ore+acc

The Calcareous Group of Lithologies

Within this group are included calc-phyllites, calc-schists, black phyllites and marbles. The phyllites and schists exhibit all gradations and it is difficult to represent them on the map as separate units (e.g. FRASL [1958, p. 400] "... the normal boundaries, at least between calc-phyllite and black phyllite, are not always very sharp in nature ... the question is then paramount, at what percentage of $CaCO_3$ should one draw the line ..."). For this reason and because the exposure in the lower parts of the area is poor and the structure complex, these individual lithologies are not distinguished on the map. At nearly all localities where this group is shown, most of the following lithologies occur.

Black phyllites

These are fine-grained dark rocks composed principally of quartz, mica, chlorite and opaque carbonaceous material, together with a variable amount of calcite. In one or two localities, they are garnet-bearing. They are gradational into calc-schists and form no recognisable outcrop pattern on a map, but would have to be shown as discontinuous patches and lenses within the calc-schists. They are therefore included in the calcareous group of lithologies.

Calc-phyllites and calc-schists

These are the dominant lithologies within the group. The calcite content varies from 15% to 70%, grading one way into phyllitic schist and the other way into marble. However, the average calc-schist probably contains 30—50% calcite. Other minerals are quartz, epidote or zoisite, and white mica. The rock rarely shows any colour banding, but layers richer in mica are frequently folded while the mica flakes themselves are oriented parallel to each other defining a good mineral foliation. Very occasionally within the calc-schists, small pods (ca. 0.5 *m.* long) of calc-silicates occur, developing tremolite or actinolite, talc and epidote. A similar development is found in the calc-schists surrounding serpentinite bodies.

Calcite marble

These rocks contain between 70% and 90% calcite, with subsidiary quartz, zoisite and white mica. They may form bands up to 10 *m.* thick, of homogeneous blue marble in the calc-schists; elsewhere they exhibit a pronounced white/pale-yellow/blue-grey colour-banding and are interlayered with calc-schists. In some localities, marble layers about 10 *cm.* thick are regularly intercalated with more schistose material. The shape of these marble bodies is not clear, however, as it is not possible to correlate them from one outcrop to another.

Dolomite marble

Very occasionally, thin bands of buff-weathering dolomite marble are found within the calc-schists. Although only of the order of one metre thick, they may be traced for one or two hundred metres laterally. They do not, however, appear to be continuous for distances much greater than this, so that their value as marker horizons is limited.

Petrography of the Carbonate Rocks

In general these rocks are very similar to those described by Exner (1964).

The following assemblages have been noted:
(1) cc+qz+mu±ep±bi±chl±ore+acc
(2) cc+qz+chl+bi+olig+acc
(3) cc+ep+olig+hb±qz±ore+acc
(4) dol+tr±mu±ep±cc±ore
(5) Mg/ank+tc

The vast majority of the calc-schists and marbles exhibit assemblage 1. Biotite and chlorite, if present, are usually accessories only. The ore minerals are generally pyrite and pyrrhotite, while other accessories include apatite and tourmaline. Calcite generally makes up between 40 and 90% of the rock. Occasionally it has an annealed fabric, but more frequently, the grains are elongated (1—2 *mm.* length) within the foliation and exhibit abundant lamellar twinning, the composition planes tending to be subparallel to the foliation. RN 398 is a sample from the calc-phyllites in the Lower Riekengraben, south of the Sonnblick Lamella; Exner (1964) attributes this series to the Unterostalpin. The sample consists of calcite, quartz and white mica. The fabric is highly strained, with large grains of quartz fractured, strained, and recrystallising into a very fine-grained mosaic. Within the rock, thin layers of fibrous calcite occur, with the fibres perpendicular to the sides of the layers (Fig. II—10; see page 166). It is possible that these represent highly deformed fossils, which have not been totally obliterated. Most of the other assemblages have only limited occurrence.

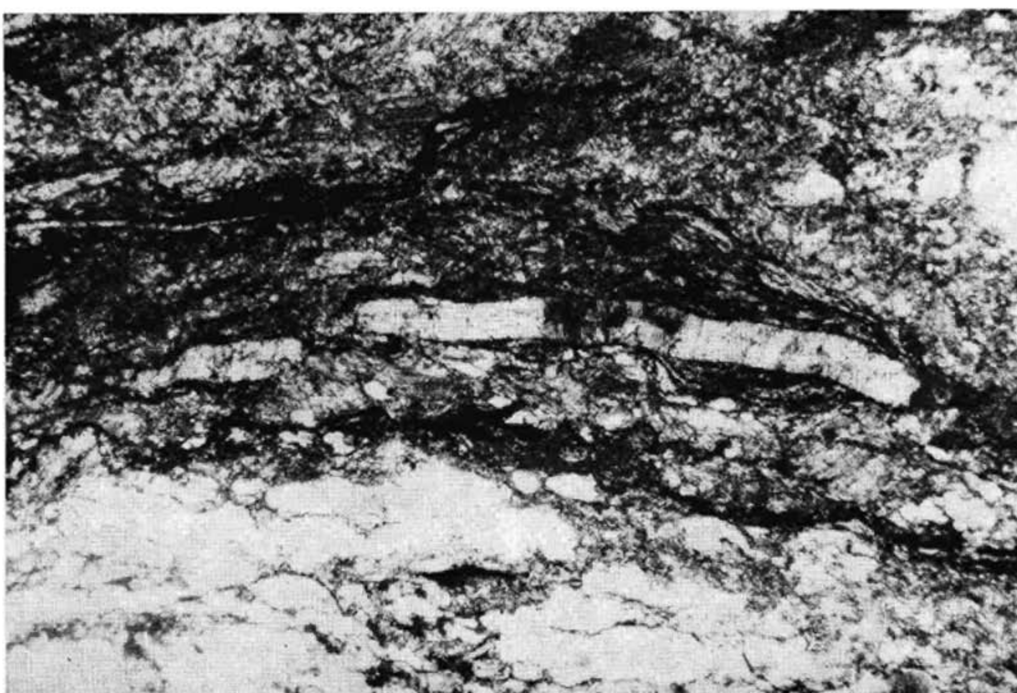


Fig. II - 10 Photomicrograph of calc-phyllite in the lower Riecken Graben; note thin layers of fibrous calcite possibly representing deformed shell debris. X 10. Peripheral Schieferhülle.

The thin dolomite marbles contain 90—95% dolomite, with occasional later veins of calcite, while the remaining 5% or so consists of tremolite, white mica and clinozoisite. This is similar to the assemblages recorded in some of the greenstones although with different proportions.

Assemblage 5 occurs in a thin (0.5 *m.*) layer in the calc-schists of the Mühldorfergraben (RN 429). Large (5—10 *mm.*) grains of the carbonate are set in a fine-grained talc matrix.

Quartzite and Dolomite-quartzite-breccia

In the Kaponiggraben there are good outcrops, on the logging road up the north flank of the valley, of a quartzite containing strongly deformed dolomite pebbles. The carbonate tends to weather out on the surface leaving cavities which give the rock a honeycomb appearance in the field. The pebbles are in the form of elongate cylinders forming a well defined linear feature parallel to the local fold axis. They vary in size from about 1 to 10 *cm.* in length and a tenth of that, or less, in diameter.

There is a tendency for the pebbles to form regular bands within the quartzite (Fig. II—11; see page 167), possibly with a gradation in pebble size from large through small into a dolomitic quartzite. However, such is the state of deformation, that in sections parallel to the lineation, one can see no clear pebbles at all, but only alternating white and buff streaks, and it is only in sections perpendicular to the long axes of the pebbles that the rock may be clearly recognised as a breccia. In thin-section the matrix consists of 95% quartz with minor amounts of white mica and sericitised feldspar, while the pebbles are formed of a fine-grained equant mosaic of dolomite.

Fig. II - 11 Dolomite-quartzite breccia near Kaponig. Note massive beds of quartzite with highly deformed pebbles of dolomite weathering out along certain horizons. Peripheral Schieferhülle.

This lithology appears to take the form of a number of large lenses floating in the calcareous lithologies. In the outcrops in the Riekengraben, there are very few recognisable dolomite pebbles although locally the rock is dolomitic and grades into calc-schists. What few recognisable patches of dolomite there are tend to be flattened discs or smears within the foliation rather than elongate cylinders. Besides the white and yellow weathering quartzite, dark grey graphitic quartzites also occur in the sequence. Some of these contain layers rich in pyrite which are folded, suggesting that the development of pyrite represents an original compositional inhomogeneity in the rock. In the Riekengraben and to the south, the calcareous lithologies in the vicinity of the quartzite contain a large proportion of black phyllite. The quartzites, however, are somewhat thinner here, and disappear completely further south.

The mineral assemblages found in the quartzites are covered by

$$qz+mu\pm plag\pm ep\pm bi\pm chl\pm cc+acc$$

They consist of between 60% and 90% quartz, so that most of the other constituents are accessories. The quartz usually forms a mosaic of slightly elongate grains (ca. 0.5 *mm.* long) with rather sutured margins; they show undulatory extinction and abundant basal deformation lamellae. Often the larger grains are partially recrystallised around the margins to smaller grains.

White mica is generally the most abundant secondary mineral, and plagioclase, in many cases, is being replaced by clinozoisite.

Mica Schists

All mica schists, apart from the distinctive "Weiss-schiefer" are mapped as one unit, although there is some variation within the group. The principal varieties are garnet-two mica-chlorite schists and chlorite-white mica-albite schists. The latter contain porphyroblasts of albite up to 5 *mm.* in diameter, although frequently oligoclase is also present in the rock. The remainder of the rock is made up of quartz, chlorite and white mica, with subsidiary epidote and calcite. The garnet-mica schists, which make up the larger part of this unit, are principally composed of quartz and abundant white mica. Some also contain chlorite, while in others, both chlorite and biotite occur. Small amounts of plagioclase are usually present, and accessory epidote or clinozoisite is ubiquitous. In some rather unusual schists in the Mühldorfergraben, large porphyroblasts of zoisite completely replace plagioclase and form a major part of the rock. The garnets range in size from 1 to 10 *mm.* in diameter. They are usually black or dark red in colour, and are irregularly dispersed throughout the rock. In much of the schist, quartz forms thin (1 *mm.* or less) lenticular laminae within the foliation. These are not quartz pods or discrete lenses, but merely zones relatively richer in quartz and low in mica. Carbonate is nearly always present as an accessory mineral, but it largely occurs along grain-boundaries and in-filling fractures, suggesting that its introduction was late.

Tourmaline is frequently present in small amounts, and in one or two localities, staurolite, kyanite and chloritoid also occur. The relationships of the mica schists to the other lithologies are difficult to establish. The group forms a continuous outcrop down the flank of the Mölltal, but the shape is very irregular along strike. Its contacts with the other lithologies are usually clear and abrupt, although it can be seen to be interfolded with the greenstones and bands of greenstone are enclosed by schist (e.g. in the Mühldorfergraben). Deformation is, however, intense and all contacts are extremely sheared; it is therefore impossible to know whether the present day relations between rock groups have any original stratigraphic significance.

In thin-section the following mineral assemblages are recorded:

(1) qz+ab+mu±ep±chl±bi+ore+acc
(2) ab+chl+cc
(3) qz+ab+gt+bi+cc
(4) qz+ab+olig+chl+mu±gt±bi±ep+ore+acc
(5) qz+olig+mu±chl±bi±ep+ore+acc
(6) qz+olig+mu+gt±chl±bi±ep+ore+acc
(7) qz+olig+mu+gt+bi+zo+cc+ore+acc
(8) olig+mu+gt+bi+zo+cc+ore+acc
(9) qz+olig+chl+bi+cc+ore+acc
(10) gt+chl+bi+mu+ore+acc
(11) qz+mu+ep+ky+ore+acc
(12) qz+mu+gt+chd+ore+acc
(13) qz+olig+mu+gt+chl+bi+ep+chd+ore+acc
(14) qz+olig+mu+gt+chl+bi+st+ore+acc
(15) qz+olig+mu+gt+chl+bi+zo+st+tr+ore+acc

Numbers 1—12 probably represent equilibrium, and numbers 13—15 disequilibrium, assemblages.

White mica, biotite, and chlorite coexist in most of the rocks, and the local absence of any one of these minerals is probably due to chemical control rather than to a change in metamorphic conditions (e.g. ALBEE, 1965). The chlorite tends to have a lower birefringence and to become optically negative in the western (higher) parts of the Peripheral Schieferhülle, but that in the lower section, near the contact with the Zentralgneis, generally has a birefringence of 0.005—0.007 +ve, suggesting a higher Al and Mg content.

The biotite generally exhibits rather pale pleochroic colours, although some may be a deeper, red-brown; green pleochroic biotites are not found. In the majority of the rocks chlorite is equal to, or greater than, biotite in abundance.

White mica is practically ubiquitous; in many rocks particularly the Weissschiefer, it has a distinct greenish tinge, suggesting a phengitic composition.

In many specimens, the micas exhibit an intense crenulation in thin section suggesting the deformation of an earlier foliation. The present grains, however, are not themselves bent but have crystallised mimetically. In one or two cases, later cross-cutting biotite grains overgrow this second foliation, suggesting three phases of mica growth. The early foliation is also defined by lines of opaque particles. These are intensely folded and overgrown by large (0.5—5 *mm.*) porphyroblasts of albite or plagioclase (Fig. II—12; see page 170). However, the included fabric may be rotated with respect to the external S, and in one case (C 212) a plagioclase grain has grown round the earlier folded fabric, been deformed itself, and then undergone further growth. It is noteworthy that the internal fabric very rarely includes large grains of mica.

A sharp break in plagioclase compositions, between An_{0-5} and An_{20}, is found in the pelitic rocks and a similar break will be described below in the basic rocks. The albite-bearing rocks occur to the southwest of those with oligoclase. In the intermediate zone, oligoclase is found rimming and replacing the porphyroblastic albite. The oligoclase may form similar porphyroblasts to the albite, but in many cases it forms smaller, elongate patches of recrystallised grains. These tend to be sericitised, and quartz frequently nucleates on grain boundaries, forming strings and pools within the plagioclase. It is possible that during grain growth quartz may become incorporated into a plagioclase grain as pools and strings, forming the myrmekite-like intergrowth frequently found in these K-feldspar-free rocks.

The composition of the plagioclase is more variable than in the greenstones described below, An_{42} being the highest anorthite content recorded. It is interesting to note that in the zoisite-rich rocks, in which there is very little plagioclase, its composition is andesine.

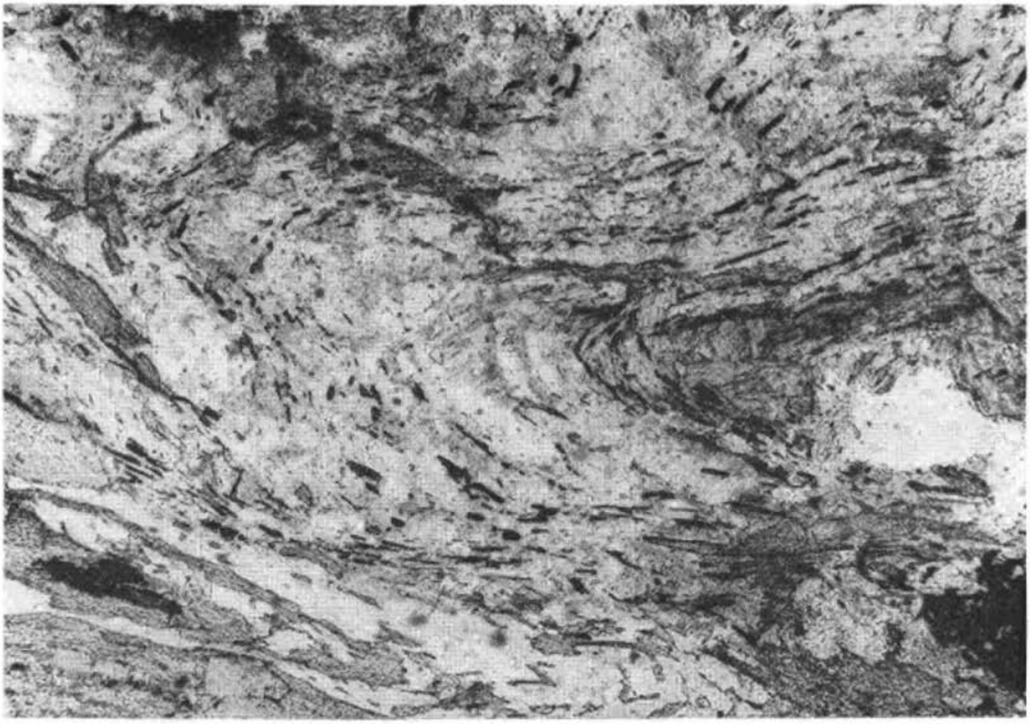


Fig. II - 12 Large grains of albite overgrowing folds in the foliation defined by small opaque particles. Note that the mica flakes define an axial plane foliation. A mica-schist from the Peripheral Schieferhülle X 10.

The garnets tend to be large (2—7 *mm.*), and rather sparsely and irregularly distributed. Many of the garnets, particularly those in rocks nearer the Möll (e.g. RN 419) have a curved internal fabric mainly of fairly large quartz grains. The amount of curvature, however, is compatible with a flattening of the foliation around the growing garnet (e.g. RAMSAY, 1963). In places the outer part of the garnet has grown along quartz grain boundaries, producing a fish-net texture (Fig. II—13). However, in several

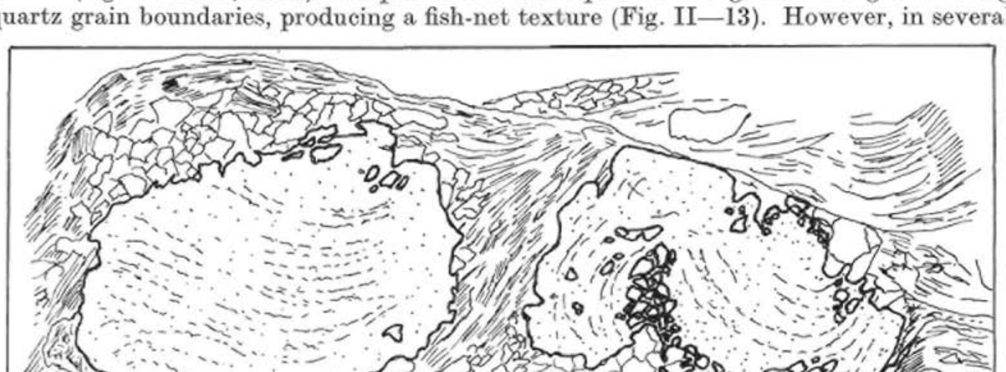


Fig. II - 13 Garnets in sample RN 414; note rotated internal S, and the late intergranular growth of garnet on the right. Matrix of quartz and white mica. Inner Schieferhülle near contact with Zentralgneis, Riecken Graben. For analysis of garnet on left see Fig II - 16.

Fig. II - 14 Zoned garnets from the Inner Schieferhülle; see Fig II - 16 for analysis of upper crystal (RN 431).

specimens, particularly from the eastern part of the Peripheral Schieferhülle, the garnets exhibit two distinct zones. The inner zone contains only fine inclusions, forming curved inclusion trails, while the outer zone contains quartz grains of a similar size to those of the ground-mass and grades into a fish-net texture. This is particularly well shown by RN 431, where there is a sharp discordance between the two zones (Fig. II—14). In other examples, the inner zone is itself bent into an S-shape (e.g. RN 441, 414, Figs. II—13, 15).

The mica fabric tends to wrap around all garnet grains, and quartz-rich pressure shadows are present, suggesting that some flattening of the foliation occurred after the growth of the garnet.

These textures suggest that the inner zone grew in a fine-grained matrix and was subjected to a rotation during growth, while the later growth occurred by intergranular diffusion in a coarser, recrystallised ground-mass, principally during flattening perpendicular to the foliation. Garnets which have grown in a plagioclase mosaic (e.g. RN 427) have not grown along the boundaries but in a regular lattice-type texture parallel to the feldspar cleavages. This suggests that in plagioclase, diffusion occurred through the grains rather than along grain boundaries as in quartz.

Many studies have been made on garnets in recent years using the microprobe (e.g. ATHERTON and EDMUNDS 1966, HENLEY 9168, BETHUNE et al 1968, BROWN 1968, DRAKE 1968, LINTHOUT and WESTRA 1968), and compositional zoning has been de-

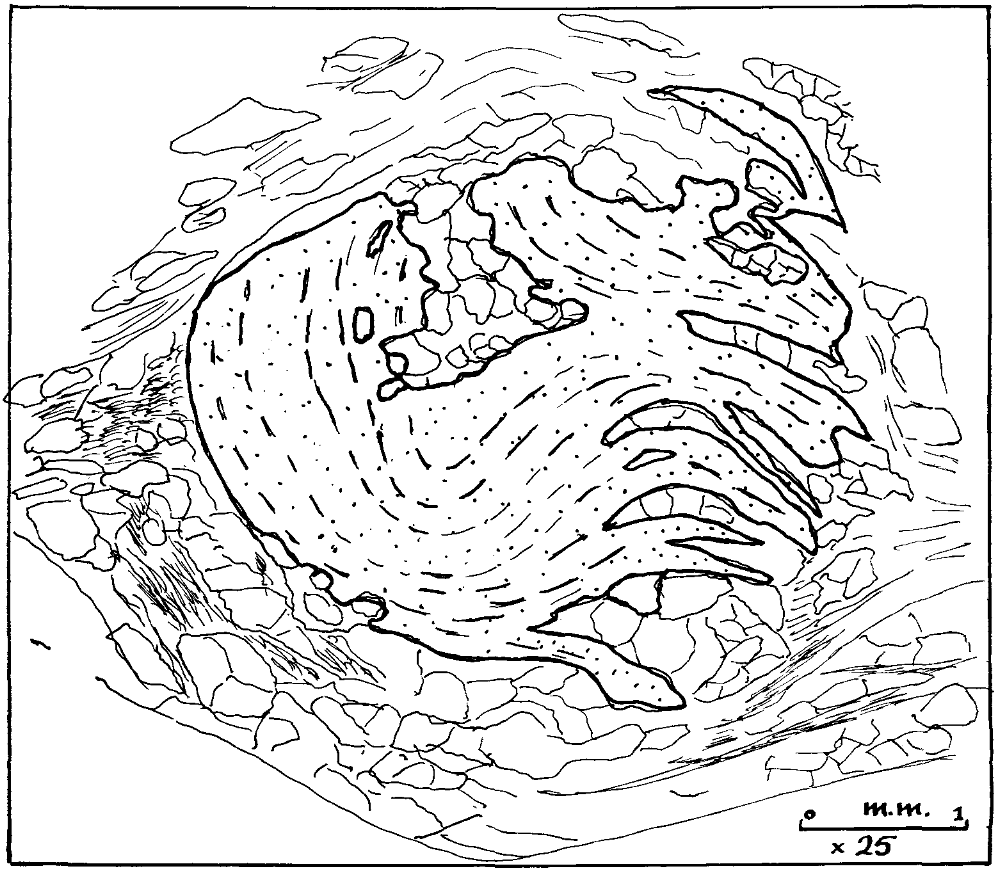


Fig. II - 15 Garnet RN 441 showing rotated internal S and late intergranular growth. Inner Schieferhülle of the Mühldorfer Graben. Analysis Fig II - 16.

monstrated in most garnets from low and intermediate grade metamorphic rocks. Garnets from only three rocks were examined in this way because of operational difficulties. The rocks were RN 414, RN 431 and RN 441 A. The first two have assemblages (14) and (15) respectively, and occur practically at the contact with the Zentralgneis, while the third has assemblage (5) and occurs some distance west of the contact. All garnets exhibit some degree of optical zoning, those of RN 431 having the greatest discordance between the two zones. Because the garnets were large and sparsely distributed, only four could be investigated. The results are shown in Fig. II—16. Note

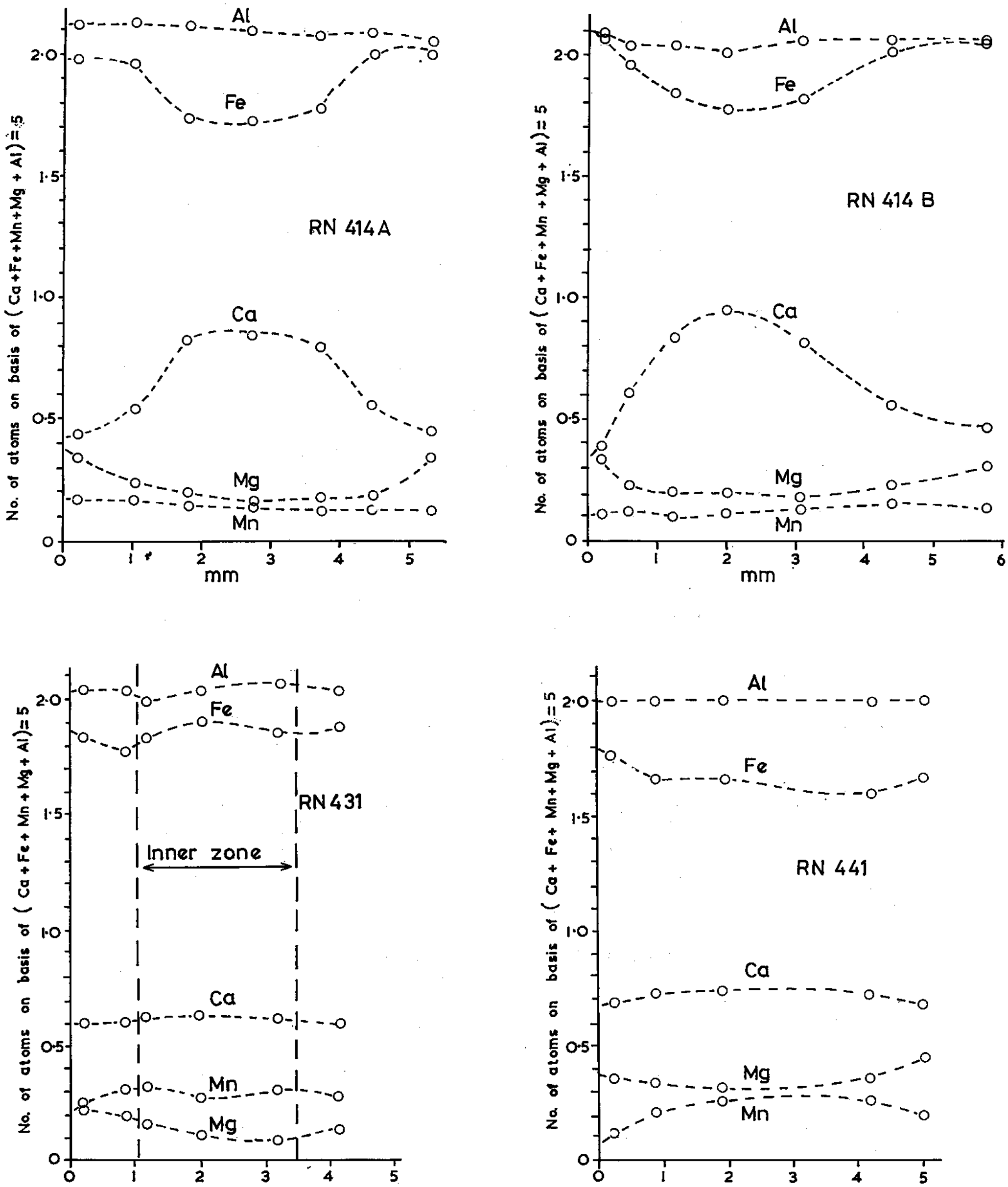


Fig. II - 16 Electron micro-probe analyses of garnets from the Peripheral Schieferhülle.

that atomic proportions of the elements on a basis of (Ca+Fe+Mn+Mg+Al) = 5, have been used. All the garnets contain between 50% and 70% of the almandine molecule. Only those from RN 414 exhibit a strong chemical zonation; note that Ca, Fe and Mg show considerable variation whereas Al and Mn (the element which usually exhibits the strongest zoning) are practically constant. The garnet in RN 441 displays a similar, but much weaker zonation, and has a lower proportion of almandine. Strangely the garnet with the most marked optical zoning (RN 431) shows practically no chemical zoning at all. The interpretation of these results will be left until the next chapter.

Epidote minerals occur in most of the rocks. Usually they form less than 5%, but in a few (e.g. RN 431, 432, 433) large crystals (up to 1 *cm.* long) of zoisite make up 40—50%. They overgrow folded trains of inclusions, and in RN 431 zoisite grains overgrow small garnets equivalent to the centre zones of the larger garnets. This suggests that the large zoisite porphyroblasts post-date most of the rotational deformation, and are broadly equivalent in time of formation to the outer zones of the garnets. Most of the epidote minerals present in the Peripheral Schieferhülle have fairly low birefringence, suggesting a composition near to clinozoisite. In most rocks they are zoned, up to three zones being present (e.g. C 111), with the outer zone generally having the lowest birefringence.

Kyanite was found only in one specimen from the Peripheral Schieferhülle (C 111). This is a quartz-mica schist (assemblage 11), about five metres thick, in a marble sequence in the lower Zwenbergergraben. The kyanite occurs as short stubby grains, about 0.5 *mm.* long, parallel to the foliation. There is no sign of retrogression and the assemblage appears to be in equilibrium. Although practically identical rocks have been found elsewhere, no kyanite was discovered. It may be that because C 111 comes from a thin layer of schist within thick marbles, the partial pressure of CO_2 depressed P_{H_2O} sufficiently to allow kyanite to form under conditions which would otherwise have given rise to hydrous alumino-silicates.

Chloritoid was recorded from three localities. In two of these assemblage (12) seems to be in equilibrium. The third occurrence (RN 427) has assemblage (13). This is the typical garnet-mica schist assemblage (6), with the addition of chloritoid. The chloritoid, however, occurs only as inclusions (0.1—0.5 *mm.* long) within the large garnets. Chloritoid may, therefore, not be a stable phase but a relict from an earlier assemblage. The first two examples of chloritoid occur close to the albite-oligoclase transition as does the kyanite-bearing rock, while the third specimen with chloritoid (RN 427) occurs in the oligoclase-bearing rocks.

Staurolite has been found at two localities (RN 414, RN 431) both very close to the contact with the Zentralgneis. RN 414 comes from the Riekengraben, and exhibits assemblage (14) while RN 431 is from the Mühldorfergraben and exhibits assemblage (15). RN 414 contains about 40% quartz and 20% of rather sericitised plagioclase. The staurolite occurs as a few small (0.1—0.3 *mm.* diam.) grains, exhibiting rather weak pleochroism (Z = Y = pale yellow; X = colourless). It appears to be undergoing alteration to biotite and chlorite. In places biotite also seems to be altered to white mica+chlorite+ore. The garnets in the rock are slightly retrogressed along fractures to biotite and chlorite. However, it is probable that assemblage (14), possibly with the exception of chlorite, represents an equilibrium assemblage, which has subsequently undergone partial retrogression.

The other staurolite-bearing rock, RN 431 is rather complex. Plagioclase which forms only 1 or 2%, is about An_{40}, and appears to have some kind of reaction relation with zoisite. This latter mineral forms 1 *cm.* square plates making up 40—50% of the rock, and is somewhat altered to calcite. Biotite forms large pale-brown flakes, while chlorite is present only as an accessory, and is probably retrogressive in origin. White mica forms about 20% of the rock, and in many places is replaced by a slightly pleochroic (very pale brown) mica-like mineral, with a higher refractive index and a low birefringence (1st order orange) (? margarite). The staurolite is very scarce, and forms larger (1—1.5 *mm.* long) grains than in RN 414; it is somewhat altered to white mica+pale biotite+quartz+ore. Again the pleochroism is rather weak. Tremolite also occurs as aggregates of small grains, rather altered to white mica. In both RN 414 and RN 431, the ore minerals are pyrrhotite and pyrite, together with a little ilmenite. In addition, some non-reflective opaque material occurs, especially within the garnets.

Pyrrhotite and pyrite are the principal ore minerals in the majority of the pelitic rocks, although in some, they are partially altered to haematite. Ilmenite also occurs and is commonly rimmed by sphene. Tourmaline is practically ubiquitous as an accessory mineral, as are apatite and sphene. Calcite occurs in all rocks, but often as a subsidiary intergranular precipitate.

Weissschiefer

The Weissschiefer is a distinctive lithology which, despite its fairly restricted occurrence, can be mapped separately. The rock contains between 70% and 90% white mica, almost certainly a phengite judging by its very pale green pleochroism. The other principal component is quartz. Tourmaline is a common accessory. On the western flank of the Kampleck there is a large body of Weissschiefer, which dies out abruptly in both directions along strike.

South of the Riekengraben, the only occurrence recorded is to the east of Kolbnitz railway station, where a small body is completely enclosed in calc-schist and marble.

The Weissschiefer seems to be restricted to zones of intense shearing and is almost certainly derived metasomatically from the other pelitic rocks and possibly impure quartzite units of the Peripheral Schieferhülle.

Amphibolites

Greenstones, consisting of pale, blue-green hornblende, chlorite, epidote, plagioclase and a little biotite, are fairly abundant within the Peripheral Schieferhülle sequence. The thicker units may be traced over distances of a kilometre or so, but do not appear to be continuous for much greater distances along strike. The thinner bodies are clearly discontinuous layers and lenses within the other lithologies. The bulk of the amphibolites are within, in contact with, or close to the mica-schists, although thin (5—10 *m.*) bands of greenstone occur within the calc-schist sequence. Some of these thin bands (e.g. in the lower Riekengraben) are interbanded with the calcareous material on the scale of a few centimetres.

Similar greenstones within the Peripheral Schieferhülle in the less metamorphosed areas of the Tauern (e.g. parts of the Glockner depression, Frasl, 1958) have been found to contain relict pyroxenes and brown hornblendes suggestive of an igneous origin. No such minerals have been found in the present area, but nevertheless, the greenstones are likely to be of a similar origin to those in the Central Tauern. A suggestion by Orville (1969) that certain amphibolites in Switzerland may represent metasomatised calcareous layers within a pelitic sequence is difficult to apply to the Tauern, as varying thicknesses of greenstone occur enclosed within both pelitic and calcareous lithologies.

Two further types of amphibolite are also found only in the upper Riekengraben. The first is a lens-shaped body of a coarse, massive amphibolite with plagioclase porphyroblasts up to 3 *mm.* diameter, and containing about 10% of biotite in flakes up to 1 *mm.* long. At its margins against the calcareous lithologies, there is apparently a metasomatic aureole, with a zone of tourmalinisation developed along the contact.

The second type is a garnet-bearing amphibolite, a few metres of which occur close to the locality described above. The garnets are less than 1 *mm.* diameter and fairly evenly distributed throughout what is otherwise a normal greenstone. In the upper Mühldorfergraben, garnets are also found in the amphibolites, but here they occur in pods up to 10 *cm.* long consisting almost entirely of red almandine garnet, with rims of clinozoisite and bright blue-green hornblende. The remainder are fine-grained, dark green rocks, the hornblendes of which are acicular in habit, either lying within a well-developed foliation or forming a prominent linear structure.

In marked contrast to the amphibolites of the Inner Schieferhülle, all amphibolites in the Peripheral Schieferhülle have an intense, penetrative, anisotropic fabric.

The mineral assemblages recorded in the basic rocks of the Peripheral Schieferhülle are:

(1) Hb+ep+olig±ab±bi±chl±qz±cc±ore+acc
(2) Hb+gt±ep+olig±bi±chl±qz+ore+acc
(3) Hb+ep+olig+mu+bi
(4) Act+hb+ep+olig+ore+sph+acc
(5) Act+olig+qz+bi+sph+acc
(6) Act+ep+ab+olig+bi+cc+sph+acc
(7) Act+cc+ore

These are thought, on textural grounds, to represent equilibrium assemblages, with the following exceptions: hornblende, biotite and garnet may be partly replaced by a green chlorite exhibiting anomalous interference colours. This appears to be a late alteration effect, however, and is possibly related to fracture-fillings which commonly contain a similar chlorite with quartz and carbonate. The relationship between plagioclase and epidote is not always clear. In some rocks, plagioclase appears to be replacing epidote, but in others, particularly those adjacent to or containing calcite-bearing lithologies, epidote and calcite appear to replace plagioclase. In the majority, however, both epidote and plagioclase appear to coexist in equilibrium.

Most of the greenstones depicted on the map exhibit assemblage (1). The names tremolite, actinolite and hornblende have been used with reference to the pleochroic scheme of the amphibole in thin-section. Tremolite has been reserved for colourless (non-pleochroic) clino-amphiboles, actinolite for those with Z = pale green, and hornblende for Z = blue-green. Brown hornblendes have not been recorded. The hornblendes are generally acicular prisms, 0.2—2 *mm.* long, with a strong preferred orientation. In one or two other samples (e.g. RN 440, C 210) there appears to be an earlier development of larger (1—4 *mm.* square), tabular, blue-green hornblendes which have subsequently partially recrystallised to the acicular habit. The latter are bent around and cross-cut the former.

In several samples (e.g. RN 445, C 412, C 249), many of the hornblende grains have distinct pale green cores with a higher birefringence than the rim. In RN 445, separate grains of colourless amphibole are intergrown with the green hornblende. A similar phenomenon, suggesting a compositional gap between actinolite and blue-green hornblende, has been described from Japan (Shido and Miyashiro, 1959).

Plagioclase generally occurs as poikilitic grains, with patchy inverse zoning and poorly developed twinning. Together with quartz, it may occupy fractures in hornblende and epidote. In most cases, the larger patches of plagioclase have partially recrystallised to an equant mosaic. Quartz tends to nucleate at triple junctions and grow along the grain boundaries producing a myrmekite-like effect. In the majority of samples, the plagioclase is oligoclase (An 22—30). In some samples, however, (e.g. RN 450, 447, 404, C 210), albite cores are present. The refractive index is much lower than Canada Balsam and determination by the other methods have shown them to be $An_{0\text{-}5}$in composition. There is a distinct break between the albite core and oligoclase rim and a change in optic orientation. The oligoclase appears to be replacing the albite, and twinned grains of the latter occur within an oligoclase 'matrix'. The degree of replacement varies; in RN 404, large porphyroblasts of albite occur with very thin rims of oligoclase, in RN 450, the relative amounts are 1 : 1, while in C 210, only a few small cores of albite remain. The albite-bearing samples occur closest to the Mölltal; i.e. structurally higher in the present succession.

Epidote occurs as small (0.1—0.3 *mm.*) grains, as larger aggregates of small grains or as larger (0.5—2 *mm.* diameter) single grains. The latter especially, are usually zoned, with more highly birefringent interiors and weakly or anomalously birefringent margins.

Chlorite is usually equal to or greater than biotite in abundance. In samples from the easterly part of the succession, the birefringence is around 0.004 to 0.007 +ve, but there is a tendency for this value to decrease towards the Mölltal, suggesting a decrease in Al or an increase in Fe, or both. Note that the later chlorites replacing hornblende, biotite or garnet, have a birefringence of 0.000 to 0.004 —ve and a higher refractive index, suggesting a lower Al and higher Fe content (Albee, 1962; Deer, Howie and Zussman, 1963).

Pyrrhotite, pyrite, magnetite and haematite occur as ore minerals. The pyrite usually occurs as cores to the pyrrhotite, which forms large (0.1—1 *mm.*) diameter grains. The haematite is usually found only as rims to the other ore minerals.

Modal analyses of typical greenstones exhibiting assemblage (1) are given in Table II—11.

TABLE II—11

Modal analyses of assemblage (1) Greenstones

Mineral	RN 404	RN 411	RN 425	RN 418	RN 440	RN 447	RN 450
Hornblende	20	58	30.5	66.5	47.5	45.5	45
Albite and Oligoclase	35	27	15	22.5	14	21	31.5
Epidote/Clinozoisite	6.5	3.5	17	2.7	16	15	3.5
Quartz	2.5	—	17.5	3.1	6	5.6	—
Biotite	30	10	1	—	0.5	—	—
Chlorite		—	1	—	11.5	2	15
Calcite	—	—	16	3.2	4	7	0.5
Ore	Acc	1	2				
Sphene	5.0	0.2	1.5	2	0.6	4	4.5
Accessories	0.4						

Garnet-bearing amphibolites (assemblage [2]) are rather restricted in occurrence and occur in the eastern (lower) half of the succession. The garnets are generally small (0.1—1 *mm.* diameter) and ragged, frequently exhibiting atoll forms. Although only a small proportion by volume, they are, with a few exceptions, evenly distributed through the rock, suggesting a homogeneous nucleation (contrast with the pelitic rocks).

The actinolite-bearing lithologies do not form major units, but frequently occur as pods and lenses in the calcareous lithologies. Actinolite may form 60—80% of the rock. Sphene is often abundant, occurring as large (0.5—1 *mm.*) grains. Biotite is pale yellow-brown in colour, suggesting a magnesium-rich variety. Two types of carbonate occasionally occur; one is normal calcite, while the other occurs as oval patches, with a higher refractive index than calcite (? dolomite) and honey-combed with a brown opaque material (siderite ?, goethite ?) and usually mantled by tremolite/actinolite or talc (e.g. RN 415, RN 440, C 200, C 110).

Serpentinites

In most of the valleys on the northeast flank of the Mölltal, serpentinite bodies are found. These can be traced out in the field, and are clearly in the form of small lenses, perhaps 100 *m.* in diameter and 10—20 *m.* thick. They are nearly always enclosed by calc-schists and develop a zone of talc, and occasionally actinolite, at the margins. One example where these are clearly seen is in the Mühldorfergraben (Fig. II—17). The serpentinite, about ten metres wide at the stream level, demonstrably wedges out upwards into the calc-schists. The contacts, which dip away in opposite directions, are zones of brecciation, with platy, sheared serpentinite and slivers of calc-schist intermixed in a talc-schist matrix. The serpentinite itself contains a considerable amount of light brown carbonate which often appears as distinct rhombs in hand-specimen. In thin-section, the serpentine is partly replaced by talc. This development of talc and carbonate is characteristic of the serpentinites in the Peripheral Schieferhülle and probably arises from their being completely enclosed by calcareous lithologies during the Alpine metamorphism.

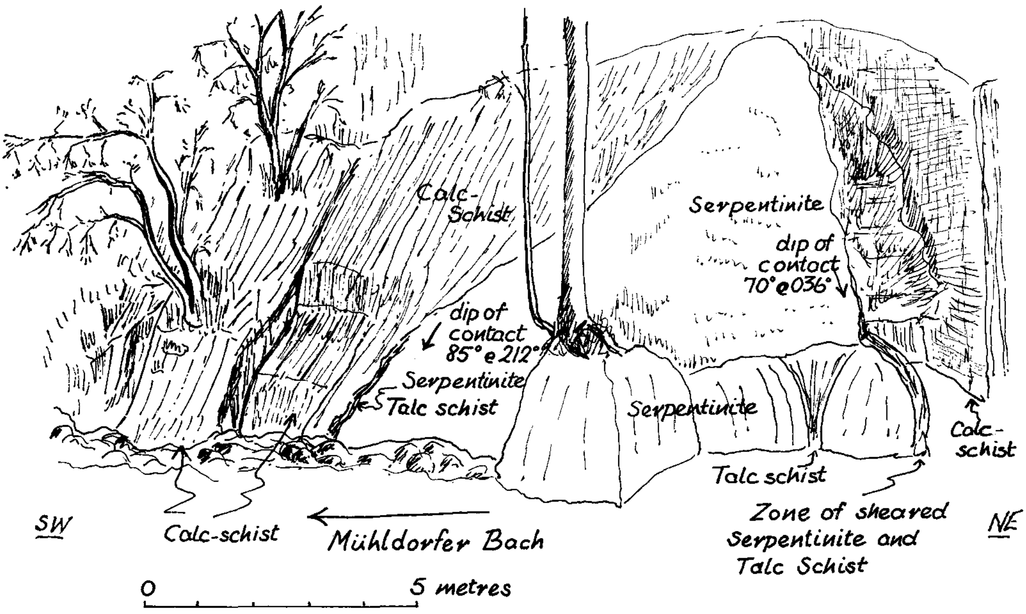


Fig. II - 17 Serpentinite body in the Mühldorfer Graben; from a field-sketch. Peripheral Schieferhülle.

The mineral assemblages recorded in the serpentinite bodies are:

(1) Antig+tc+dol+mt
(2) Liz+antig+tc+dol+mt
(3) Liz+antig+tc+dol+chl+tr+mu+mt

None of these are equilibrium assemblages, as in every case, the serpentine minerals are being replaced by talc+dolomite.

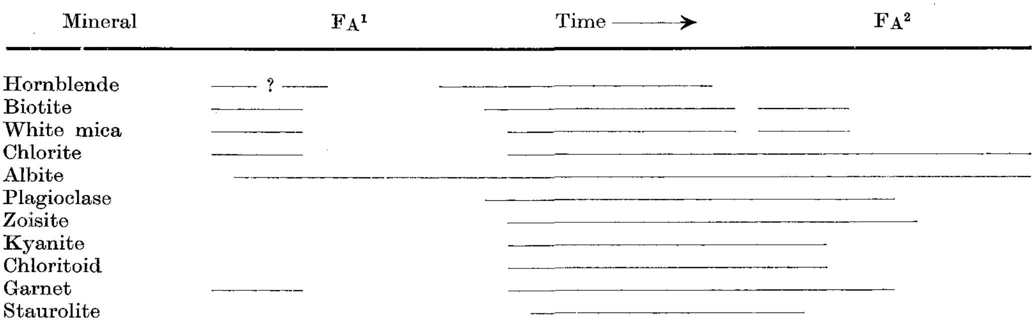
These latter minerals usually make up over 50% of the rock. In assemblage(3) (specimen C 201, upper Kaponiggraben) the tremolite has been largely replaced by talc+dolomite, as have the serpentine minerals. However, in part of the thin-section, the serpentine is being replaced by Mg-chlorite+white mica+carbonate. The lizardite occurs as parallel ribbons, with the length-slow fibres perpendicular to the margins of the ribbons. It appears to be altering to an antigorite mesh. Most of the serpentinites, however, have assemblage (1).

