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STRUCTURAL, METAMORPHIC AND GEOCHRONOLOGICAL STUDIES IN THE SOUTH-EAST TAUERN

by

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ABSTRACT

An area of c. 50 km^2 within the Altkristallin Sheet (Upper East Alpine Sheet sensu lato) south of the Tauern Window has been mapped at a scale of 1:10,000. The rocks have been tilted southwards by uplift of the Hohe Tauern area, and the mapped area represents a section from deep tectonic levels characterised by early Alpine (80 my) K-Ar mineral ages, through to higher levels with Hercynian (220-300 My) ages.

The rocks are mainly metasediments with minor amounts of metamorphosed igneous material. Three principal schist types are distinguished.

(i) in the south: phyllitic garnet-muscovite and quartzose schists. (Kreuzeck Schists).

(ii) in the north: medium- to coarse-grained mica-schists and paragneisses (Polinik Schists).

(iii) in a central zone: schists intermediate in character between the above (Intermediate Schists).

The tectonic sequence in the map area has been determined by structural analysis and correlation, and the K-Ar mineral age provinces correlated with episodes of mineral growth. The area underwent a major Hercynian deformation and metamorphic event (F_1 and M_1) of amphibolite facies, with pelitic assemblages dominated by muscovite and garnet, now preserved in the Kreuzeck Schists. Mineral textures indicate syntectonic growth. Relict eclogites in the Polinik Schists are also assigned to this phase. An F_2 deformation increases northwards (i.e. downwards in the Altkristallin Sheet) in both intensity and ductile nature. F_2 structure has been synthesised across the map area largely by minor fold correlation. Early Alpine metamorphic recrystallisation also increases downwards, obliterating M_1 textures and defining a new foliation, axial-planar to F_2 folds. Although

mineral growth is post-tectonic, it is suggested that M_2 and F_2 are part of the same tectonic event. An F_3 deformation, involving microfolding, flexuring and local chevron-folding, was followed by late Alpine retrogression (M_3) which is patchily distributed over most of the area, but more uniformly developed in greenschist facies in the basal kilometer of the Altkristallin Sheet.

The Altkristallin Sheet is broken up into blocks by shear zones of (at least) two generations, distinguished on lithological grounds.

Strain in quartz fabrics is related to late structures, probably to late arching of the Altkristallin Sheet (F_4), and locally to the development of shear zones.

Petrograhic mapping revealed the progressive overprint of M_1 by M_2 assemblages in the Kreuzeck and Intermediate Schists, with the following features in approximate order of appearance, south to north:

(i) growth of biotite and plagioclase porphyroblasts

(ii) gradual increase in muscovite fabric recrystallisation

(iii) growth of staurolite and breakdown of old garnet

(iv) breakdown of M_2 staurolite and growth of Al-silicates (sillimanite, andalusite and kyanite)

Three isograds may be constructed in the Intermediate Schists:

(i) an isograd marking the first appearance of M_2 staurolite.

(ii) an andalusite-sillimanite isograd

(iii) a garnet-Al silicate-biotite isograd

Isograds (ii) and (iii) have been mapped over an outcrop length of 8 km and a vertical relief of c. 1 km, and are approximately parallel to the M_2 regional foliation and the compositional banding.

The Polinik schists lie within the stability field of staurolite + quartz, but entirely within the kyanite field, indicating that there are two distinct areas of differing thermal gradient during M_2.

Evidence for the conditions of M_2 metamorphism is drawn from:

(i) experimentally determined equilibria in pelitic compositions

(ii) the Na content of muscovites

(iii) calcite-dolomite geothermometry at a locality in the Polinik schists.

(iv) the pressure-sensitive equilibrium

3 anorthite = grossular + 2 kyanite + quartz

(v) equilibria in the iron pelite system

K_2O-FeO-(MgO)-Al_2O_3-SiO_2-H_2O, recalculated with self-consistent thermochemical data.

The Richardson et al. (1969) Al-silicate triple point is preferred, on the basis of field evidence and (iv) and (v) above.