Summary

A detailed discussion of the metamorphism will be presented in Chapter III. The mineral assemblages suggest, however, an Alpine metamorphism ranging from greenschist to amphibolite facies, followed by a slight retrogression. The table below relates the phases of mineral growth to the main phases of movement discussed in Chapter IV.

TABLE II—12

Growth of Minerals during Alpine Metamorphism, in Peripheral Schieferhülle, with respect to phases of deformation



Fissure Mineralisation

Along joint planes and fissures, a late stage mineralisation has occurred. The minerals have usually grown from the fissure walls without being deformed. The fissures range in width from a few millimetres to a metre. The wider examples usually contain a fault breccia, with the minerals growing around the rock fragments.

The following assemblages were recorded:
Zentralgneis and Inner Schieferhülle

(1) Quartz
(2) Quartz+chlorite+epidote±specular haematite±adularia±zeolites
(3) Quartz+dolomite
(4) Calcite+dolomite
(5) Calcite+specular haematite+clouded plagioclase
(6) Albite+chlorite
(7) Quartz+white mica
(8) Quartz+biotite
(9) Quartz+pyrite-goethite
(10) Quartz+tourmaline
(11) Calcite+tourmaline

Peripheral Schieferhülle

(1) Calcite
(2) Quartz
(3) Quartz+chlorite±tourmaline±pyrite

In the older rocks, quartz is almost ubiquitous as a fissure mineral; the quartz-deficient assemblages have been recorded in very few localities. Assemblage (4) occurs along a fault which cuts the main ridge to the south of the Reisseck. Dolomite surrounds the fragments of fault breccia while calcite occurs as small rhombohedral grains in cavities. The albite-chlorite assemblages generally occur in the amphibolites. The chlorite is a green pleochoric, anomalously birefringent variety (? high Fe, high Si) and occurs as a fine-grained, infilling material, as a thin, platy covering on fissure walls, or most commonly, as small (1 *mm.*) sheaves of radiating flakes. Specular haematite is the commonest ore mineral found. The quartz-muscovite assemblage is usually confined to the granitic rocks; the fissure walls are often bleached, with sericitisation of the plagioclase and the replacement of the biotite by white mica. Assemblages (10) and (11) occur only in the Randgneis and outer parts of the Basement Complex.

The assemblages found in the Peripheral Schieferhülle are considerably different. Quartz and calcite are the most abundant filling materials, but chlorite, tourmaline and pyrite are also frequently found. The restriction of tourmaline to the Peripheral Schieferhülle and margins, and specular haematite to the older rocks is particularly noteworthy.

This mineralisation indicates the existence of hydrothermal fluids at a late stage in the history of the area. The difference in the assemblage from place to place, and particularly the differences between older and younger rocks suggests that much of the material was locally derived. The nature of the mineralisation is very similar to that described from the Western Alps (PARKER, 1960; FAGNANI, 1960; GRIGORIEV, 1960) and other areas of Alpine deformation. This fissure mineralisation is widespread in the Tauern, and gold-bearing quartz veins have been recorded in the Sonnblick group, where they were once worked economically, and in the present area (DAMM and SIMON, 1966). At other localities iron ores and magnesite have been worked in the past.

III. Metamorphism

by R. A. CLIFF, R. J. NORRIS and E. R. OXBURGH

Introduction

This chapter elaborates on those aspects of the petrography and mineralogy that provide information on the variation of metamorphism in time and space. Some of the reactions inferred from the petrographic evidence have been the subject of hydrothermal experiments. Using the results of such experiments, limiting values for pressure and temperature are estimated and the possible role of additional variables is examined. In a later chapter the PT conditions deduced in this way are related to the geochronological results in order to propose a possible thermal history of the southeast Tauern.

In considering the metamorphism of the map-area the possibility that there were several independent phases of metamorphism, possibly under differing conditions, must be taken into account. An Alpine metamorphic event is well established both in the map-area and in the rest of the Tauernfenster, since the Mesozoic lithologies of the Peripheral Schieferhülle were affected. In much of the Tauernfenster this metamorphism produced greenschist facies assemblages, but lower amphibolite facies assemblages also occur, the most extensive occurrence being in the southeast. Relatively little attention has in the past been given to the possibility that there was more than one independent phase of metamorphism within the Peripheral Schieferhülle of the Tauernfenster. However such features as the sporadic occurrence of glaucophane indicate that a closer investigation is called for; in the Western Alps, outside the zone of highest grade metamorphism, several phases have been recognized (BEARTH, 1962; VAN DER PLAS, 1959). In the Inner Schieferhülle and parts of the Zentralgneis, Variscan metamorphic assemblages may be partially preserved; the possibility that a significant part of the amphibolites may comprise Variscan minerals is strengthened by the structural analysis, and the small number of old K/Ar dates on hornblendes.

The discussion deals first with the main lithological groups that were sensitive to metamorphism; subsequently the variation in composition of plagioclase, which occurs in virtually all lithologies, is discussed.

Amphibolites

In the Peripheral Schieferhülle the mineral assemblage in the amphibolites to within a short distance of the lower contact is albite-epidote-hornblende. In the lowest part of the Peripheral Schieferhülle and in the Inner Schieferhülle the assemblage is plagioclase-epidote-hornblende $\pm$ almandine.

The amphibole of both the Peripheral and Inner Schieferhülle is typically a blue-green hornblende. In the Peripheral Schieferhülle actinolite may also be present either as distinct acicular crystals or as more highly birefringent cores within rims of blue-green hornblende. In the Inner Schieferhülle the blue-green hornblende tends to be more deeply coloured than in the Peripheral Schieferhülle; in a few cases grains have inner zones packed with minute opaque inclusions surrounded by somewhat paler inclusion-free rims, and in one case a patchy core of brown hornblende is surrounded by actinolite. No systematic variation of the hornblende optical properties with grade was noted; this is consistent with general experience that amphibole composition in the lower amphibolite facies is controlled mainly by whole rock composition (ENGEL and ENGEL, 1962; SHIDO and MIYASHIRO, 1959).

Chlorite is present in most of the Peripheral Schieferhülle amphibolites but is absent from those in the Inner Schieferhülle. Within the Peripheral Schieferhülle the birefringence increases toward the contact with Basement Complex, indicating increased Mg and Al content. The absence of chlorite in the Inner Schieferhülle is probably not associated with the upper stability limit of chlorite. ALBEE (1965) has shown that magnesian chlorite can persist well into the amphibolite facies. More likely a reaction of the type

(1) amphibole + water + $CO_2 \rightleftharpoons$ chlorite + calcite + quartz

was driven to the right by high P_{CO_2} in the Peripheral Schieferhülle, but not in the carbonate-poor Inner Schieferhülle.

Biotite occurs in all the amphibolites; its pleochroism changes from yellow-brown in the Peripheral Schieferhülle, to green or green-brown in the Inner Schieferhülle. This suggests a higher Fe^{3+} in the latter, reflecting higher oxygen fugacity in the Inner Schieferhülle.

Epidote shows a similar effect with greater amounts of pistacite in solid solution in the Inner Schieferhülle epidotes, in contrast to the clinozoisite of the Peripheral Schieferhülle. However the picture is complicated by ubiquitous zoning from pistacitic cores to clinozoisite rims. The observations of KEITH et al (1968) in the Salton Sea geothermal system suggest that this may indicate a change in P_{O_2} with time rather than a temperature effect as suggested by MIYASHIRO and SEKI (1958). LAMBERT (1963) has attributed differences in epidote composition between Lewisian inliers and enclosing Moine schist to variation in P_{O_2}. The opaque mineral assemblages add further supporting evidence: in the Inner Schieferhülle both magnetite and haematite occur and rutile is quite common, whereas in the Peripheral Schieferhülle pyrrhotite, pyrite and ilmenite predominate, suggesting a higher P_{O_2} in the basement.

Serpentinites

Within the serpentinites of the Peripheral Schieferhülle, serpentine is partially replaced by talc + dolomite. JOHANNES (1967, 1969) has investigated the following reactions experimentally:

(2) 2 forsterite + 2 H_2O + CO_2 = 1 serpentine + 1 magnesite
(3) 4 forsterite + 1 H_2O + 5 CO_2 = 1 talc + 5 magnesite
(4) 2 serpentine + 3 CO_2 = 1 talc + 3 magnesite + 3 H_2O
(5) 1 talc + 3 CO_2 = 4 quartz + 3 magnesite + H_2O

He demonstrates that serpentine is stable only at low P_{CO_2} and that at CO_2 contents of the fluid phase exceeding 2—6 mol%, reaction (4) will occur. Fig. III—1 is taken from JOHANNES (1967) and shows the relationship between the reaction curves, tem-

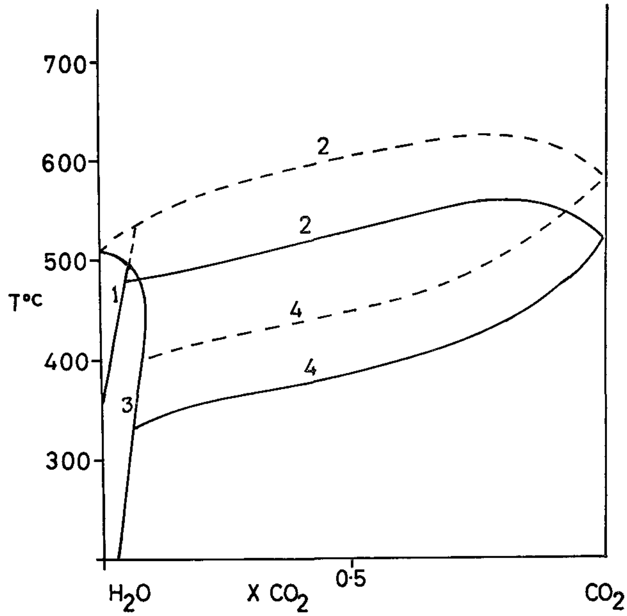


Fig. III - 1 Experimentally determined equilibria after JOHANNES (1967, 1969)

(1) 2 forsterite $+ 2H_2O + CO_2 \rightleftarrows$ 1 serpentine $+$ 1 magnesite

(2) 4 forsterite $+ H_2O + 5CO_2 \rightleftarrows$ 1 talc $+$ 5 magnesite

(3) 2 serpentine $+ 3CO_2 \rightleftarrows$ 1 talc $+$ magnesite $+ 3H_2O$

(4) 1 talc $+ 3CO_2 \rightleftarrows$ 4 quartz $+$ 3 magnesite $+ H_2O$
Solid curves: P_f = 4kb; dashed curves: P_f = 2kb.

perature and composition of the vapour phase at $P_f = 2$ kb. In his 1969 paper, he presents experimentally determined equilibrium curves for reactions (3) and (5) at a number of different pressures. The dashed lines in the diagram represent these two curves at $P_f = 4$ kb. From these relationships, some broad temperature limits can be placed on the reaction, for although the reaction is incomplete; it is mainly governed by P_{CO_2} rather than by temperature. With a confining pressure of 4 kb and X_{CO_2} (partial molar volume) from 0.05 to 0.5, the temperature of reaction probably lies between 430° and 500° C. However, it is not known when either the formation of the serpentinite or its subsequent metasomatism occurred. This range of temperatures is in accordance with the predominantly upper greenschist facies assemblages in the surrounding rocks. The relationships in the serpentinites of the Inner Schieferhülle would suggest a low and fairly constant P_{CO_2} as serpentine+talc+carbonate occur without any obvious replacement textures (cf. TURNER, 1968).

The alteration of serpentinites to talc+carbonate is a fairly common phenomenon (e.g. NALDRETT, 1966), and is particularly widespread in the Peripheral Schieferhülle of the Tauern (ANGEL, 1939; JOHANNES, 1967).

Calc-silicates and Marbles

The stable assemblages in the calcareous lithologies of the Peripheral Schieferhülle are:

cc+qz
cc+qz+tr
cc+dol+tr

The reactions involving these assemblages have been studied fairly extensively but their interpretation in terms of temperature and total pressure is complicated by the role of CO_2 as a third variable. From the data available, the general form of the equilibrium curves over a range of pressures, temperatures and fluid phase compositions for reactions involving siliceous carbonate lithologies is shown in Fig III—2 (based on WINKLER, 1967; METZ, 1967; and JOHANNES, 1969). The reaction calcite+quartz = wollastonite+CO_2 is very sensitive to P_{CO_2} (cf. TROMMSDORF, 1968), but even at a P_{CO_2} as low as 500 b, the assemblage calcite+quartz would be stable up to 600° C. At 4 kb the reaction dolomite+quartz+H_2O = talc+CO_2 would take place at temperatures between 450° and 500° C for X_{CO_2} between 0.1 and 0.9. It is probable that the equivalent reaction producing tremolite would occur at similar temperatures (WINKLER, 1967).

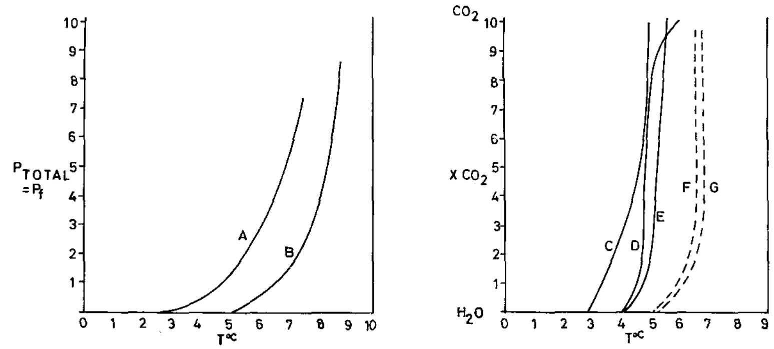


Fig. III - 2 The system MgO-CaO-SiO$_2$-H$_2$O-CO$_2$ (after WINKLER, 1967; METZ, 1967; JOHANNES, 1967)
A: tremolite+calcite+quartz $\rightleftarrows$ diopside+H_2O; for $P_{H_2O} = P_{CO_2}$
B: calcite+quartz $\rightleftarrows$ wollastonite+CO_2; for $P_{CO_2} = P_{total}$
C: dolomite+quartz+H_2O $\rightleftarrows$ talc+calcite+CO_2; for Pf = 2 kb.
D: magnesite+quartz+H_2O $\rightleftarrows$ talc+CO_2; for Pf = 4kb.
E: tremolite+calcite+quartz $\rightleftarrows$ diopside+CO_2+H_2O; Pf = 1kb.
F, G: estimated positions of curve E for Pf = 4kb and 5 kb respectively.

The reaction tremolite+calcite+quartz = diopside+H_2O+CO_2 has been used by WINKLER (1967) to define the base of the amphibolite facies in carbonate lithologies. At high fluid pressure and high P_{CO_2}, however, the stability field of tremolite +quartz+calcite may be extended. For instance, TROMMSDORFF (1966) has described the progressive metamorphism of siliceous carbonate rocks from the Swiss Alps, and finds that the appearance of the assemblage tremolite+calcite is coincident with the appearance of staurolite and kyanite in pelitic rocks, whereas the reaction to produce diopside does not take place until almost into the sillimanite zone. He suggests that the diopside-calcite isograd in these rocks lies between 570° and 620° C, which is well above the lower boundary of the stability field of staurolite+quartz. If $P_{H_2O} = P_{CO_2}$, the temperature of the tremolite/diopside reaction would be about 650° C at 5 kb. Thus high P_{CO_2} would explain the sequence of facies recorded by TROMMSDORFF (note that if this also obtained in the pelitic rocks, the concomitant lowering of P_{H_2O} would also lower the temperature at which staurolite became stable [see later discussion]). In the Peripheral Schieferhülle of the present area, tremolite+calcite assemblages appear before staurolite, but do not give way to diopside-bearing assemblages, suggesting a P_{CO_2} greater than about 500 b total pressure (Fig. III—2). However, the curve has not been determined over a wide range of values of P_{CO_2} and total pressure, and from the results at 1000 b, it would seem that the reaction is not very sensitive to X_{CO_2} between 0.1 and 0.9 mole %. If this holds true at higher pressures, values of P_{CO_2} as low as 250 b may be sufficient.

Diopside occurs in four samples from the Inner Schieferhülle. In three of them reaction to produce actinolite from pyroxene and epidote from calcic plagioclase is observed; this may be accounted for by the reaction

(6) 5 diopside+9 anorthite+4 water = 6 clinozoisite+tremolite+2 quartz.

But an older brown hornblende is rimmed with actinolite and may be involved in the reaction.

In the fourth sample the diopside is also unstable, but here the reaction products are tremolite+quartz+calcite, suggesting that the reaction was

(7) 5 diopside+water+3 CO_2 = tremolite+2 quartz+3 calcite.

In this rock the texture also suggests that two generations of amphibole are present, one of which may have coexisted with the diopside.

The prominent reaction texture, coupled with the assemblages in the associated rocks indicate that pyroxene was unstable in the main phase of Alpine metamorphism. There is no direct evidence to show whether the pyroxene crystallised in an early Alpine metamorphism or still earlier in the Variscan. However the arguments presented in favour of partially preserved Variscan assemblages in the amphibolites suggest that the latter is more probable.

Pelitic Rocks

The pelitic lithologies of the Schieferhülle show a sequence of mineral assemblages corresponding broadly to the Barrovian zones. The southernmost part of the Peripheral

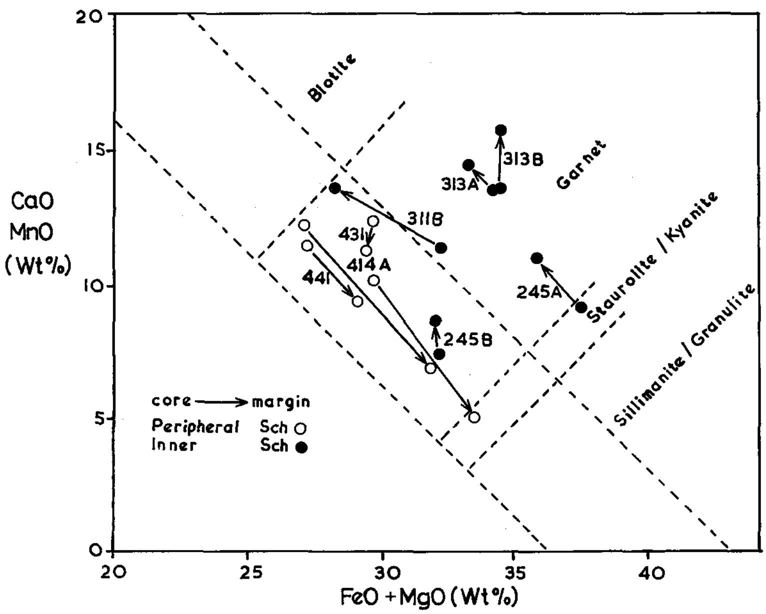


Fig. III - 3 Plot of garnet compositions from Schieferhülle by method of STURT (1962). Dashed lines indicate STURT's garnet fields.

Schieferhülle contains chlorite-white mica-quartz-albite schists; approaching the contact with the Basement Complex these are succeeded by two-mica schists, garnet-mica schists and at a few isolated localities kyanite- and staurolite-bearing assemblages occur. Chloritoid is also found in a few samples.

Garnetiferous schists are among the most abundant pelitic rocks, and chemical study of the garnets in selected samples demonstrates variations in composition both within single grains and from one sample to another. In all cases the garnet is predominantly almandine.

MIYASHIRO (1953) suggested that with increasing metamorphic grade, the Mn^{+2} content is replaced by Fe^{+2} and Mg^{+2}, and suggested that these substitutions may be controlled by the decreasing molar volume of the garnets with increasing grade, ions with smaller ionic radii being preferred at higher grades.

LAMBERT (1959) found that the CaO content of garnets in pelitic schists also shows a systematic decrease with increasing grade. Similar results have been reported by other workers (e.g. HIETANEN, 1969). STURT (1962) has suggested that there is a regular pattern of substitution of (Fe+Mg) for (Ca+Mn) with increasing grade of metamorphism (ionic radii 0.83, 0.78, 1.06, 0.91 Å respectively) and that Fe/Mg and Ca/Mn ratios are largely determined by bulk composition whereas the Fe+Mg/Ca+Mn ratio depends on grade. He plots some 60 published garnet analyses from pelitic schists at various grades of metamorphism on a graph of (FeO+MgO) against (CaO+MnO) and demonstrates that they form a linear array in order of increasing grade. The microprobe analyses of garnets quoted in chapter II have been recalculated as weight-percent oxides and are set out in table III—1. Only the centre and edge values of the zoned garnets have been taken.

TABLE III—1

Garnet compositions measured with the microprobe; weight % oxides

Inner Schieferhülle

	RN 311 B (migmatite, ass. 2)		RN 313 A (migmatite, ass. 6)		RN 313 B		RN 245 A (migmatite, ass. 2)		RN 245 B (2nd generation gt.)	
	centre	edge	centre	edge	centre	edge	centre	edge	centre	edge
Al_2O_3	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Mg O	2.6	1.7	2.1	2.7	4.9	4.9	3.3	3.3	2.3	2.3
FeO	29.6	26.5	32.0	30.6	29.4	29.4	34.4	32.8	29.6	29.6
CaO	3.4	2.0	1.4	1.1	3.7	2.4	1.3	1.1	6.7	7.6
MnO	8.0	11.7	12.3	13.7	9.9	14.2	6.7	9.9	1.0	0.7

Peripheral Schieferhülle

	RN 414 A (Gt-mica schist, ass. 14)		RN 414 B		RN 431 (Gt-mica schist, ass. 15)		RN 441 (Gt-mica schist, ass. 6)	
	centre	edge	centre	edge	centre	edge	centre	edge
Al_2O_3	22.0	22.0	22.2	22.1	21.8	21.8	21.3	21.5
Mg O	1.2	2.9	1.7	2.8	0.9	2.0	2.65	3.0
FeO	26.0	29.0	28.0	30.6	28.75	27.4	24.9	26.2
CaO	9.8	5.1	8.1	3.2	7.6	7.1	8.5	8.1
MnO	2.5	1.8	2.0	1.8	4.7	4.1	3.0	1.4

ass. = assemblage quoted in chapter II. n.d. = not determined

Figure III—3 is a similar diagram to that of STURT, with the analyses of table III—1 plotted. The two dashed lines represent the boundaries of the analyses plotted by STURT, with the approximate metamorphic grades indicated. The garnets from the Peripheral Schieferhülle fall largely within the garnet grade, with the centres plotting close to the biotite grade field and the edges towards the higher grades. The margin of one garnet (RN 414) actually falls within the staurolite grade field; this rock happens to be one of the few samples from the area which contain staurolite.

The garnets from the Inner Schieferhülle are completely different. Only the small, second generation garnet (RN 245 B) plots close to those from the Peripheral Schieferhülle. The large first generation garnets plot above the field defined by STURT, and their margins are at a lower apparent grade than the centres. In every case, the centres have a higher content of FeO+MgO, for the same CaO+MnO content, than those of the Peripheral Schieferhülle. The mean value of (FeO+MgO+CaO+MnO) for the latter is 39.7 whereas for the Older Schieferhülle garnets it is 44.5. The microprobe does not differentiate between ferrous and ferric iron, and one possible explanation is that the garnets from the Inner Schieferhülle contain an appreciable amount of ferric iron substituting for Al; unfortunately, Al analyses were not made for these garnets.

The most commonly recorded chemical zoning in garnets is a systematic decrease in Mn content towards the margin. HOLLISTER (1966) has suggested that this does not necessarily imply an increase in grade of metamorphism, but may also be explained by a progressive local impoverishment in Mn during the growth of the garnet. He demonstrates that concentration profiles of Mn across garnets correspond fairly closely to theoretical profiles calculated assuming a Rayleigh fractionation model. Mn is particularly sensitive to this type of process because it is concentrated in the garnet. HENLEY (1968) has demonstrated two types of zoning pattern in garnets from the Sulitjelma metamorphic complex. In both, MnO decreases outwards, but in one, CaO is constant with FeO increasing outwards, while in the other FeO is constant with CaO increasing outwards. HENLEY suggests that the core composition is determined largely by the Ca and Mn contents of the rock, while the margin composition is controlled mainly by metamorphic grade for Fe, Mg and Mn, but by host rock composition for Ca. DRAKE (1968) has measured the zonation profiles of 67 garnets, together with the compositions of coexisting minerals, from pelitic schists in the garnet and staurolite zones of regional metamorphism. He concludes that in the garnet zone, garnet grows according to a depletion model with chlorite acting as an Mn reservoir, but that in the staurolite zone, the depletion model is no longer applicable, and the Mn content increases with grade. BROWN (1968) has also studied this problem, and concluded that two controls on garnet composition operate; one is the depletion of Mn from the environment during garnet growth while the other is the increase in the amount of almandine stable in the garnet with increased grade. However, he suggests that the second is the more important at grades above the biotite isograd. Thus there is no simple interpretation of garnet zoning. The suggestion that Mn increases at high grades does not seem to be borne out by published analyses of high grade garnets (cf. STURT, 1962). The most reasonable conclusion seems to be that the initial Ca and Mn ratios are determined by the rock composition, and their total content by metamorphic grade, that the Mn content is a function both of its depletion in the immediate environment and of metamorphic grade while Fe and Mg increase with increasing grade.

The garnets in the present study do not have zonation profiles like those recorded in the literature. In the Peripheral Schieferhülle, Mn shows little or no variation. Calcium zonation does occur, but as the garnets occur in rocks containing epidote or zoisite, plagioclase, calcite and other calcium-bearing minerals, it is unlikely that this zonation is due to a depletion process (cf. BUTLER, 1967): it is concluded that the zonation profiles of the garnets in the Peripheral Schieferhülle reflect increasing grade of metamorphism.

The garnets from the Inner Schieferhülle are more difficult to interpret. They have very high Mn contents, and Mn increases towards the margin while Fe and Ca decrease. One explanation is that the garnets grew during a decrease in grade. However, three of the four specimens show continuous Mn zonation, and it seems unlikely that garnets of this size would grow entirely during falling temperature without any growth having occurred during the prograde period. Another possibility is that during a later phase of metamorphism at a lower grade, further growth occurred together with a partial homogenisation of Mn, so that it is now largely distributed according to a diffusion gradient. The difference in Ca and Mn contents between the one small second generation garnet investigated and the larger ones is considerable, and would suggest a much lower availability of Mn during the growth of the former. A third possibility is the release of Mn into the system during metamorphism, through the breakdown of some other Mn-bearing phase, or possibly by its introduction from outside.

The rocks in which these Inner Schieferhülle garnets occur have a complex metamorphic history. The larger garnets have probably experienced three phases of metamorphism; the first during which they originated, the second associated with the introduction of granitic material, and finally the main Alpine metamorphism. Zonation profiles established during the first or second phases may have been considerably modified during the third by diffusion although further growth may not have occurred. Clearly more data are required for the solution of this problem, but this must be obtained with the microprobe, as bulk chemical analysis of garnet separated from these rocks would furnish meaningless results.

Chloritoid was recorded in three rocks and in two it is apparently in equilibrium; in the third it is enclosed in a garnet. Although rare, chloritoid occurrence is important in view of the narrow stability range demonstrated by experimental studies (HOSCHEK, 1967; 1969; GANGULY and NEWTON, 1968; GANGULY, 1968). The major uncertainty in the application of the experimental results is the sensitivity of the reactions involved to P_{O_2}, although this problem may be alleviated to some extent by the presence of graphite in the assemblages concerned. The reaction curve suggested by HOSCHEK for

$$\text{(8) chlorite} + \text{hydrous Al-silicate} \rightleftharpoons \text{chloritoid} + \text{quartz} + \text{vapour}$$

is close to the pyrophyllite breakdown curve of KERRICK (1968). This is in agreement with the field occurrence of kyanite and chloritoid in adjacent localities.

Staurolite occurs at two localities in the Peripheral Schieferhülle, although probably not in equilibrium assemblages. It does appear, however, that at some stage during the Alpine metamorphism quartz and staurolite were in equilibrium together. This probably represents the highest temperature mineral assemblage recognized within the Peripheral Schieferhülle. It is, however, difficult to derive absolute temperatures from it because of the uncertain effects of P_{O_2} (GANGULY, 1968; HOSCHEK, 1969) and uncertainties of compositional control in the staurolite (RICHARDSON et al., 1968).

Kyanite occurs in five samples, four from the Inner Schieferhülle and one from the Peripheral Schieferhülle. It has also been recorded at many other localities in the Tauernfenster and is the only aluminium silicate polymorph found.

In one of the samples from the Inner Schieferhülle the kyanite has been extensively replaced by plagioclase. An original assemblage comprising quartz, biotite, kyanite and, possibly, chlorite has been replaced by a new assemblage in which kyanite is embayed and locally completely replaced by oligoclase with trails of white mica running through it. It is possible that the chlorite first crystallised with the plagioclase and white mica at the expense of biotite. However it is difficult to envisage conditions under which this reaction would occur at the same time as calcic plagioclase was forming at the expense of kyanite. It seems more probable that the reaction involved only the reaction of kyanite with metasomatically introduced sodium and calcium.

In this case all the evidence is consistent with Alpine metamorphism within the stability field of kyanite. The wide range of P T conditions determined experimentally for the aluminium silicates in early experiments has been reduced in the most recent work and it now seems possible to place reasonable limits on the location of the triple point (ALTHAUS, 1967 a; MATSUSHIMA et al., 1967; FYFE, 1967; RICHARDSON et al., 1968; HOLDAWAY, 1968; RICHARDSON et al., 1969; ANDERSON and KLEPPA, 1969); ROSENFELD (1969) has presented indirect evidence for the pressure of the triple point. The role of minor elements such as Fe^{3+} may account for much of the remaining scatter of results; ALTHAUS' (1969) demonstration of differing results using his own and RICHARDSON's starting materials is of great significance in this context.

The evidence available indicates the triple point lies at $600° \pm 30°$ C and 6.0 ± 0.5 kb (Fig. III—4).

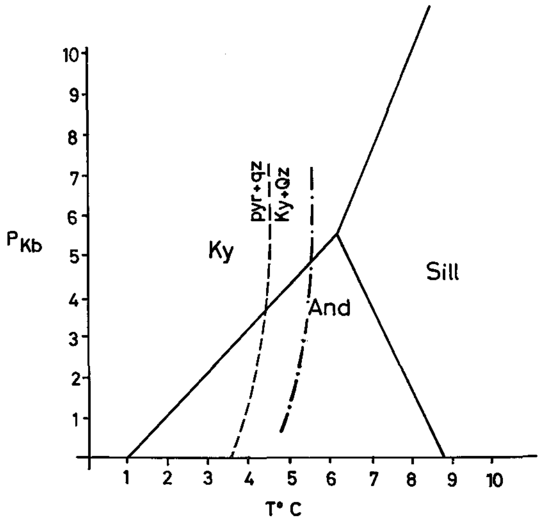


Fig. III - 4 Stability fields of the Al_2SO_5 polymorphs after RICHARDSON et al. (1968, 1969) and the pyrophyllite break-down curve after KERRICK (1968) — dashed line, and ALTHAUS (1966, 1967) dot-dashed line. Temperature in hundreds of degrees C.

The lower temperature limit of anhydrous aluminium silicates is, under hydrous conditions, determined by the upper limit of pyrophyllite. This has been investigated by KERRICK (1968) and ALTHAUS (1966, 1967 b); their results differ by about 100° C (Fig. III—4). ALTHAUS' results (Fig. III—4) suggest that for $P_{tot} = P_{H_2O}$ the kyanite field is completely masked at 4.5 kb total pressure — the value suggested below for Peripheral Schieferhülle metamorphism. For the moment accepting KERRICK's results, we conclude that metamorphism of the Peripheral Schieferhülle occurred within temperature limits of 430° and 550° C. Using ALTHAUS' results would clearly raise the lower limit. If, however, P_{H_2O} was less than P_{tot}, the pyrophyllite field would have been reduced and the lower temperature limit lowered. There is some indication that the latter may have been the case as the kyanite-bearing schist from the Peripheral Schieferhülle is intercalated with marbles, and a high local P_{CO_2} may have lowered the partial pressure of water.

If the P_{tot} for the metamorphism of the Inner Schieferhülle was 6.5 kb, a maximum temperature of 670° C is indicated.

Quartzofeldspathic Lithologies

General Statement

Granitic rocks are not particularly sensitive to regional metamorphism in the sense that few new major phases form. However, metamorphism does produce changes in texture and in the composition of some minerals. In the Zentralgneis the textural changes were extensive, and reactions involving the feldspars were widespread.

The granites that formed the Zentralgneis are probably all pre-Alpine (Chapter V), and some of them may have already been deformed and metamorphosed to granite-gneisses during Variscan metamorphism. In general, however, the relation of the deformational fabric to mesoscopic structures (Chapter IV) indicates that Alpine recrystallisation was responsible for the present textures in the Zentralgneis.

The nature of the recrystallisation of the Zentralgneis varies greatly with the intensity of deformation, and the latter, in part at least, seems to be controlled by the modal composition of the rock being deformed. For this reason certain distinctive textures are restricted to specific lithologies.

Accompanying the textural modifications the plagioclase throughout the Zentralgneis acquired a near uniform composition. In most of the K-feldspar bearing rocks the formation of myrmekite was also common. In some rocks phengitic micas crystallised at the expense of K-feldspar.

Development of metamorphic textures

Discussions of metamorphic textural evolution and its analysis in terms of a general tendency to approach a thermodynamic equilibrium, characterised by minimum interfacial free energy, have been presented by several authors (Voll, 1960; Rast, 1965; McLean, 1965; Kretz, 1966 and Vernon, 1968). These workers were, however, concerned mainly with metasediments in which metamorphism was accompanied by a general increase in grain-size. The Zentralgneis presents a rather different problem since the original textures seem to have been close to equilibrium and the net result of metamorphism and deformation has been a general decrease in grain size. Other factors being equal, the smaller the grain size the higher the interfacial free energy.

The complex nature of the quartz-feldspar textural relations has been recognised throughout the Tauernfenster and many authors have described the occurrence of several textural varities of each mineral. Sander (1912) and Karl (1959) have employed a simple two-fold division into an older generation, representing relict grains from the original granitic texture, and younger grains crystallised during Alpine metamorphism. Exner (1949, 1957, 1964) and his students have adopted a three-fold classification, in which porphyroblastic Alpine feldspars are also recognised.

The striking variations in intensity of foliation in the Zentralgneis were described in Chapter II. Almost unfoliated facies are found in the tonalite, leucogranite, granodiorite, and fine porphyritic granite.

In the tonalite, plagioclase is nearly always partly recrystallised to a fine-grained polyhedral mosaic but a few examples were found in which the original large, subhedral tablets were almost completely preserved. In many cases plagioclase recrystallisation is complete even though no foliation is apparent in hand specimen. As the intensity of foliation increases the aggregates of plagioclase which in the less deformed rocks retain the general form of the original subhedral grains, become flattened in the plane of foliation. In addition, the characteristic clots of biotite are destroyed and the mica becomes evenly distributed through the fabric, oriented parallel to the foliation plane. Grain size is reduced as the foliation becomes more intense and in the extreme, along shear zones, the tonalite can be traced continuously into a feldspathic biotite schist.

The sequence of changes in the granodiorite is very similar to that in the tonalite, except that the K-feldspar remains as megacrysts even when the ground-mass has become strongly foliated, thus giving rise to an augen-gneiss. Many of the augen have retained euhedral outlines, but in other cases they have been deformed and their ends recrystallised and drawn out as 'tails' in the direction of a mineral lineation. The development of fine-grained augen-gneiss from the fine porphyritic granite follows the same course. In both cases there is a tendency for the quartz and plagioclase to occur in monomineralic streaks; these are most pronounced in the coarser augen-gneiss derived from the granodiorite and are probably produced by deformation from large original grains.

In the coarse leucogranite the response to progressive deformation was different. As in the tonalite, the mineral most susceptible to recrystallisation was plagioclase; in the more intensely foliated facies, however, the quartz has remained coarse-grained, whereas the K-feldspar has recrystallised to a fine-grained polyhedral mosaic, similar to the plagioclase. This different response may reflect the different initial texture and mineral composition. The K-feldspar in the granodiorite occurs as large grains set in a relatively mica-rich matrix and deformation should have readily been accommodated by intergranular slip in the matrix. In the leucogranite on the other hand, the K-feldspar was originally no coarser than the quartz and plagioclase and, in addition, the mica content is too low to have been very effective in facilitating deformation.

In addition to these lithologies which occur both in highly deformed and relatively undeformed facies, there are several granite-gneiss lithologies that have no identifiable undeformed equivalents. These include some augen-gneisses which show the same textural relationships as those already described, and the fine-grained granite-gneiss which in hand-specimen has an even-grained texture with oriented flakes of biotite and white mica scattered throughout the rock. In thin-section this apparently simple texture is seen to be extremely complex. Unlike the other granitic lithologies there seems to be no tendency for quartz-plagioclase interfaces to be less common than quartz-quartz. Quartz occurs in two distinct types of grain: as irregular grains similar in size to the majority of the feldspar grains, and as tiny droplets entirely enclosed in feldspar. The K-feldspar occurs as crudely polyhedral grains but the grain boundaries are generally concave and have finger-like extensions reaching out between adjacent quartz and plagioclase grains. It is not clear how this texture developed since its original nature is unknown. In some places it seems that extreme deformation of the augen-gneiss can lead to this kind of texture, but the fine-grained granite gneiss locally contains relatively undeformed xenoliths suggesting that this is not a general explanation.

K-feldspar Reactions

The K-feldspar augen have been regarded by some authors as evidence of potash metasomatism accompanying the Alpine metamorphism, while others have concluded they are relict phenocrysts from porphyritic granites. Frasl (1954, 1957) has argued that the frequent occurrence of rings of oriented plagioclase inclusions is proof of a magmatic origin, and his argument was subsequently accepted by Exner (1957, 1964). However, similar rings of inclusions are common in other areas and can also be explained as the result of reactions in the solid state. Voll (1960) and Riederer (1967) have suggested possible mechanisms, which do not, however, explain the great abundance of inclusions in the present case.

A pre-metamorphic origin of the K-feldspar augen is supported by their deformation during the main Alpine movements, as shown by the way the ends of certain augen are drawn out in spindles parallel to the mineral lineation. Also the reaction to produce phengite, discussed below, indicates that K-feldspar became unstable during the Alpine metamorphism. It is possible the K-feldspar augen did form originally as porphyroblasts by potash metasomatism but if so, this is more likely to have been associated with Variscan metamorphism.

The formation of phengite from K-feldspar is observed in most of the augen-gneisses of the upper structural levels of the Zentralgneis; it seems to be less widespread in the lower levels of the Gössgraben Kern. Phengite occurs as sheaths surrounding the augen so that a reaction involving K-feldspar seems to be well established. On the other hand there is no textural evidence, in the majority of cases, that biotite was involved in the reaction. In shear zones, however, the amount of phengite is very great and here direct replacement of biotite by phengite is seen. The general increase in the amount of phengite with degree of deformation (i.e. with the intensity of the foliation) suggests that the reaction may depend upon mechanical granulation of K-feldspar and accessibility to a fluid phase.

Partial analyses of three phengite samples show 2.0 to 2.5% MgO and 3.5 to 4.0% Fe as Fe_2O_3; these are very similar to compositions reported by other workers for lower amphibolite facies rocks (LAMBERT, 1959; BUTLER, 1967; SCHWANDER et al., 1968). The problem in the phengite-bearing augen-gneisses where there is no evidence for reaction of biotite, is the source of the iron and magnesia for the phengite reaction. Presumably a certain amount of biotite must have reacted despite the lack of textural evidence for this. Possibly a reaction occurred in which biotite participated both as reactant and product, e.g.

(9) K-feldspar + 2 biotite + quartz + water = 2 phengite + biotite

This reaction balances if oxidation occurs in going from left to right.

Phengite is the typical white mica formed during Alpine metamorphism throughout the Tauern (EXNER, 1965; FRISCH, 1970). In other areas it has frequently been associated with low-temperature, high-pressure metamorphism in the glaucophane schist facies (ERNST, 1963). Several studies have recently shown that it is also the typical white mica of Barrovian-type greenschist facies metamorphism (LAMBERT, 1959; GRAESER and NIGGLI, 1967; HUNZIKER, 1966; SCHWANDER et al, 1968). BUTLER (1967) presented analytical evidence for a general decrease in the phengite content of white mica with increasing metamorphic grade, although GUIDOTTI (1969) showed that BUTLER's data could in part be explained by the control of white mica composition by bulk rock composition. In the Ticino Alps the high grade limit of phengite corresponds approximately to the calcite-diopside isograd in calc-silicate rocks (TROMMSDORFF, 1968).

The only hydrothermal data available (VELDE, 1965) serve chiefly to demonstrate the complexity of the system. The phengite structure apparently discriminates against ferrous iron so that iron and magnesia operate as independent components; also ferric iron can readily enter the phengite lattice so that phengite-producing reactions are likely to be affected by oxygen fugacity. A further complication is apparent from some of the analyses of SCHWANDER et al (op. cit.) which show appreciable paragonite solid solution with phengite. In general, therefore, the stability field of phengite found by VELDE (op. cit.) for 30% MgAl celadonite should be affected by Na and Fe contents and P_{O_2}. This work indicated an upper stability limit at 6 kb of about 500° C.

As noted above, in shear zones the amount of phengite present increases appreciably. This is the first stage in the development of Weissschiefer from quartzofeldspathic rocks. Weissschiefer are composed of quartz and one or more mica phases which vary from one locality to the next. Typically the mica is phengite but this may be accompanied by a very pale biotite (phlogopite) or a second non-pleochroic white mica, which may be paragonite. At one locality in the Badgastein area a meter thick zone of pure paragonite Weissschiefer occurs.

These rocks have developed by complete reaction of plagioclase and K-feldspar with the fluid phase to produce mica+quartz. The reaction between feldspars and a fluid phase containing alkali chloride has been discussed by GRESENS (1967); mica formation is favoured by high pressures (P_{H_2O}) and low pH (H^+/K^+ ratio). In the shear zones of the present map-area a reaction of the following form seems to have occurred

(10) K-feldspar+2 albite+2 anorthite+6 H^+ = 3 muscovite+10 quartz+ 2 Ca^{++}+2 Na^+.

Structural State of K-feldspar

Optically the K-feldspars in all the granitic rocks are intermediate between microcline and orthoclase. Individual crystals are frequently inhomogeneous; extinction varies patchily from place to place and some parts show indistinct cross-hatch twinning whereas others are apparently untwinned. The presence of both monoclinic and triclinic domains within single crystals was confirmed by x-ray diffraction of powders prepared from single augen; these showed peaks corresponding to Or (131) as well as to microcline (131) and ($1\bar{3}1$). The triclinicity values indicated ranged up to 0.74 (MACKENZIE and SMITH, 1960) but the values are only approximate since possible apparent wavelength shift resulting from the adjacent orthoclase peak was not taken into account.

In the coarse leucogranite some interstitial K-feldspar has better developed cross-hatched twinning and may approach maximum microcline.

Maximum microcline with well-developed cross-hatch twinning is observed only in the gneiss lamellae in the outer zones of the Peripheral Schieferhülle.

Plagioclase

The distribution of plagioclase compositions is shown in Fig. III—5. In much of the Peripheral Schieferhülle the plagioclase is albite; passing inward toward the Zentralgneis there is a discontinuous transition to oligoclase with anorthite contents in excess of 20%. The break is observed in basic, and pelitic lithologies; in granitic rocks too, a distinct break would be expected but there are no outcrops of granitic rocks between the albite-bearing gneiss lamellae and the oligoclase-bearing main mass of Zentralgneis. In some rocks albite and oligoclase coexist, with oligoclase rimming the albite. Within the oligoclase zone there is no systematic change of composition with grade, but there is a tendency for the anorthite content in carbonate-bearing lithologies to be higher.

Similar changes in plagioclase composition have been observed in other metamorphic regions (e.g. DE WAARD, 1959; WENK, 1962; WENK and KELLER, 1969). The sudden change in composition has been attributed to the effect of the peristerite solvus, and several workers have demonstrated the coexistence of albite and oligoclase in the same rock (EVANS, 1964; CRAWFORD, 1966); CRAWFORD suggests that the two phases are in equilibrium and has shown that the composition of both phases varies systematically with grade. On the basis of her results from New Zealand and Vermont she proposed a general form of the peristerite solvus; others (e.g. SMITH and RIBBE, 1969) doubt the existence of a true solvus and favour a kinetic compositional control. The general increase in anorthite content has usually been attributed to the breakdown reactions of calcite and epidote (RAMBERG, 1952; NOBLE, 1962; KRETZ, 1963).

Of the many possible reactions the only one for which textural evidence is observed is:

(11) $2\,ep+CO_2 = 3\,an+cc+H_2O$

This reaction will be strongly influenced by the relative values of P_{CO_2} and P_{H_2O}. In the carbonate-bearing rocks the anorthite content is generally higher than in the surrounding rocks, presumably because high P_{CO_2} has driven the above reaction to

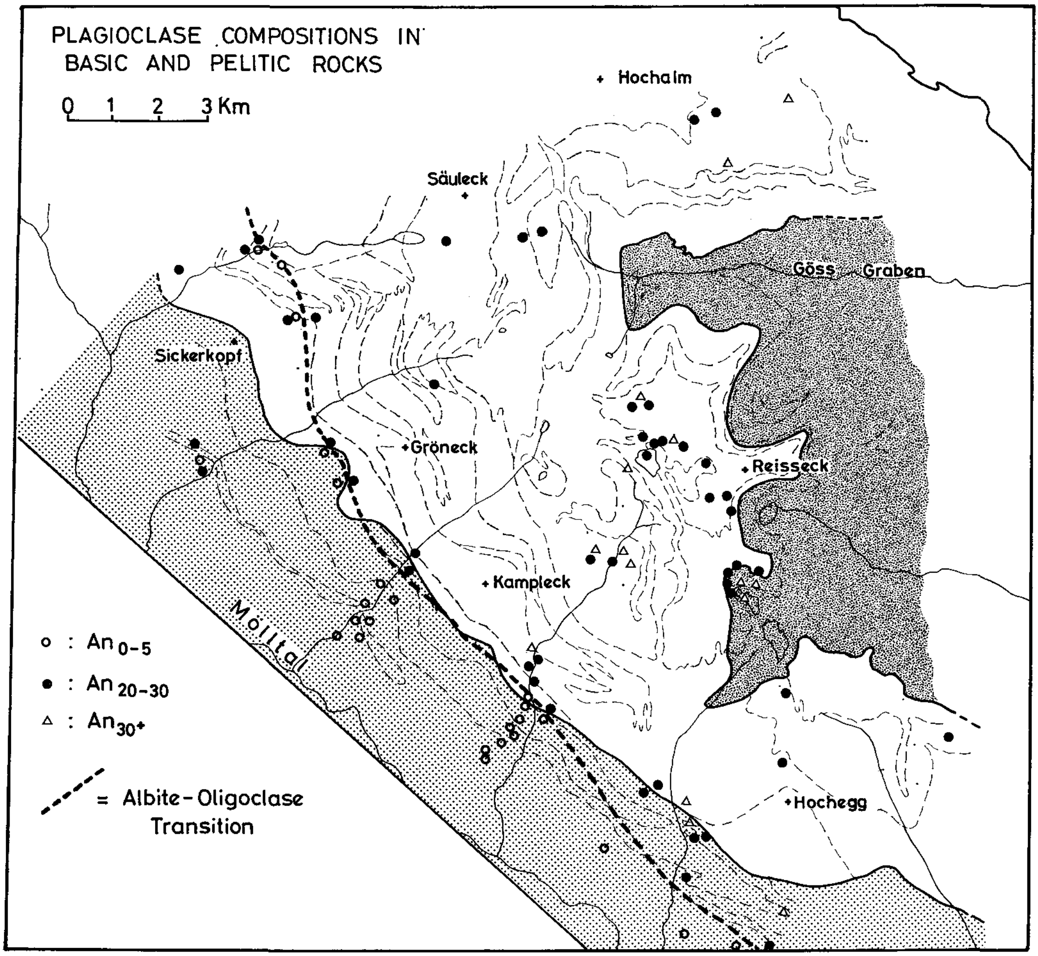


Fig. III - 5

the right. A similar effect was noted by WENK (1962) and WENK and KELLER (1969) in the Ticino Alps. In many cases the textural evidence indicates the reaction has proceeded in the opposite direction; this may reflect a drop in P_{CO_2} during the course of metamorphism.

The lack of textural evidence for reactions producing oligoclase in the majority of rocks may be explained if reactions were primarily between the solid reactants and a fluid phase. Evidence for this type of reaction has been presented by CARMICHAEL (1968) and the experimental work of ORVILLE (1963) and WINKLER (1967) shows that reaction between feldspars and a chloride bearing fluid phase may be rapid.

Thus reaction (9) above might occur as

(12) $2\,ep = 3\,an + Ca^{++} + 2\,OH^-$

coupled with

(13) $CaCO_3 \xrightleftharpoons{aq} Ca^{2+} + (HCO_3)^-$

Reaction of preexisting albite to produce oligoclase may be described by

(14) $2\,ab + Ca^{2+} = an + 4\,SiO_2 + 2\,Na^+$.

ORVILLE has shown that the composition of alkali feldspar varies continuously with the composition of the fluid phase, except for one composition, unique at a given temperature, at which two alkali feldspars coexist. He also predicts the effect of varying K/Na ratio on feldspar compositions in the ternary system. There are no experimental data relating fluid Ca/Na ratio to that of the coexisting plagioclase. In general, a high Ca/Na ratio may be expected to yield a higher anorthite content, but the effect is more complicated than in the alkali feldspars since the reaction is not a simple substitution.

As noted above the plagioclase composition throughout the Zentralgneis is remarkably uniform. In part this may reflect an originally uniform composition, but in the case of the tonalite and granodiorite, it is clear that oligoclase has been produced by reaction of an originally more calcic plagioclase. In those samples of the tonalite in which the original large subhedral tablets of plagioclase are preserved, the interior of each grain is packed with microlites of white mica and clinozoisite forming the well-known "gefüllte Plagioklas" texture that is found throughout the Tauernfenster. Both in the present area and elsewhere (FRASL, 1953; KARL, 1959), the microlites of clinozoisite are often arranged concentrically, strongly suggesting an original oscillatory zoning; KARL demonstrated the striking resemblance between these rings of microlites and the oscillatory-zoned plagioclase from the Periadriatic tonalites. This feature suggests that the reaction has occurred without strongly disrupting the lattice and probably without a large volume change. The reaction usually suggested is

$$(15)\quad \underset{404\ cc}{4\ CaAl_2Si_2O_8} + K_2O + \tfrac{1}{2}Al_2O_3 + SiO_2 + 2\ H_2O \rightleftharpoons 2\ \underset{264\ cc}{Ca_2Al_3Si_3O_{12}(OH)} + \underset{141\ cc}{KAl_3Si_3O_{10}(OH)_2}$$

Clearly this involves extensive introduction of extraneous reactants and is unlikely to preserve the subtleties of original zoning. A more probable reaction is

$$(16)\quad \underset{606\ cc}{6\ Ca\ Al_2Si_2O_8} + K^+ + 2\ H_2O + (OH)^- = \underset{396\ cc}{3\ Ca_2Al_3Si_3O_{12}OH} + \underset{141\ cc}{KAl_3Si_3O_{10}(OH)_2}$$

$(\Delta V = 11\%$ for An_{100}; 5% for $An_{50})$.

After completion of this reaction in most of the tonalite and granodiorite, the plagioclase recrystallised to a fine-grained mosaic. The composition of this mosaic was compared with that of the earlier large grains by means of the microprobe. The results (Table II—1) demonstrate that there was no large-scale change in composition accompanying the recrystallisation. However, the plagioclase in the mosaic shows quite strong inverse zoning. The unrecrystallized plagioclase has a composition of $An_{23\pm2}$ whereas the fine mosaic is usually zoned from $An_{22\pm4}$ in the cores to $An_{28\pm3}$ at the rims. Similar inverse zoning is observed throughout the Zentralgneis. In the past it has generally been interpreted in terms of recrystallization under conditions of rising temperature or falling P_{H_2O}, the composition being controlled by reaction with epidote. Considering the relationship to metamorphic grade demonstrated by CRAWFORD (1966), other interpretations are possible: if the observed plagioclase compositions do delimit a true solvus, then inverse zoning could result from crystallisation under falling temperature; alternatively it could equally well develop at constant temperature as a result of changing fluid composition. In nature neither parameter is likely to be constant and plagioclase zoning will depend on epidote and other solid phases reacting with the fluid phase, and on temperature.

As noted earlier the plagioclase in the Zentralgneis is of rather uniform composition; oligoclase is found both in lithologies with little or no epidote, such as the coarse leucogranite, as well as in the epidote-rich tonalite and granodiorite. This uniformity of composition is readily understood if the plagioclase composition was controlled by the fluid phase composition, since the fluid phase in the small leucogranite bodies may have approached equilibrium with that in the enclosing tonalite. Thus plagioclase in the leucogranite may have become more calcic in the course of metamorphism, through the operation of reaction (14); if the silica released by this reaction were precipitated in situ the anomalously high quartz content of some of the leucogranites could be explained. The principal objection to this explanation is the lack of textural evidence for crystallisation of new quartz in the leucogranite although it is possible that this occurred as overgrowths on pre-existing grains. In the fine-grained granite gneiss on the other hand there is textural evidence for new quartz in the form of tiny droplets inside plagioclase grains but these rocks are not among those with unusually high modal quartz. The plagioclase of the Peripheral Schieferhülle and Randgneis contains abundant quartz droplets and myrmekitic intergrowths which would be consistent with an origin by reaction of albite with a fluid phase.

Supposing that plagioclase composition is indeed controlled through the fluid phase, it is important to establish what reactions may buffer the composition of this phase. As discussed above, the epidote-breakdown reactions are likely to be important; in addition, carbonate equilibria and the alkali feldspar/fluid phase equilibrium will also affect the Ca/Na ratio of the fluid phase. Several epidote reactions have been investigated experimentally: NEWTON (1965, 1966) has investigated the following equilibria:

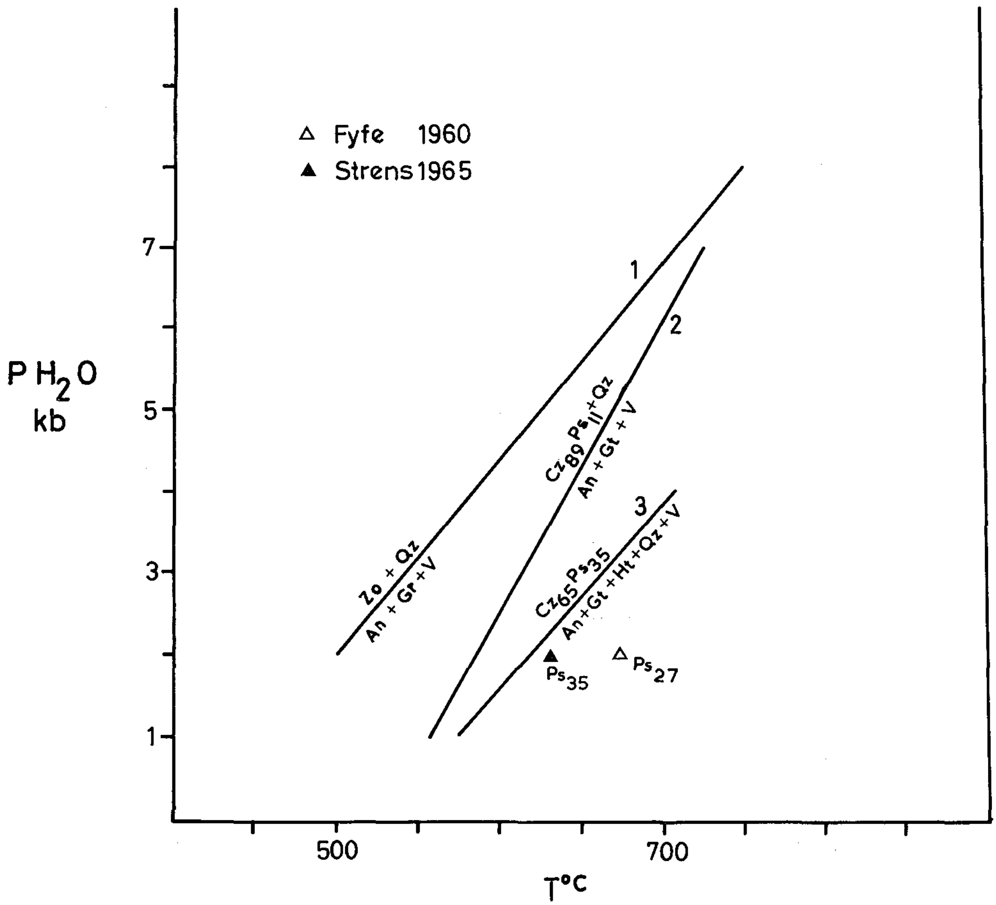


Fig. III - 6 Experimental data on epidote stability; (1) NEWTON (1966); (2) HOLDAWAY, 1965 and (3) 1967.

(17) 8 zoisite + 2 quartz = 10 anorthite + 2 grossular + 4 water vapour
(18) zoisite + sillimanite = anorthite + grossular + corundum + vapour
(19) zoisite + sillimanite + quartz = anorthite + corundum + vapour

In addition the breakdown of some iron-bearing epidotes has been studied by FYFE (1960), HOLDAWAY (1965, 1967), and STRENS (1965). A synopsis of these results is presented in Fig. III—6. None of the reactions studied is likely to have occurred in the rocks of the present map-area since grossular garnet was not recorded, but the data emphasise the strong pressure dependence of these reactions involving epidote and a vapour phase by comparison with other dehydration reactions; the reaction curves in Fig. III—6 have slopes of about 40° *C/km.* which would make a low angle intersection with possible geothermal gradients. This suggests that the temperature of the epidote break-down might vary rapidly from one area to another with slight fluctuations in the thermal gradient. On the other hand the albite-oligoclase reaction seems to occur at a fairly constant temperature as shown by its successful use to define the base of the amphibolite facies in basic rocks (FYFE et al, 1958; TURNER, 1968; WINKLER, 1967). At lower pressures the plagioclase reaction might be inhibited from occurring if the fluid phase Ca/Na ratio, which increases through the breakdown of epidote, is held below the critical value for oligoclase crystallisation by a third reaction possibly involving garnet or amphibole, although amphibole compositions generally change little at this facies boundary (ENGEL and ENGEL, 1962). The commonly observed reaction in amphibolites

(20) ilmenite + Ca^{2+} + SiO_2 = sphene

would also tend to lower the Ca/Na ratio.

Myrmekite

Myrmekitic intergrowths of quartz and plagioclase are widespread in all lithologies of the Zentralgneis. Recent literature has contained many discussions of the formation of this texture and has increasingly favoured exsolution of both quartz and plagioclase from a supersilicated feldspar phase. This theory was originally proposed by SCHWANTKE (1909) and found support in the experimental work by CARMAN and TUTTLE (1963). Several features of the myrmekite in the Zentralgneis, however, are inconsistent with this explanation.

a) In several cases myrmekite occurs in K-feldspar along cracks which clearly developed after the exsolution of perthite.

b) The amount of myrmekite in a single grain is highly variable and often in excess of what might reasonably be contained in solid solution. In a few cases K-feldspar grains have been almost completely replaced by myrmekite.

In addition to these features, a satisfactory theory of myrmekite formation must explain its general association with K-feldspar and its rare occurrence independently; the similarity of composition of myrmekite and other ground-mass plagioclase, and the vermicular form of the quartz. Various theories have been reviewed by SHELLEY (1964, 1967); BECKE's (1908) theory of metasomatic attack of alkali feldspar, however, accounts for most features observed in the Zentralgneis and in addition BECKE was able to demonstrate the appropriate relationship between plagioclase composition and the relative amounts of quartz and plagioclase demanded by his proposed reaction

(21) 2 or + CaO = an + 4 SiO_2 + K_2O.

In common with other theories, however, it does not explain the vermicular form of the quartz. This could result from the interfacial free energies associated with quartz/plagioclase and quartz/K-feldspar grain-boundaries. As noted in the general discussion of metamorphic textures above, there is a pronounced tendency in the augen-gneisses for quartz and plagioclase to occur as monomineralic segregations, and it was suggested that the rarity of quartz/plagioclase interfaces might be explained by a relatively high interfacial energy. Furthermore, a relatively low interfacial energy for interfaces involving K-feldspar can be inferred from the concave boundaries and intergranular, almost reticulate habit of K-feldspar, noted in the fine-grained granite-gneiss. Given a plagioclase/K-feldspar interface at which quartz is crystallising the shape of the growing quartz crystal may therefore be expected to maintain the area of quartz/K-feldspar interface at a maximum and the area of quartz/plagioclase interface at a minimum. If the myrmekitic plagioclase is An_{25} the reaction is

$$(22)\quad \underset{543\ cc}{5\ KAlSi_3O_8} + 3\ Na^+ + Ca^{2+} = \underset{300\ cc}{3\ NaAlSi_3O_8} + \underset{101\ cc}{CaAl_2Si_2O_8} + 5\ K^+ + \underset{91\ cc}{4\ SiO_2}.$$

Since approximately four volumes of plagioclase crystallise for every volume of quartz a grain of quartz once nucleated will be constrained to maintain an interface with the K-feldspar, rather than grow to a more equant form. As suggested above, the droplets of quartz occurring in the plagioclase of the fine-grained granite-gneiss might be the result of the analogous reaction with sodic plagioclase, where, assuming anisotropy to be insignificant, the lowest interfacial energy would be achieved by spherical grainse This reaction would also explain the frequent myrmekitic character of the oligoclase in the Peripheral Schieferhülle as having developed during the replacement of albite.

Conditions of Alpine Metamorphism

In most of the map-area the metamorphic assemblages can be attributed to the amphibolite facies. Greenschist facies rocks, defined by presence of albite in basic lithologies, are largely restricted to the Peripheral Schieferhülle. The transition from albite- to oligoclase-bearing rocks is narrow and in the northwest, falls within the Randgneis of the Basement Complex, suggesting that the metamorphic grade of the Peripheral Schieferhülle increases to the southeast. This is consistent with the retrogression of garnets in the Randgneis in the north, and with the restriction of staurolite-bearing lithologies to the southern part of the area.

Estimates of total pressure during the Alpine metamorphism can be based on the supposed thickness of the overburden. The Peripheral Schieferhülle could have been overlain by three major tectonic units at the time of its metamorphism — the Unterostalpin thrust sheet, the Altkristallin (metamorphic) rocks of the Ober-(Mittel-)ostalpin sheet, and the Northern Calcareous Alps possibly with some of the Grauwackenzone rocks beneath them; the thickness of all of these units is rather uncertain. The northern Calcareous Alps and Grauwackenzone are unlikely to have exceeded 4 *km.*; the Unterostalpin is several kilometers thick in the Radstädter Tauern but appears to thin southwards, say 1 *km.* Addition of 6 *km.* for the metamorphic rocks of the Altkristallin gives a figure of 11 ± 4 *km.*, which is roughly equivalent to 3 ± 1.5 kb pressure at the top of the Peripheral Schieferhülle.

The thickness of the Peripheral Schieferhülle itself could have been very variable because it was presumably deformed by the over-riding Ostalpin thrust masses. There is evidence today for a thickness of 5 *km.* or more in tectonic depressions (e.g. Glockner-depression, Fig. I—2) but such great thicknesses may not be typical of all the Tauern if formation of the depressions began before the conclusion of the overthrusting.

We therefore assume the following pressures during the main Alpine metamorphism

Top of the Peripheral Schieferhülle ....—3 kb
Base of the Peripheral Schieferhülle ....—5 kb
Deepest present day exposure of the Zentralgneis ....—6 kb

These estimates are also consistent with pressure-estimates from the occurrence of kyanite (Fig. III—4).

Temperature estimates are rather difficult to make because of uncertainty about the variation of the partial pressures of water, carbon dioxide and oxygen, and the magnitude of the effects these would have on stability relations. A relative lowering of P_{H_2O} was deduced in the carbonate lithologies of the Peripheral Schieferhülle and in the non-calcareous lithologies conrained in them. The partial pressure of oxygen was of P_{H_2O} was deduced in the carbonate lithologies of the Peripheral Schieferhülle and in the non-calcareous lithologies contained in them. The partial pressure of oxygen was lower in the Peripheral Schieferhülle than in the Inner Schieferhülle, as indicated by the preponderance of sulphides among the opaque minerals of the former; in the latter, oxides predominate and higher ferric iron contents in the epidote also suggest higher P_{O_2}. These differences persisted throughout the metamorphism and are reflected in the fissure mineral assemblages.

In the serpentinites of the Peripheral Schieferhülle, a temperature between 430° and 500° C is indicated. The occurrences of kyanite, as described earlier, probably require temperatures of over 400° C. The staurolite and chloritoid assemblages probably require rather higher temperatures but are subject to large uncertainties; possibly a temperature close to 550° C is reasonable for the formation of staurolite at the base of the Peripheral Schieferhülle. For conditions of $P_{H_2O} = P_{tot}$, a maximum temperature could be established from the absence of partial melting in the Zentralgneis, which would begin at approximately 650° C.

There is little indication of more than one main phase of crystallisation during the Alpine metamorphism. The majority of porphyroblastic minerals crystallised or recrystallised after rotational deformation had ceased although in the higher-grade rocks two stages of garnet growth are inferred. Some of the mica must also have originally crystallised before the folding. In the Zentralgneis the two-stage crystallisation of plagioclase, especially in the tonalite suggests that metamorphism may have started before the main deformation but the uniformity of plagioclase composition argues against a significant change in conditions between the two stages.

In the discussion of feldspar compositions the possibility of relatively high mobility of calcium, potassium and sodium was suggested on the basis of very similar feldspar compositions in widely differing lithologies. The widespread occurrence of myrmekite, and the general occurrence of inversely-zoned oligoclase can be explained by reactions between plagioclase and a fluid phase analogous to those established for alkali feldspars by ORVILLE (1963). ORVILLE'S results might also explain the antipathetic relation between chess-board albite and myrmekite recorded by EXNER (1961). In the deeper parts of the Zentralgneis the fluid phase composition was probably mainly controlled by epidote-breakdown reactions, resulting in a low Na/Ca+K+Na ratio; in the highly sheared upper parts of the Zentralgneis and particularly in the gneiss-lamellae, on the other hand, the fluid phase composition was more likely to have been influenced by the composition of the fluid phase in the adjacent metasediments. Within the stability field of epidote the latter would be characterised by a high Na/K+Ca+Na ratio so that reaction of K-feldspar to albite would be expected.

In conclusion it is important to note that from the present work there is no evidence that in this part of the Tauern the main Alpine metamorphism was associated with higher pressures than the typical Barrovian series (ANGEL, 1939; DEN TEX, 1965); nor is there any indication of an earlier low temperature/high pressure metamorphism, such as occurs in the eastern part of the Swiss Pennine Alps and might be inferred for the Tauern on the basis of occurrences of glaucophane and eclogitic bodies recorded elsewhere within the Peripheral Schieferhülle.

IV. Structure

by R. J. NORRIS, E. R. OXBURGH, R. A. CLIFF and R. C. WRIGHT

Introduction

For the purpose of a structural analysis of the area, three principal units may be considered. These are:

(a) The Peripheral Schieferhülle. As this is thought to be largely Mesozoic in age, by correlation around the Sonnblick Kern with the type localities in the Glockner depression, it may be expected to exhibit only Alpine structures.

(b) The Inner Schieferhülle. As this is a pre-Mesozoic metamorphic series, and was intruded by Variscan granites (see chapter V), it may exhibit several phases of deformation, both Alpine and pre-Alpine in age.

(c) The Zentralgneis. Much of the Zentralgneis is late Variscan in age (Chapter V), although some parts may be rather older. Hence it may in places exhibit pre-Alpine deformation, while not in others.

The latter two units comprise the pre-Mesozoic basement and will be discussed together. Because of the age relationships discussed above, folds of various ages may be present in some parts of the area and not in others. Because of this, the usual system of numbering fold phases (i.e. F_1, F_2, etc.) is difficult to apply, as some fold episodes may predate some of the rocks. Faced with a similar problem in the Western Alps, CHADWICK (1968) used a letter of the alphabet as a qualifying subscript. The disadvantage with this system is that the subscripts give no immediate indication of the relative age of the folds. The system employed here is to use a letter to signify the major period of deformation and a number to signify the particular phase within that period. Thus pre-Alpine folds will be F_V and Alpine folds F_A. Two phases of Alpine folding, say, will be designated as $F_A{}^1$, $F_A{}^2$ etc. Foliations produced during a phase of folding will be identified by the same subscripts as the fold phase.

In the area mapped, the Peripheral Schieferhülle exhibits two phases of deformation, $F_A{}^1$ and $F_A{}^2$ while the Inner Schieferhülle exhibits three or more phases; at least one of these is pre-Alpine, and they are termed F_V, $F_A{}^1$ and $F_A{}^2$. These symbols will be introduced into the discussion after the evidence for their recognition and correlation has been described.

Peripheral Schieferhülle

Mesoscopic Scale

The nature of the principal planar feature in the Peripheral Schieferhülle varies according to the lithology; the mica-schists and calc-schists exhibit a mineral foliation while the foliation in the marbles is defined largely by a colour banding, although any mica flakes lie parallel to it. Lineations are not strongly developed in the Riekengraben and Mühldorfergraben, although further north, in the Kaponiggraben, strongly lineated dolomite-quartzite breccias occur. A mineral lineation is occasionally present, especially in the amphibole-bearing lithologies, but in the southern part of the area, the planar element is dominant. Another linear feature more frequently seen is an intense crenulation of the foliation. In the mica-schists, it is generally only visible in thin-section. In the calc-schists and marbles, however, it clearly corresponds to the hinge-lines of tight to isoclinal folds which are occasionally found. From evidence such as this it becomes clear that the principal planar feature is parallel to the axial surfaces of these folds, the limbs being rotated into parallelism, and the hinges gradually destroyed by shearing to produce a new foliation (e.g. BALK, 1936).

Figure IV—1 (see page 200) shows a large mesoscopic fold in banded marbles in the Taborgraben. The fold profile becomes tighter upwards along the axial surface, while the hinge of the synform on the left has become sheared out. In the core of the

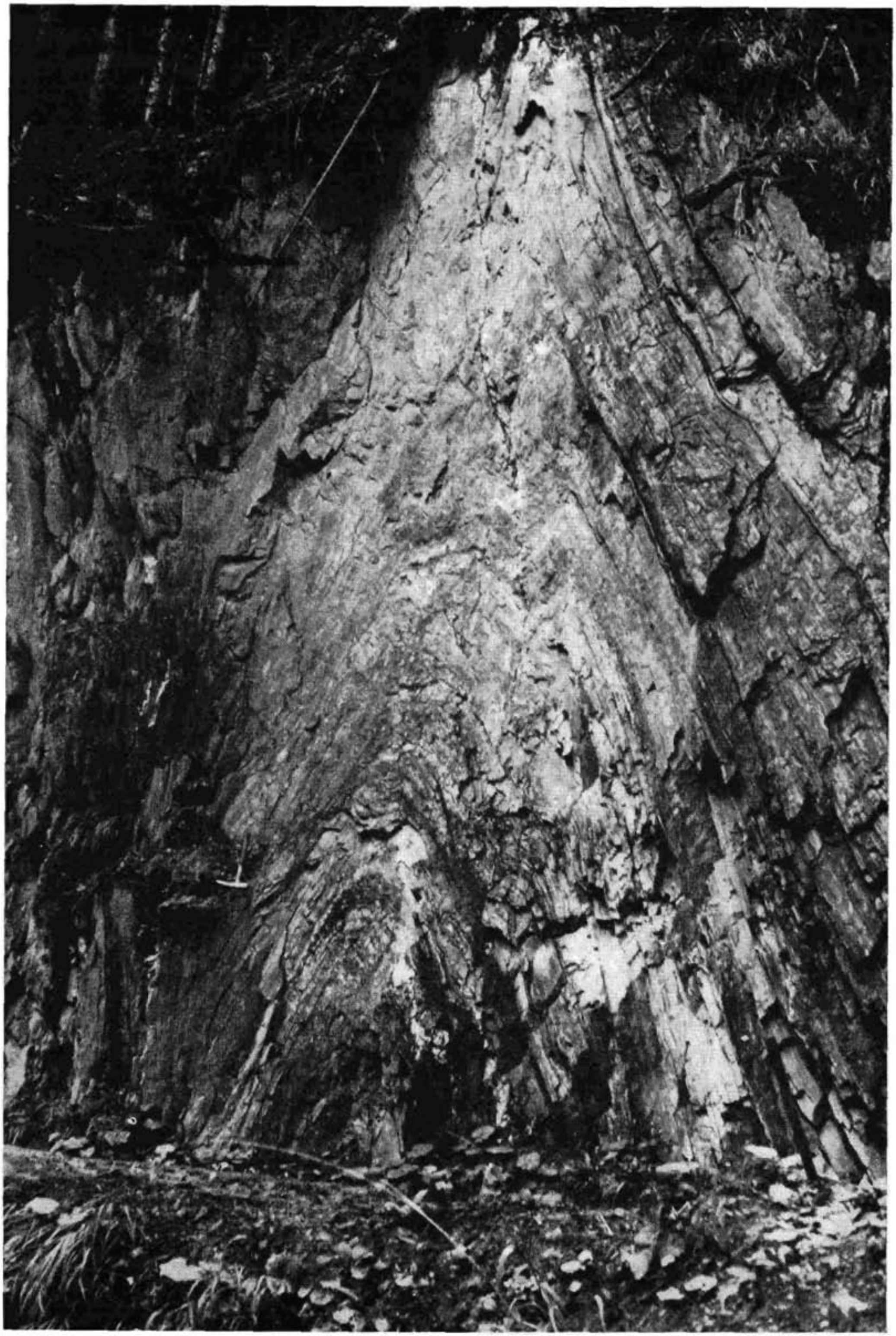


Fig. IV - 1 Large subisoclinal fold in banded marbles of the Peripheral Schieferhülle; Tabor Graben; hammer for scale. For detail see Fig. IV - 2.

fold a small-scale crenulation of the finer bands occurs which in many places becomes so intense as to define a new planar feature parallel to the axial surface (Fig. IV—2; see page 201).

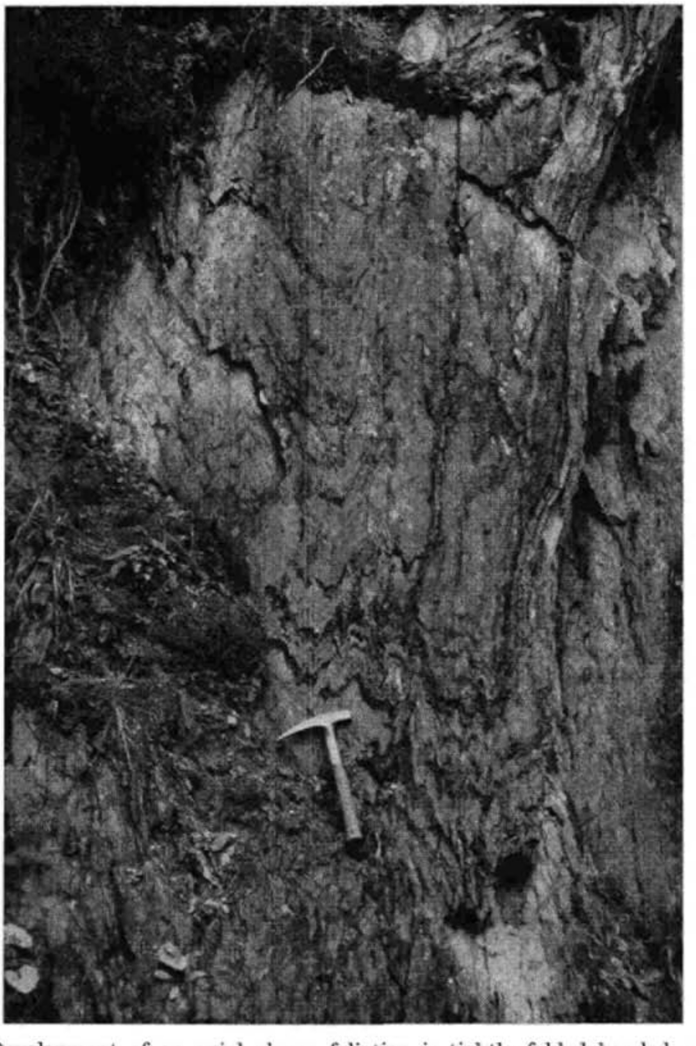


Fig. IV - 2 Development of an axial planar foliation in tightly folded banded marbles and calc-schists; Tabor Graben; hammer for scale. Peripheral Schieferhülle.

Within the calc-schists, the folds are commonly less well preserved and the principal planar feature is parallel to the axial surfaces (Fig. IV—3; see page 202). The flakes of white mica are parallel to this foliation, fold hinges being defined by layers richer in this mineral. In the schists and amphibolites, folds are rarely seen, although occasionally, quartz laminae may be intensely crenulated. The banding in the marbles and the micaceous layers in the calc-schists probably represent original sedimentary bedding although re-oriented and repeated by the deformation.

Folds are tight to isoclinal (i.e. with an interlimb angle of less than 30° [Fleuty, 1964]) and have an amplitude/wavelength ratio greater than 2/1. The profiles of the folds approach the perfect similar model (constant thickness measured parallel to the axial surface — class 2 of Ramsay, 1967). However, there are small deviations from

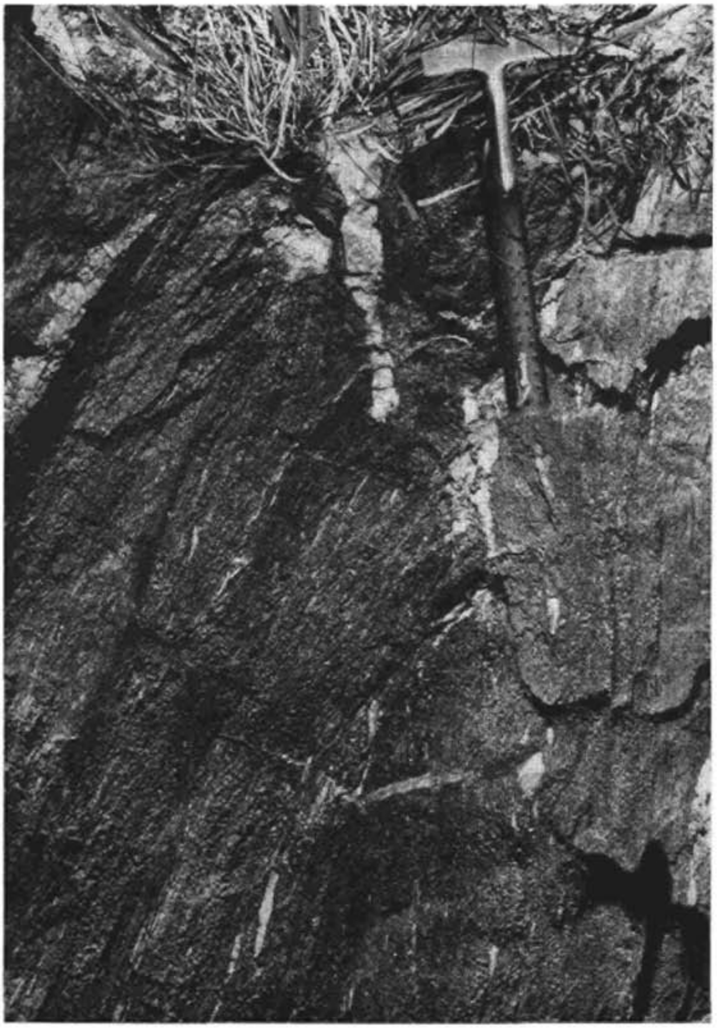


Fig. IV - 3 Tightly folded calc-schists near Kaponig; calcite segregations parallel to S; the tip of the hammer handle is a few *cm* to the right of a fold-core. Peripheral Schieferhülle.

this style which may be demonstrated by constructing isogons (lines joining points of equal slope on the two folded surfaces [RAMSAY, 1967]) for a series of folded layers, whereby a pattern of alternately diverging and converging isogons is obtained. In many of the folds, the limbs are of unequal thickness, the thinner one frequently passing into a shear zone. A sense of asymmetry is rarely observable as few folds may be traced over sufficient distance, but those most frequently encountered are Z-folds (RAMSAY, 1962 b) when viewed in a northwesterly direction (e.g. the greenstones of the Riekenbach).

A later set of structures is related to the development of fractures, faults and joints. In the Peripheral Schieferhülle, such features are much more abundant than in the older units. The calc-schists frequently exhibit closely spaced conjugate shear surfaces, breaking the rock up into small rotated packets. (Fig. IV—4.) Tension gashes and related boudinage develop in the banded marbles, whereas kink-bands are plentiful in the micaceous schists. Locally, these are sufficiently closely-spaced to define a new foliation. Flexures, associated with movement on joint planes, are abundant; their axes are steep as both the foliation and the fractures are sub-vertical. Quartz and carbonate commonly occur within the fractures.

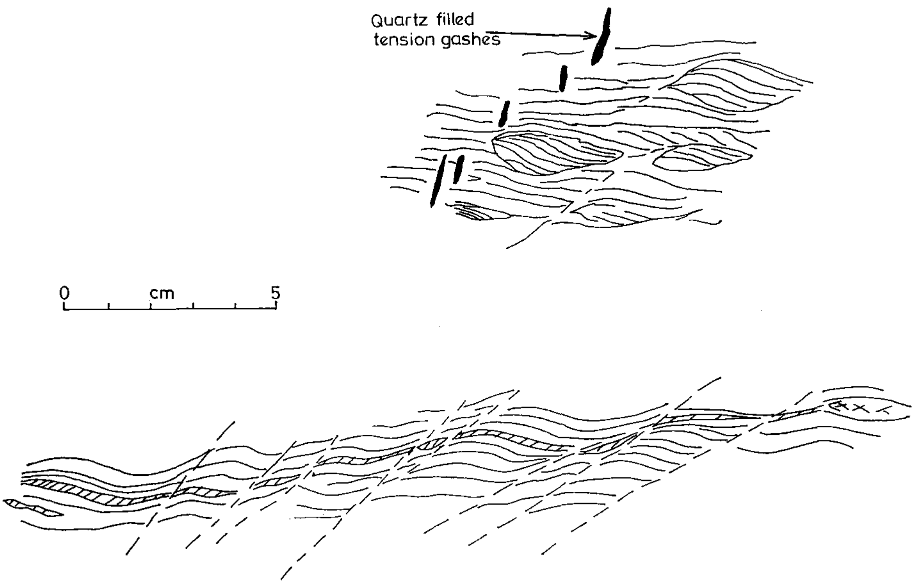


Fig. IV - 4 Closely spaced shear planes associated with joints in calc-schists of the Mühldorfer Graben; Peripheral Schieferhülle.

From specimens of the dolomite-quartzite breccia in the Kaponiggraben, the axial ratios of a number of pebbles were measured (see Appendix II for details). The mean axial ratios (determined by calculating the geometric mean [cf. Ramsay, 1967; Dunnet, 1969; Gay, 1969]) are 9.6 : 2.1 : 1. The long axes of the pebbles are parallel to the local fold axis, and this particular locality is probably situated in the hinge zone of a large asymmetric fold. The pebble measurements may represent an elongation parallel to the fold axis of 270% (natural strain [ε] = 1.4). However, other rocks nearby exhibit a predominantly planar fabric, suggesting that such a nearly uniaxial elongation may be of local significance only. Various mechanisms of producing elongate pebbles parallel to a fold axis are discussed in Appendix II. From this, limiting values of $\varepsilon = 1.5$ and 2.5 may be placed on the total flattening perpendicular to the foliation in these rocks. Despite the difficulty of interpretation, these measurements indicate that the Peripheral Schieferhülle has undergone a very great strain, in excess of 75% flattening. A strain of this magnitude would also be likely to modify initially concentric folds to such an extent that they would be scarcely distinguishable from perfect similar folds.

Microscopic Scale

In thin-section, most rocks from the Peripheral Schieferhülle possess a metamorphic fabric, in which the long axes of the mineral grains lie within the foliation. Folding on a microscopic scale is generally restricted to the phyllites and schists. In some of these rocks folds in thin-section are commonly defined by trains of opaque particles. Occasionally, flakes of mica also follow the fold mimetically. The axial surfaces of these small-scale folds lie parallel to the dominant foliation, as do those of the larger folds. They are overgrown by porphyroblasts of albite, oligoclase, and zoisite. In many cases, however, these minerals have themselves been somewhat deformed, suggesting that their growth began after the initiation of folding but continued during deformation.

In some of the higher grade rocks, an early growth of garnet is indicated, as many garnets exhibit cores containing rotated trains of inclusions. These inclusions are generally small and comprise quartz, calcite, mica or an opaque mineral (usually graphite). Marginally, the garnets have grown along the grain-boundaries of a coarser quartz fabric producing a fish-net texture. This suggests that garnet growth began during the rotational movements, and continued after the development of a coarse metamorphic quartz fabric. In some rocks (e.g. RN 431), a distinct 'unconformity' separates the inner and outer zones of the garnet (Fig. II—14). In some of the lower grade rocks (e.g. RN 419), the garnets have a curved internal fabric of fairly large quartz grains. The amount of curvature, however, is compatible with a flattening of the foliation around the growing garnet (cf. RAMSAY, 1962 a). The outer part of this garnet has grown along quartz grain boundaries, to give a fish-net texture.

Randomly oriented biotite and chlorite grains occur in some rocks and are probably related to the late-stage mineralisation. This has occurred principally along the joint planes, which are frequently filled with quartz, calcite and chlorite. Thus growth of metamorphic minerals continued until a very late stage in the structural development of the area.

Measurement of quartz c-axis and calcite c-axis orientations from five specimens within the Peripheral Schieferhülle, were made in order to determine the relationship between the microscopic and the mesoscopic fabric. Equal numbers of measurements were made in two sections cut approximately at right angles.

Two of the specimens investigated (RN 480, RN 484) were strongly lineated rocks from the Kaponiggraben, whereas the other three were strongly foliated rocks from the Riekengraben (RN 405, RN 406) and Mühldorfergraben (RN 436); see Fig. IV—19, 20. Specimen RN 480 is from the strongly lineated (Fig. IV—5; see page 205, dolomite-quartzite breccia in the Kaponiggraben, the lineation being defined by the elongation of dolomite fragments. The quartz c-axes are symmetrically oriented with respect to this lineation forming what appears to be a crossed girdle or large-radius, small-circle distribution. Sub-basal lamellae are fairly abundant in the quartz grains in this specimen, and these have a strongly preferred orientation with two maxima symmetrical to the weak foliation. Sub-basal lamellae have been produced experimentally by CHRISTIE et al. (1964) and HOBBS (1968), and shown to be due to slip parallel to the basal plane. If such lamellae form late in the deformation history (NAHA, 1959) and represent only an infinitesimal amount of strain, they may be used to determine the principal stress directions in the rocks (e.g. FRIEDMAN, 1964; CARTER and RALEIGH, 1969; SYLVESTER, 1969). In the present example, the orientation of the lamellae would suggest that σ_1 was oriented perpendicular to the weak foliation and σ_2 was parallel to the lineation. Lamellae are quite common in quartz from rocks near to the Mölltal, while they are rare within rocks belonging to the basement. As lamellae tend to be annealed at fairly low temperatures, this suggests that the rocks near to the Mölltal underwent a low temperature deformation at a late stage in their metamorphic history.

Specimen RN 484, also a strongly lineated rock from the Kaponiggraben, is a calcite marble with elongate lenses of quartz. The quartz c-axis orientation is somewhat similar to that of RN 480, but with a number of concentrations within the small circle distribution. Calcite c-axes from the same rock exhibit a maximum at a low angle to the pole to the foliation, and a considerable spread in a girdle normal to the prominent lineation.

Specimen RN 405 is from the Sonnblick lamella in the Riekengraben, and RN 406 from a quartzite occurring immediately southwest of the railway viaduct. Both are strongly foliated rocks, and both exhibit similar preferred orientations of quartz c-axes. These are concentrated into a broad zone perpendicular to the foliation; within the zone, two maxima occur which are the normals to two planes, the acute angle between which is bisected by the foliation.

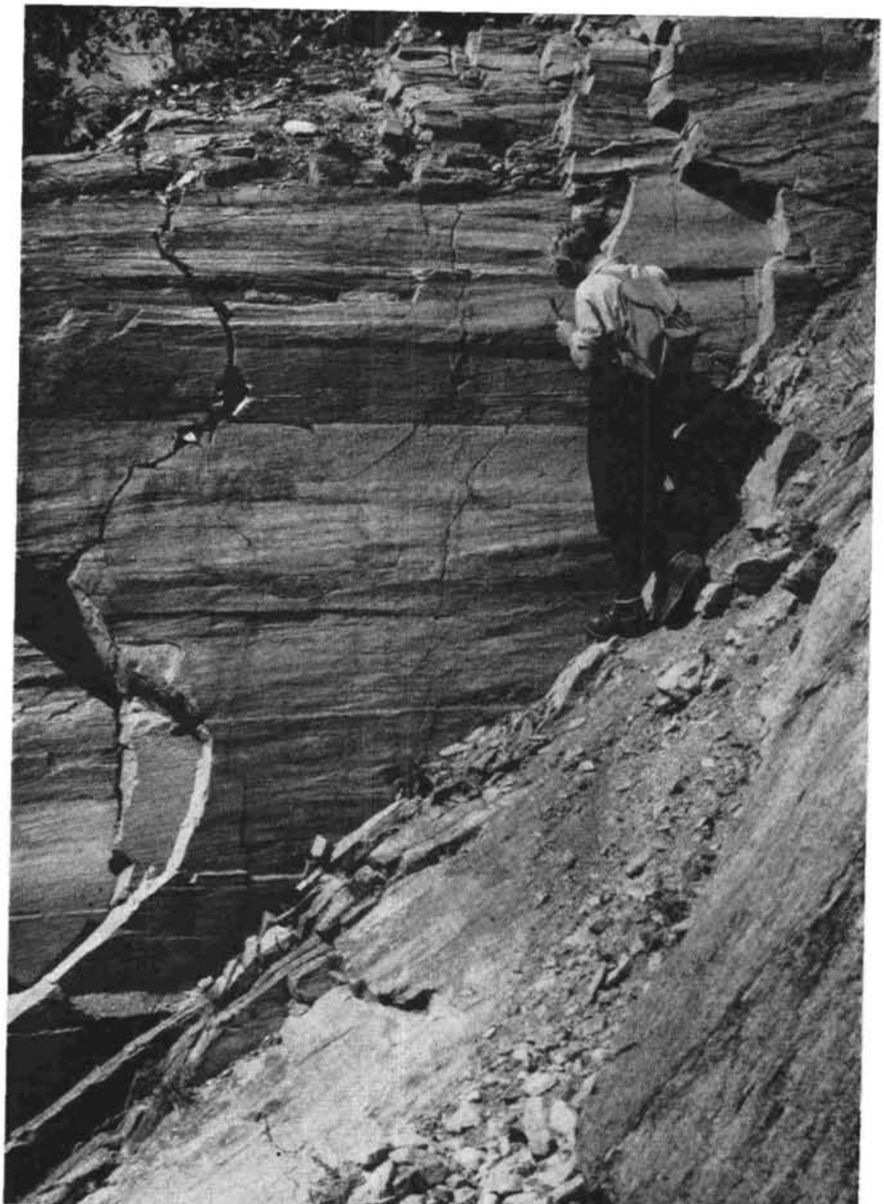


Fig. IV - 5 Strong lineations (microfolds) in dolomite quartzite breccia; Kaponig Graben; see also Fig. IV - 7. Peripheral Schieferhülle.

Specimen RN 436 is from a strongly foliated calcite marble in the Mühldorfergraben. The c-axes exhibit a strong maximum at approximately 30° to the normal to the foliation. Poles to e-twins within these grains have a similar preferred orientation. Such twins have been used to determine principal stress directions within rocks (e.g. CARTER and RALEIGH, 1969). However, TURNER and WEISS (1963) have pointed out that as e-twins form late in the deformation period, their orientation may be controlled by a previously developed grain orientation as well as by the deforming stresses. In the present example, calculated compression and tension axes do not fall into two clearly defined fields, suggesting that this is the case here.

Macroscopic Scale

The Peripheral Schieferhülle forms an elongate southwestwards dipping belt along the southwest side of the map area. The principal planar feature in the Peripheral Schieferhülle, which in most instances is a foliation parallel to the axial surfaces of the mesoscopic folds, steepens southwestwards until it becomes vertical and in places dipping to the northeast. This is the case as far south as Mühldorf, where the strike swings round to the east, and the dip becomes shallower, until in the Hinterregengraben, it is only 35° to the south.

The foliation in sub-areas 3 and 5 (Plate 4) has a mean dip of 50° to the northeast, whereas that in the adjoining sub-areas 2 and 4 is approximately vertical. The former two sub-areas cover the steep southwest sloping flanks of the Mölltal, where the outcrops are usually small crags projecting through the cover of boulder clay and vegetation, whereas in the latter two sub-areas exposures are principally found in the Rieken- and Mühldorfergrabens. In some of the larger exposures in sub-areas 3 and 5, there is clear evidence of extensive hill-creep, and it is suggested that this is the most likely explanation of the discrepancy between the mean attitude of the foliation on the valley sides and in the valley bottoms.

In only three places were sufficient minor structures seen to suggest the probable geometry of large-scale folds. In the Kaponiggraben, a section along the logging road was recorded, particular attention being paid to the sense of asymmetry of the mesoscopic folds. A suggested interpretation of the structure is shown in Fig. IV—6. The

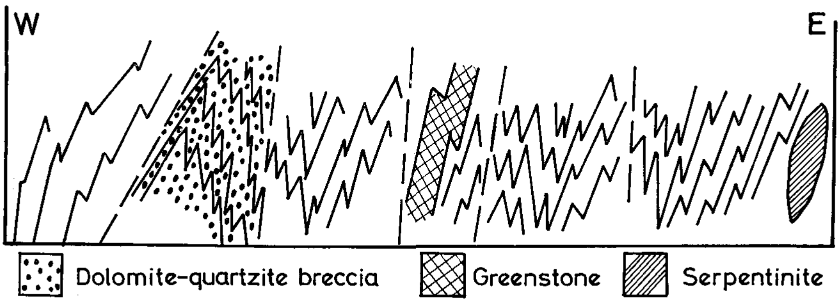


Fig. IV - 6 Interpretation of the structure of the northwest side of the Kaponig Graben; Peripheral Schieferhülle. That part of the section which is unornamented is calc-schist. The section is about 800 *m.* long. Peripheral Schieferhülle.

second place of interest is in the Riekengraben, where the thick band of greenstones crossing the river is folded into a series of asymmetric folds climbing to the northeast. The garnet-mica schists also exhibit this folding, but at the northeast margin of the latter, a thin dolomite horizon occurs which can be traced northwards up the hillside for two hundred metres or so without any evidence of folding. To the south, however, it is no longer found. The garnet-mica schists end abruptly at this junction and their place is taken by calc-schists and marbles, suggesting that the dolomite occurs along a major structural discontinuity. An interpretation of the structure is shown in Fig. IV—7.

In the calcareous lithologies along a logging road in the Taborgraben, the same pattern of asymmetric folds climbing to the northeast and cut by strain discontinuities is seen (Fig. IV—8). It is interesting that in all three cases where there is sufficient exposure of mesoscopic structures to indicate the character of the larger structures, these are similar.

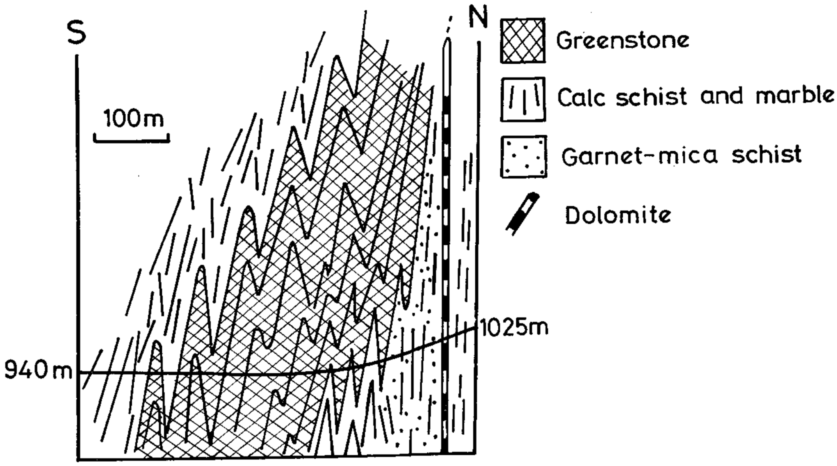


Fig. IV - 7 Section showing the structural relations of greenstones in the Rieken Graben; horizontally-trending line indicates height of the river-bed. Peripheral Schieferhülle.

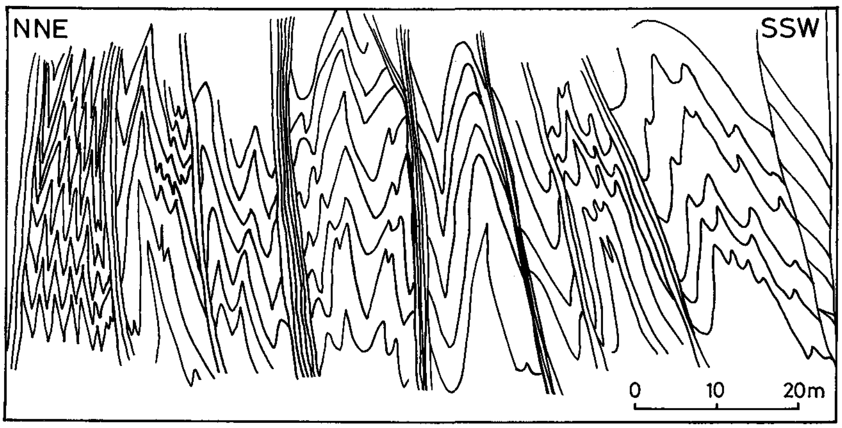


Fig. IV - 8 Structures in calc-schists and marbles along the east side of the Tabor Graben; note vertical shear zones, Peripheral Schieferhülle.

Lithological mapping within the Peripheral Schieferhülle has failed to reveal the character of any large scale structures. Only two lithological groups exhibit much lateral continuity; these are the calcareous group and the mica-schist group, which may be traced the length of the Mölltal. However, these groups are themselves composed of a variety of lithologies, such as marbles, black phyllites and different types of mica-schist, which individually are laterally discontinuous. Also, the outcrop patterns of these two units, although semi-continuous, are very difficult to relate to any large-scale structures which are consistent with the observed small-scale structures. The other lithologies are discontinuous laterally. On a large scale the dolomite-quartzite breccia and associated quartzites form a "pinch and swell" structure, breaking up further south into a series of lenses. The greenstones occur as a series of layers and lenses within both calc- and mica-schists. The Weissschiefer clearly shows rapid variations in thickness along strike having the form of large lens-shaped bodies. Although the serpentinites tend to occur along strike from each other at about the same position in the sequence (see later discussion), they are clearly in the form of small pods, and may have been introduced tectonically. The continuity of strata within the Peripheral

Schieferhülle of the map area therefore, appears to have been severely disrupted. A similar conclusion may be drawn for many of the lithological units in the Peripheral Schieferhülle of the Mallnitzer Mulde in the Sonnblick Group (EXNER, 1964). FRASL (1958) and FRASL and FRANK (1966) have put forward a stratigraphic scheme for the Mesozoic Schieferhülle based on work in the Glockner Depression. They recognise a basal Permo-Trias unit comprising phengite quartzites, dolomite and Rauhwacke. These lithologies, however, are not found at the base of the Peripheral Schieferhülle in the present area. Here the contact with the basement has clearly been the site of much shearing and is characterised by a somewhat mineralised zone of phyllonite up to several meters thick, and containing occasional slivers of basement rocks. At some localities, however, thin discontinuous layers of dolomite do occur within lithologies typical of FRASL's Jurassic Bündnerschiefer series. FRASL (1958) has advised caution in interpreting all lenses of dolomite as tectonically emplaced slivers of Trias, and suggests that many such bodies of dolomite within the Bündnerschiefer could be due to a resedimentation of Triassic dolomite during the Jurassic. In the present area, the amount of deformation is so great that sedimentary relationships are not preserved on a mesoscopic scale, and it is generally not possible to determine whether the dolomite slivers are autochthonous or allochthonous with respect to the surrounding rocks. The contact between the mica-schists and calc-schists in the Riekengraben (see previous discussion), where one such sliver of dolomite occurs, is, however, almost certainly a zone of relative movement. In the Mallnitzer Mulde in the Sonnblick Group, EXNER (1964) has mapped similar thin layers of dolomite which are in this case associated with gneiss lamellae. If the gneiss lamellae are accepted as thin slivers of basement, then these dolomite occurrences are also associated with major zones of movement.

In the lower Mölltal, we have the succession from the Hochalm Kern to the Sonnblick Lamella. To the north, this sequence contains up to four other subsidiary gneiss lamellae. Although these pinch out progressively southwards, it is difficult to see how the tectonic dislocations they represent can do likewise. This suggests that major dislocations occur within the Peripheral Schieferhülle on a macroscopic scale, in addition to the abundant structural discontinuities observed on a mesoscopic scale.

Taking the lower Mölltal succession as a whole, its range of lithologies most closely corresponds to the Bündnerschiefer in Glockner Facies (EXNER, 1964; FRASL and FRANK, 1966). According to FRASL (1958), the calc-schists, dolomite-quartzite breccia, and marbles occur in the lower part of the succession, the breccia being Liassic in age. The greenstones and mica-schists form the upper part and probably represent the Upper Jurassic. The serpentinite pods could have been emplaced at practically any time after this, but FRASL (1958) notes that they always seem to occur at approximately the same 'horizon', i.e. within the calc-schists, above the dolomite-quartzite breccia and just below the greenstones. However, if this is the case, then the Peripheral Schieferhülle of the Lower Mölltal must at the very least, be totally inverted. The same applies to the sections across the Mallnitzer Mulde in the Sonnblick Group (EXNER, 1964) where again, serpentinite and mica-schist underlie the calcareous lithologies. In the Sonnblick area, however, a large amount of greenstone is associated in a most complex fashion with the calc-schists; this is inconsistent with FRASL's scheme in which the major part of the greenstones is later than the calcareous lithologies.

During deformation, an original stratigraphy may become increasingly obliterated. Folding which does not involve a large amount of shearing and disruption of beds will still preserve the stratigraphic sequence, although it may possibly invert or repeat it. Deformation leading to a disruption of the sequence may be considered as having two components. One of these is a non-translational homogeneous strain, i.e. a stretching and flattening accompanied by boudinage of the more competent horizons within the less competent lithologies. The other component is a translational inhomogeneous strain involving disruption of the sequence into a series of sub-units, between which differential movement occurs. The degree to which this latter process destroys the stratigraphic relationships depends upon the size of the sub-units, the magnitude of the movements between them, and the scale at which the stratigraphic sequence is being considered. In the case of the Peripheral Schieferhülle we are dealing with a section from Permo-Trias to Cretaceous and extending about 40 kilometres laterally. Figure IV—9 is an attempt to show diagrammatically a possible sequence for the pro-

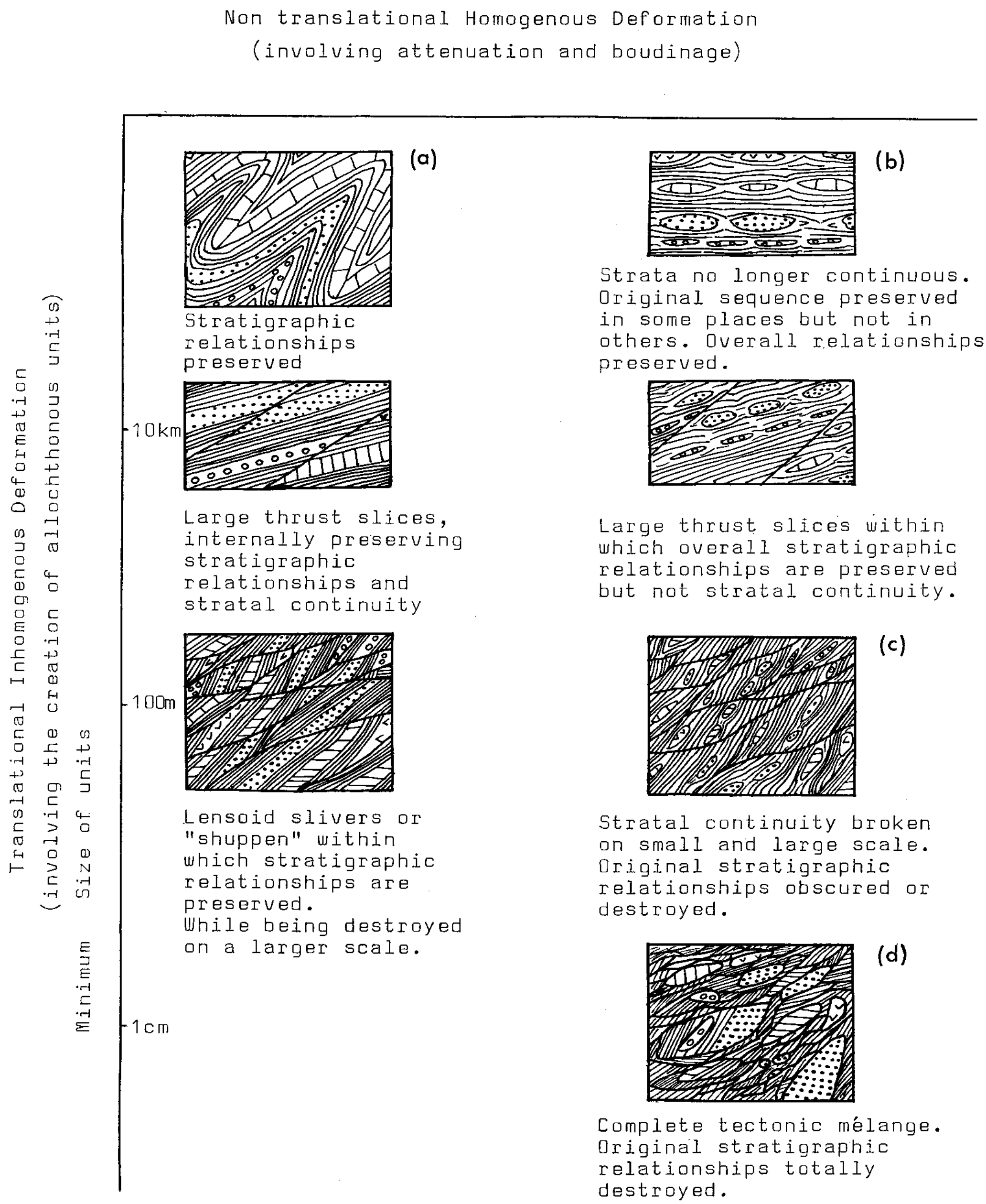


Fig. IV - 9 Diagrammatic representation of the disruption of original stratigraphic relationships. See text.

gressive disruption of stratigraphic relationships. States a, b, c, d, from the diagram will be considered.

(a) The original stratigraphic relationships are preserved.

(b) The original stratigraphic relationships are modified by deformation, so that the more competent rocks are boudinaged within a less competent matrix producing a series of broken strata. On a small scale, the sequence may be disrupted, but the overall relationships are still preserved.

(c) The competent strata are boudinaged and broken as in (b), but in addition, major dislocations break the sequence up into a number of large lensoid units. These units are tectonically intercalated, so that the sequence is no longer continuous on a large scale. Due to the homogeneous strain, the sequence is also discontinuous within the units, so that the original stratigraphic relationships are obscured or obliterated.

(d) Large and small lensoid fragments of the original rocks are tectonically intermixed within a ductile matrix, producing a tectonic mélange in which the original stratigraphy has been totally destroyed (as described by HSU (1968) and HSU and OHRBOM [1969] in the Franciscan rocks of California).

In the first two cases, the usual rules of original stratal continuity and stratigraphic superposition still apply. Where such assumptions are valid, stratigraphic mapping is useful in determining the overall structure even in complicated areas (e.g. the Moine series of Scotland [LAMBERT, 1959; LAMBERT and POOLE, 1964]). In the case of (c) and (d), however, these 'laws' are not valid, as pointed out by HSU (1968).

The evidence discussed earlier would suggest that the Peripheral Schieferhülle within the area has suffered a large homogeneous strain leading to boudinage and disruption of the more competent lithologies, and in addition, structural discontinuities are abundant on both macroscopic and mesoscopic scales, so that in the Lower Mölltal at least, it no longer exhibits stratal continuity or normal stratal superposition. A distinction between cases (c) and (d) would be difficult to make due to the large amount of deformation; however, the Peripheral Schieferhülle of the Lower Mölltal cannot be viewed as an autochthonous sequence folded and deformed *in situ* while preserving the original stratigraphic relationships.

Basement

Mesoscopic Scale

As described previously, the major part of the Inner Schieferhülle exhibits a compositional banding of one sort or another, and this forms the principal planar feature in outcrop. In most places, the basal planes of the platy minerals and the long axes of quartz and epidote pods lie within this foliation, so that the mineral foliation and compositional banding are parallel. At a few localities, isoclinal hinges are seen within the banding (e.g. Fig. II—6; see page 152). The amplitude of these folds is generally many times the wavelength. In most cases, however, the hinges have become isolated from the limbs thus producing the present banding parallel to their axial planes (cf. BALK, 1936). In most places, therefore, the principal planar feature is a compositional banding, formed in its present orientation by the rotation of an earlier banding in the limbs of isoclinal folds. The isoclinal folds range from a few centimetres in amplitude to several metres.

The most abundant folds seen on a mesoscopic scale are asymmetric tight to open folds with subhorizontal axes (Figs IV—10, 11; see page 211). The profiles of the folds frequently vary considerably along the axial surface, and layers are often folded disharmonically suggesting a buckling mechanism. This conclusion is supported by measurements of the orthogonal thickness of the folded layers in the manner suggested by RAMSAY (1967); on this basis, the majority contain layers belonging to class 1C of RAMSAY, suggesting that such folds contain a buckling component. The shallow limbs

Fig. IV - 10 $F^1{}_A$ folds in banded gneisses with amphibolites near the head of the Göss Graben; note that the lower limb of the fold is sheared out, along the line of the hammer head; note also disharmonic elements. Inner Schieferhülle.

Fig. IV - 11 $F^1{}_A$ fold in fine banded gneisses near the Schönangersee; note the marked change in fold amplitude along the axial surface; the field of view is about 10 *m. high.* Inner Schieferhülle.

are thinner than the steep limbs, so that the axial surface, defined as the locus of the hinge lines, does not bisect the interlimb angle but is closer to the shallow limb (Fig. IV—12). In the shallow limb, therefore, the banding is close to the axial plane

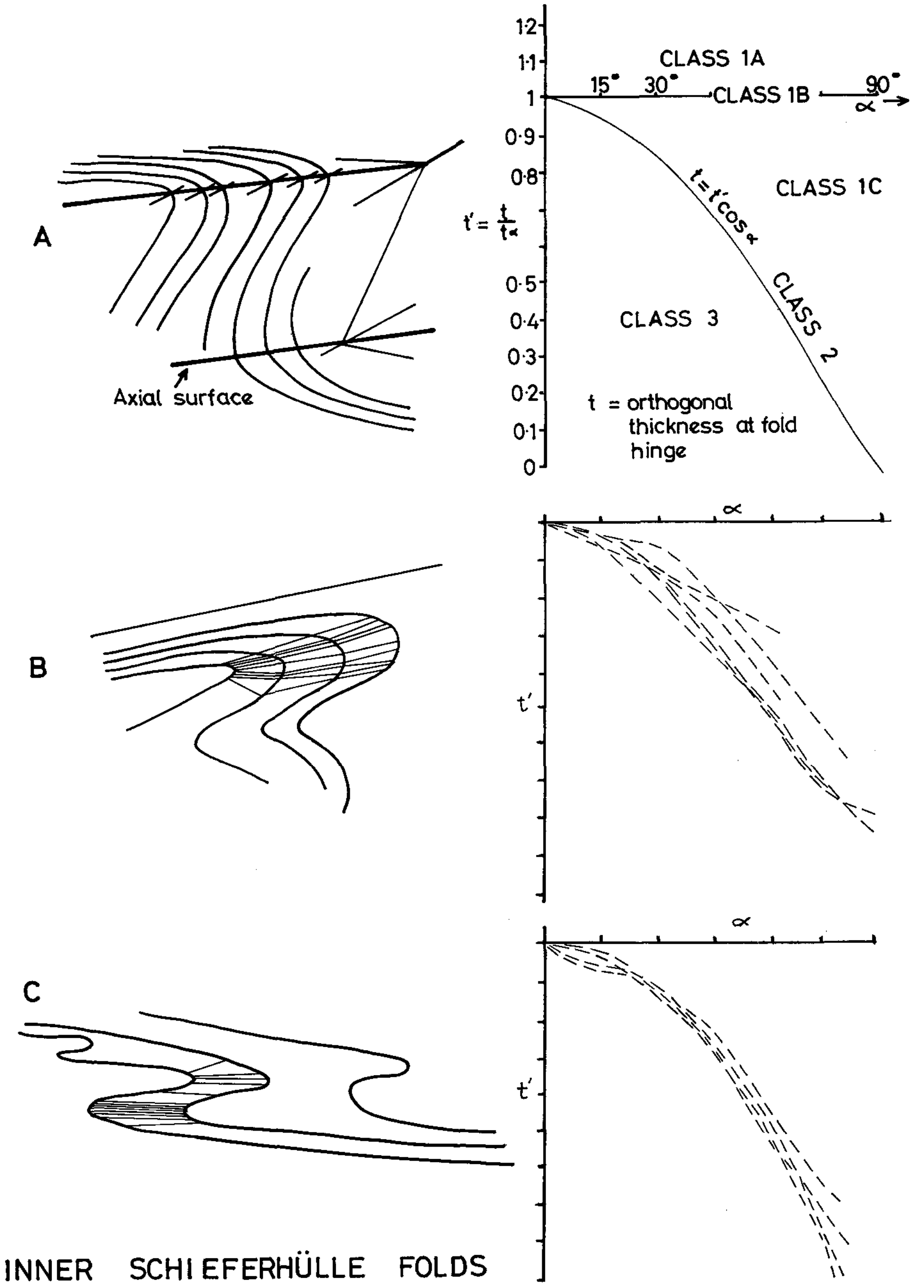


Fig. IV - 12 Forms of $F^1{}_A$ folds in the Inner Schieferhülle. A shows the departure of the axial surface from the line bisecting the limbs (left) and on the right shows the fold classes of RAMSAY (1967); B and C (left) show examples of specific folds with isogons constructed and (right) plots of t′ against α after RAMSAY (1967) for these folds. $t\alpha$ = the orthogonal thickness of a bed at a point where the tangent to the bed makes an angle, α, with the normal to the axial surface of the fold.

in orientation; in the steep limb, however, it is usually folded into minor folds (M-folds of RAMSAY, 1962 a) and crenulations. In the micaceous lithologies, a crenulation or strain-slip foliation is formed parallel to the axial surfaces of the micro-folds, and a penetrative linear feature is generated by the hinge lines. With increasing deformation, the micas are rotated and recrystallised into the new foliation and the finer banding is disrupted. This intense crumpling in the steep limbs and hinge zones of the folds produces a mullion structure or rodding parallel to the fold axis (cf. WILSON, 1953; TURNER and WEISS, 1963). The hinge zones of the larger folds are frequently characterised by a strong development of this structure, due to a disruption of bands and a transposition of the fragments into the later foliation.

In several localities, these later folds are seen to refold the early isoclines within the principal foliation, producing an interference pattern of Type III (RAMSAY, 1962 b; 1967). This suggests that the orientations of the two sets of fold axes were sub-parallel. Good examples of fold superposition are found on the south side of the Riekener Sonnblick, and in the amphibolites and banded lithologies to the west of the Kleine Mühldorfersee. There is occasionally a linear element (e.g. a mineral lineation) associated with the early folds. In the amphibolites especially, amphibole prisms sometimes exhibit a preferred orientation which is folded by later folds. These early lineations are poorly preserved, and it has not been possible to collect sufficient measurements from any one outcrop to establish their geometry completely; it is clear, however, that the deformed lineations do not lie within a plane, but tend to scatter around the later axial direction (Fig. IV—13). The earlier folds are generally best preserved within the steep limbs of the later folds where the respective axial surfaces are at a high angle so that the early folds are refolded, rather than merely tightened and sheared out. In a few examples, two generations of early isoclinal folds, both refolded by the later asymmetric folds, are noted. Thus the early group of folds may contain folds of several generations; however, because they are of a similar style and are rarely observed together, it is impracticable to distinguish them.

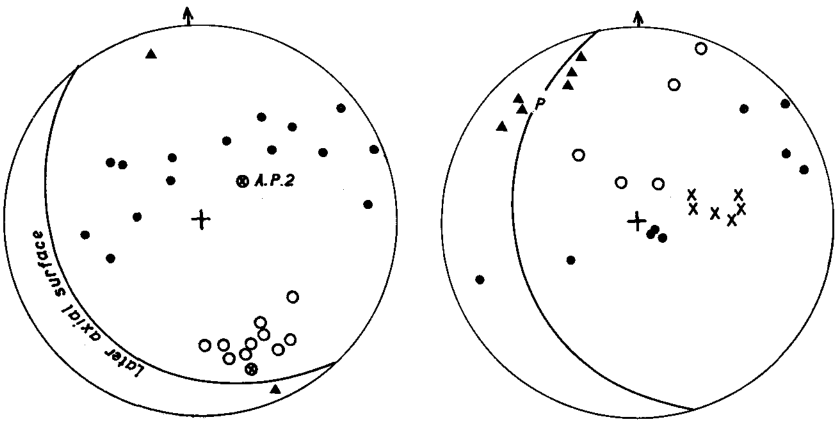


Fig. IV - 13 Equal area stereographic plots of refolded mineral lineations in the Inner Schieferhülle. Old lineations, Lv, are shown by circles; the axial surfaces of the F^1A folds folding Lv are shown by solid lines, and in the plot on the left the pole to the axial surface is marked A. P. 2; L^1A lineations and fold axes are shown by triangles; poles to the pre-Alpine compositional banding are shown by dots, and poles to the F^1A axial plane schistosity by crosses. For the fold on the left, Lv is close to $L^1A = \beta$ (circle with cross). In that on the right, Lv is much more scattered.

The Zentralgneis occurs both as large and small bodies. The principal fabric elements on a mesoscopic scale are a mineral foliation and a lineation of varying intensity. The foliation is defined by a preferred orientation of the micas and by flattened aggregates of quartz and feldspar. A lineation is developed where the mica flakes are variable in attitude but preserve a common line of intersection. Thus all gradations between lineated and foliated rocks occur. Mica lineations of this kind may be associated with mineral aggregates of a more elongate shape.

In the larger areas of granite gneiss (e.g. Pfaffenbergerkar, Gössgraben Kern and Villacherhütte area), the rocks chiefly contain a shallow to moderately dipping mineral foliation, together with an associated lineation. On a regional scale, this foliation is parallel to the axial surfaces of the later folds in the surrounding areas of Inner Schieferhülle, and the associated lineation to their axial direction. In very few places is there any evidence of an earlier phase of deformation in the granite gneisses, and the same fabric orientation is shared both by strongly deformed gneisses, and rocks still preserving a granitic texture. Occasional patches of augen-gneiss within the granite-gneisses, however, show clear evidence of a well-developed earlier foliation, which is absent in the surrounding rock, and which is rippled or microfolded by the later fabric. The margins of many of the smaller bodies, veins and lenses of leucocratic granite-gneiss and leucogranite intruding the Inner Schieferhülle, are discordant to the compositional banding in the latter and veins and apophyses intrude across it. In one or two examples, they are also seen to be discordant to the early isoclines within this banding. Nowhere have veins from the granite-gneisses been seen defining early folds within the foliation of the banded lithologies. However, smaller intrusive bodies and veins are clearly folded together with the banded lithologies into the asymmetric folds with horizontal axes representing the later phase of folding in the Inner Schieferhülle. On a larger scale also, the bodies of granite-gneiss are demonstrably affected by this set of folds. These granite-gneisses are known on isotopic evidence to be of pre-Alpine and probably Variscan age (Chapter V), and on field evidence to be younger than the early set of folds; the latter, which will be hereafter termed F_V, are therefore pre-intrusion and hence pre-Alpine, whereas the younger set of buckle folds (termed $F_A{}^1$ after the principles discussed at the beginning of this chapter) are post-intrusion, and thus may be Alpine in age. The penetrative fabric developed in most of the Zentralgneis may be correlated with this later phase and together with related planar and linear features in the Inner Schieferhülle, will be termed $S_A{}^1$ and $L_A{}^1$ for foliation and lineation respectively.

In the smaller bodies and veins of granite-gneiss, the foliation tends to form parallel to the margins. Where these are folded, the foliation in the outer arc of the fold develops parallel to the margin, passes into a lineation parallel to the fold axis in the centre of the hinge zone, and becomes oriented parallel to the axial surface in the core of the fold. If mineral foliations develop parallel to the plane of maximum elongation (i.e. the XY plane of the strain ellipsoid), then the strain distribution closely resembles that determined theoretically for a buckled viscous layer in a less viscous medium (Ramsay, 1967; Dieterich and Carter, 1969; Dieterich, 1969) and established from the results of experiments on model materials (Mukhopadhyay, 1965).

Discontinuities and inhomogeneities in the pattern of deformation are abundant within the Zentralgneis on all scales. These may take a variety of forms, but in general terms they are all represented by shear zones along which there is an intensification of the foliation, a shearing out of veins, bands, etc. in the adjacent lithologies, and frequently growth of a new mineral assemblage (Chapter III). In some cases such zones may separate areas which have very similar deformational fabrics, e.g. zones with a very intense development of foliation frequently transect the granite-gneiss terranes. In other cases, however, the deformation pattern may be radically different on either side of the zone, e.g. such a sheared zone is usually present at the contacts of the granite-gneisses with the Inner Schieferhülle; the granite- gneiss may be homogeneously deformed with little or no folding whereas the Inner Schieferhülle lithologies may be highly crumpled and sheared. Within the latter, some shear zones may be clearly recognisable as thrust faults (Fig. IV—14; see page 215); in other cases, any indications of relative movement may be obscured. In all of these various cases, the

Fig. IV - 14 Shear zone with "floating" isoclinal hinges forming the lower limb of an F^1. fold in banded gneisses of the Göss Graben. Hammer for scale. Inner Schieferhülle.

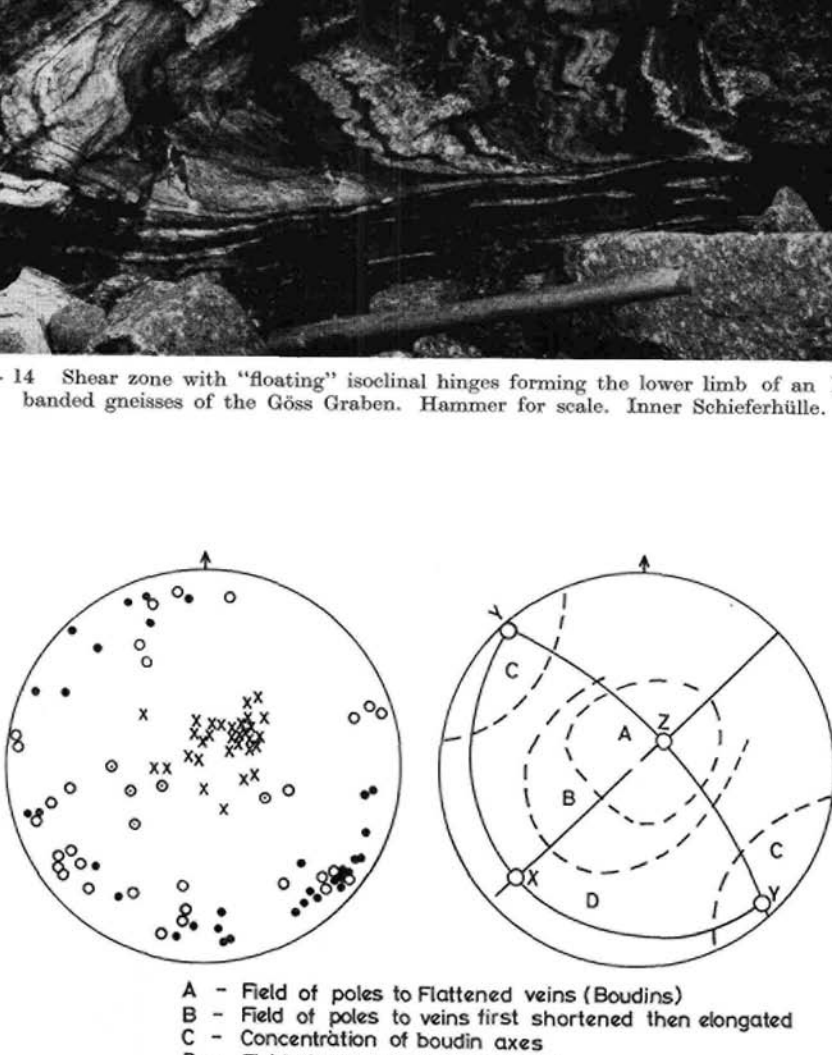


Fig. IV - 15 Left (a): an equal-area stereogram of deformed veins from the Reisseck synform; crosses are poles to boudinaged veins; circles are poles to buckled veins; dots within circles are poles to boudinaged and buckled veins and dots are boudin axes. Right (b): suggested interpretation of vein orientation observations; X, Y and Z are the inferred regional strain axes.

thrust plane, shear zone, crush zone, etc., represents a break or discontinuity in the strain pattern. Thus the term "strain discontinuity" will be used for all such zones, irrespective of their precise mode of origin. Consideration of such discontinuities is important in assessing the overall distribution of strain within the area.

The strain within the basement is clearly inhomogeneous on a large scale, as igneous textures and contact relationships are preserved within the Zentralgneis in some places and more or less obliterated in others. The Inner Schieferhülle on the other hand usually shows abundant mesoscopic $F_A{}^1$ folds and is strongly deformed, and much of the strain within the basement as a whole seems to have been taken up by the Inner Schieferhülle. Within the larger masses of Zentralgneis, the deformation is usually concentrated in the marginal areas and along irregularly distributed strain discontinuities; thus in the centre of such bodies (e.g. the Pfaffenbergerkar) intrusive relationships and igneous textures may be preserved relatively undeformed.

TABLE IV—1

Strain measurements on buckled veins (See Appendix II)

Locality	Rock Types		(natural strain = $\log \sqrt{\lambda.}$)		R (η/η)	
	vein	host	without allowance for initial shortening	after allowance for initial shortening	Before	After
(1) Hochkedl ridge	Aplite	Bi-qz schistose gneiss	—0.75	—1.29	2.3	ca. 12
(2) Obere Mooshütte	Aplite	Bi-qz gneiss	—0.4	—0.86	4.7	ca. 12
(3) -do-	Aplite	-do-	—0.3	—0.84	3.3	ca. 12
(4) -do-	Aplite	-do-	—0.42	—0.92	2.8	ca. 12
(5) Stapniksee	Aplite	Bi-qz gneiss	—0.39	—1.17	ca. 1	ca. 5
(6) Zwenberger-kar	Aplite	leucogranite gneiss	—0.62	—1.20	ca. 1.8	ca. 8
(7) Hochedl ridge	Aplite	Bi-qz schistose gneiss	—0.52	—1.14	ca. 1.8	ca. 10
(8) -do-	Aplite	-do-	—0.57	—1.14	ca. 1.8	ca. 9
(9) -do-	Aplite	-do-	—0.86	—1.20	ca. 6	ca. 14
(10) Villacher-hütte area	Pegmatite	Tonalite	—0.61	—1.21	ca. 1.8	ca. 9
(11) Stapniksee	Aplite	Hornblende mica schist	—0.67	—1.13	ca. 9	ca. 20—25
(12) Riekener Sonnblick ridge	Qz rich layer	Banded gneiss	—0.84			
(13) Top of Hinterregen Graben	Qz rich layer	Grey banded gneiss	—1.05			
(14) Riekener Sonnblick	Qz vein	Amphibolite	—0.72			
(15) Stapnik summit	Aplite	Hornblende schist	—0.74			

Few potential strain indicators, such as deformed pebbles, are present within the basement. There are, however a number of deformed veins of aplite and microgranite, particularly within the banded gneisses and amphibolites. Several authors have suggested ways in which such veins may be analysed in order to provide information about the total strain of the rock (e.g. RAMBERG, 1959; FLINN, 1962; WATTERSON, 1968; RAMSAY, 1967). Veins may be divided into those which have only been shortened, those which have been shortened and then elongated, and those which have only been elongated. By plotting poles to these veins, the fields of shortening and elongation may be determined, and from this the orientation of the finite strain ellipsoid derived. This method can only be applied with any precision for volumes of rock within which the strain is homogeneous at the scale of the strain indicator. In the present area, as discussed earlier, the strain within the basement is not homogeneous. However, in the same way that geometrically inhomogeneous structures may be statistically homogeneous if analysed over a large area, strain axes may exhibit a preferred orientation over a larger area, although in detail they may be inhomogeneous. Figure IV—15 a is an equal area plot of all deformed veins measured in the synformal structure (Reisseck synform) extending from the Hinterregen Graben to the Zwenbergertörl. The poles to buckled veins are clearly concentrated around the margins of the net, the poles to boudinaged veins in the centre, and the extension directions of boudinaged veins perpendicular to the regional fold axis. The suggested strain axes for the F_A^1 deformation are shown in Fig. IV—15 b. The mean axial surface orientation is perpendicular to the Z-axis, and the mean fold axial direction parallel to the Y-axis. Boudinaged veins may be extended in various directions within the XY plane, showing that extension may occur parallel to the Y-axis. However, the boudins within the Inner Schieferhülle are usually axially symmetric structures, about an axis parallel to the local fold axis, suggesting that the strain was not an uniaxial flattening.

TABLE IV—2

Strain measurements on boudinaged veins (See Appendix II)

Locality	Rock types	Natural strain, ε, perpendicular to plane of boudinage	
		less homogeneous strain	with homogeneous strain
(1) Hochkedl ridge	Qz boudins in Qz/Bi schistose gneiss	—0.76	—0.84
(2) Reisseck	Aplite in banded gneiss	—0.83	—1.06
(3) Riedbock	Aplite boudins in Grey banded gneiss	—0.44	—0.54
(4) Above Hohenbahn Terminal	Aplite boudins in banded gneiss	—0.69	—0.85
(5) Riekener Sonnblick	Qz boudins in Amphibolite	—0.62	—0.76
(6) N. of Gamolnig Spitze	Aplite boudins in Amphibolite	—0.77	—0.95
(7) Riekengraben Randgneis	Qz boudins in Amphibolite		—2.30
(8) Riekengraben Randgneis	Aplite boudins in fine banded gneiss	—1.38	—1.70
(9) —do—	—do—	—1.57	—1.93
(10) —do—	—do—	—1.70	—2.11
(11) —do—	—do—	—1.96	—2.40

Detailed localities for observations given in Norris (1970).

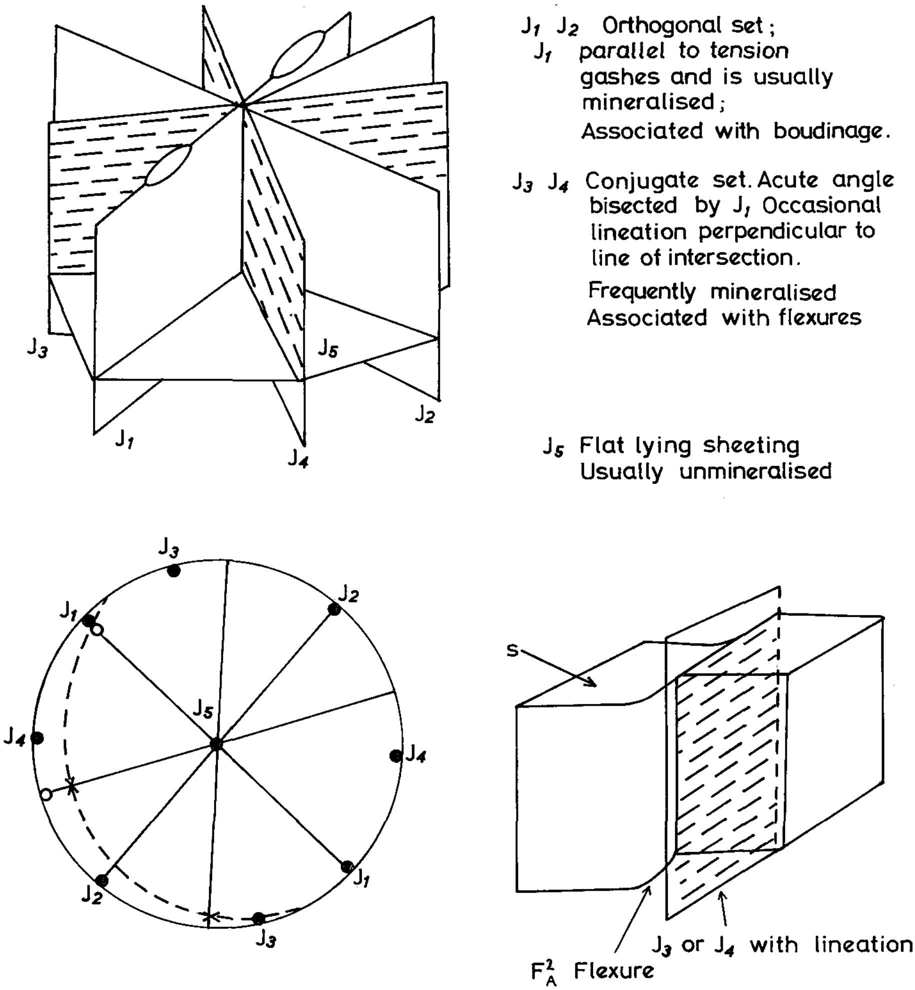
In addition to studying the orientations of deformed veins, it may also be possible to calculate the amount of shortening or elongation parallel to a vein. These studies require a number of assumptions, so that the results cannot be regarded as accurate measurements of strain, but nevertheless, give some idea of the order of magnitude. The mechanics of buckling layers of contrasting rheological properties have been discussed by RAMBERG (1959, 1964), BIOT (1961), CHAPPLE (1968, 1969) and SHERWIN and CHAPPLE (1968). Although several theories of the formation of ptygmatic veins have been put forward (KUENEN, 1968), it is now generally accepted that they form by compression of a more competent layer within a less competent matrix. Details on the method and results of strain calculations on buckled and boudinaged veins within the area are given in Appendix II. Tables IV—1,2 show the range of values obtained. Despite the uncertainties and the inhomogeneous nature of the strain, the results are surprisingly consistent, and suggest that the banded lithologies and amphibolites of the Inner Schieferhülle have undergone a flattening of between 40 and 70% ($\varepsilon = 0.5$ to 1.3) in the central area, and in excess of 80% ($\varepsilon = 1.7$ to 2.4) in the marginal zone near to the contact with the Peripheral Schieferhülle (Fig. AII—1).

A further set of mesoscopic-scale structures are found in both the Inner Schieferhülle and the Zentralgneis. These are a series of gentle flexures and kinks related to the development of joints and fractures (Fig. IV—16). Normally three or four vertical or nearly vertical fractures are developed, three of which are frequently quartz-filled. Evidence of movement, in the form of flexuring of the foliation and near horizontal slickensiding on the joint surfaces, is associated with two of the joints (J_3, J_4). These are usually at an angle of 60° to 90° to each other, and the acute angle is bisected by a third joint which is frequently in the form of quartz-filled tension gashes (J_1). The fourth joint (J_2) forms perpendicular to these gashes. A fifth joint, in the form of a shallowly dipping sheeting, is also fairly common, particularly in the Zentralgneis. This generation of structures will be referred to as FA^2.

Microscopic scale

In thin section, the basement rocks show more evidence of annealing than the Peripheral Schieferhülle. Deformation lamellae are very rarely found in quartz, and the plagioclase usually exhibits an equilibrium polygonal mosaic. The K-feldspar megacrysts in the augen-gneisses show abundant evidence of deformation and partial recrystallisation along shear zones within the grains. The larger amphibole grains in the amphibolites are also strained and partially recrystallised. In general, the amphiboles in the Inner Schieferhülle have a short prismatic habit and exhibit much weaker preferred orientation than those in the Peripheral Schieferhülle.

A number of analyses of lattice orientations of quartz and mica from granitic rocks of the Zentralgneis were made in order to determine their relationship to the mesoscopic structures. The results are shown in Figs IV—17 to 21. In most of the samples, the quartz exhibits a partially recrystallised habit (cf. CARTER et al., 1964; HOBBS, 1968) with smaller, equant, unstrained grains apparently recrystallising from the larger strained grains. Where possible, these have been plotted separately. The small, recrystallised grains have a weaker preferred orientation than the strained grains although the overall pattern is similar. However, individual maxima for the strained grains tend not to coincide with the c-axes of the later grains. This behaviour was noted in naturally deformed quartz crystals by PHILIPS (1965) and in experimentally recrystallised quartz by HOBBS (1968).



SUMMARY OF JOINT GEOMETRY ON A MESOSCOPIC SCALE

Fig. IV - 16 Joint relationships shown diagrammatically and on a stereogram. The numbering system J_1, J_2 etc. is intended only to distinguish different joint directions and not to imply an age sequence. See text for discussion.

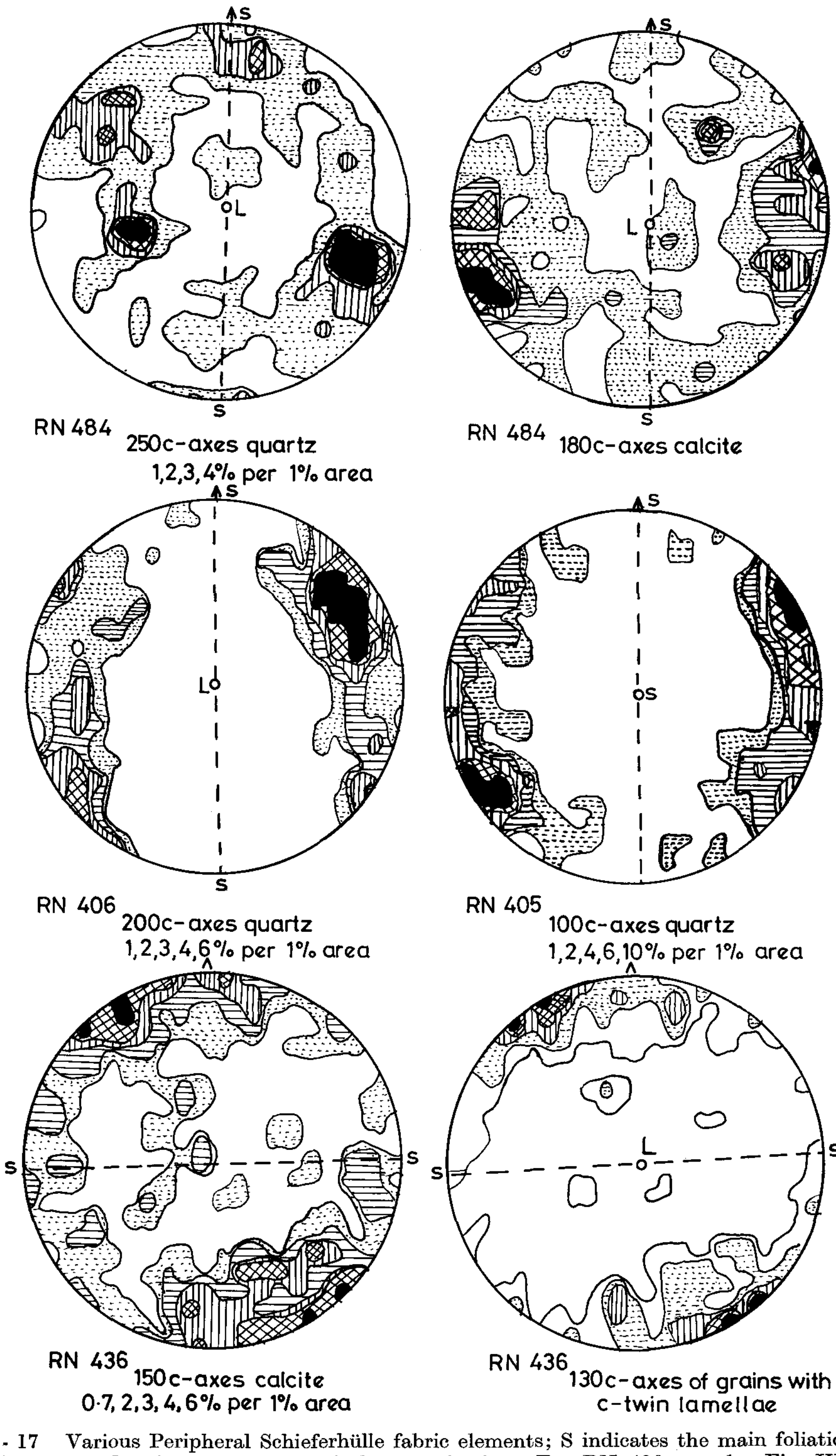


Fig. IV - 17 Various Peripheral Schieferhülle fabric elements; S indicates the main foliation and L the lineation. Equal area lower hemisphere projection. For RN 436 see also Fig. IV - 21.

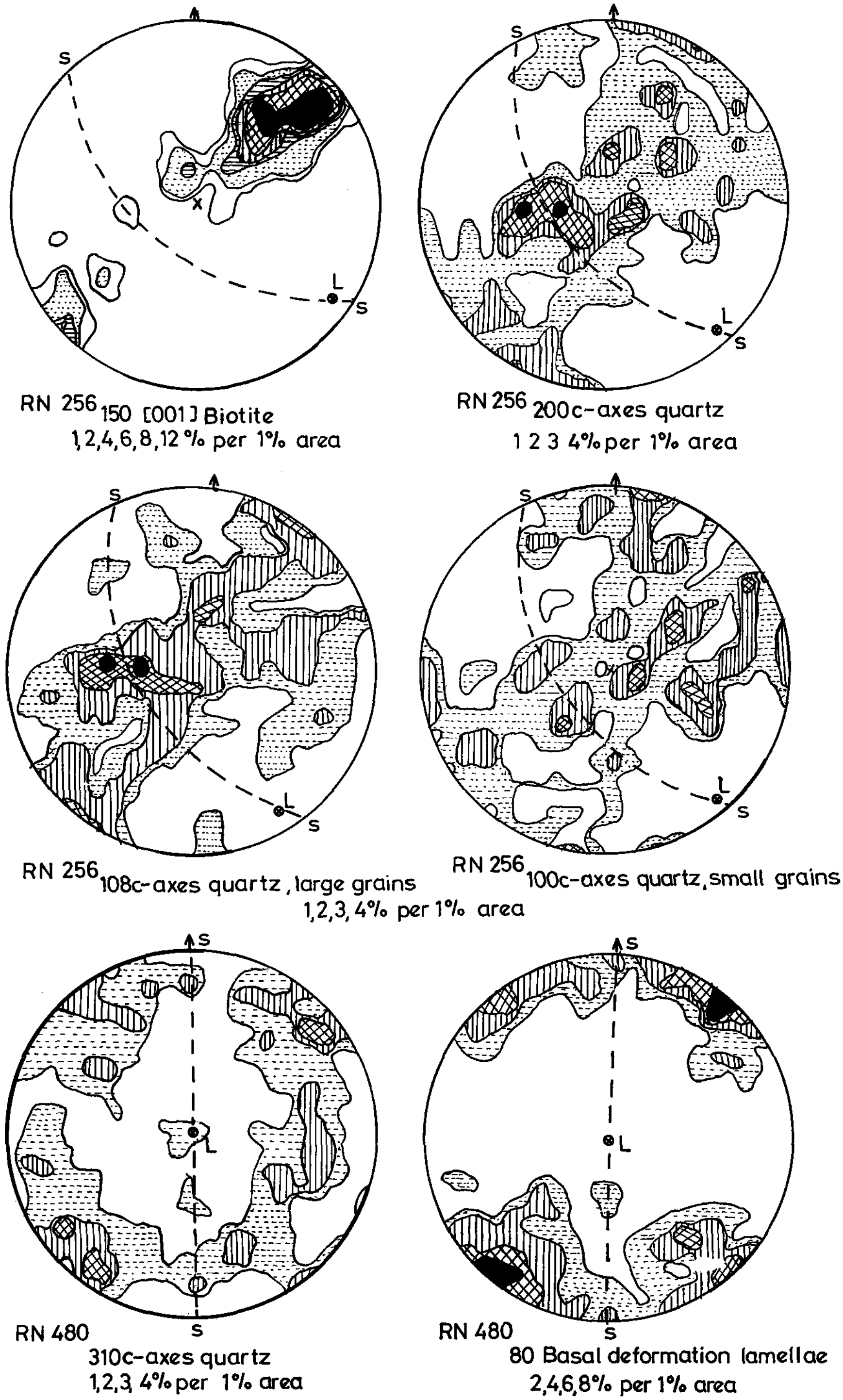


Fig. IV - 18 Various Zentralgneis (RN 256) and Peripheral Schieferhülle (RN 480) fabric elements; S indicates the main foliation and L the lineation.

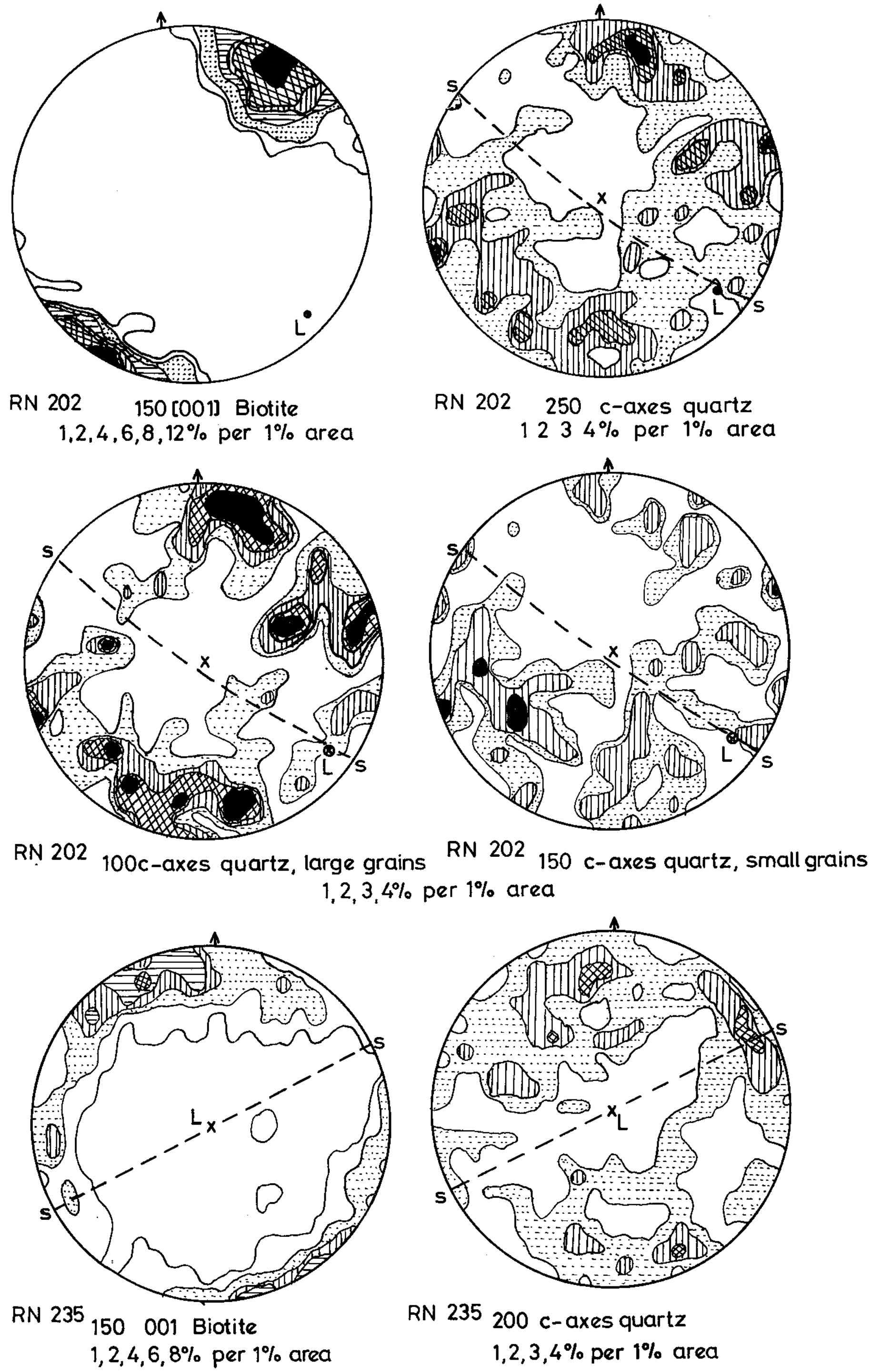


Fig. IV - 19 Various Zentralgneis fabric elements; S indicates the main foliation and L the lineation.

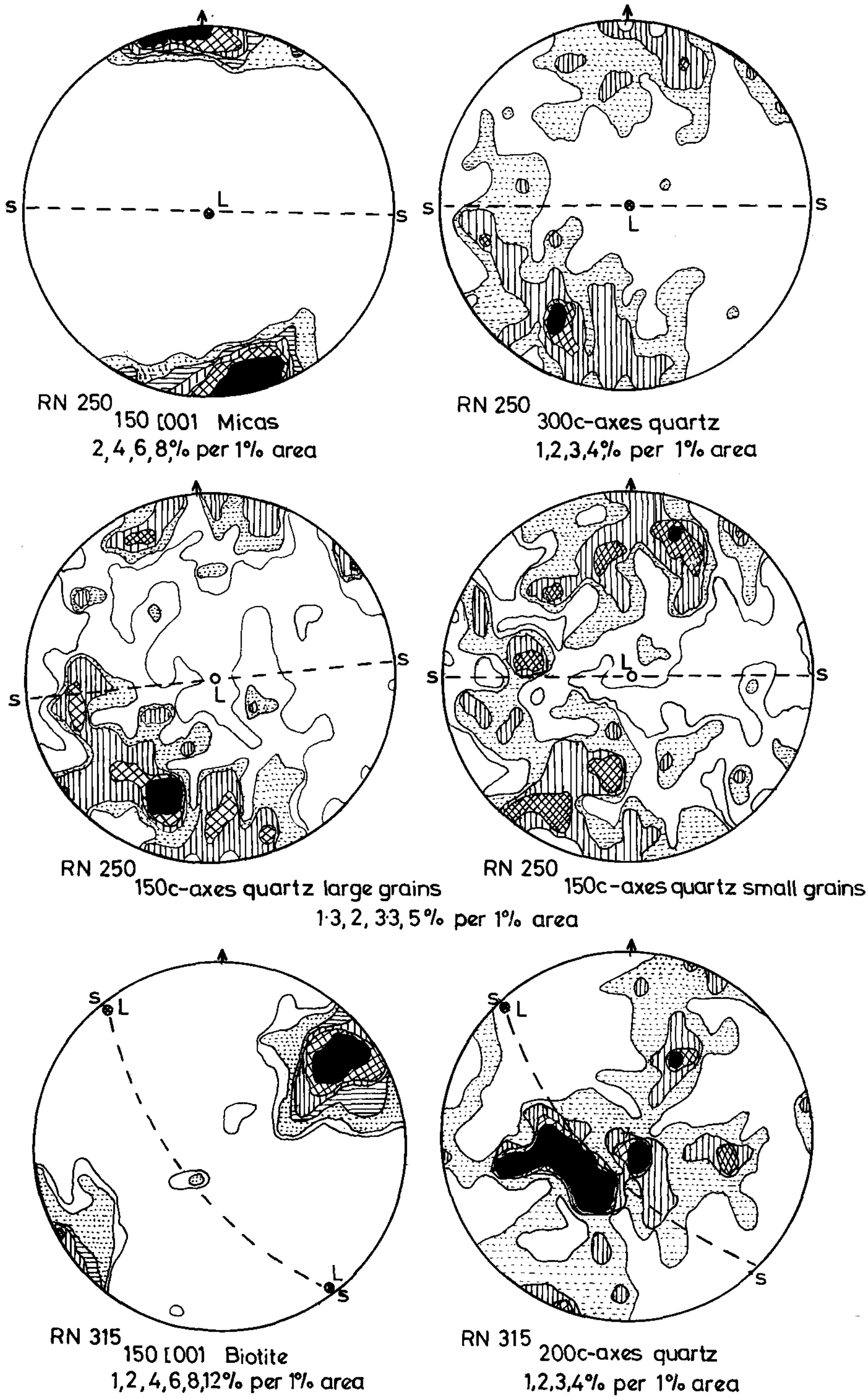
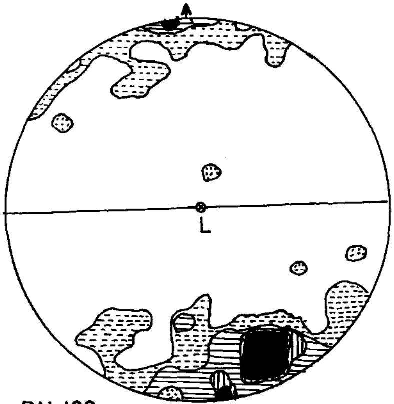
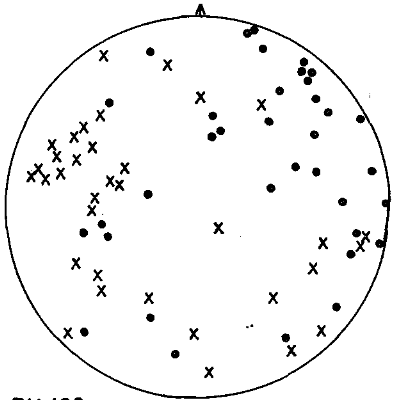


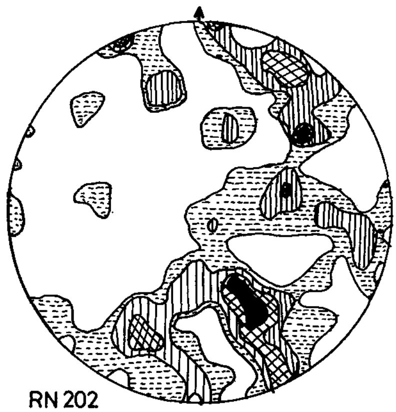
Fig. IV - 20 Various Zentralgneis fabric elements; S indicates the main foliation and L the lineation.



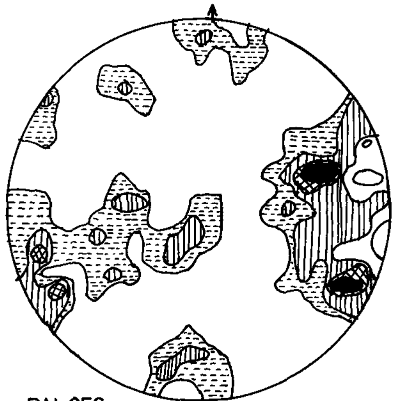
RN 436 97c-lamellae, calcite
2 4 6 8% per 1% area



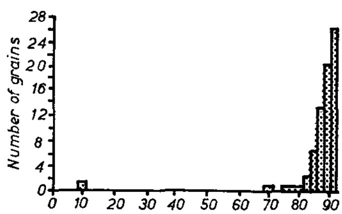
RN 436 Calculated compression(•) and
tension (x) axes from 40 grains



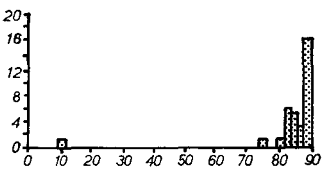
RN 202
77 poles to low angle sub-boundaries,
quartz 1·3, 2·7, 4, 5·3% per 1% area



RN 256
55 poles to low angle sub-boundaries,
quartz 2, 4, 6, 8% per 1% area



RN 202. Angle between c-axis
and pole to sub-boundary



RN 256. Angle between c-axis
and pole to sub-boundary

Fig. IV - 21 Various Zentralgneis (RN 202, RN 256) and Peripheral Schieferhülle (RN 436) fabric elements; S indicates the main foliation and L the lineation.

The c-axes are generally symmetrically related to the mesoscopic fabric elements. RN 235, 256, and 315 are rocks with a fairly well-developed mineral lineation. The c-axes are preferentially oriented in the plane perpendicular to this lineation (L). A maximum tends to form perpendicular to L within the foliation defined by the (001) planes of the micas. It is possible to interpret the patterns as crossed girdles, with maxima at the girdle intersections (cf. SYLVESTER and CHRISTIE [1968]), but it would be equally possible to describe them as large-radius, small-circles about the lineation, or simply as poorly defined single girdles.

The c-axis distribution for RN 250 is not quite symmetrical to L or S, as defined by the mica subfabric. A similar case was described from Anglesey by WEISS (1955) and ascribed to a slip direction lying oblique to the foliation. A similar explanation, not involving a slip mechanism, would be a situation in which the direction of maximum elongation (X-axis of the strain-ellipsoid) was oblique to the axis of rotation (the fold axis).

RN 202 has an S-fabric on a mesoscopic scale. The principal concentrations fall approximately on small circles about the poles to the foliations. There is no girdle distribution about the local fold axial direction, and the only girdle-like spread of the c-axes is about a direction within the foliation at right angles to this.

The results suggest that the quartz in the granitic lithologies was strained and recrystallised during and after the $F_A{}^1$ phase of folding.

The general pattern of preferred orientation developed in foliated and lineated rocks resembles those found in the Peripheral Schieferhülle, although they are less pronounced.

Macroscopic Scale

The principal macroscopic structural feature of the area is a dome-shaped structure with its culmination under the Reisseck. The centre of this structure is composed mainly of granite-gneisses of various types; biotite augen-gneiss is predominant in the area around the Mühldorferseen and the Radlsee, but is replaced by leucocratic granite-gneiss in the Gössgraben. This dome-shaped body of gneiss has been termed the Gössgraben Kern (EXNER, 1954). Included in the Gössgraben Kern is the thin unit of banded gneisses immediately overlying the granite-gneisses. This shows a similar dome-shaped structure and is not folded up into the overlying structures. The foliation in the gneiss and banded lithologies dips northwards in the Gössgraben, southwestwards around the Mühldorfersee area, southwards in the Hinterregengraben and eastwards on the Tandlspitze, thus defining the dome. Beneath the Reisseck, the foliation is horizontal.

The trend of the contact between augen-gneiss and the overlying banded gneisses is broadly parallel to the mineral foliation in the gneiss, but locally, it can be seen to be more irregular (e.g. on the west side of the Radlmauer. Tight $F_A{}^1$ minor folds are found within the banded gneisses, although they are not) well developed on a large scale. This is also illustrated by the plots of poles to foliation (Plate 4, subareas 12 and 13) which show very poorly developed girdles, but distinct maxima coincident with the poles to the axial surfaces of the $F_A{}^1$ folds. The $F_A{}^1$ fold axes and associated linear features are also rather dispersed within their axial planes.

The margins of the Gössgraben Kern are marked by a structural discontinuity. This is clearly seen at the south-east end of the Grosse Mühldorfersee (Fig. IV—22) and west of the Hochkedl, where augen-gneiss of a higher unit is thrust over the Kern, folding the marginal leucogranite and banded lithologies into a gently inclined synform, the thrust-plane taking the place of the upper limb. Minor structures in the rocks on either side of this thrust-plane suggest a relative movement of the overlying rocks to the north-east. On the north-west side of the Mühldorfer Graben, the thrust is

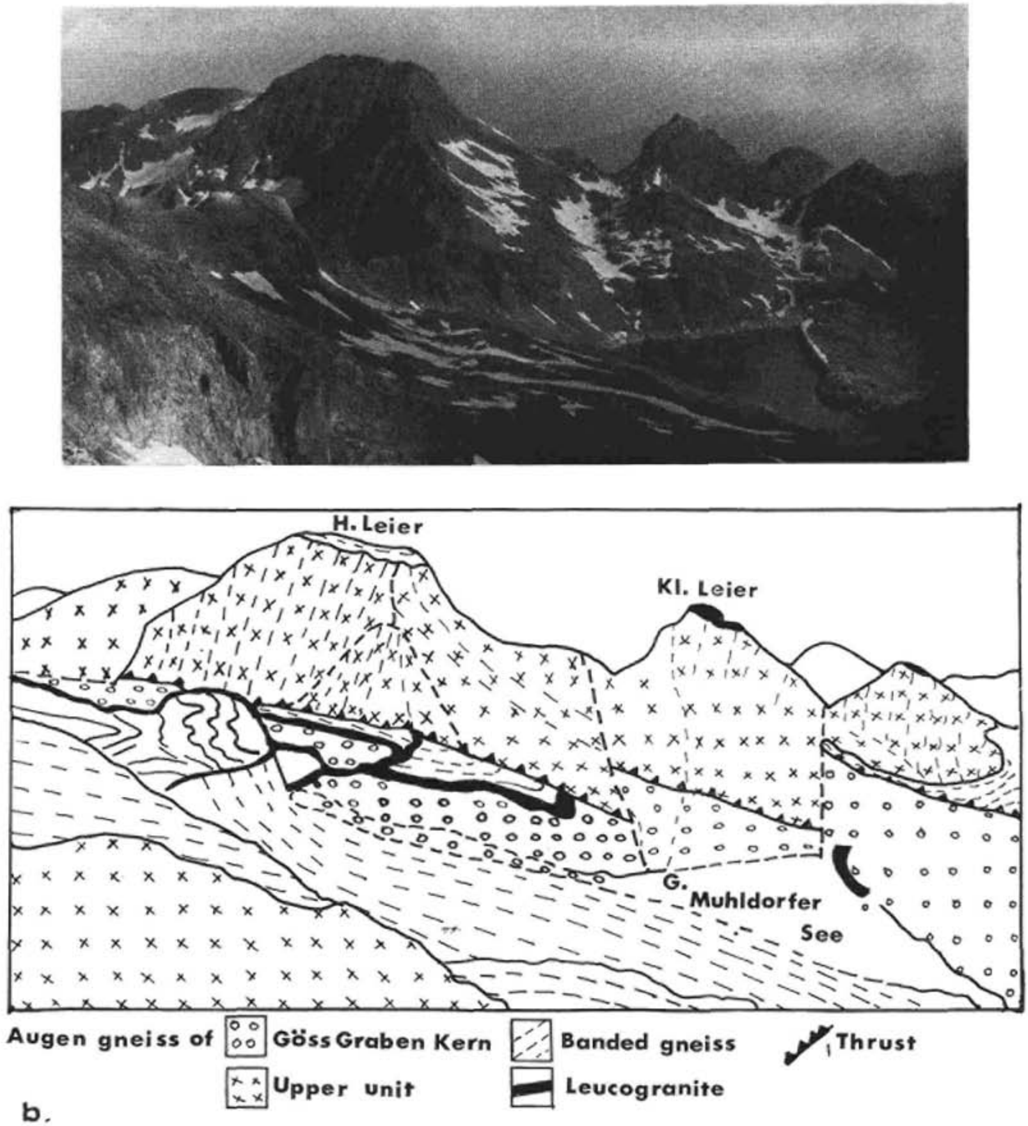


Fig. IV - 22 (a) — view of the Hohe Leier showing a shear-zone at upper contact of Göss Graben Kern; (b) — a geological sketch of the view in (a).

found at the base of the leucocratic granite-gneiss unit; further north, however, although a structural discontinuity is still present, it is no longer clearly identifiable as a thrust fault.

In the overlying rocks, a zone of Inner Schieferhülle extends from the Hinterregen Graben to the Villacher Hütte; this has been termed the Reisseck Mulde (cf. Exner, 1954). Within this unit, a zone of asymmetric $F_A{}^1$ folds, overturning to the north-east, runs northwards from the Upper Hinterregengraben to the Zwenberger Törl. To the east, the lithological contacts are flat-lying and show little macroscopic folding, as may be seen on the north face of the Reisseck and on the Tandlspitze. The folded zone corresponds to the splitting up of the large body of granite-gneiss forming the Kammwand and Gamolnigspitze, into a number of tongues and lensoid bodies intruding the Inner Schieferhülle, and here has the overall form of an asymmetric synform closing to the south-west. Because of the irregular shape of the granite bodies, and because they have behaved in a more competent fashion than the country rocks, the resulting fold geometry is inhomogeneous. The diagrams in Plate 4 show that although a girdle is defined by the poles to the foliation, there is a considerable scatter of fold axes even within the subareas. However, the general trend of the fold axes is between north-west and north and with axial surfaces dipping to the southwest and west at angles of between 0° and 35°.

This synform passes westwards into a complimentary antiform. On the west limb of this structure (subareas 6, 17 & 19, Plate 4), folds are seen rather infrequently, and tend towards being isoclinal so that the rocks develop a strong foliation parallel to the axial surfaces of the mesoscopic folds. This foliation steepens to to the west and at the contact with the Peripheral Schieferhülle, it may locally be close to vertical. The name Reisseck synform and Hochalm antiform are proposed for these structures. The steepening of the south-west limb of the Hochalm antiform also warps the axial surfaces of the $F_A{}^1$ folds and therefore presumably occurred after the initial development of these structures.

South of the Mühldorfer Graben (sub-areas 8, 9, Plate 4), the cascade of asymmetric folds representing the continuation of the Reisseck synform begins to die out and it is here partly defined by a large body of augen-gneiss, which appears to thin to the southwest and south (Fig. IV—23). The orientations of the minor fold axes gradually change from southeast to east, and the strike of the axial surfaces does likewise. The southwest limb of the structure is overlain by a large sheet of granite-gneisses. At the base of this unit and within it, numerous strain discontinuities occur, suggesting that it represents a thrust slice or series of slices, thrust over the Reisseck synform.

North of the Zwenberger Törl, the banded lithologies of the Reisseck Mulde are overlain and replaced by tonalite and granite-gneisses. Their foliation dips consistently 40° to the west, and the large scale antiformal and synformal structures appear to die out. In the Villacherhütte area, banded gneisses and mica-schists appear below the tonalite and are folded on east-west axes into a northwards-facing set of asymmetric folds. Clearly, the wedge of Inner Schieferhülle, the Reisseck Mulde, is not a simple synform. It is perhaps better thought of as a zone of folded, inhomogeneous, Inner Schieferhülle within which structures may die out or appear on a variety of trends. The direction of dip of the axial surfaces swings to the northwest in the Hohes Gösskar and to the north in the Villacherhütte area (Plate 4) consistent with a doming over the Gössgraben Kern.

At the contact of the schists and the tonalite in the Hohes Gösskar, an extremely complex zone resembling a tectonic mélange occurs, and is thought to represent a thrust contact. The lower contact of the tonalite in the Zwenbergergraben is perfectly sharp and concordant to the foliation, suggesting that it too is tectonic in origin. At the contact with the schists below the Villacherhütte a few metres of phyllonite occur and the tonalite becomes extremely sheared. Similar zones occur within the tonalite, and at its upper contact below the Preimlspitze, it is separated from the augen-gneiss by about 30 *m.* of intensely deformed, banded gneisses. Discordant veins in the tonalite become sheared out at the margins of the zone. These strain discontinuities are evidence of discontinuous deformation, and possibly thrusting, but the magnitude of the displacement cannot be ascertained. One difficulty is that the configuration of the original contact between the granitic rocks and the Inner Schieferhülle was probably rather complex; as there is an intrusive relationship between the two, there is no reason to believe that the latter everywhere initially overlay the former. Therefore although a considerable amount of shearing has occurred at the margins of the tonalite, it does not necessarily represent an allochthonous thrust-sheet.

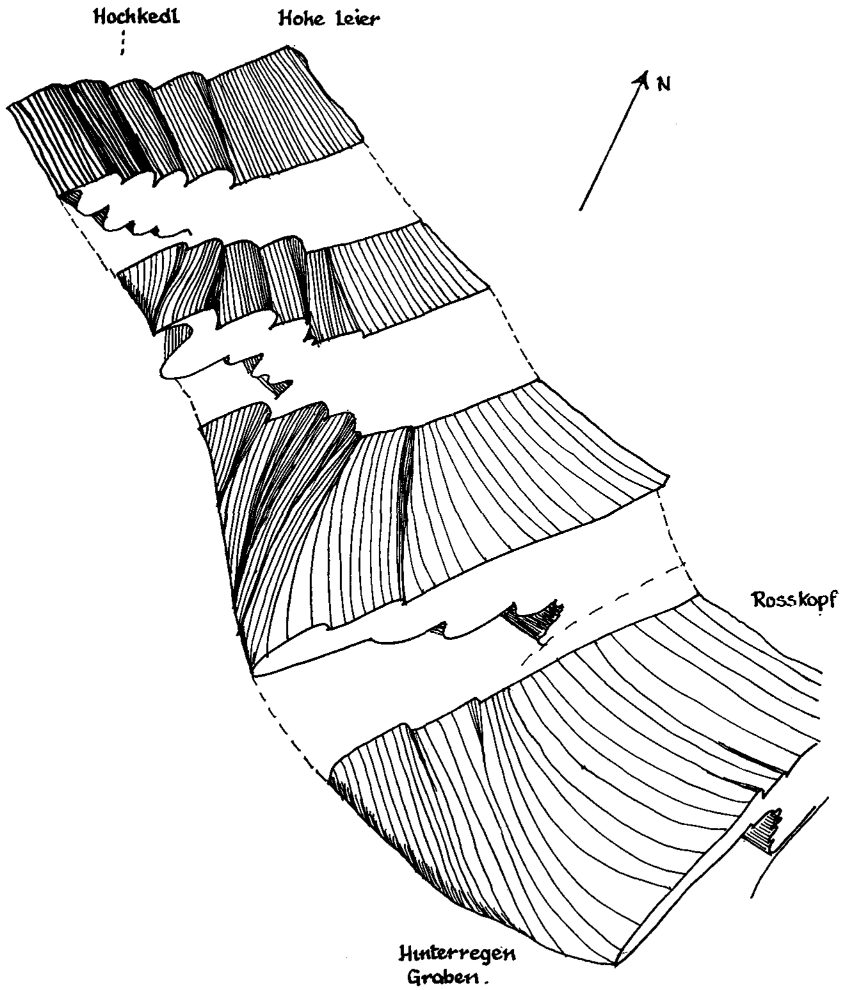


Fig. IV - 23 A three-dimensional drawing of the biotite augen-gneiss body which trends northwest from the upper Hinterregen Graben.

These considerations emphasise the fundamental difference between the "Mulden" involving Peripheral Schieferhulle, such as the Mallnitzermulde, and those involving Inner Schieferhülle. In the former case, the basement/cover contact can be assumed to have been initially sub-horizontal, so that any present day relief on this interface is attributable to the Alpine deformation. In the latter case, however, the contact was an intrusive one and probably highly irregular, so that the present shape represents the superposition of the Alpine deformation onto an initially complex configuration. Within such a "Mulde", structures may occur which are not continuous with each other and whose forms are largely governed by the initial complexities of the contacts. Used in this way, the term "Mulde" simply refers to a zone of Inner Schieferhülle without necessarily implying anything about the shape of its margins.

To summarise, the basement is a deformed intrusive complex, within which lithologies of different physical properties are irregularly distributed, and whose contacts were irregular before the onset of the $F_A{}^1$ deformation. The granitic rocks appear to have behaved in a more competent fashion than the Inner Schieferhülle and this has led to an inhomogenous strain and a complex geometry. Discontinuous deformation is prominent and strain discontinuities are abundant on all scales, particularly within the granitic rocks and around their margins.

Descriptive Synthesis

Plate 4 shows the variation in geometry of the mesoscopic features over the area.

The latest phase of deformation in both the basement and cover is related to the development of joints and faults. All but sub-areas 10, 11, 12 & 13 & 24 show the same joint pattern, except that in some sub-areas, certain directions are less well developed than in others. Two maxima (J_3 & J_4) represent planes intersecting at an angle of 60° to 90° and correspond to fractures in the field having evidence of displacement along them (such as flexuring, faulting, sub-horizontal slickensiding, etc.). These are bisected by another plane (J_1) corresponding to tension gashes. A fourth maximum (J_2) represents a plane at right angles to this last direction. This latter fracture is only rarely filled with quartz, and there is some field evidence of vertical movements along this fracture. The reason for its apparent non-development in the Peripheral Schieferhülle is probably that it is sub-parallel to the compositional banding and is not easily recognised. The same geometry is shared by both basement and cover and these structures may be confidently correlated as the $F_A{}^2$ phase of deformation. The pattern of fractures is very similar to that described by several workers from other areas of folding (e.g. MUECKE and CHARLESWORTH, 1966; PRICE, 1966).

Because jointing affects rocks exhibiting brittle properties, and because the magnitude of the total finite strain is relatively small, experimental work on the fracturing of rocks (e.g. BRACE, 1964; PRICE, 1966) is relevant to this aspect of rock deformation, and a reasonable attempt may be made to relate a single set of fractures to the stresses responsible. The Coulomb-Mohr criterion of failure predicts that the principal compressive stress should bisect the acute angle between a conjugate pair of shear fractures. In the present area, J_3 and J_4 are thought to be conjugate shear joints and are bisected by J_1, a tension joint. These three joints are frequently mineralised; their line of intersection is vertical, suggesting that the intermediate principal stress was also vertical.

The joints from the Gössgraben Kern (sub-areas 10, 11, 12 & 13) have a slightly different pattern, exhibiting only two maxima representing orthogonal planes approximately parallel to the J_1 and J_2 joints. In the field, however, mesoscopic fractures with these orientations exhibit horizontal lineations and are associated with flexures. That is, they have the appearance of shear joints. The angle between them is nearly 90°. By analogy with experiment (BRACE, 1964), this may mean that σ_2 (least principal stress) was greater in this area.

If stress trajectories are calculated for the area on the basis of the joint pattern, the pattern obtained (Fig. IV—24) is very similar to that figured by ODÉ (1957) for the stress trajectories around a domical uplift superimposed on a regional stress field. Since the joints are developed late in the deformation history, it is suggested that they formed during the uplift of the Gössgraben Kern. Joint J_2 is only infrequently mineralised, and probably formed a little later than the other three; evidence of vertical movement on this fracture suggests that it may have been initiated when the principal compressive stress was vertical. The low-angle sheeting joint (J_5) is probably a true relaxation joint formed during the final stages of uplift and unloading.

On the basis of style and orientation, it has been possible to identify the latest phase of deformation, $F_A{}^2$ in both basement and cover rocks. This leaves the $F_A{}^1$ phase in the basement to be correlated with the main phase of deformation in the Peripheral Schieferhülle, which is of Alpine age. This correlation is based on the age relationships, and on the similarity in orientation of the axial surfaces. Within the basement, a pro-

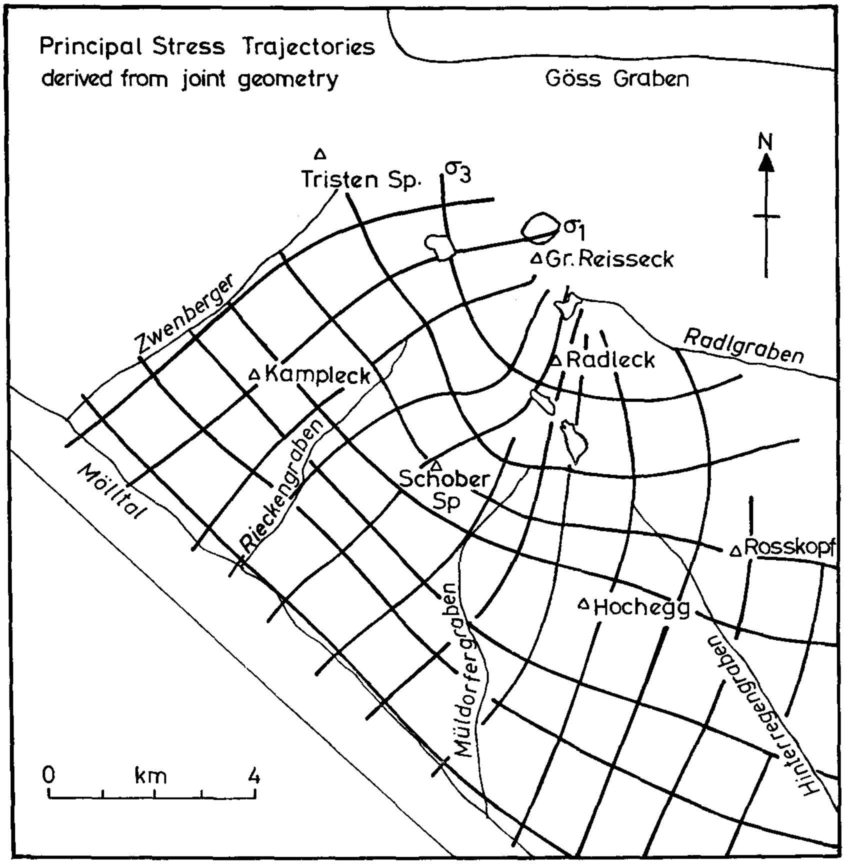


Fig. IV - 24 for discussion see text.

gression from the more open $F_A{}^1$ folds, characteristic of the Inner Schieferhülle, to sub-isoclinal folds typical of the Peripheral Schieferhülle is seen as the contact with the latter is approached.

The few strain measurements available suggest that the strain associated with the sub-isoclinal folding and strongly developed axial surface foliation of the most south-westerly parts of the Inner Schieferhülle (the Randgneis) and Peripheral Schieferhülle is very much greater than that associated with the more open folds and weak crenulation cleavage of the Reisseck Synform. This is also indicated by the degree of flattening of the folds and other qualitative strain markers. Thus it would appear that in the map area, the cover rocks have suffered a much greater strain than the basement rocks (except the Randgneis), which have deformed in an inhomogeneous and discontinuous fashion.

Both mesoscopic and macroscopic $F_A{}^1$ folds exhibit a strong sense of asymmetry, suggesting an important rotational component of the $F_A{}^1$ movements. HANSEN (1966) has suggested that such a rotational component may be described by relative movement of rocks on one side of the axial surface with respect to those on the other. The sense and direction of this movement may be determined by plotting the axes of the folds and their senses of rotation. The points will define a partial great circle girdle corresponding to the axial surface, while their rotations will be opposite on either side of a mirror plane of symmetry. The pole to this plane is the rotation axis; its line of intersection with the axial surface defines the direction of movement while its sense

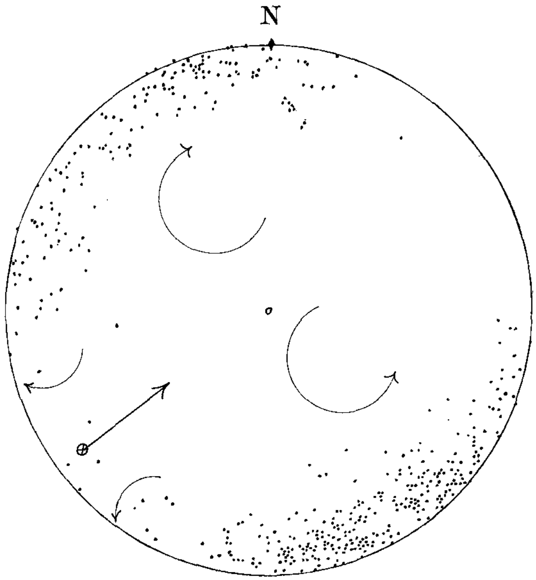


Fig. IV - 25 A stereographic plot of 450 $F^1{}_A$ fold axes from the basement rocks; senses of rotation shown; note that the axes define a plane dipping gently southwest; senses of rotation are derived from fold asymmetry; axes in the northwest half of the diagram show clockwise rotation (viewed down axis) and those in the southeast anticlockwise. The straight arrow lies in the plane of symmetry and has as its origin (circle with cross), the direction of tectonic transport; see text.

is defined by the sense of asymmetry of the folds (cf. HOWARD, 1968). Figure IV—25 is a plot of $F_A{}^1$ fold axes from the Inner Schieferhülle and Zentralgneis. Despite the large variation in orientation, a unique mirror plane of symmetry is exhibited defining a northwest-trending axis of rotation and a northeast direction of tectonic transport. This direction is also indicated by the sense of movement on thrusts and strain discontinuities described previously.

The steepening of the south-west limb of the Hochalm antiform appears to have occurred between phases $F_A{}^1$ and $F_A{}^2$ and was presumably associated with uplift on

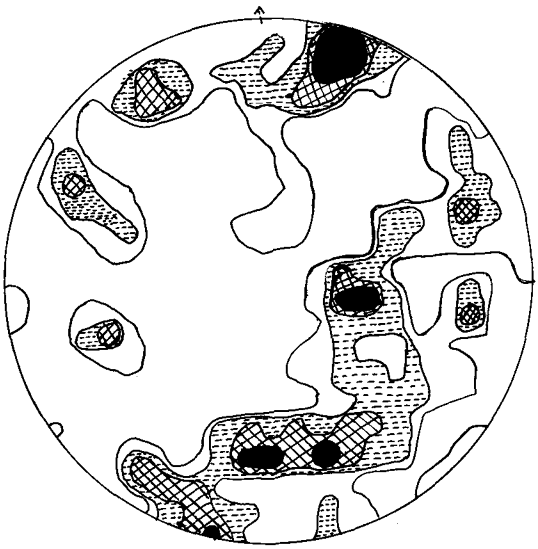


Fig. IV - 26 A stereographic plot of 93 Fv fold axes and lineations from different parts of the map area; contours 4, 3, 2, 1% area.

the north-east side of the Mölltal; this movement was probably responsible for the non-annealed fabrics of the Peripheral Schieferhülle immediately adjacent to the Mölltal.

Not a great deal can be deduced about the original form of the pre-Alpine F_V folds in the Inner Schieferhülle, as the granitic intrusions have largely obliterated any large-scale structures. Figure IV—26 is a plot of F_V fold axes from the Inner Schieferhülle. There is a marked tendency for the points to define a small circle distribution about the *predominantly north-west trending* $F_A{}^1$ axis. *Figure* IV—27 gives plots of F_V

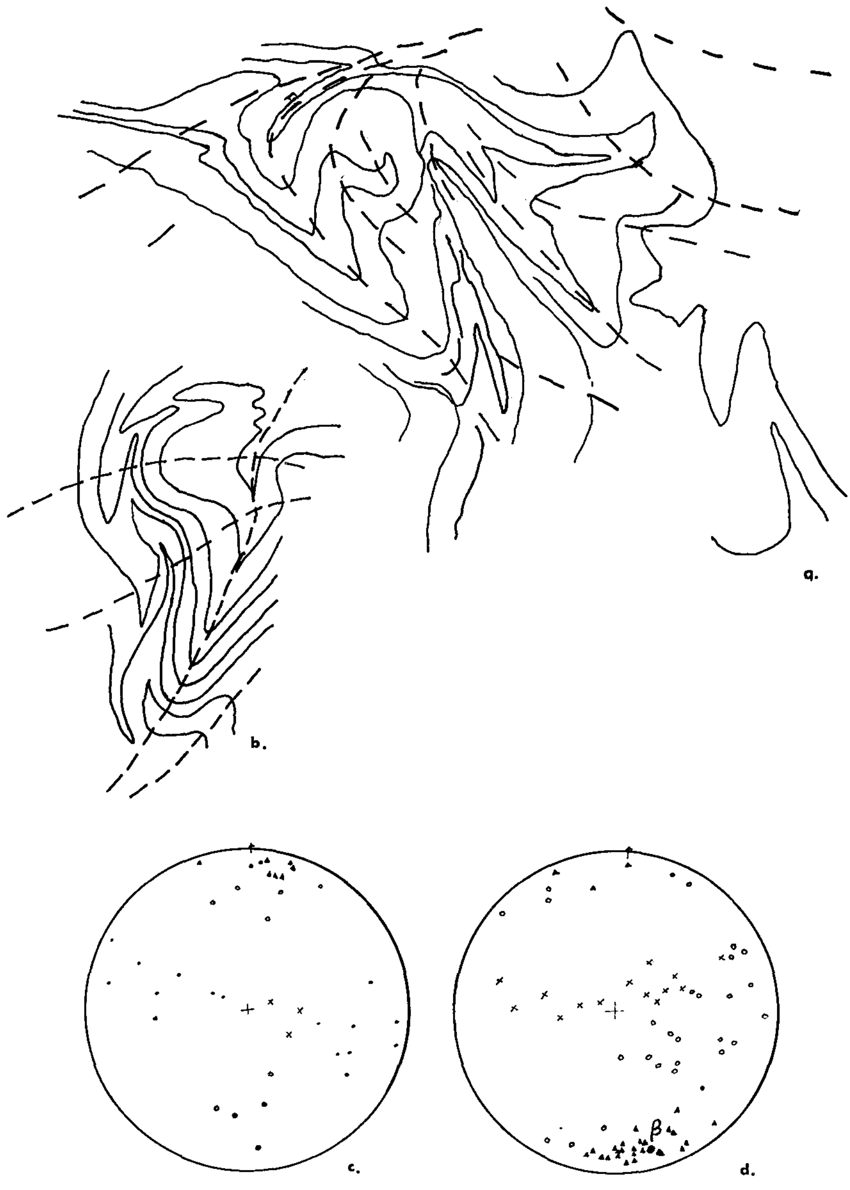


Fig. IV - 27 Two mesoscopic scale fold interference patterns from the Riekener Sonnblick; dashed lines show axial surfaces; c is an equal-area plot of observations made on a and b while d is a plot of similar information from a fold-complex west of the Kleine Mühldorfersee. Dots give poles to compositional banding; crosses give poles to the axial surfaces of the second folds; circles give the axes of the earlier folds; triangles give the orientation of rodding parallel to the later fold axes.

fold axes at two localities of refolded F_V folds, the Riekener Sonnblick east ridge, and west of the Kleine Mühldorfersee. At the former locality, where the local $F_A{}^1$ axis is approximately north-south, the F_V axes form a small-radius spread about this direction. In the latter area, where the dominant $F_A{}^1$ axis is northwest-southeast, they form a larger-radius circle. The orientation common to both distributions is north-northeast, and a strong maximum occurs in this position in Fig IV—26. It is tentatively suggested that this represents the most likely orientation for the Fv fold axes; a general northerly direction is also suggested by the type of interference pattern observed.

To summarize, the banded gneisses and amphibolites of the Inner Schieferhülle, or their pre-orogenic equivalents, were folded during at least two phases of deformation prior to 250 my (Middle Permian). Some granitic intrusion may have occurred before this time. The folds produced during these periods of deformation are not generally distinguishable, at present, and are grouped together as F_V, or pre-Alpine, structures. The latest phase of F_V folds was probably developed on northnortheast trending axes. The Inner Schieferhülle was then intruded by a plutonic suite ranging from tonalite to leucogranite in composition, and forming a series of irregular bodies and veins both concordant and discordant to the pre-existing structures. These were subsequently uplifted and eroded, and presumably formed a basement for Mesozoic sedimentation.

The $F_A{}^1$ phase affected both basement and cover and produced folds on a variety of axes, but predominantly northwest-southeast in orientation. These folds have the same sense of asymmetry on all scales, and therefore, the strain has an important rotational component. This rotational component may be described by a north-east relative movement on the upper side of the axial surface. The basement, especially the granitic lithologies, appears to have behaved relatively more competently than the cover rocks, and the strain associated with the Alpine deformation increases considerably within the latter. Following the development of the $F_A{}^1$ structures, but possibly beginning before these movements were complete (see Chapter VI), there began a phase of differential basement uplift. This resulted in the development of an irregular pattern of basement domes and depressions. Figure IV—28 is a tentative attempt to delineate this "basement topography" and contours show the present heights (in some cases projected) of the cover/basement interface; this interface was presumably sub-horizontal before the F_A movements began. The development of the Gössgraben Kern as a domical structure occurred as part of these vertical movements. It is also possible that uplift occurred along the northeast side of the Mölltal at this time causing a steepening of the $S_A{}^1$ foliation on the southwest flank of the Hochalm antiform although this could have occurred a little later.

During the later stages of uplift, the $F_A{}^2$ structures, comprising flexures, joints and faults, developed. Thus the $F_A{}^1$ structures are related to rotational movements developing a regional pattern of antiforms and synforms, whereas the $F_A{}^2$ structures are related to essentially vertical movements.

It is, however, important to distinguish two distinct kinds of vertical movements: those which on a local scale produced the pattern of domes and depressions in the basement (Fig. IV—28) and those which gave rise to the regional uplift of the Tauern. As shown in Chapter V there had been time for local thermal equilibriation between basement and cover after the development of the basement relief, and before the regional uplift and denudation of the Tauern as a whole.

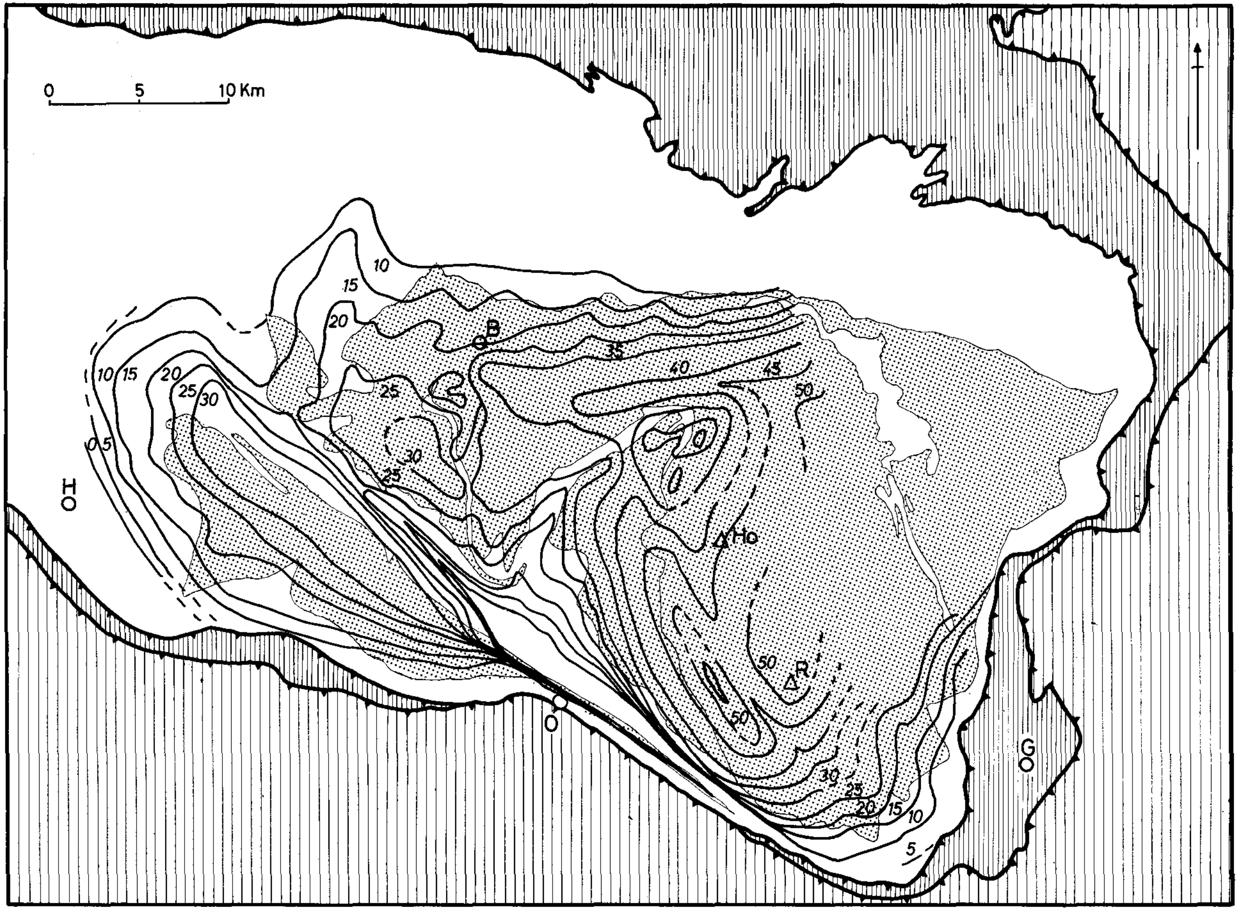


Fig. IV - 28 The eastern part of the Tauernfenster: contours showing the height or projected height of the basement-cover interface in hundreds of metres above sea-level. Basement, stippled; cover, blank; Unterostalpin, close ruling; Oberostalpin, wide ruling; H Heiligenblut; B — Badgastein; O — Obervellach; G — Gmünd; Ho — Hochalm; R — Reisseck.

V. Geochronology

by R. A. Cliff and E. R. Oxburgh

Introduction

Isotopic studies on the geochronology of the map area have been presented in several papers [Oxburgh et al. (1966), Cliff (1968 a and b and in preparation), and Brewer & Oxburgh (1972)]; in addition Lambert (1964, 1970) and Brewer (1969, 1970) have presented data from the surrounding area. The interpretation of these and other unpublished data are here reviewed, and an attempt is made to use them along with other geological results to establish a coherent thermal history for the south-east part of the Tauernfenster.

The isotopic data are summarised in Tables V—1, 2, 3 and 4; the Rb/Sr whole rock data from Table V—1 are also presented graphically in strontium evolution diagrams in Figs V—1 and 2. The mineral data are subdivided into Rb/Sr results, K/Ar results on micas and K/Ar results on other samples. Standard analytical techniques were used and are described fully in the original papers; rubidium and strontium were determined by isotope dilution; the 87/86 Sr ratio was measured on an unspiked aliquot as well as in an 84 Sr-spiked run in the majority of cases; argon was measured by isotope dilution and potassium by flame photometry. The constants used to correct the analytical results and calculate the ages are indicated in the tables.

For interpretive purposes the results can be divided into two groups:

1. Results bearing on the Zentralgneis — mainly Rb/Sr whole rock data, but also certain mineral ages.

2. Results which provide information on the Alpine metamorphism and subsequent cooling — mineral ages, chiefly on micas.

The age of the Zentralgneis

Rubidium-strontium whole rock ages have been measured on two granitic lithologies from the map-area — the fine-grained leucocratic granite-gneiss and the coarse leucogranite. In addition reconnaissance measurements were made on the fine-grained leucogranite.

The fine-grained leucocratic granite-gneiss has a well established position in the sequence of intrusions that make up the Zentralgneis and is particularly important in view of its well defined relation to the major folding episodes (Chapter IV). The results of Cliff (1968 a) are reproduced in Fig. V—1; unfortunately the Rb/Sr ratios are rather low and with the large analytical uncertainty associated with this suite of measurements, the data do not allow precise calculation of the age. An estimate is given by the best fit straight line, which corresponds to an age of 208 ± 60 my.

Despite the high error the results demonstrate the pre-Alpine age of the fine-grained granite-gneiss. Since this is one of the youngest lithologies recognised in the Zentralgneis (only the fine grained leucogranite is younger) the greater part of the Zentralgneis is clearly pre-Alpine. As discussed elsewhere this result also demonstrates that the two earliest fold episodes recognised in the map-area are pre-Alpine.

The reconnaissance analyses of two samples of the fine-grained leucogranite showed that the Rb/Sr ratios in this lithology are too low to allow precise determination of its age; the results obtained indicate an age of 150 ± 50 my assuming that the isochron model can be applied to the two samples. There is no evidence that this assumption is correct and the significance of the results is uncertain — they are, however, more easily reconciled with a Variscan age for the intrusion of the fine leucogranite than with an Alpine age.

TABLE V—1

WHOLE ROCK DATA

Coarse Leucogranite

Sample No.		Distance from Contact	uMRb^{87}	uMSr^{86}	Rb^{87}/Sr^{86}	Sr^{87}/Sr^{86}	
6702		500 *cm.*	0.607	0.00685	88.6 ±2.4	1.022 ±0.001	
6701		500 *cm.*	0.533	0.00627	85.0 ±2.3	1.008 ±0.002	
6703	i	500 *cm.*	0.563	0.00750	75.1 ±2.0	0.9922 ±0.0014 *0.9932 ±0.0009	0.9927
	ii		0.549	0.00756	72.6 ±2.0	0.9883 ±0.0018	
6712	i	70 *cm.*				*0.8466 ±0.0002	
	ii		0.470	0.0122	38.4 ±1.0	0.8497 ±0.0016 *0.8467 ±0.0017	0.8474
	iii		0.453	0.0122	37.6 ±1.0		
6707		35 *cm.*	0.409	0.0168	24.3 ±0.7	0.8053 ±0.0016	
6706		—	0.518	0.0325	15.9 ±0.4	0.7700 ±0.0019 *0.7697 ±0.0009	0.7699
6713		10 *cm.*	0.419	0.0482	8.60 ±0.2	0.7406 ±0.0014 *0.7402 ±0.0009	0.7404
6709		30 *cm.*	0.496	0.0762	6.52 ±0.2	0.7316 ±0.0014 *0.7320 ±0.0007	0.7318
6711	i	5 *cm.*	0.408	0.0727	5.61 0.15	0.7280 ±0.0014 *0.7295 ±0.0010	0.7284
	ii		0.375			*0.7275 ±0.0007	
	iii		0.392				
6708	L 4						
6708	L 1	1 *cm.*	0.602	0.184	3.28 ±0.1	0.7230 ±0.0016 *0.7209 0.0007	0.7216

*) indicates ratio measured on unspiked aliquot

TABLE V—1 (continued)

Fine-grained Leucocratic Granite-gneiss

Sample No.	uM Rb^{87}	uM Sr^{86}	Rb^{87}/Sr^{86}	Sr^{87}/Sr^{86}
A 310	0.880	0.115	7.67	0.7344±0.0036
A 312	0.802	0.118	6.77	0.7271±0.0036
A 341	0.687	0.182	3.77	0.7190±0.0036
A 337	0.466	0.606	0.75	0.7116±0.0035
		Fine Leucogranite		
A 414	0.016	0.361	0.04	0.7097±0.0010
A 417	0.693	0.216	3.2	0.7169±0.0010

TABLE V—2 Rb-Sr MINERAL DATA

Sample	Mineral	Lithology	Rb ppm	Sr ppm	Rb^{87}/Sr^{86}	Sr^{87}/Sr^{86}	Age
A 379	Mu	pegmatite	1832	0.89	5932.0	21.5	206 ±15
C 6712	Mu	leucogranite	542	5.07	310	0.8932	15.1±2
	All		n.d.	n.d.	—	0.8244	
	WR		142	10.8	38.0	0.8474	
C 6711	Mu	leucogranite	393	17.9	63.5	0.7447	17 ±4
	Bi		644	14.0	132.7	0.7564	14 ±3
	Plag		17.7	55.9	0.9	0.7250	
	WR		120	64.6	5.61	0.7284	
C 6710	Bi	tonalite	418	37.2	32.5	0.7169	14.5±3.6
	Ap		0.7	739	0.002	0.7105	
	Plag		17.7	55.9	0.15	0.7086	
C 6716	Bi	tonalite	339	62.3	15.8	0.7125	15.9±8
	Ap		n.d.	143	—	0.7094	
	Ep		3.9	2201	0.005	0.7074	
	Plag		11.0	323	0.1	0.7096	

WR = whole rock

The coarse grained leucogranite has a wide spread of Rb/Sr ratios and yielded a much more precise age. Preliminary data were published by CLIFF (1968 b); subsequent analyses on new and larger samples confirm an age within the error limits quoted in the earlier work, although the new results require a radically different interpretation of the detailed response of strontium isotopes to the Alpine metamorphism (CLIFF in preparation). The best-fit isochron calculated with the exclusion of high-Sr samples (CLIFF, 1970) by the method of YORK (1966) has an age of 244±6 my, with an initial ratio of 0.7141±0.0033 my; the data are plotted on Fig. V—2. The data show a reasonably close fit to the isochron, in spite of resetting of the mineral ages during the Alpine metamorphism. The coarse leucogranite is younger than most of the Zentralgneis and the results thus confirm the less precise ages discussed above. In particular the coarse leucogranite is demonstrably younger than the tonalite which had been suggested by KARL (1959) as a possible Alpine intrusion.

TABLE V—3 K-Ar DATA ON MICAS

Mesh Fraction 40—80

Mineral	Sample	Lithology	At. % K	Ar^{40}* cc STP/g $\times 10^{-5}$	%radio-genic	Age
Biotite	Gr	mu-bi-augen gn.	6.47	0.493	50	19.0±0.5
Muscovite			7.87	0.744	74	23.5±0.5
Biotite	Tr	augen gneiss	6.75	0.683	84	25.2±0.5
Muscovite			8.70	0.751	67	20.9±0.5
Biotite	A 346	amphibolite	7.59	1.491	85	48.6±1.0
Biotite	C 308	bi-augen gneiss	5.94	0.892	75	37.0±2.0*
Biotite	C 213	coarse augen gneiss	6.69	0.596	84	22.0±0.5*
Biotite	Z 162	bi gneiss	6.51	0.570	72	21.9±0.5
Biotite	Z 160	bi gneiss	7.08	0.609	76	21.5±0.4
Biotite	C 18	aplitic gneiss	7.12	0.553	82	19.5±1.0*
Biotite	C 212	bi pegmatite	7.10	0.558	81	19.5±0.5*
Biotite	Z 150	bi gneiss	7.41	0.517	79	17.5±0.2
Biotite	Z 151	bi gneiss	7.19	0.462	72	16.1±0.2
Biotite	Z 159	bi gneiss	7.23	0.461	72	16.0±0.3
Muscovite	C 422	mu. sch.	8.48	0.746	58	22.0±1.0*
Muscovite	C 55	mu. sch.	7.99	0.704	79	22.0±0.5*
Muscovite	C 495	mu.qz.sch.	8.74	0.755	51	21.5±1.0*
Muscovite	O VI	mu.dol.qtzt.	7.67	0.646	82	21.0±0.5*
Muscovite	C 315	mica sch.	8.49	0.721	61	21.0±1.0*
Muscovite	C 31	gt.mu.sch.	6.97	0.486	78	17.5±0.5*

*) from Oxburgh et al. 1966; other data unpublished results of M. S. Brewer and E. R. Oxburgh.

TABLE V—4 K-Ar DATA ON AMPHIBOLES AND WHOLE ROCKS

Mesh Fraction 40—80 unless otherwise indicated

Mineral	Sample	Lithology	At. % K	Ar^{40}* cc STP/g $\times 10^{-5}$	%radio-genic	Age my
Amphibole	A 346†	biotite amphib.	0.312	0.789	94	597 ±12
		40—80		0.874	86	546 ±10
		80—120	0.365	0.741	89	450 ±9
Amphibole	A 233	amphibolite 80—120	0.21	0.161	64	183 ±4
Amphibole	A 300	amphibolite 80—120	0.853	0.344	88	98.3±2.1
Amphibole	A 225	amphibolite 80—120	0.796	0.177	69	54.9±1.3
Amphibole	C 250	amphibolite 80—120	0.27	0.046	33	42.9±2.2
Amphibole	C 170	amphibolite 80—120	0.244	0.032	21	32.7±2.8
WR	C 523	kyanite schist	4.02	0.300	24	18.5±2.0*
WR	L 122	phyllite	1.68	0.122	53	18.2±0.5**
WR	L 52	calc-phyllite	3.47	0.163	69	11.7±0.2**

*) from Oxburgh et al. 1966.
**) unpublished analyses by M. S. Brewer;
other data, unpublished results of E. R. Oxburgh and J. G. Simons.
†) (Note coexisting biotite age: 48.6 my).

The data from the map-area are consistent with the bulk of the Zentralgneis being Variscan in age, although the presence of older rocks is not excluded. Data from other parts of the Tauernfenster support this inference: LAMBERT (1964) determined an isochron of 243±11 my for the Zentralgneis in the Badgastein area. JÄGER et. al. (1969) report a similar age for the augen-gneisses of the Venedigergruppe, and the tonalite from the same area has yielded K/Ar hornblende ages between 292 and 358 my (BESANG et al. 1968). A single sample from the Granatspitzekern indicates a Rb/Sr whole rock age of 296±15 my assuming 0.708 for the initial ratio (CLIFF, unpublished result).

At present there is no indication from isotopic studies that there was any Alpine granitic magmatic activity within the Tauernfenster. The nearest Alpine intrusions are the Rieserferner tonalite and the small granodiorite stock of Wöllaner Kopf; the latter has yielded K/Ar biotite ages of 31—44 my (OXBURGH unpublished results).

It would still be possible for some of the late undeformed aplite and pegmatite veins in the Zentralgneis to be of Alpine age; however, a single Rb/Sr age on a pegmatitic muscovite gives 206±25 my (CLIFF, unpublished result).

The Alpine Metamorphism

With few exceptions even those minerals which originally crystallised before the Alpine orogeny have had their isotopic ages modified by the Alpine metamorphism. The most resistant systems (e.g. K/Ar in hornblendes and Rb/Sr in pegmatitic muscovite) have suffered incomplete daughter loss during the metamorphism. The micas, which account for the bulk of the samples analysed from the map-area, either crystallised or recrystallised during the metamorphism and continued to lose their daughter isotopes for some time afterward.

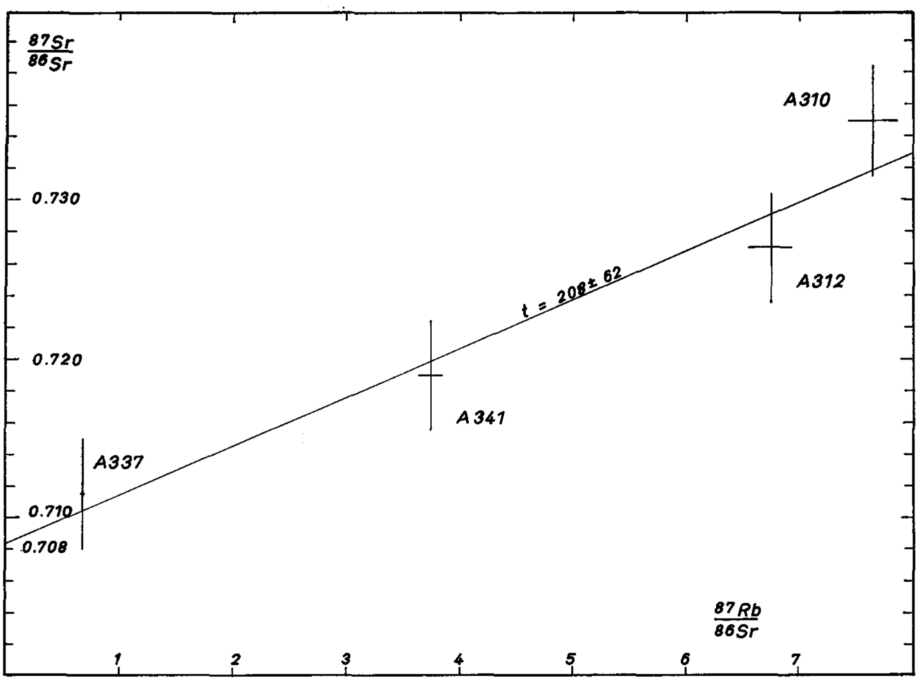


Fig. V - 1 A strontium evolution diagram for the fine-grained leucocratic granite-gneiss.

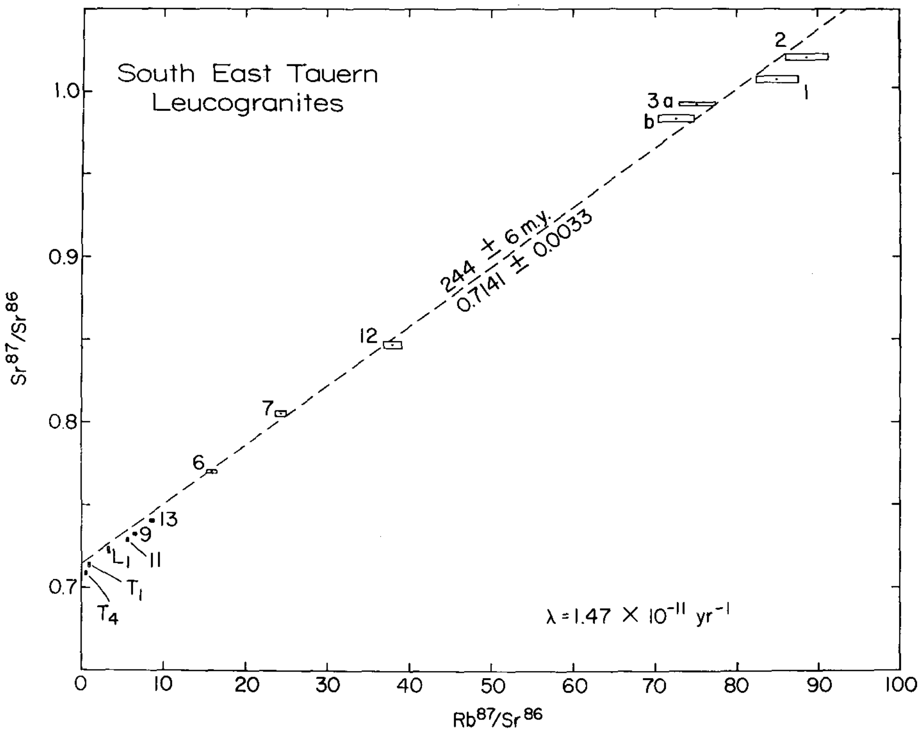


Fig. V - 2 A strontium evolution diagram for the coarse leucogranites. Samples T 1 and T 4 are from a tonalite across a leucogranite contact while L 1, 9, 11 and 13 are within the leucogranite but contaminated by strontium from the tonalite. All of these samples are excluded from the calculation of the isochron. A more detailed discussion will be published by R. A. C. elsewhere.

Only a few micas have been analysed with the Rb/Sr method and the following discussion is concerned principally with the results of K/Ar analyses on biotites and muscovites by Oxburgh et al. (1966), Oxburgh (1968 unpublished results) and Brewer and Oxburgh (1972). The geographical distribution of the analysed samples is shown in Figs V—3, 4. The muscovite analyses within the area mapped show a fairly narrow range of ages, around 22±1.5 my, with one exception (17.5 my), and there is no apparent correlation between the variation in age and tectonic position.

The biotite ages on the other hand show a much wider spread. Disregarding the two oldest samples, which are discussed individually below, the ages range from 16 to 25 my. There are too few samples to establish any systematic variation of age with tectonic position. However, there is a suggestion that the biotite ages from the northern part of the map-area are older than those from the south: in the north the biotites give ages of about 22 my, (muscovites from the same area give about 21 my); by contrast the biotites in the south range from 16 to 19 my while the muscovites are again about 21 my. Unfortunately there are only two coexisting biotite-muscovite pairs among the analyses — one from the southern area confirming the general pattern with the biotite 4.5 my younger than the muscovite (note, however, that a second biotite sample from very close by yielded the anomalous 37 my age); the second pair, from the northern part of the map-area shows the inverse relationship. In the case of the suite of samples from the Gössstollen the transition from young to high biotite ages is apparently abrupt, and can be tentatively correlated with a structural break across a fault that cuts the tunnel.

The anomalous 37 my biotite age of sample C 308 is very difficult to explain. Oxburgh et al. (1966) suggested there might possibly be an outlier of Altkristallin (Oberostalpin) on the peak of the Gröneck but this has not been substantiated by subsequent field work and biotite from a second sample collected later from approximately the same locality yielded an age of 19 my.

The highest biotite age measured was 48.6 my from biotite in amphibolite A 346. No obvious explanation for this high age is apparent. The temperature reached during

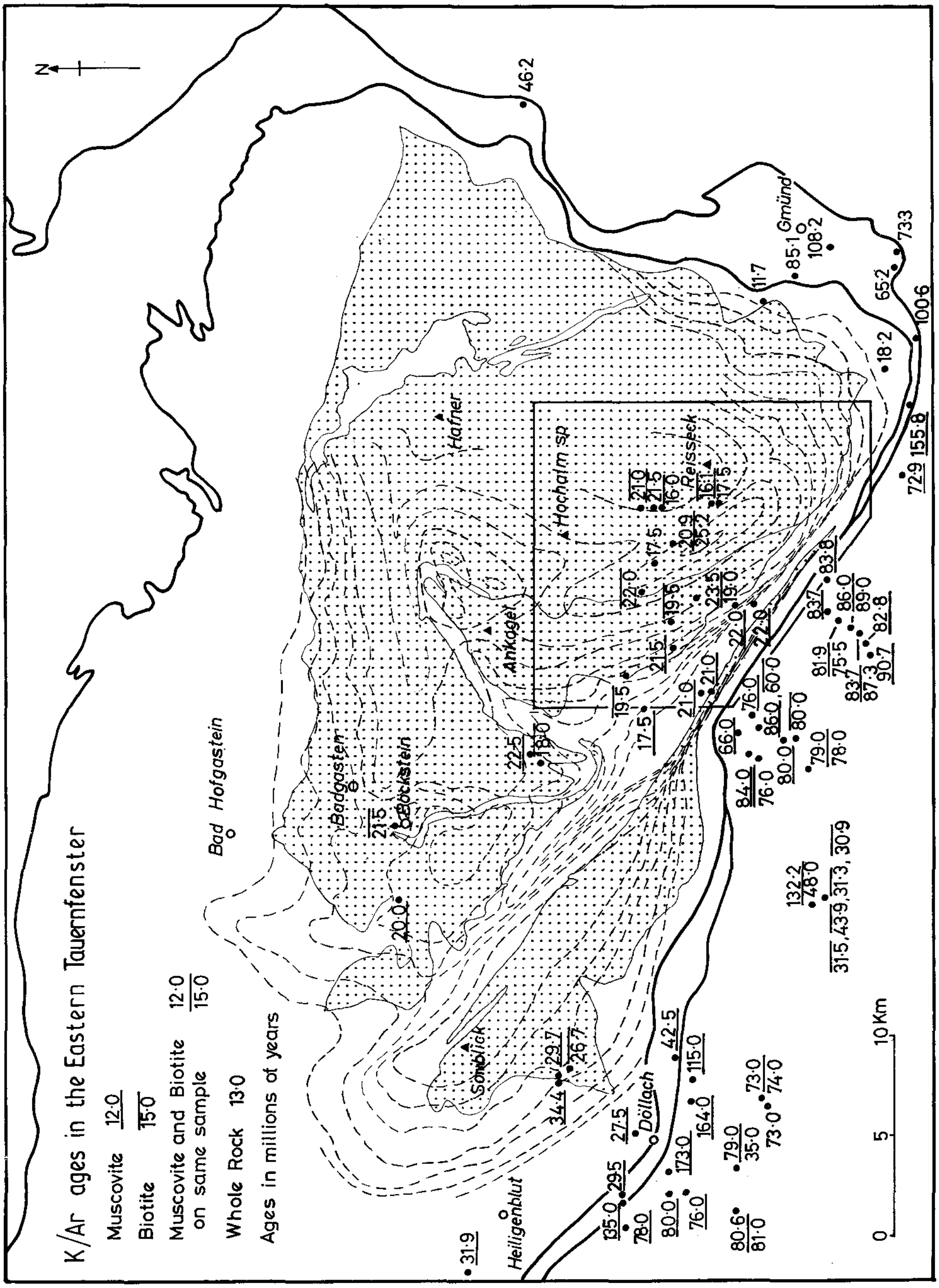


Fig. V - 3 For discussion see text.

the Alpine metamorphism of this part of the map-area was certainly higher than that normally necessary to outgas biotite. Although the other hornblende data are discussed below, it is appropriate to mention here the hornblende from A 346 which also has an anomalously high age: repeated determinations on two size-fractions gave an age of 597 ± 12 my (546 ± 10 my on second aliquot) on the coarse fraction, while the finer fraction gave an age of 450 ± 9 my. There is petrographic evidence of two hornblende generations in the sample and it seems likely that the size-fractions represent mixtures of these two generations in different proportions. Petrographically the younger generation seems more abundant so that it is likely that both generations are pre-Alpine. The high age of the biotite might be explained by its enclosure in more resistant amphibole, and its occurence as large (15 *mm.*) poikiloblastic grains.

The six hornblende ages (Table V—4, Fig V—4) range from 597 my to 33 my. In view of this wide scatter and the small amount of data, the results are not very significant, especially if, as suggested above, there is more than one generation of pre-Alpine hornblende. It is likely that the majority of the hornblendes crystallised in the Variscan metamorphism; this suggestion is supported by their alignment parallel to the Variscan fold axes at several localities. It appears, however, that none of the hornblende ages records the date of the Variscan metamorphism: either the Variscan hornblendes were only partially outgassed in the Alpine metamorphism or the analysed samples represent mixtures of young and old hornblendes. Apart from A 346 which it is thought may contain pre-Variscan hornblende, none of the samples show petrographic evidence for more than one generation, so that partial outgassing seems more likely. In this case a wide spread of ages in a small area may provide a maximum limit for the age of the Alpine metamorphism, since if some hornblendes from a single small area were incompletely outgassed it is unlikely that any samples from the same area could have remained above the critical temperature range for very long. Thus the youngest measured age in such a suite of hornblende ages should be greater than the age of the metamorphism. Samples A 346 and A 233 were collected from the same amphibolite outcrop, while A 225 comes from a locality approximately 2 *km.* further south. This group of samples would therefore indicate a maximum age for the Alpine metamorphism of 55 my.

Three isolated whole-rock K/Ar ages add no critical information: the two higher ages are not greatly different from biotite ages in the area; the significance of the 11.5 my age is impossible to evaluate in the absence of more data.

Several Rb/Sr mineral ages were determined in connection with the detailed study of the coarse leucogranite (Cliff, in preparation). The biotite ages are appreciably younger than the K/Ar results from the same area: C 6711 yields an age of 14 ± 2 my and within the limits of error the two biotites from the tonalite are consistent with this age. The muscovite from C 6711 is older at 17 ± 4 my, but that from C 6712 has a calculated age of 14 ± 3 my (see Fig. V—4).

Interpretation of Mineral Ages

The accumulation of daughter isotopes in K/Ar and Rb/Sr mineral systems may be interrupted and previously accumulated strontium and argon lost, either as a result of reaction during metamorphism, or by continuous volume diffusion if the temperature becomes sufficiently high. The accumulation of the daughter isotopes does not begin again until the metamorphic reaction is complete and/or the temperature has fallen sufficiently that rates of diffusive loss are small. Several studies have indicated that the temperatures at which such diffusion becomes important lie somewhat below the maximum temperature believed to have been reached in the course of regional metamorphism in the Tauern. They also show that the activation energies are such

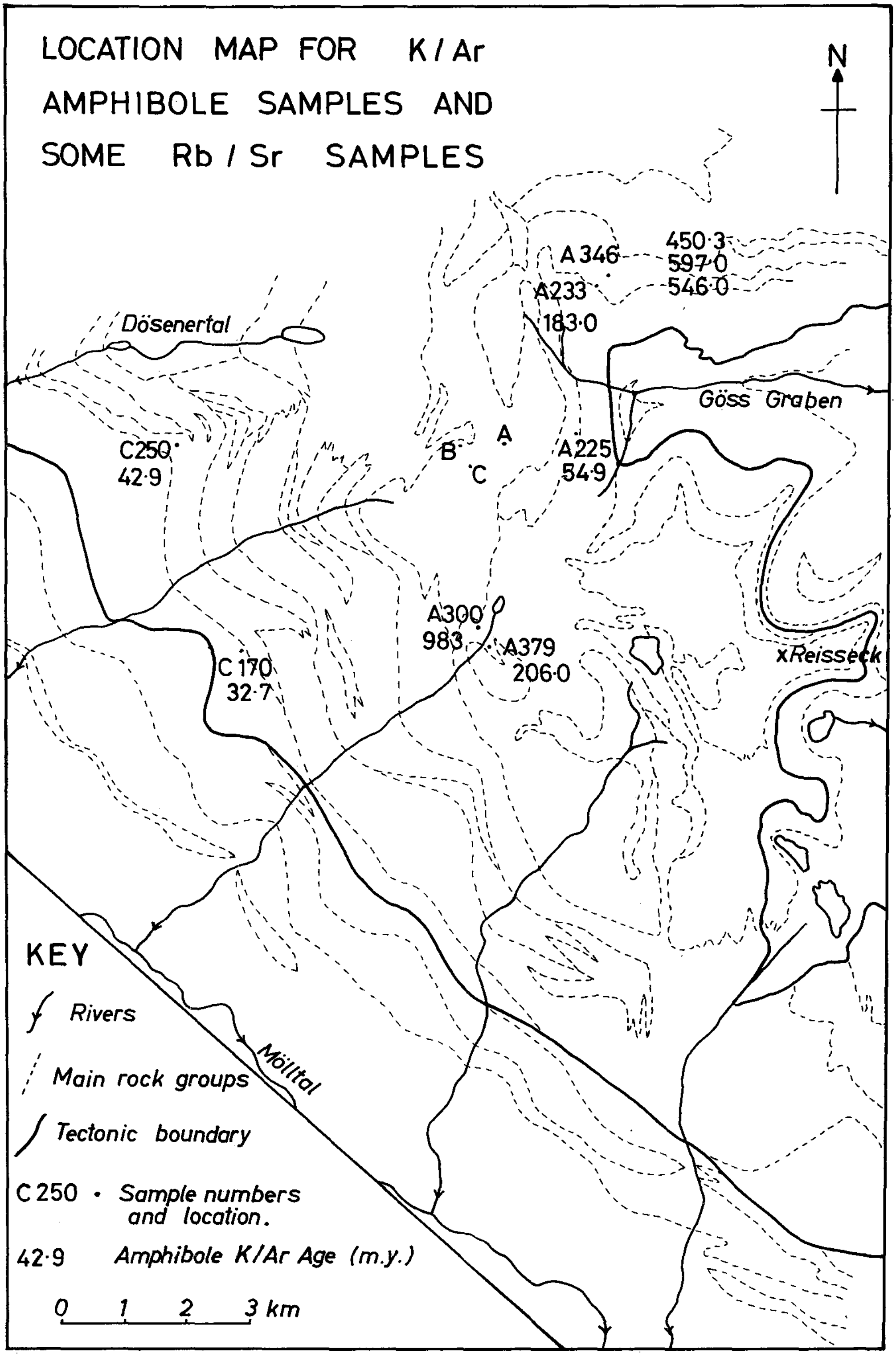


Fig. V - 4 For discussion see text; A = locality for 67 - 1 to 5; B = 67 - 7 to 9; C = 67 - 6, and 10 to 16.

that the transition from a state where diffusion is negligible to one where it is important, should occur in a fairly narrow temperature interval (TILTON & HART, 1963; HURLEY et al., 1962; EVERNDEN and CURTIS, 1965; and JÄGER et al., 1967), sometimes loosely called the critical temperature. Uncertainties about the transport mechanisms operating both under natural conditions and in experiments, require that extrapolated values for diffusion coefficients at low temperature, based on experimental values for activation energies, be applied with caution. Nevertheless the studies cited above indicate that a cooling biotite crystal does not become a closed system with respect to the diffusion of radiogenic argon and strontium until a temperature somewhere in the range 200° to 350° C is reached, while for muscovite the equivalent temperature interval may be as much as several hundred degrees higher. In most field studies the critical temperatures for strontium and argon appear to have been similar, but in some cases other external factors e.g. high $P_{Ar^{40*}}$, have been more important than temperature in controlling diffusion; such factors will in general operate differently on argon and strontium.

To summarise, mica ages in high or medium grade metamorphic rocks do not generally record the actual time of metamorphism but rather a stage in the subsequent cooling. The measured mica ages for the map area, which concentrate around 20 my thus indicate a minimum age for the Alpine metamorphism of the area and at the same time indicate a period in which cooling and therefore presumably uplift were occurring.

It is to be expected therefore that these mica ages place a lower limit of 25 my on the age of the Alpine metamorphism while an upper limit of 55 my is provided by the hornblende ages discussed above.

Isotopic Data from Other Parts of the Tauernfenster and its Surroundings

Mineral ages have been measured in several other areas within the Tauernfenster, and the results conform in general with those from the map-area. In the Badgastein area OXBURGH et al. (1966) found K/Ar mica ages between 18 and 26 my and from the same area LAMBERT (1964) reported an approximate mineral Rb/Sr isochron age of 20±10 my. In the Döllach area on the south-west flank of the Sonnblick Kern LAMBERT (1970) determined four muscovites from 28 to 34 my and a single biotite at 27 my. Further west in the Venedigergruppe BESANG et al (1968) found K/Ar ages in the range 19 to 30 my on biotite and muscovite from the Zentralgneis; Rb/Sr analyses on the same minerals gave generally concordant values although some results were higher and some lower than the K/Ar values. Amphiboles from two samples of tonalite gave K/Ar ages ranging from 392 my to 293 my depending upon the size fraction analyzed. Coexisting biotites gave 341 (K/Ar) and 30 (Rb/Sr), and 317 (K/Ar) and 39 my (Rb/Sr) respectively.

These mica data are significant in setting a higher minimum age for the metamorphism of the Peripheral Schieferhülle; the higher ages found by LAMBERT could indicate either that his samples come from a relatively high tectonic level in the Peripheral Schieferhülle and thus, on uplift, cooled earlier than the deeper zones, or that uplift and cooling actually occurred earlier in the Döllach region than further east. JÄGER et al. (1967) found similar results in the Swiss Alps.

The area immediately surrounding the Tauernfenster can also provide information on the conditions during the Alpine metamorphism in the window, in particular in establishing how much additional cover was transposed on to the Tauern area as a result of overthrusting of the Ostalpin, and when this occurred. For the area immediately adjacent to the map-area, OXBURGH et al. (1966), BREWER (1969, 1970), and BREWER and OXBURGH (1972) have presented a large number of K/Ar mica analyses; the general pattern of distribution of these results is indicated on Fig. V—3. Their interpretation is complicated by the established occurrence of excess argon, at least in the biotites, in part of the area, (BREWER, 1969). Elsewhere around the window both mica species record ages close to 80 my; in general the muscovite ages are near 86 my whereas the biotites are around 80 my old. Three Rb/Sr determinations on biotite indicate an age of 65 ± 5 my.

In any case the most striking feature of this pattern is that there is a sharp discontinuity in the age pattern corresponding exactly to the edge of the Tauernfenster along the Mölltal. This means that the present structural relationship between the rocks inside the window and higher units south-west of the Mölltal cannot be more than about 20 my old. Had the present relationship existed before the rocks within the window cooled, at least the lower parts of the higher units should have experienced partial outgassing and there would have been a gradation in the K/Ar ages across the edge of the window. As discussed elsewhere there is other evidence suggesting significant differential vertical movement along the Mölltal associated with the main uplift of Tauern in Miocene times.

Further away from the map-area, SCHMIDT et al. (1967), MILLER et al. (1967), HARRE et al. (1968), GENTNER et al. (1963) and LAMBERT (1970) have all determined ages on minerals from the Ostalpin close to the edge of the Tauernfenster. In every case ages close to 80 my were found and a cooling at this time seems well established; on the basis of BREWER's Rb/Sr results it seems that cooling may have continued from this time until after 65 my ago.

A Model for the Thermal History of the Tauernfenster

The geochronological data discussed above are consistent in indicating a regional metamorphism within the Tauernfenster before 36 my ago, and possibly later than 55 my ago. During this metamorphism temperatures were attained at which greenschist facies assemblages crystallised in much of the Peripheral Schieferhülle and part of the basement complex. Where temperatures were higher, however, in the south-east, amphibolite facies assemblages were stable. This suggests that temperatures may have been in the vicinity of 450° C possibly reaching 550° C in the Reisseck area.

Between 25 and 15 my ago the region cooled through critical temperatures for argon and strontium diffusion in both muscovite and biotite. The absence of a well-defined difference between the muscovite and biotite ages suggests that cooling was rapid and that by 15 my ago the temperature had probably fallen to approximately 250° C. The difference between the K/Ar and Rb/Sr ages suggests that external environmental factors may have influenced diffusion in one or both of these systems, since in cases where volume diffusion into an infinite reservoir might be postulated the ages are generally very close (e.g. ARMSTRONG et al., 1966).

It is known that about 15 my ago (see stratigraphic evidence below) the Alps underwent relatively rapid uplift and erosion. Such erosion exposes or brings nearer to the surface rocks previously deeply buried and in consequence they cool much more rapidly than had they simply experienced conductive loss of heat from their original depth of burial. This means that in an area which is undergoing differential uplift, the distribution of isotherms and thus the pattern of mineral cooling ages will be controlled by rates if vertical movement and their variation from place to place. It is very interesting that the distribution of ages within the eastern part of the window suggests a general uplift and that there seems to be little evidence for a difference in time of cooling between the Zentralgneis domes and the intervening depressions occupied by Schieferhülle as predicted by OXBURGH et al. (1966). For instance the similarity of mica ages in the Badgastein (near the centre of the window) area and at the edge of the window is surprising if this were the case. This implies that there was sufficient time for thermal equilibration to take place after the elevation of the gneiss domes and before a more general uplift which was responsible for the cooling of the Tauern.

The difference in mica ages of similar grade rocks between the map-area and Döllach area (Lambert, 1970) is more significant (Fig. V—3) and perhaps indicates earlier uplift of the Sonnblick dome — this suggestion is supported by structural evidence (Chapter IV) that this uplift occurred while northward translational movements were still continuing.

In contrast to the rocks within the Tauernfenster the Ostalpin around the edge of the window appears to have undergone a cooling some 80 my ago. Little is known of this cooling or of the heating which presumably preceded it. At the western end of the Tauernfenster (the Brenner area) a late Mesozoic metamorphic event affected the Ostalpin, and the Mesozoic cover rocks reached biotite grade. It is not known whether the 80 my cooling in the present area was immediately preceded by a metamorphic event, because its effects would be difficult to distinguish from those of the pre-Mesozoic metamorphism. If there was such an event, the almost unmetamorphosed state of the Unterostalpin (e.g. in the Radstädter Tauern) suggests that it occurred before translation of the Ostalpin thrust sheets over the Tauern area. It is suggested following Brewer (1970) that the overthrusting took place about 65 my ago with consequent rapid cooling of the base of the Ostalpin thrust sheet (see also below).

A possible model for the Alpine metamorphism in the Tauern thus supposes burial of the area under the Ostalpin thrust masses about 65 my ago, followed by an increase in the temperature of the buried rocks sufficient for greenschist to amphibolite facies regional metamorphism between 55 and 36 my, and subsequent cooling to the critical temperature for biotite between 25 and 15 my ago. This model does not preclude the possibility of continuing horizontal movements within the Ostalpin or between it and the underlying Pennine Zone. The recent stratigraphic synthesis of Oberhauser (1968) suggests that overthrusting of the Tauern began as early as 80 my and was completed only in the Eocene (55 my — Table V—5).

With the increasing availability of information on the physical properties of rocks, the conditions under which metamorphic reactions can occur, and the time intervals available for metamorphism, it should become possible to extend the suggested model and draw up a thermal balance sheet for this segment of the earth's crust during the Alpine orogeny. Indeed model calculations for the post-metamorphic cooling stage in the Tauern Railway Tunnel at the western edge of the map area, have already been presented in a most interesting paper by Clark and Jäger (1969). These calculations were based on the assumption of a relatively simple initial temperature distribution, and subsequent heat transfer by conduction. Neither of these important assumptions, can, however, be accepted without reservation.

As discussed in the next chapter, large-scale horizontal movement can disturb the geothermal gradient in complex ways and may even cause it to change sign (this almost certainly happens after regional overthrusting). In consequence the assumption that the Tauern geothermal gradient was a monotonic function of depth during the Tertiary is somewhat doubtful.

Conduction may not have been the only significant mechanism of heat transfer. The abundance of open fissures with hydrothermal mineral fillings, including high temperature gold mineralisation, suggest that convective heat transfer may have been important. Various localities in the eastern Tauern are today noted for their hot springs.

Clark and Jäger have clearly separated the many variables upon which the evolving thermal structure of an orogenic region depends, and their analysis could relatively easily be extended to take account of the first problem (the initial thermal gradient) if it proved necessary — this would depend upon the extent of thermal equilibration after the main horizontal movements and before the final episode of uplift. The problem of convective heat transfer however, is less tractable because the geological evidence is less complete; it amounts to introducing a thermal conductivity which is a function both of depth and time.

TABLE V—5

my		Time Scale (Geological Society of London, 1964)	Events based on Stratigraphic Evidence (after Oberhauser, 1968)	Events based on Radiometric Evidence	Observed Radiometric Ages
0	TERTIARY	Pliocene			
10					Bi Rb/Sr ages in Tauernfenster
		Miocene	Main Molasse deposition		
20				Uplift of Hohe Tauern, cooling of Pennine rocks	Bi K/Ar ages in Tauernfenster Mu K/Ar ages in Tauernfenster
30		Oligocene	Increased sedimentation in Molasse		
40		Eocene	Oldest Molasse	Main Alpine Metamorphism	
50					A 225 Hb K/Ar age in Tauernfenster
60		Palaeocene	Tauern area overthrust		
	CRETACEOUS			Tauern area overthrust	Bi Rb/Sr age in Ost-alpin-Altkristallin
70		Maestrichtian Campanian	Ultrapienidische Rücken overthrust		
80		Santonian		Beginning of overthrusting, uplift & cooling of Ostalpin-Altkristallin	
		Coniacian			Bi K/Ar ages in Ost-alpin-Altkristallin
90		Turonian	Thrusting within N. Calcareous Alps		Mu K/Ar ages in Ost-alpin-Altkristallin
		Cenomanian		? Metamorphism in Altkristallin	
100		Albian			

VI. Conclusions

by E. R. OXBURGH, R. J. NORRIS and R. A. CLIFF

In this chapter we attempt to synthesise the observations and conclusions drawn in the earlier chapters, and to reconstruct the outlines of the geological history of the region mapped, within the broader setting of the Southeastern Tauern.

Relatively little may be said of the pre-Mesozoic history. Parts of the Inner Schieferhülle show evidence of one or more pre-Mesozoic episodes of metamorphism and deformation; the garnet-mica schists and amphibolites had both undergone at least one phase of folding before the intrusion of leucogranites at about 244 my. At that time the schists had already undergone regional metamorphism — isolated, partially digested, fragments of garnet schist occur as xenoliths within the leucocratic igneous intrusives. The age and character of this early phase of regional metamorphism is not well established; coarse retrogressed kyanite occurs in some of the schists and pyroxene is occasionally found in the amphibolites; the former could, however, be the product of Alpine metamorphism and the latter could be a relict igneous pyroxene from an original basalt or dolerite. At least some of the schists pre-date the tonalites which are themselves older than the main group of leucocratic igneous intrusives.

There is some indication from the limited K/Ar data available on amphiboles that the amphibolites could be of Cambro-Ordovician age and whereas this would not be inconsistent with other information, the data are too few and too ambiguous to be taken very seriously.

It is, however, clear that in Permian times there was an episode of intrusion of granodioritic magma on a regional scale, closely followed by the intrusion of leucogranites. Some parts of the granodiorite may be pre-Permian for they already possessed a mica foliation before the Permian intrusive episode ended. At the level of present-day erosion, rocks of generally granodioritic composition make up about two thirds of the area of outcrop of the basement complex. The banded gneisses, schists and amphibolites (and possibly a small part of the granodiorite and tonalite) represent screens and pendants of roof rock into which the main suite of granodioritic intrusions penetrated. Before the Alpine deformation, this Hercynian intrusive complex may have resembled some of the major Mesozoic batholithic complexes of western North America (e.g. Sierra Nevada, BATEMAN et al., 1963). Mapping in the present area has established that the intrusive relationship between the Inner Schieferhülle and Zentralgneis was geometrically complex. Relationships were in many places not stratiform and were produced by several different intrusive phases. Variation in thickness, orientation and distribution of the igneous bodies have exerted a strong influence on the orientation and intensity of later structures.

From the evidence available within the area mapped, the early Mesozoic history of the basement complex described above is rather uncertain. The Mesozoic, metasedimentary Peripheral Schieferhülle sequence which overlies it today has undergone very intensive deformation. Nowhere is it interfolded with the basement rocks on a mesoscopic scale; rather it occupies the major structural depressions within the basement surface and any original stratigraphic relationship between basement and cover has been obliterated by shearing. Thus here, as in much of the Tauern, there is no direct evidence of autochthonous Mesozoic deposition, although on general palaeogeographic considerations this is generally assumed to have occurred (e.g. KOBER, 1955; TOLLMANN, 1963; EXNER, 1964; OBERHAUSER, 1968). The possibility that in the present area the Peripheral Schieferhülle is entirely allochthonous, cannot be eliminated.

Whether autochthonous or allochthonous, the Peripheral Schieferhülle represents deposition of sediment in a wide variety of sedimentary environments varying from shallow-water shelf conditions which were presumably necessary for the deposition of the limestones and quartzites (a few of these might, however, represent meta-cherts) to somewhat deeper water for the formation of black, calcareous shales associated with mafic and ultramafic igneous rocks. Estimates have been made for the stratigraphic thickness of the Peripheral Schieferhülle, but in view of the very high degree of internal disorder of these rocks and the extreme flattening which was discussed in Chapter IV, these estimates cannot be regarded as very reliable. TOLLMANN (1963) has estimated a stratigraphic thickness of 2 *km*. In the Glockner depression the present vertical thickness, with tectonic repetition, is over 5 *km*.

The next major event in the geological history of the area must have been the overriding of the Hercynian basement and its autochthonous cover by the relatively northeast-moving Ostalpin thrust sheets. The time at which this took place, and whether these movements were episodic or continuous, rapid or relatively slow, are at present all matters for speculation. The stratigraphic evidence is ambiguous; TOLLMANN (1963) and CLAR (1965) have favoured a late Cretaceous age for the movements while OBERHAUSER (1968) has presented arguments for the movement of the Oberostalpin over the Tauern during the Eocene. Observations made during the present study have not contributed to the solution of this problem. Radiometric ages in Ostalpin rocks surrounding the Tauernfenster have provisionally been interpreted as indicating that the main northward movements occurred at about 65 my (BREWER, 1969; BREWER and OXBURGH, 1972). It is also of considerable importance to establish the age of the detrital metamorphic minerals which have been recognized in the unmetamorphosed Mesozoic and Tertiary formations on the flanks of the Alps (WOLETZ, 1967). In particular OBERHAUSER (1968) draws attention to the widespread occurrence of detrital glaucophane in the Walserberg sandstone (Cenomanian of Salzburg); if this were the product of some Mesozoic metamorphic episode in the Eastern Alps, it would have far-reaching implications for the history of the Tauern.

The thickness of the overriding Ostalpin thrust sheets was probably of the order of 11 *km*. (Chapter III). The Peripheral Schieferhülle which underlies them shows evidence of only one main episode of folding; on a regional scale a single direction of rotation (to the northeast) is recognized and the amount of strain is much greater than in the basement; in so far as it is difficult to imagine the northward movement of the Ostalpin not having an important deformational effect on the rocks beneath, we conclude that this movement was responsible for the folding of the Peripheral Schieferhülle, i.e. that the thrusting of the higher units was responsible for the $F_A{}^1$ deformation as whole, flattening existing folds in the Inner Schieferhülle, locally generating new ones, and in many places imprinting a foliation on the Zentralgneis. The orientation of $F_A{}^1$ fold structures show that the movements of the Ostalpin were towards the northeast rather than the north. Pre-Alpine structures are commonly preserved within the Inner Schieferhülle, modified to varying degrees. Less commonly they are found in the Zentralgneis.

The Alpine structures developed in the basement rocks are very variable both in their orientation and in the intensity of their development. As discussed above this results from the irregular distribution of Inner Schieferhülle and Zentralgneis within the basement and the differences in their deformational behaviour. Even within the larger Zentralgneis bodies it is clear that they did not everywhere undergo a thoroughgoing, penetrative Alpine ($F_A{}^1$) deformation; certain zones, behaving almost like the augen within augen-gneiss, persist today as relatively undeformed masses surrounded by zones of much more intense Alpine deformation. The basement in the area mapped is furthermore cut by important low-angle shear zones and it may be that the basement in the Eastern Tauern is a series of imbricated slices, piled upon each other. There is little direct evidence, but the similarity in lithology and pre-Mesozoic geological history throughout this part of the basement suggests that the movement between such slices is unlikely to exceed a few kilometres.

It is notable that nowhere within the area mapped is the Peripheral Schieferhülle interfolded with the basement complex on a small scale, although mesoscopic and smaller-scale F_A^1 structures are found both in the Peripheral Schieferhülle and in the basement; wherever the contact between the two is seen, it is a relatively smooth surface and a zone of shearing, with the foliation of units both above and below the contact, parallel to the contact and to each other. The contact zone is apparently one of intense movement. On a larger scale the contact defines the complex three-dimensional form of domes and depressions already discussed in Chapter IV (Fig .IV—26).

The Sonnblick Kern is a northwest-trending basement antiform protruding from beneath the Peripheral Schieferhülle. It has the appearance of a "proto-nappe" — an incipient nappe which never developed an allochthonous nose; its southeastern, tapering extension, however, the Sonnblick lamella of the Mölltal, is extremely flattened and could have had a thin tongue-like allochthonous continuation northeastwards over the Tauern.

Although the basement and Peripheral Schieferhülle are not interfolded, certain large-scale intercalations do occur. Perhaps the most interesting are those represented by the gneiss lamellae within the Schieferhülle of the Mallnitzer Mulde (Chapters I and IV) which appear to represent thin, allochthonous basement slices derived from the south and carried northwards along with the Peripheral Schieferhülle. Indeed their existence is a powerful argument in favour of the allochthonous nature of at least all the Peripheral Schieferhülle which overlies them. Any northward continuations of these lamellae over the Tauern have today been removed by erosion.

Other major intercalations of cover rocks within the basement occur within the Eastern Tauern. These support the concept of allochthonous gneiss units. A good example is the gneiss mass forming the most northeasterly basement outcrop within the Tauernfenster (Fig. I—2). This mass, which has been termed the Mureck Decke by Exner (1940), dips northwards and eastwards beneath cover rocks, but is apparently also underlain by the cover rocks of the Silbereck Mulde which separate it from the main basement units of the present map area. Exner's mapping suggests that the Mureck Decke is the northern part of an allochthonous sheet which previously extended southwards over the Tauern.

As was noted earlier the gneiss lamellae are folded down into the Mallnitzer Mulde synform, indicating that they had been transported over the area of the eastern Tauern before the development of the present pattern of zones of elevation and depression.

We thus derive the following generalized sequence of events:

(1) Mesozoic deposition upon an eroded, inhomogeneous Variscan basement probably containing pre-, syn- and post-tectonic intrusive elements.

(2) In the late Mesozoic or early Tertiary, horizontal movements of the Ostalpin allochthonous thrust-masses over the Tauern. These were possibly accompanied by the partial or complete "tectonic erosion" of any autochthonous sedimentary cover in the Tauern, and the "tectonic deposition" of an intensely deformed and stratigraphically disordered Mesozoic sequence derived from the southwest. It appears that the intense folding of this Peripheral Schieferhülle sequence was associated with horizontal movements which transported it to its present situation. Subordinate 'basement slices' were also dragged northward (or rather northeastward) at this time. The main phase of metamorphic crystallization did not occur until these movements were largely complete.

(3) Differential vertical movements within the basement led to the formation of the present irregular pattern of domes and intervening depressions. In the present area these movements must largely postdate the horizontal movements because on a regional scale the F_A^1 axial surfaces are somewhat rotated; associated small scale structures do not seem to have been developed. It is, however, noteworthy that the

trend and sense of asymmetry of the Sonnblick Kern antiform are the same as those of the $F_A{}^1$ folds in the Peripheral Schieferhülle. This could indicate that the elevation of the Sonnblick antiform had begun before the conclusion of the horizontal movement of the Ostalpin. Even in the present area vertical movements may have begun before the horizontal movements were complete; if this is the case, greater thicknesses of Peripheral Schieferhülle may have accumulated tectonically in tectonic depressions than above rising domes. Because of the absence of associated smaller scale structures, the relationship of these vertical movements to the main metamorphism is unclear. Further information on the form of the albite-oligoclase isograd surface should clarify this problem.

So far the discussion has centred on the transfer and movement of material; we now turn to the problems of temperature distribution and the transfer of heat. In this context there are several observations of importance.

(1) It was shown in Chapter V that the K/Ar age pattern is apparently not related to the tectonic relief on the basement/cover interface (Fig. V—3). This implies that there was time for some degree of thermal equilibration after the development of the gneiss domes etc.

(2) This conclusion is supported by the behaviour of the albite/oligoclase isograd (Fig. III—5) which does not parallel the basement/cover interface but in the north-west part of the area swings out of the Peripheral Schieferhülle into basement rocks.

(3) There is no evidence of Mesozoic or Tertiary intrusive igneous activity within the Tauernfenster; thus, at the present level of erosion, magmas did not play any part in the processes of heat transfer. The late mineralization along joints in both basement and cover must, however, have involved the passage of hydrothermal fluids.

(4) South of the Tauernfenster igneous intrusions of Tertiary age are found within the Ostalpin; the most northerly of these is the small granodiorite plug at Wöllaner Kopf, just four kilometres south of the edge of the window.

In a region as strongly affected by large-scale movements as the eastern Tauern, the thermal history is likely to be complex. The temperature profile of the region at any instant in time reflects a shifting balance between, on the one hand, the general tendency of processes of thermal conduction to establish an equilibrium in which isotherms are broadly parallel to fixed-temperature boundaries (e.g. the surface of the earth) and thermal gradients parallel to such boundaries are small; and on the other hand, departures from this state caused by large-scale and relatively rapid, tectonic movements and any localized transfer of heat by magmas within the crust.

We consider first how the vertical temperature profile within the area mapped, may have varied with time (Fig. VI—1). Let (a) represent the situation in late Cretaceous times immediately before the overriding by the Ostalpin; a linear thermal gradient of $15°/km.$ (a "normal" geothermal gradient) is assumed. Immediately after burial of the basement by the Ostalpin and by any other material carried along beneath or above it, the profile appears as in (b); the profile contains a sharp minimum at the interface between the allochthonous material and the basement. With time, this minimum decays by thermal conduction (c) both upwards and downwards (note that initially the temperature of the lower part of the allochthon may fall). Assuming uniform conductivities throughout, and a constant flux of heat to the base of the system, the profile will slowly evolve through stages t_1, t_2, t_3, t_4 and t_5. Ultimately, if the system undergoes no further perturbations the mean temperature of the autochthon will have risen about 200° C and the temperature at the base of the allochthon will have returned to its initial value (case b [II]) or exceded it (case b [I]).

It has been shown, however, that shortly before the main horizontal movements were complete, differential vertical movements within the basement were intiated which finally gave rise to more than 5 *km.* of relief on the interface between basement and

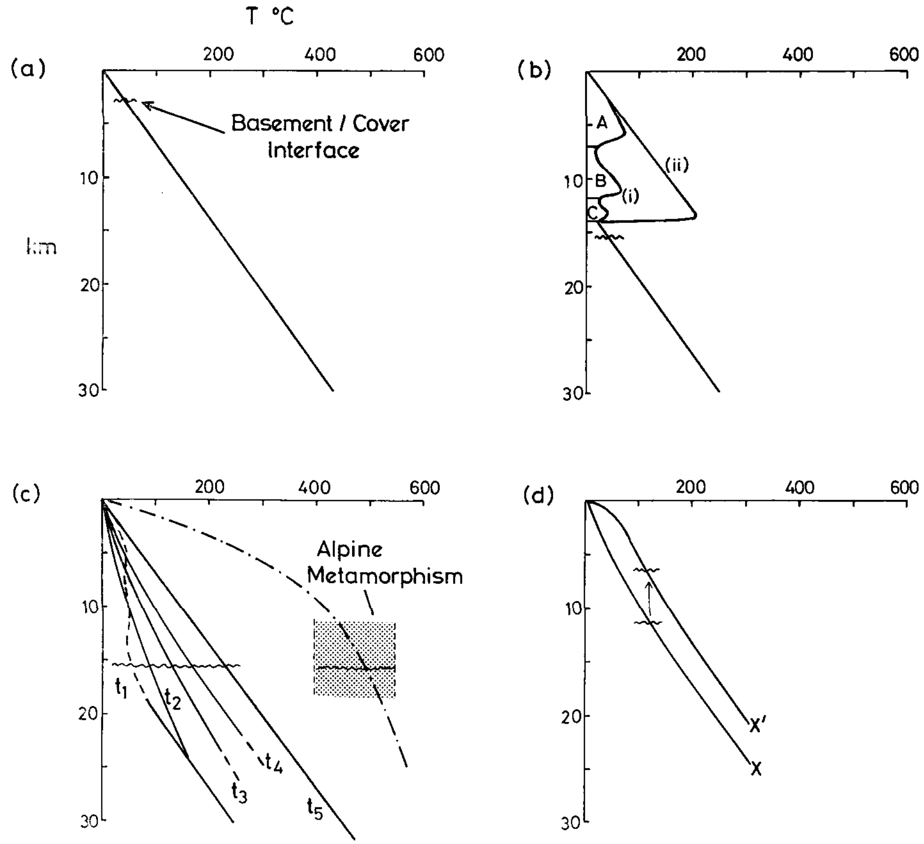


Fig. VI - 1 (a) Temperature distribution with depth in cover and basement before overthrusting, for a geothermal gradient of 15° C/*km*. (b) Temperature distribution immediately after overthrusting by (i) a series of three thrust-sheets (A, B, C) and (ii) a single thick sheet; prethrusting thermal gradient in allochthonous material 15° C/*km*.; the temperature in the autochthon, i. e. material below C, is the same in either case. (c) a schematic representation of the stages (t_1—t_5) by which the crust might return to conductive equilibrium after overthrusting (b); t_2 is the earliest stage and follows from b(i); the earliest stages would be somewhat different for b(ii). In either case the final equilibrium gradient is t_5, i. e. 15° C/*km*. in both allochthon and autochthon. (d) the effect of uplift in a hypothetical situation; provided that uplift occurs at a rate which is high by comparison with the conductive transfer of heat (say, more than a few *mm*./yr) the thermal gradient (X) is, in the short term, unaffected at depth, and is simply displaced upwards (X′) and "beheaded" giving relatively higher temperatures at shallow depths.

cover (Fig. IV—26). These movements may have had their expression on the upper surface of the overlying Ostalpin and brought about erosion of the elevated areas, resulting in the "beheading" the thermal gradient in the elevated areas (d). It does appear, however, that these movements either occurred very slowly so that departures from conductive equilibrium in the near surface areas were rather small or were at any rate followed, by a period of relative quiescence long enough for such equilibrium to be restored. That such an equilibrium had been attained by the Miocene, i.e. the time of uplift of the Tauern, is demonstrated by the pattern of K/Ar ages; as discussed in Chapter V the absence of systematic differences in cooling ages between domes and intervening depressions, indicates that before the final phase of uplift there was time for the restoration of thermal equilibrium after the tectonic movements which produced this relief. The same conclusion is suggested by the behaviour of the albite/oligoclase isograd as discussed earlier.

One implication of the model presented so far is that the Peripheral Schieferhülle never reached a temperature higher than about 250° C and attained that temperature only after a very long time (ca. 20 my). For the regional metamorphism of the Peripheral Schieferhülle, however, the combination of temperature and pressure indicated in Fig. VI—1 c is required (Chapter III). Our model suggests that this situation will never be attained; the model is therefore in some way deficient. There are broadly three possibilities (1) the mean thermal gradient for the whole region during the late Mesozoic and Tertiary was higher than assumed i.e. 30—40°/*km.* rather than 15°/*km.* (2) that the depth of burial and the amount of post-metamorphic erosion was much greater than assumed. (3) that there was an additional and transient source of heat available to the system during the mid-Tertiary (e.g. Exner, 1954).

Detailed calculations will be presented elsewhere but it can be shown that although (1) is theoretically possible, rates of attainment of thermal equilibrium by conduction are so low that unless the gradient assumed is unreasonably high, in the time available it is not possible to attain the appropriate temperatures at the required depth. For the reasons given in Chapter III we regard (2) as unlikely. We therefore favour alternative (3). This more or less implies the emplacement of magmatic bodies beneath the Tauern during the Tertiary. Although they are not exposed at the present level of erosion, the local occurrence of thermal springs, the mineralization along steep joints and the high heat-flux measured at the surface (Clark, 1961; Boldizsar, 1968), may reasonably be attributed to them. Our conclusions may be compared with those of Clark and Jäger (1969) discussed in Chapter V.

So far little has been said of horizontal variations in temperature. Less information is available but there is some suggestion of a general increase in temperature laterally towards the south. This is indicated by (1) the presence of the Peri-Adriatic tonalitic intrusives in the Ostalpin south of the Tauern; these intrusives were emplaced after the main movements northwards and at about the time of the main metamorphism in the Tauern (Chapters III and V). Even though a much deeper tectonic level is exposed in the Tauern than further south, no such intrusives are found; this implies that during the mid-Tertiary magmatic transfer of heat to shallow depths was taking place only south of the Tauern and therefore that there the temperature at any given depth was likely to be higher in the south than in the north.

This is consistent with the variation in metamorphic grade within the Peripheral Schieferhülle. Figure VI—2 shows the distribution of the main metamorphic minerals recorded in the eastern Tauern. It appears that only at the base of the Peripheral Schieferhülle on the south side of the Tauern have temperatures high enough for the growth of staurolite been attained.

A lateral variation in temperature may also be reflected in the differences in behaviour of the basement complex. From northeast to southwest three different types of deformational behaviour may be recognized:

(i) the relatively inert basement massifs of the north within which our map area lies; the basement is cut by shear zones and may be imbricated; in most places an Alpine foliation is seen but locally it is very weak and virtually only the quartz fabric is affected.

(ii) the Sonnblick antiform: here too Exner (1964) has shown that some pre-Alpine features may remain relatively undeformed, but the form and asymmetry of the antiform as a whole suggests that its physical state allowed it to participate more fully in the Alpine movements; at its southeastern end the Sonnblick lamella appears to have been derived from the main antiform by more extreme penetrative deformation.

(iii) the gneiss lamellae of the Mölltal: these must have been derived from south of the Tauernfenster beneath the Ostalpin; they have suffered intense internal de-

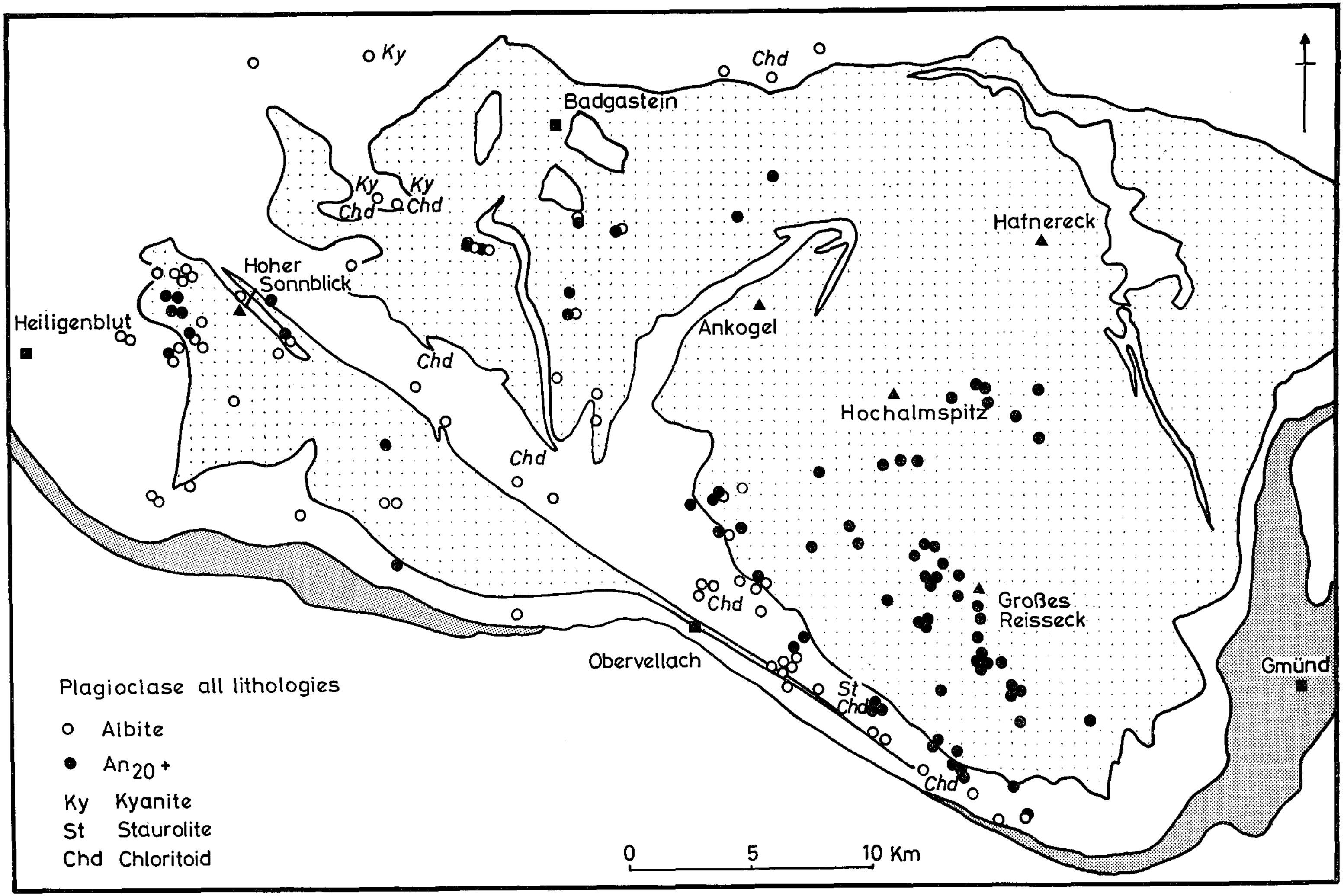


Fig. VI - 2 The distribution of some significant metamorphic minerals in the eastern Tauernfenster from the authors' observations and the literature. In addition there are kyanite and chloritoid occurrences 4 *km* due east of Obervellach and a staurolite locality close to the most southeasterly chloritoid occurrence.

formation and could represent the attenuated cores of nappes derived from basement highs similar to the Sonnblick antiform, but situated further south, and today buried beneath the Ostalpin.

These apparent differences in response of the basement to the horizontal movements of the Ostalpin could reflect a general increase in ductility southwards corresponding to an elevation in temperature; alternatively they could simply reflect the earlier formation of basement 'highs' in the south and their consequent greater involvement in horizontal movements.

One might speculate that the equivalent of the Lepontic zone of Central Alps (characterized by very high temperature regional metamorphism) ist to be found in the Penninikum buried beneath the Ostalpin rocks of the Kreuzeck Group and the Drauzug. Such high temperatures were not reached in the Tauern and at the thermal maximum between 55 and 35 my ago mineral assemblages varying from the greenschist facies of regional metamorphism to amphibolite facies were developed, depending upon depth.

Some ten million years later, in the Miocene, important vertical movements began to affect the Tauern. The pattern of K/Ar ages suggests that these began earlier to the west of the present map area and extended eastwards. Over much of the Tauern the uplift may have been accomplished by a broad regional warping. This, however, was not the case along the easternmost part of the southern margin of the Tauernfenster where the edge of the window corresponds to the Mölltal line. This part of the window margin is delimited by a steep vertical fault which appears to bring a relatively high-level in the Ostalpin rocks on the south side of the Mölltal against Peripheral Schieferhülle. This steep fault probably extends some distance to the southeast as the Mölltal line, but once outside the Tauernfenster it merely causes a displacement within the Ostalpin; although it is difficult to demonstrate such a displacement from geological observations, Brewer (1970) has shown that an important break in the Ostalpin pattern of K/Ar ages occurs along the line.

These late vertical movements took place in rocks at relatively low temperatures, and unannealed basal strain lamellae are common in quartz grains from the Mölltal Schieferhülle.

As the Tauern rose and were domed up, a regional pattern of joints developed facilitating the passage of mineralizing fluids. The detritus from the rising mountain chain was shed into shallow water basins to both north and south and the changing heavy mineral content of the sediments with time reflects the progressive unroofing of deeper and deeper tectonic levels within the Tauern.

Appendix I

Aerial Photograph Interpretation

by R. C. Wright

The map-area is covered by an overlapping mosaic of 210 aerial photographs (scale approximately 1 : 17,000). In this study, aerial photographs have been used for four main purposes: location whilst in the field, interpretation of geology, study of lineaments, and mapping of geomorphological features.

The value of using aerial photographs as a location technique, especially on high ground with local topographical change, has already been pointed out (Chapter I). The use of aerial photographs to interpret the geology has been limited, chiefly because few of the rock-units have sufficient contrasting photographical tone to be recognised individually. However, some success was achieved by using aerial photographs in conjunction with fieldwork to map structures in the Peripheral Schieferhülle of the steep and heavily-forested Mühldorfer Graben—Plankogel—Taborgraben area in the extreme southern part of the map.

The main value of the aerial photographs has been in the mapping of the lineament pattern and the geomorphology. Detailed analysis has been made on stereoscopic pairs of photographs, covering the map-area. Prominent linear depressions in areas of bare rock, areal stream and gully alignments in areas of dissected topography, and major changes in slope on the main valley sides have been mapped as indicators of the lineament pattern. On the sides of the Mölltal, changes in vegetation were of limited value as lineament indicators. In mapping from the photographs, however, areal constancy of lineaments, and possible lineaments, was always a very important consideration. The study of the geomorphology using the aerial photographs involved the demarcation of glaciers and permanent snow-fields, moraine, boulder fields, scree chutes, stream debris-cones and -fans, and river alluvium.

Some 948 lineaments have been recognized from the aerial photograph cover of the map-area, and their lengths range from 100 *m.* to 4 *km.* (e.g. the major lineament running northeast-southwest from the head of the Radl Graben to the Mühldorfer Graben). The results of the study are shown on Fig. A I—1, and have been summarised in the form of rose-diagrams (Fig. A I—1, A I—2).

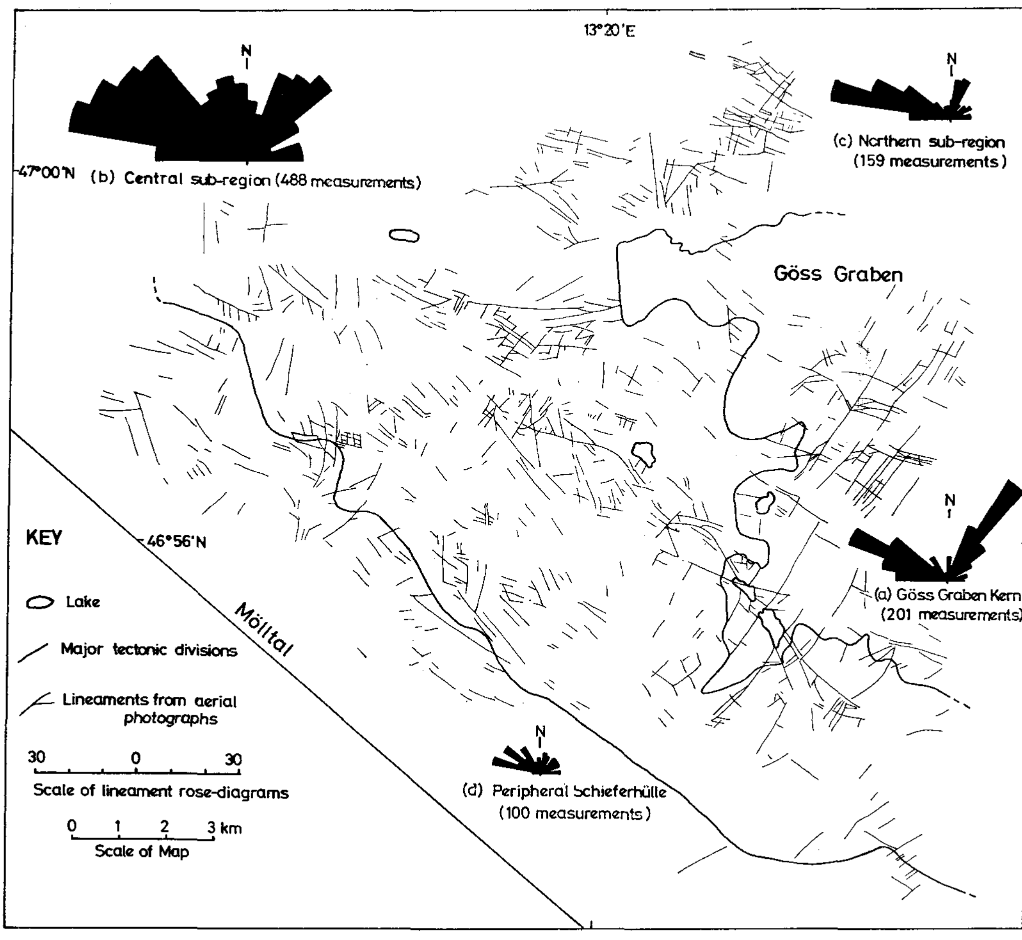


Fig. AI - 1 Lineament distribution pattern and rose-diagrams for major sub-regions.

The results of the geomorphological study are incorporated in the main geological map (Plate 1); the areas of glacier and snowfield, Mölltal alluvium, and moraine, boulder fields, scree and debris-fans are distinguished by a series of dashed and dotted lines. Although it was generally possible to distinguish moraine, boulder-scree, debris-cones etc. on the aerial photographs, their relationships were in many areas complex and it proved impractible to show these distinctions on an already complex map. The majority of superficial deposits have, therefore, been grouped together. The map does show, however, the major areas where the solid geology is concealed by superficial deposits.

Interpretation of the results of the lineament study is based upon comparisons of rose-diagrams of the different sub-regions. The sub-regions were chosen to correspond to the major tectonic units, (i.e. Göss Graben Kern, Hochalm Kern and Reisseck Mulde, and Peripheral Schieferhülle). In addition, the Hochalm Kern/Reisseck Mulde unit was arbitrarily subdivided across the head of the Göss Graben to the Dösener See, to give a northern sub-region (Hochalm Sp. and Villacher Hütte area) and a central sub-region (Dösenertal to Hintereggen Graben).

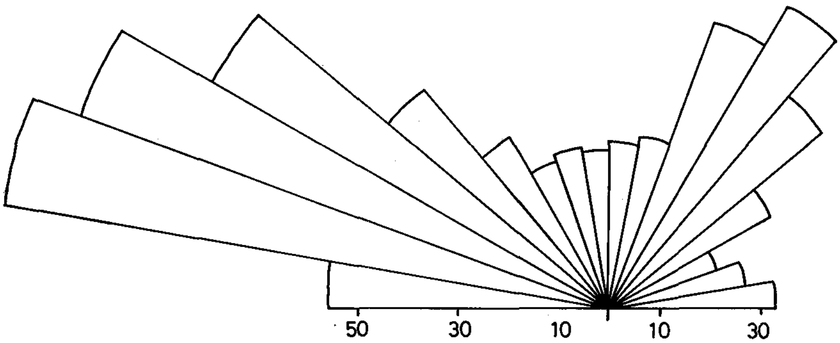


Fig. AI - 2 Rose-diagram showing total lineament pattern (948 measurements).

The rose-diagram of the lineament pattern showing total measurements (Fig. A I—2) indicates a broad overall westnorthwest-eastsoutheast and northeast-southwest conjugate set of lineaments. Within the Göss Graben Kern (Fig. A I—1 [a]), this conjugate set is well-developed, but there is a dominance of northeast-southwest lineaments. The Hochalm Kern and Reisseck Mulde of the central sub-region (Fig. A I—1 [b]), on the other hand, show a dominance of westnorthwest-eastsoutheast lineaments over the northeast-southwest trend, and in addition, a set of northnorthwest-southsoutheast lineaments is developed. The northern sub-region (Fig. A I—1 [c]) shows the westnorthwest-eastsoutheast and northeast-southwest conjugate set of the overall lineament pattern, with a marked dominance of the former trend. Lineaments in the Peripheral Schieferhülle (Fig. A I—1 [d]) show a very broad westnorthwest-eastsoutheast and northeast-southwest pattern, with an important development of lineaments which parallel the Mölltal Line.

Thus, all tectonic units in the map-area show a development of the westnorthwest-eastsoutheast and northeast-southwest conjugate set of lineaments. The development of this basic set of lineaments presumably postdates virtually all other movements, and certainly lineaments are seen to cut straight across the major low-angle faults and shear zones marking the main tectonic divisions. In local areas, one lineament trend can appear to truncate other trends; elsewhere it seems this trend is itself offset by the others; but very often neither trend appears to be affected at a crossing-point: thus there is no consistent truncation of one trend by another over any large area.

Within the different tectonic units, the two main trends are developed to different degrees. Within the Göss Graben Kern, high, bare rock plateau areas (e.g. Tröska Alm and Hohenkar) tend to record a dominance of northeast-southwest lineaments, whilst very similar topography in the northern sub-region of the Hochalm Kern (e.g. Hochalm Kar, southwest of the Villacher Hütte) records a marked dominance of westnorthwest-eastsoutheast lineaments. From a general consideration of the map showing the lineament distribution pattern (Fig. A I—1), it can be seen that the observed northeast-southwest lineaments dominate the whole eastern and south-eastern area of the map, and that elsewhere this trend is only apparent to a lesser degree. In some areas, other lineament trends overprint the basic set to give a more complex rose-pattern, e.g. the development of lineaments parallel to the Mölltal Line in the Peripheral Schieferhülle (Fig. A I—1 [d]), or the development of northnorthwest-southsoutheast trends in the central sub-region (Fig. A I—1 [b]).

In any study of lineament pattern, consideration of large-scale topographical alignment is very important. In the map-area it is of interest to note that the Dösen, Kaponig, Zwenberger, Rieken, Mühldorfer, Ritter and Tröska Alm valleys all tend to parallel the northeast-southwest lineament direction, whereas the Radl and, to a lesser extent, the Hintereggen valleys parallel the westnorthwest-eastsoutheast lineaments and the Mölltal Line.

The study of the lineament pattern from the aerial photographs provides information about the most recent stress environment to which the rocks of this part of the South-East Tauern have been subjected. The development of a westnorthwest-eastsoutheast and northeast-southwest conjugate set of lineaments can probably be related to the final uplift of the Tauern, with local stress field variations causing local lineament pattern variations (e.g. trends paralleling the Mölltal Line). It will be of considerable interest to compare these lineament trends with those developed in the higher tectonic units which overlie the Pennine Zone and surround the Tauernfenster.

Appendix II

Methods of calculation of rock strain

by R. J. Norris

Dolomite Quartzite Breccia

Ramsay (1967) and Dunnet (1969) have criticised calculations of strain based on arithmetic means of axial ratios of deformed pebbles. They consider the deformation of a set of generally ellipsoidal, randomly oriented particles, and show that if the axial ratio is plotted against the angle between the long axis of the particle and some chosen direction within the reference plane, a symmetrical figure will result. The line of symmetry of the figure is parallel to the long axis of the strain ellipse within the plane, and the true axial ratio of the strain is given by

$$R_t = \sqrt{R_T \max R_T \min} \qquad \text{(Ramsay, 1967. p. 210)}$$

where R_t = Strain axial ratio $\frac{X}{Y}$ within plane

R_T = pebble axial ratio

If the ellipsoids initially had a preferred orientation, the distribution will be asymmetric.

Dunnet (1969) concludes that at least 40 measurements are required to define this distribution adequately, but unfortunately in the present study, it was not possible to collect this number of measurements. GAY (1969) shows, however, that even a crude calculation based on a small number of measurements may give results surprisingly close to those obtained after a much more detailed analysis.

Despite the paucity of measurements, it was considered that a rough calculation based on geometric means might indicate the order of magnitude of strain. For the dolomite quartzite breccia of the lower Kaponigtal the axial ratios of the pebbles were measured on sections sawn parallel and perpendicular to the pebble elongation and the strain axial ratios calculated from the geometric means. Each one was then calculated from the other two, the average value taken, and the process repeated until agreement to within 0.1 was reached. In fact, very little adjustment was necessary, suggesting that the measured values were fairly representative. The results are:

	Number of measurements	Arithmetic mean	Std. Dev.	Geometric Mean
Y/Z	17	2.05	0.56	2.1
X/Y	9	4.9	1.36	4.6
X/Z	12	10	1.85	9.6

Axial ratio of strain ellipsoid $= 9.6 : 2.1 : 1$

$$K = 3, \text{ where } K = \frac{Z(X-Y)}{Y(Y-Z)}$$

This result is in marked contrast with other strain indicators in the Peripheral Schieferhülle. The fabric suggests that the strain has been predominately an unilateral flattening. Two methods of producing elongated pebbles parallel to the fold axis have been considered. During a rotational strain, the axes of each succesive strain increment are not parallel to the axes of the finite strain ellipsoid. It is possible for the long axis of the latter, during progressive deformation, to be rotated through the surface of no infinitesimal strain and so be shortened again. In such a situation, for instance in the short limb of an asymmetric fold, the net result will be an elongation parallel to the rotation axis. This is what happens to an irregularly shaped lump of putty when it is rolled beneath the hand against a hard flat surface. Consider a flattening type strain (K = O, FLINN, 1962) in which the principal quadratic elongations are $\lambda_x, \lambda_x, \frac{1}{(\lambda_x)^2}$. If a rotation parallel to the Y axis is occurring, only in this direction will the strain be the same as in the case of no rotation. However the principal compression, $\frac{1}{(\lambda_x)^2}$ may still be calculated from this. In the present case, the elongation parallel to the fold axis may represent the effect of rotation of the short limb of an asymmetric fold within a unit undergoing a flattening type strain. In this case $\lambda_2 = \frac{1}{(\lambda_y)^2} = 0.064$ or $\varepsilon = -2.50$ (in logarithmic strain).

An alternative mechanism is suggested by RAMSAY (1967, p. 220), namely that the pebbles were initially disc-shaped, lying within the plane of the bedding, and subsequent compression parallel to the bedding has produced elongated pebbles.

Consider a disc-shaped pebble, radius r and thickness t, let the axes of the pebble before strain be A, B, C, and after strain be A′, B′, C′, where C is perpendicular to the bedding and B is parallel to the long axis of the strain ellipse within the bedding. The pebble is now deformed by a flattening-type strain, with $\lambda_x = \frac{1}{(\lambda)^{1/2}} = \lambda_y$, $\lambda_z = \lambda$ where λ_x is parallel to C, λ_y to B and λ_z to A.

Then,

$$(1.)\qquad A = r; \qquad A' = \frac{r}{(\lambda)^{1/2}}$$

$$B = r; \qquad B' = \frac{r}{(\lambda)^{1/4}}$$

$$C = t; \qquad C' = \frac{t}{(\lambda)^{1/4}}$$

The calculated extension parallel to B′, assuming (mistakenly) an initially spherical article, will be

$$(2.)\qquad (\lambda_B)^{1/2} = \left(\frac{r}{t}\right)^{1/3} \cdot \frac{1}{(\lambda)^{1/4}};$$

Let $p = \frac{C'}{A'}$, then

$$(3.)\qquad \left(\frac{r}{t}\right)^{1/3} = \frac{1}{(p)^{1/3}} \cdot \frac{1}{(\lambda)^{1/4}}$$

$$(4.)\text{ from eqn. (2),}\qquad (\lambda_B)^{1/2} = \frac{1}{p} \cdot \frac{1}{(\lambda)^{1/2}}$$

$$(5.)\text{ i.e.}\qquad (\lambda)^{1/2} = \left(\frac{1}{p}\right)^{1/3} \cdot \frac{1}{(\lambda^B)^{1/2}}$$

In the present case, $\frac{C'}{A'} = 2.1$, and $(\lambda_B)^{1/2} = 3.55$

so that $(\lambda)^{1/2} = 0.22$ ($\varepsilon = —1.50$).

Thus limiting values of $\varepsilon = —1.50$ (in the case of RAMSAY's theory) and $\varepsilon = —2.50$ (in the case of the rotation theory) may be set.

Strain Calculations from Buckled and Boudinaged Veins

In the veins measured in this investigation, the thickness at the hinges of the folds was greater than on the limbs. This is due to a component of homogeneous strain. By measuring the orthogonal thickness (s) of the vein in various directions at an angle, φ_1, to a given direction, the initial thickness may be derived by plotting the polar co-ordinates (s, φ). The resulting figure is symmetrical about two orthogonal axes, the axes of the homogeneous strain ellipse, and is described by the relation

(6.) $s^2 = p^2 (\lambda_1 \cos^2 \varphi + \lambda_2 \sin^2 \varphi)$ (RAMSAY, 1967 p. 412; NORRIS, 1970) where p = initial thickness, φ = angle between thickness vector and X axis of strain ellipse, λ_1 and λ_2 = principal strains. In cartesian co-ordinates, (7.) $x = p\ (\lambda_1 \cos^2 \varphi + \lambda_2 \sin^2\varphi)^{1/2} \cos\varphi$

$y = p\ (\lambda_1 \cos^2 \varphi + \lambda_2 \sin^2 \varphi)^{1/2} \sin\varphi$

when $\varphi = 0$, $x_0 = p(\lambda_1)^{1/2}$; when $\varphi = 90°$, $y_0 = o\ (\lambda_2)^{1/2}$

so that $(x_0 y_0)^{1/2} = p(\lambda_1\lambda_2)^{1/4}$

if the strain is a pure shear, $\lambda_1 = \dfrac{1}{\lambda_2}$

(8.) i.e. $(x_0 y_0)^{1/2} = p$

Measurements were made on photographs taken at right angles to the fold axes. All the veins examined (from within the Reisseck synform) showed little or no folding or boudinage parallel to the fold axis. All were approximately perpendicular to the $S_A{}^1$ foliation, and as they are related to the granite intrusions, record only the subsequent strain. The cross-sectional area was measured (A), and the original length calculated from $l_0 = \dfrac{A}{p}$. The present orthogonal length of the vein (l_1) was measured, and the strain, $(\lambda)^{1/2} = \dfrac{l_1}{l_0}$, calculated.

However, the viscosity ratios calculated from the Biot equation (Biot, 1961) were very low, suggesting that a considerable amount of layer shortening during the early stages of buckling has taken place. This problem has been discussed by Sherwin and Chapple (1968) and use was made of their method of calculating this component. The measurements of dominant wavelength, thickness and amplitude substituted into their equations, were first recalculated to remove the later homogenous strain. The results are given in Table IV—1.

Boudinaged veins may also be used to estimate the amount of elongation. The cross-sectional areas (A) of each boudin may be measured together with the maximum thickness (h), and the "initial" length (l_0) found. By adding these together, and measuring the present length of the vein, the elongation parallel to it may be calculated from $(\lambda)^{1/2} = \dfrac{l_1}{\Sigma l_0}$ where $l_0 = \dfrac{A}{h}$.

This does not allow for a homogeneous stretching of the layer prior to failure or for any homogeneous flattening of the boudins after rupture. It is not possible to allow for any initial stretching without a detailed knowledge of the behaviour of the rocks during deformation, their physical properties and other parameters such as the rate of strain. With regard to a homogeneous flattening of the boudins after rupture, Gay (1968) has investigated the behaviour of viscous ellipsoidal particles during a pure shear deformation, and derived the following relationship:

(9.) $$\ln\left(\frac{a}{b}\right) = \ln\left(\frac{a_i}{b_i}\right) + \left[\frac{5}{(2R+3)}\right] \ln\left(\frac{\lambda_1}{\lambda_2}\right)^{1/2}$$ (Gay, 1968, eqn. 16)

where a_i, b_i, a and b are the initial and final semi-axial lengths of the particle cross-section, $R = \dfrac{\eta}{\eta_1}$ = viscosity ratio, and λ_1 and λ_2 are the principal quadratic elongations.

Let $\dfrac{l_1}{\Sigma l_0} = (\ \lambda_s\ )^{1/2}$; then

(10.) $\ln \lambda_t = \ln \lambda_v + \ln \lambda_s$

where λ_t = total elongation, λ_v = homogeneous strain since onset of necking, λ_s = strain calculated from separation, also, $\frac{a}{a_i} = \frac{b_i}{b} = (\lambda_v)^{1/2}$, and for pure shear, $\lambda_1 = (1/\lambda_2) = \lambda_t$

Eqn. (9) then becomes:

$$\ln \lambda_v = \left[\frac{5}{(2R+3)}\right] \ln \lambda_t = \left[\frac{5}{(2R+3)}\right] (\ln \lambda_v + \ln \lambda_s)$$

$$\ln \lambda_v = \left[1 - \frac{5}{(2R+3)}\right] = \left[\frac{5}{(2R+3)}\right] \ln \lambda_s$$

$$\ln \lambda_v = \left[\frac{5}{(2R-2)}\right] \ln \lambda_s \qquad (11.)$$

$$\text{or} \qquad \varepsilon_v = \left[\frac{5}{(2R-2)}\right] \varepsilon_s \qquad (\varepsilon = \tfrac{1}{2} \ln \lambda)$$

Thus by assuming a value for R, the viscosity ratio, a value for the homogeneous strain since necking may be calculated. A value of R = 12 has been assumed for the present calculations, based on the values obtained from the measurement of buckled veins. In most of the Reisseck synform, as already mentioned, the boudins are elongate structures parallel to the axial direction. Measurements were made in the plane perpendicular to this direction, and a plane strain $\left(\lambda_1 = \frac{1}{\lambda_2}\right)$ assumed in calculating the principal compression. Towards the contact with the Peripheral Schieferhülle, measurements were made in two directions at right angles, and in the marginal zone, an uniaxial compression $[\lambda_1 = 1\ (\lambda_2)^{1/2}]$ was assumed as the boudins here are disc-shaped. The principal compression was calculated for comparison between these localities. The results on boudinaged veins are given in Table IV—II and all results are shown in Fig. AII—I. It is clear that the strains in the marginal zone are considerably greater

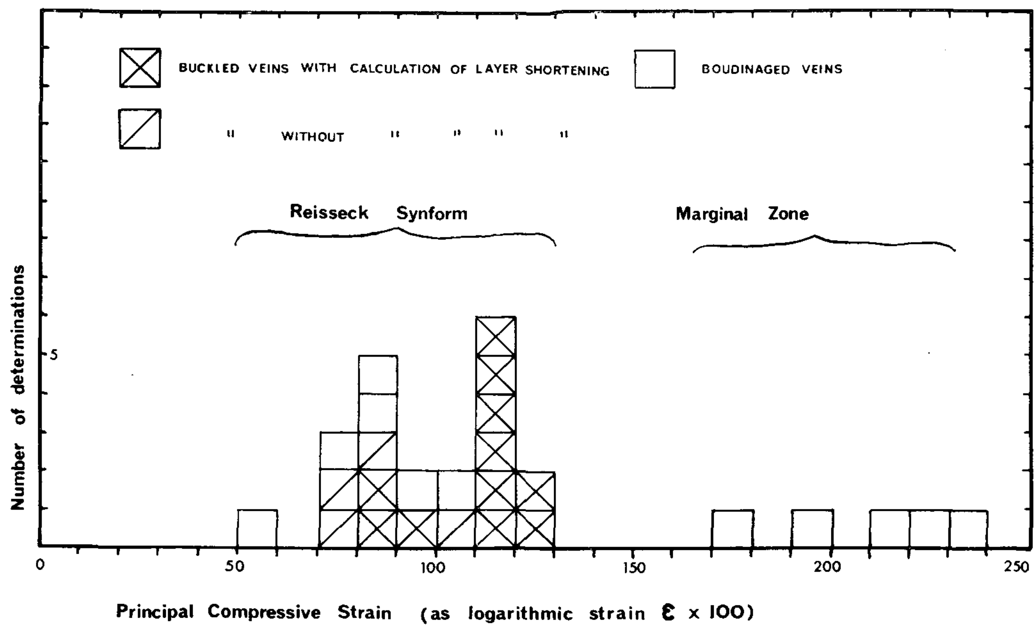


Fig. AII - 1 Histogram of compressive strains measured on the Inner Schieferhülle by different methods.

than those in the Reisseck synform. Within the Reisseck synform, the measurements are fairly consistent, those of boudinaged veins overlapping those of buckled veins. As a group, however, the boudin measurements tend to be lower in value. The explanation for this is probably twofold:

(1.) the strain is not planar, so thar there is some extension perpendicular to the plane of measurement which has not led td boudinage.

(2.) homogeneous elongation of the layer occurred prior to necking, analagous to the layer-shortening during the early stages of buckling. This element of strain is probably the main reason for the discrepancy, and thus the measurements of boudinaged veins must be considered as minimum estimates only.

The limits of accuracy of these measurements are difficult to estimate as the method depends upon certain assumptions which, if incorrect, would seriously diminish the accuracy of the results.

With regard to buckled veins, these assumptions are:

(1.) that the veins were of uniform thickness and have been deformed by a buckling process. Initial irregularities of thickness have, however, been partly allowed for by the method of thickness calculation.

(2.) that the materials followed a linear viscous model (as is assumed in the layer shortening equations). However, CHAPPLE (1969) has investigated the behaviour of viscous-plastic materials (which are more probable analogues for natural rock materials) and concludes that the validity of the layer shortening equations is unlikely to be affected, as the dominant wavelength is established at low strains while the material is behaving in a linear viscous fashion.

(3.) That the strain within the vein approximated to a plane strain. The initial thickness actually measured is $p(\lambda_1 \lambda_2)^{1/4}$, so that only if $\lambda_1 = \frac{1}{\lambda_2}$ will this be strictly correct. In the veins measured, there was no inhomogeneous deformation parallel to the fold axis. Note that the wave number is unaffected by a non-planar strain.

(4.) that the calculated wave number, which is substituted into the layer-shortening equations, is the dominant wave number. SHERWIN and CHAPPLE (1968) suggest that every fold within a layer should be counted in calculating the wave number. This assumes a normal distribution of individual wavelengths about the dominant wavelength, and is the best approach for layers containing a large number of folds. In veins such as those in the present study, with relatively few complete wavelengths, a single fold which is strongly atypical may completely upset the calculation and this part of the calculation becomes a serious potential source of error. Because of this, small, low-amplitude buckles were not counted, so that the strains calculated here may be too low. As this will increase the viscosity ratios, it will also have the effect of reducing the values calculated from the boudinaged veins for which this ratio was used. However, a genuine minimum as presented here is probably of more value than one which could be considerably too large.

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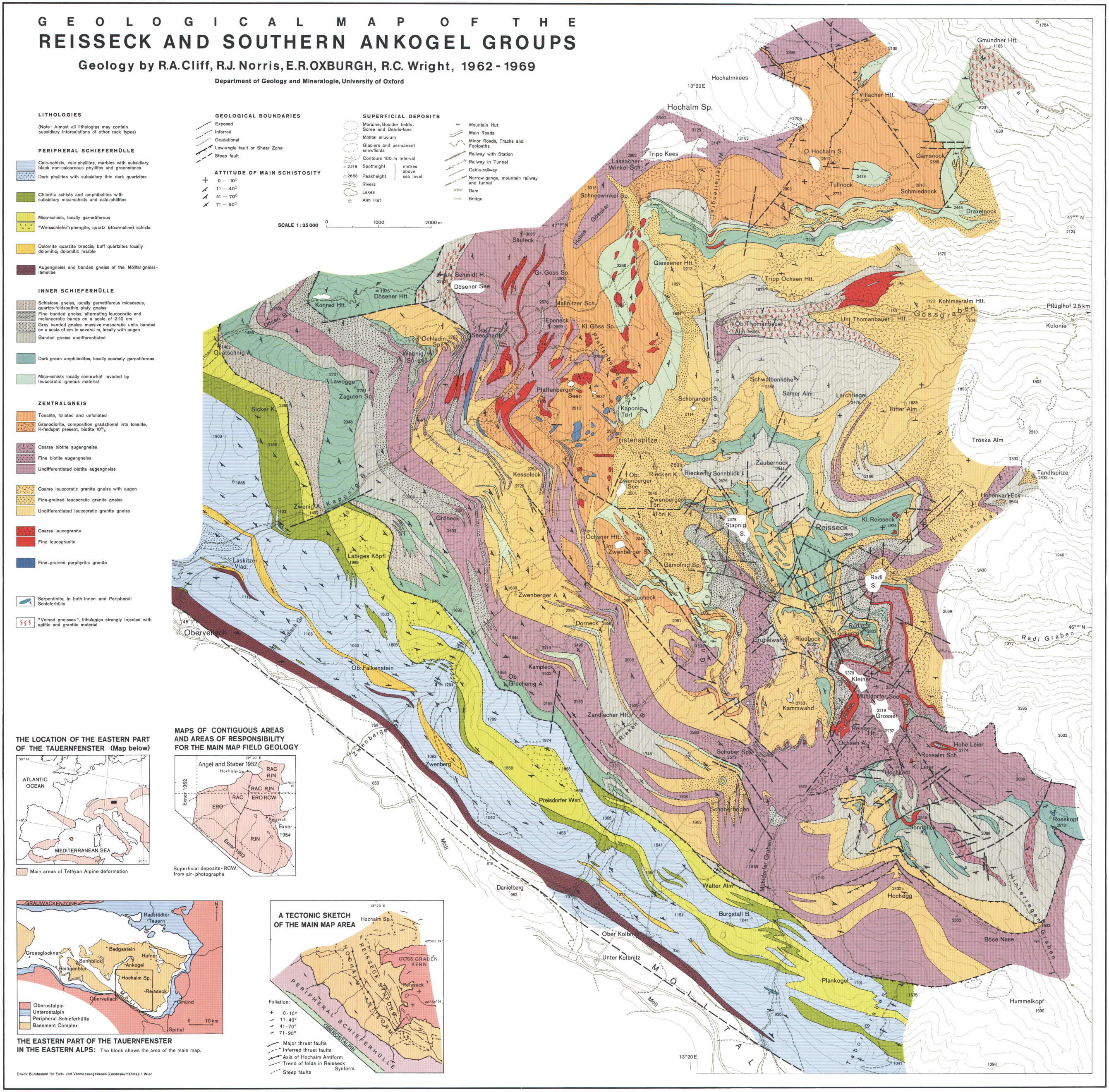

GEOLOGICAL MAP OF THE
REISSECK AND SOUTHERN ANKOGEL GROUPS
Geology by R.A.Cliff, R.J.Norris, E.R.OXBURGH, R.C. Wright, 1962 - 1969
Department of Geology and Mineralogie, University of Oxford
LITHOLOGIES
(Note: Almost all lithologies may contain subsidiary intercalations of other rock types)
PERIPHERAL SCHIEFERHÜLLE
Calc-schists, calc-phyllites, marbles with subsidiary black non-calcareous phyllites and greenstones
Dark phyllites with subsidiary thin dark quartzites
Chloritic schists and amphibolites with subsidiary mica-schists and calc-phillites
Mica-schists, locally garnetiferous
"Weisschiefer": phengite, quartz (±tourmaline) schists
Dolomite quarzite breccia; buff quartzites locally dolomitic; dolomitic marble
Augengneiss and banded gneiss of the Mölltal gneiss-lamellae
INNER SCHIEFERHÜLLE
Schistose gneiss, locally garnetiferous micaceous, quartzo-feldspathic platy gneiss
Fine banded gneiss, alternating leucocratic and melanocratic bands on a scale of 2-10 cm
Grey banded gneiss, massive mesocratic units banded on a scale of cm to several m, locally with augen
Banded gneiss undifferentiated
Dark green amphibolites, locally coarsely garnetiferous
Mica-schists locally somewhat invaded by leucocratic igneous material
ZENTRALGNEIS
Tonalite, foliated and unfoliated
Granodiorite, composition gradational into tonalite, K-feldspat present, biotite 10%
Coarse biotite augengneiss
Fine biotite augengneiss
Undifferentiated biotite augengneiss
Coarse leucocratic granite gneiss with augen
Fine-grained leucocratic granite gneiss
Undifferentiated leucocratic granite gneiss
Coarse leucogranite
Fine leucogranite
Fine-grained porphyritic granite
Serpentinite, in both Inner- and Peripheral-Schieferhülle
"Veined gneisses", lithologies strongly injected with aplitic and granitic material
GEOLOGICAL BOUNDARIES
Exposed
Inferred
Gradational
Low-angle fault or Shear Zone
Steep fault
ATTITUDE OF MAIN SCHISTOSITY
0 — 10°
11 — 40°
41 — 70°
71 — 90°
SUPERFICIAL DEPOSITS
Moraine, Boulder fields, Scree and Debris-fans
Mölltal alluvium
Glaciers and permanent snowfields
Contours 100 m interval
2219 Spotheight
2658 Peakheight
metres above sea level
Rivers
Lakes
Alm Hut
Mountain Hut
Main Roads
Minor Roads, Tracks and Footpaths
Railway with Station
Railway in Tunnel
Cable-railway
Narrow-gange, mountain railway and tunnel
Dam
Bridge
SCALE 1:25 000
0
1000
2000 m
Hochalm Sp.
Hochalmkees
Tripp Kees
Gmündner Htt.
Villacher Htt.
O. Hochalm S.
Tullnock
Gamsnock
Schmiednock
Draxelnock
Säuleck
Dösener See
Dösener Htt.
Konrad Htt.
Gr. Göss Sp.
Giessener Htt.
Tripp Ochsen Htt.
Kohlmayralm Htt.
Gössgraben
Pflüglhof 2,5 km
Kolonie
Mallnitzer Sch.
Kl. Göss Sp.
Ob.Thomanbauer Alm
Unt.Thomanbauer Htt.
Quatschnig A.
Lawogge
Zaguten Sp.
Pfaffenberger Seen
Kaponig Törl
Schönanger S.
Schwalbenhöhe
Samer Alm
Larchriegel
Ritter Alm
Tristenspitze
Tröska Alm
Zaubernock
Rieckener Sonnblick
Tandlspitze
Kesseleck
Hohenkar Eck
Gröneck
Zwenig A.
Stapnig S.
Reisseck
Kl. Reisseck
Labiges Köpfl
Laskitzer Viad.
Ochsner Htt.
Zwenberger S.
Gamolnig Sp.
Radl S.
Zwenberger A.
Jocheck
Dorneck
Obervellach
Riedbock
Radlgraben
Kleiner
Mühldorfer See
Ob. Falkenstein
Kampleck
Ob. Grechenig A.
Zandlacher Htt.
Kammwand
Grosser
Reisseck Htt.
Ochsen A.
Hohe Leier
Rossalm Sch.
Hochkedl
Kl. Leier
Schober Sp.
Zwenberg
Preisdorfer Wsn.
Schoberboden
Rosskopf
Sonnblick
Danielberg
Walter Alm
Hochegg
Burgstall B.
Ober Kolbnitz
Unter Kolbnitz
Böse Nase
Plankogel
Hummelkopf
MÖLLTAL
Möll
Tabor Gr.
13°20'E
47°N
46°50'N
THE LOCATION OF THE EASTERN PART OF THE TAUERNFENSTER (Map below)
ATLANTIC OCEAN
MEDITERRANEAN SEA
Main areas of Tethyan Alpine deformation
MAPS OF CONTIGUOUS AREAS AND AREAS OF RESPONSIBILITY FOR THE MAIN MAP FIELD GEOLOGY
Angel and Staber 1952
Hochalm Sp.
RAC
RJN
ERO
RCW
Exner 1962
Exner 1954
Reisseck
Superficial deposits - RCW. from air-photographs
GRAUWACKENZONE
Radstädter Tauern
Grossglockner
Sonnblick
Heiligenblut
Badgastein
Hafner
Ankogel
Hochalm Sp.
Reisseck
Obervellach
Gmünd
Spittal
Oberostalpin
Unterostalpin
Peripheral Schieferhülle
Basement Complex
0 10 km
THE EASTERN PART OF THE TAUERNFENSTER IN THE EASTERN ALPS: The block shows the area of the main map.
A TECTONIC SKETCH OF THE MAIN MAP AREA
Hochalm Sp.
GÖSS GRABEN KERN
Reisseck
HOCHALM ANTIFORM
REISSECK SYNFORM
PERIPHERAL SCHIEFERHÜLLE
OBEROSTALPIN
Foliation:
0-10°
11-40°
41-70°
71-90°
Major thrust faults
Inferred thrust faults
Axis of Hochalm Antiform
Trend of folds in Reisseck Synform.
Steep faults
Druck: Bundesamt für Eich- und Vermessungswesen (Landesaufnahme) in Wien

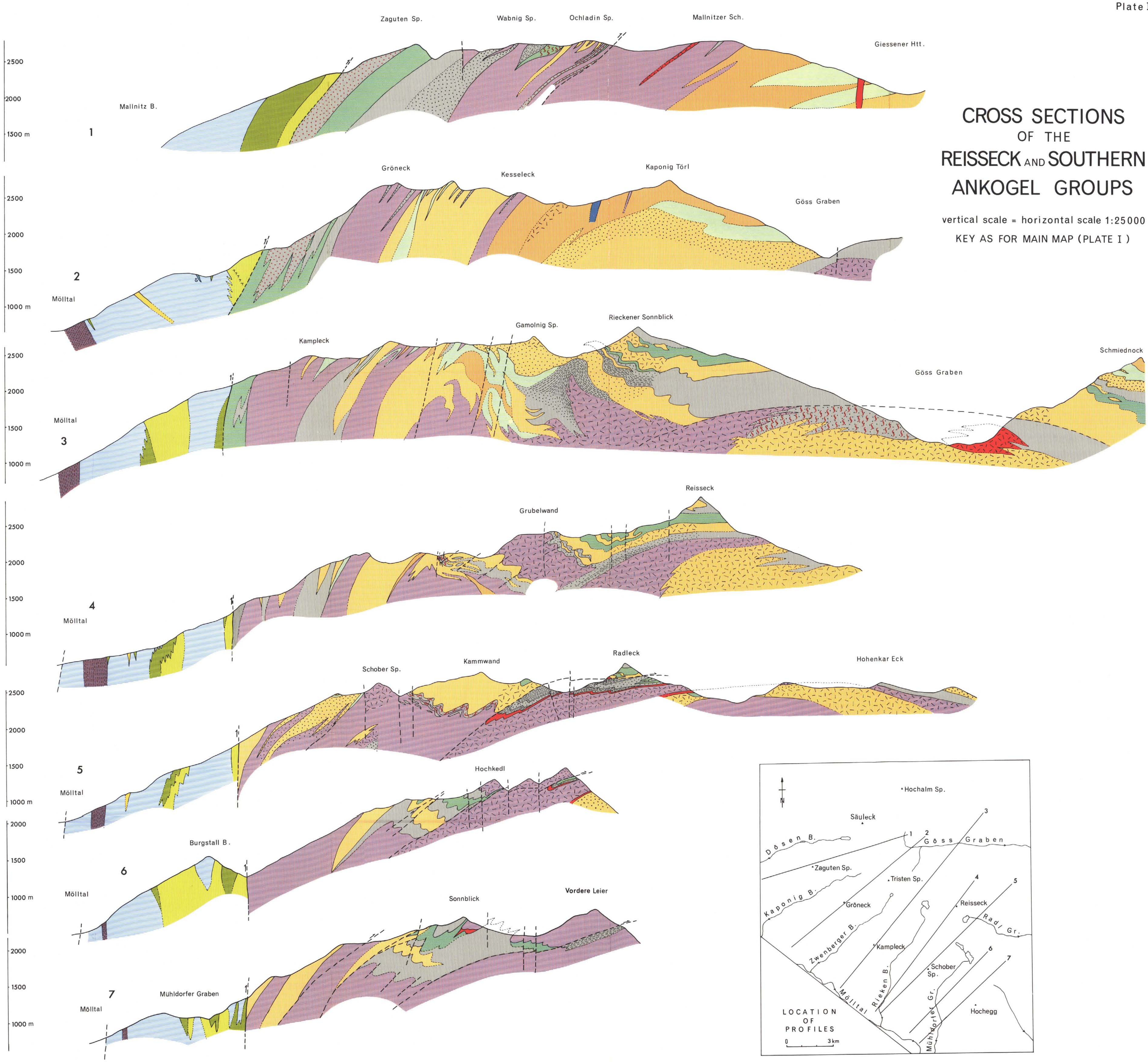
Jahrbuch der Geologischen Bundesanstalt, 114. Band, 1971 — Beilage 8
R. A. CLIFF, R. J. NORRIS, E. R. OXBURGH, R. C. WRIGHT — Plate 2

Plate II

CROSS SECTIONS OF THE REISSECK AND SOUTHERN ANKOGEL GROUPS

vertical scale = horizontal scale 1:25000

KEY AS FOR MAIN MAP (PLATE I)



Druck: Bundesamt für Eich- und Vermessungswesen (Landesaufnahme) in Wien

Figure Caption for Plate 3.

PLATE 3—Thin section sketches of Zentralgneis lithologies. The arrows indicate genetic relationships. The intensity of foliation increases from top to bottom of the plate, the colour index from right to left.

Ornament as follows:

Quartz and plagioclase—white; see individual descriptions.
"Gefüllte" plagioclase—white with random small dashes.
Twinned plagioclase—white with faint parallel ruling.
K-feldspar—dotted ornament.
Biotite—very dark elongate patches with light streaks.
White mica—light elongate patches with ornament of parallel dashes.

Larger areas left blank are very fine grained plagioclase mosaic with subsidiary quartz and K-feldspar. The insets are at 5× greater magnification than the main stetches which are at a 8× magnification.

a) Tonalite

Large subhedral grains of gefüllte oligoclase showing fine albite twinning; subsidiary amounts of fine grained plagioclase (1) Quartz occurs in anhedral aggregates with irregular grain boundaries (2) Clots of biotite and accessory minerals are without obvious preferred orientation.
(A330/AS17161—mode: qz-29%; sodic oligoclase-47%; K-feldspar-3%; biotite-17%; accessory clinozoisite, allanite, white mica, sphene and apatite.)

b) Granodiorite

Megacrysts of K-feldspar, showing uneven extinction or, rarely, indistinct crosshatch twinning; plagioclase as small gefüllte grains (2) and in a fine polyhedral mosaic (1); small clots of biotite with weak preferred orientation.
(A275/AS17167-mode: accessory quartz-33%; oligoclase-32%; K-feldspar-26%; biotite-10%; clinozoisite, sphene and apatite.)

c) Coarse grained Biotite Augen Gneiss

Megacrysts of K-feldspar, partially enclosed in white mica "envelopes"; plagioclase entirely as small untwinned inverse-zoned grains in a polyhedral mosaic (2) also a little K-feldspar in the mosaic; parallel orientation of biotite flakes, rather than clots, elongation of quartz aggregates (1) parallel to mica foliation.
(A395/AS17189-mode: quartz-35%; oligoclase-35%; K-feldspar-23%; biotite-5%; accessory white mica, clinozoisite, sphene and zircon.)

d) Biotite Augen Gneiss

Rounded megacrysts of K-feldspar with white mica envelopes; small biotite flakes scattered through fine-grained plagioclase mosaic; quartz as long monomineralic streaks, composed of strings of single grains (arrow), parallel to the mica foliation and wrapping around the augen.
(A159-mode: quartz-∼35%; oligoclase-∼40%; K-feldspar-17%; biotite-6%; accessory white mica, sphene, apatite, zircon.)

e) Phengite Augen Gneiss

Very strong foliation defined by phengite (1) orientation and prominent quartz streaks (2); the latter in contrast to (d) are several grains thick; biotite is very rare but some chlorite may represent altered biotite.

f) Fine-grained Porphyritic Granite

Subhedral tablets of K-feldspar showing Carlsbad twinning occur in a fine grained groundmass composed of quartz (grain boundaries drawn in), oligoclase (left blank on sketch) and scattered flakes and small clots of biotite with no preferred orientation.
(A383/AS17176-mode: quartz-25%; oligoclase-35%; K-feldspar-30%; biotite-10%.)

g) Fine-grained Augen Gneiss

Small rounded K-feldspar augen set in a groundmass made up of oligoclase mosaic and irregular quartz aggregates with scattered, well oriented flakes of biotite.

h) Coarse Leucogranite

Irregular granular fabric made up of single anhedral grains of K-feldspar, aggregates of quartz grains with jagged grain boundaries (1) and small patches of fine grained plagioclase mosaic (2); minor mica.
(Ċ6712-mode: quartz-45%; K-feldspar-30%; oligoclase-25%; minor biotite, altered partly to chlorite, accessory allanite.)

i) Coarse Leucogneiss

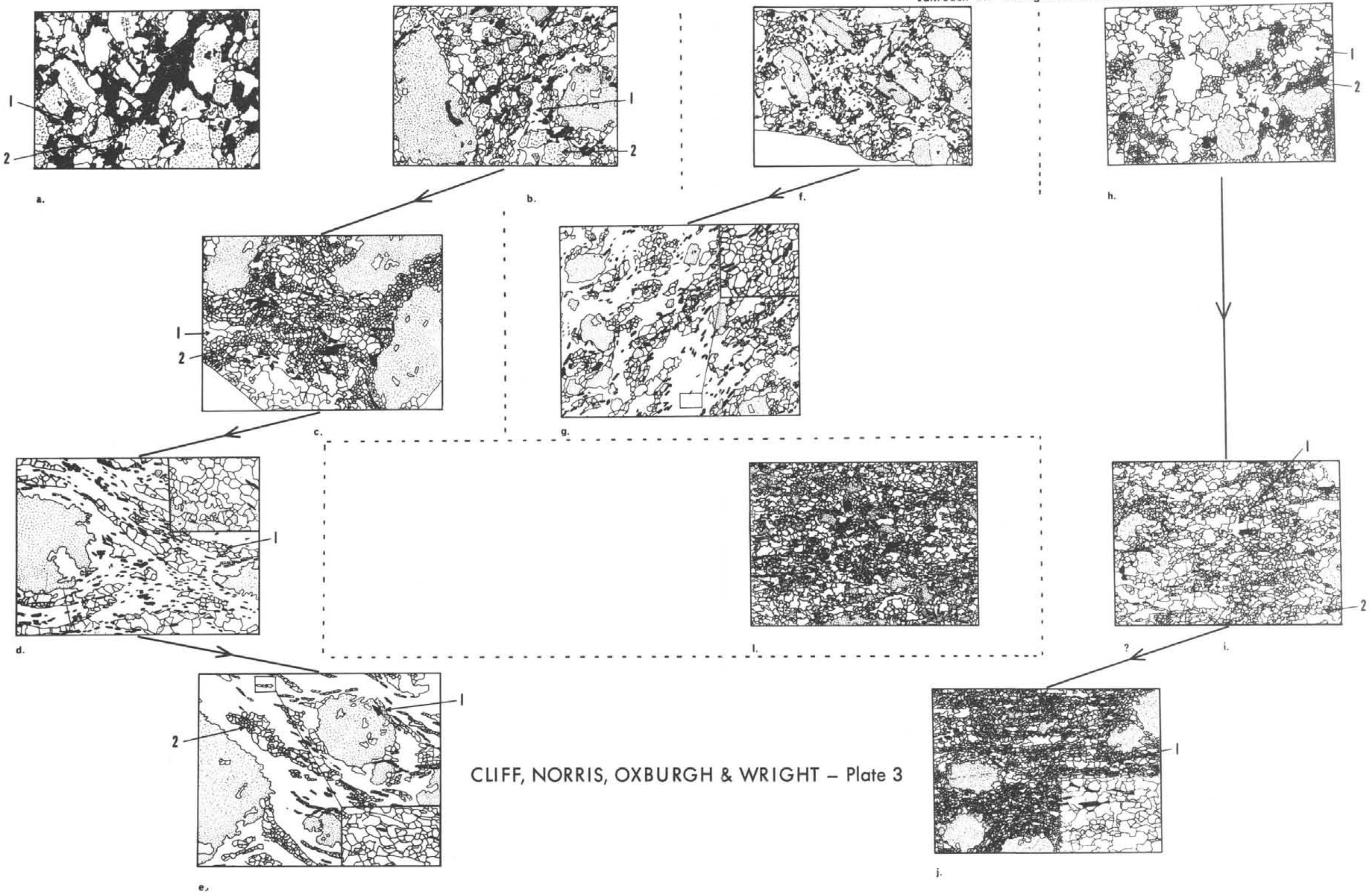
Elongate aggregates of quartz (1) and patches of plagioclase mosaic parallel to the mica foliation; K-feldspar partly as large grains and partly as a fine mosaic (2).
(C6709-mode: quartz-40%; K-feldspar-30%; plagioclase-25%; biotite-minor.)

j) Fine-grained Leucogneiss

Foliation defined by alternate strings of quartz grains (1) and an irregular mosaic with about equal amounts of oligoclase and K-feldspar; also parallel orientation of minor biotite flakes. A few larger K-feldspar crystals.

l) Fine-grained Leucocratic Granite-Gneiss

Fairly large flakes of biotite and muscovite define the mica foliation; irregular quartz aggregates (faint parallel ruling) oriented in the same plane. Oligoclase and K-feldspar are distributed in an irregular mosaic of variable grain size.
(A337/AS17195-mode: quartz-29%; K-feldspar-26%; oligoclase-40%; biotite-3%; muscovite-2%; accessory chlorite, apatite and clinozoisite.)



CLIFF, NORRIS, OXBURGH & WRIGHT – Plate 3

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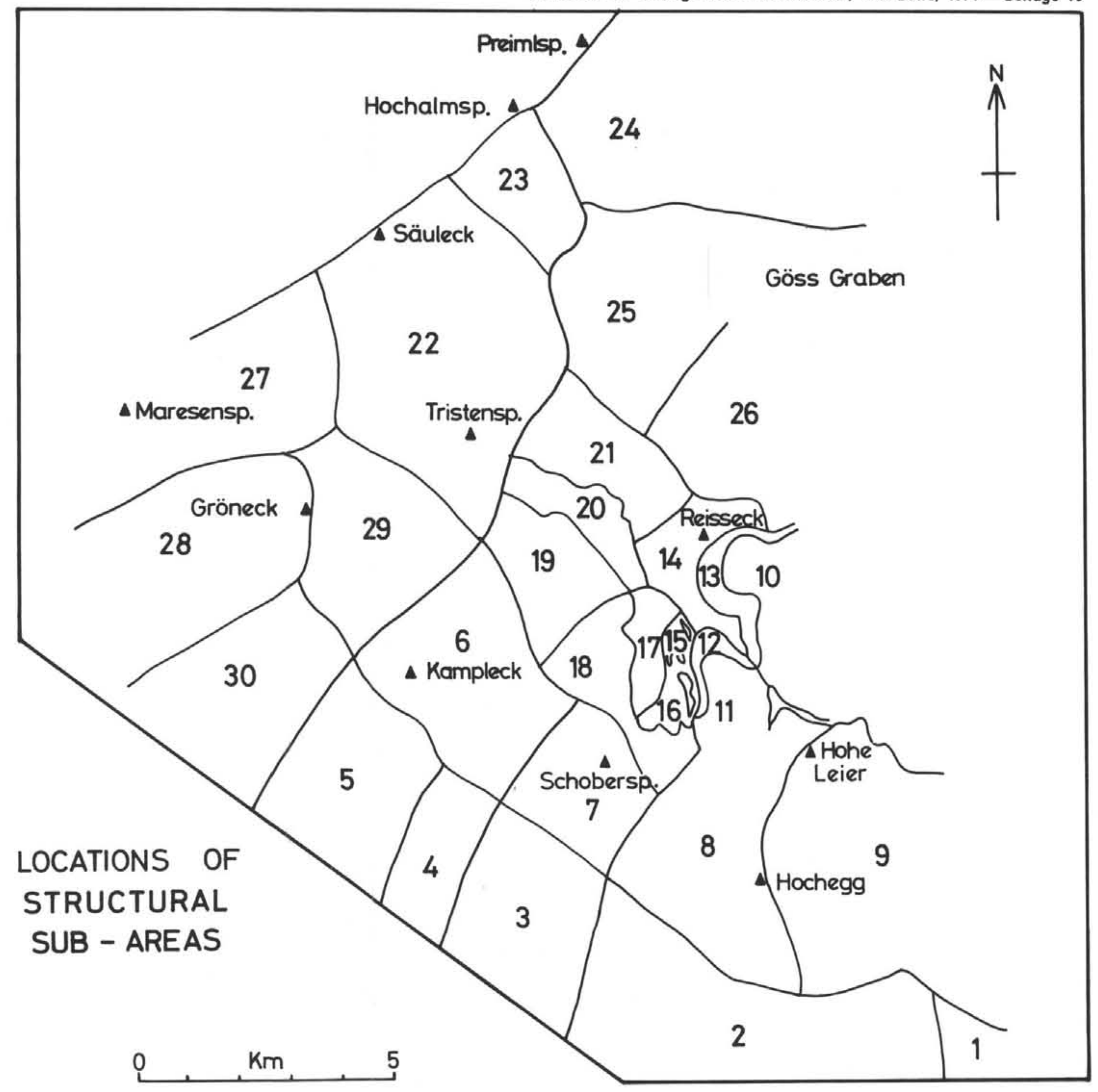
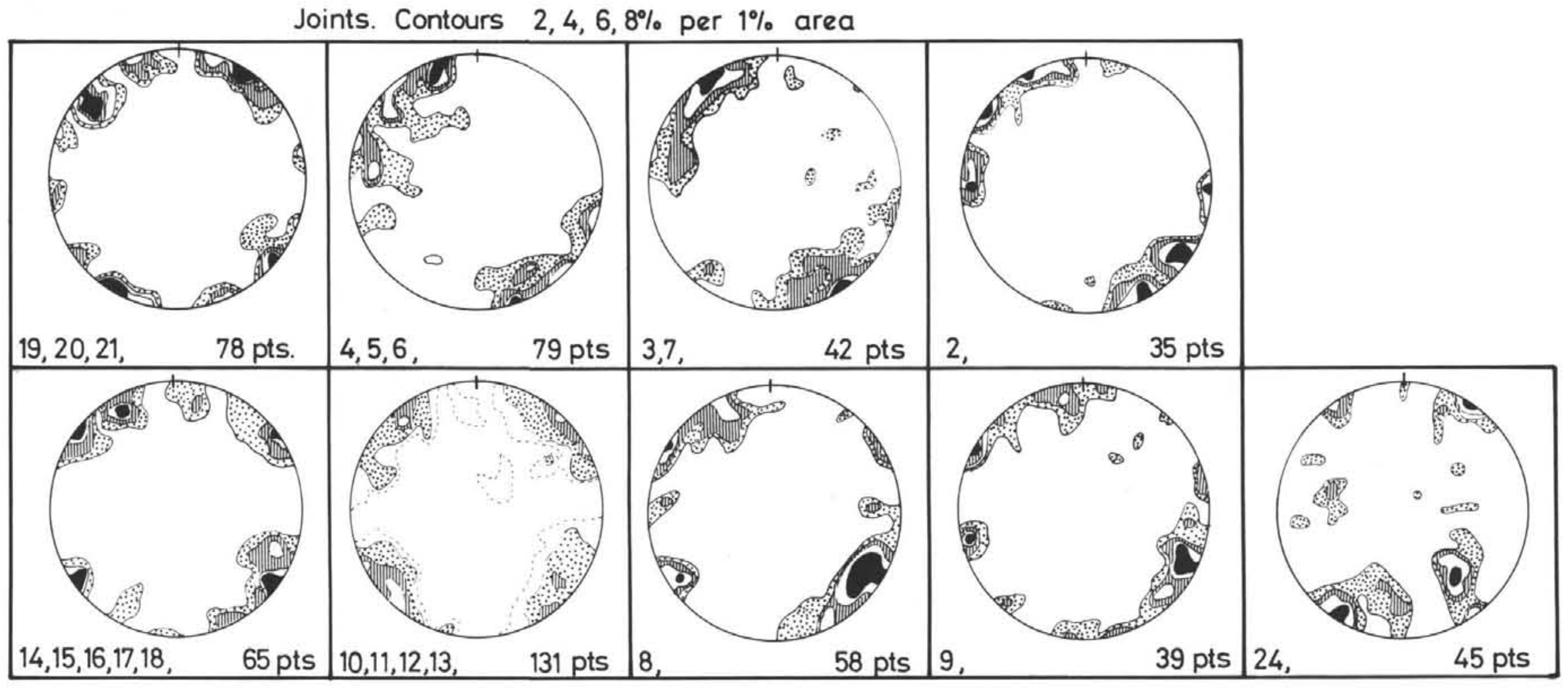
STRUCTURAL OBSERVATIONS for the REISSECK and SOUTHERN ANKOGEL GROUPS



Joints. Contours 2, 4, 6, 8% per 1% area



Equal area lower hemisphere projections

Column A: Principal Planar Feature

Column B: F Fold axes (·) and Axial Planes (×)

Column C: Other Linear Elements ∘ mineral lineation · Ripple axis, mullion, etc ▴ Boudin axis