The metamorphic conditions indicated are:

(i) for the Polinik schists: $640 \pm 25^{o}C$ at 6 ± 1kb pressure

(ii) for the andalusite-sillimanite and garnet-Al silicate-biotite isograds of the Intermediate Schists: $640 \pm 20^{o}C$ at 5 ± 1 kb.

In the higher-grade areas there is evidence for a free fluid phase, whose composition was locally buffered by maximum phase pelitic assemblages in the Polinik schists, but was not controlled in the Intermediate Schists.

45 new K-Ar mineral ages are presented from the southern Glockner Group and Kreuzeck Group. Anomalous K-Ar ages, which do not fit the three-phase cooling model, may be qualitatively understood if it is assumed:

(i) that in certain circumstances the rate of escape of argon from whole rock systems, on a scale of metres or larger, may be slow compared with escape from millimetre-scale mineral systems.

(ii) that rapid diffusion in whole rock systems requires the presence of abundant fluids, and that assumption (i) is applicable to fluid-deficient rocks.

Corollaries of this approach, which may be tested with evidence from the study area, are:

(i) that redistribution of argon will occur between minerals which are open to diffusion within a larger rock system which is closed, or partly closed.

(ii) that growth of new minerals in a partly closed whole rock system leads to occlusion of argon, yielding an apparent age greater than the true age.

(iii) that the closure of mineral systems in rock systems which were partially open may allow the recognition of "fossilised" concentration gradients of argon out of bodies of fluid-deficient rock.

Anomalous ages due to excess argon in the base of the Altkristallin Sheet and in the M_2 staurolite zone of the Intermediate Schists are believed to be caused by the relative concentration of argon due to the removal of H_2O and other components of the metamorphic fluid phase by retrograde reactions.

The recognition of a high temperature, relatively low pressure metamorphic event, distributed in an elongate east-west trending belt in the Altkristallin, suggests that the two contrasted early Alpine metamorphisms in the Alps constitute paired metamorphic belts indicating southward subduction under the Altkristallin during the Cretaceous. The absence of large volumes of igneous rocks may be explained if the subduction was relatively short-lived.

The original base of the Altkristallin (below about 22 km depth) together with the lower lithosphere, was removed before or during the overthrusting of the East Alpine over the Pennine units. Detachment was probably assisted by thermal softening of the lower regions during early Alpine metamorphism.

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CHAPTER 1

Introduction

A. The Eastern Alps, and their major tectonic subdivisions

The Alps form part of a system of relatively young orogenic belts in the Mediterranean region, and stretch from the Ligurian Sea in an arc through Switzerland and eastwards to Vienna. The Alps fall naturally into four geographical and geological divisions:-

(i) The Western, or French Alps, the north-south trending mountains between the Mediterranean coast and Geneva.

(ii) The Central, or Swiss Alps.

(iii) The Eastern Alps, whose western boundary is approximately taken at the Swiss/Austrian frontier, and which reach eastwards to the Vienna basin. They are bounded to the north by the thick Tertiary deposits of the Molasse basin, and to the south by the Southern Alps.

(iv) The Southern Alps, which lie to the south of the Eastern and Central Alps, and are separated from them by the topographic and tectonic lineament called, in its different segments, the Insubric, Judicarian, or Gailtal line. Further south, the rocks of the Southern Alps disappear under another zone of thick Tertiary deposits in the Po basin.

The geology of the Eastern Alps has been summarised in English by Oxburgh (1968). In this study the English nomenclature of tectonic units largely follows Oxburgh (1968) and Cliff et al. (1971).

For some seventy years it has been recognised that the Eastern Alps, like the French and Swiss Alps, consist of a pile of overthrust sheets.

Oxburgh (1968) divides the Eastern Alps into two east-west trending zones:

(i) a Northern Zone, which is composed mostly of sedimentary rocks and is a zone of net basement depression;

(ii) an Axial Zone which consists largely of metamorphic and igneous rocks and is a zone of net basement elevation.

The Northern Zone includes the Helvetic, Klippen and Flysch tectonic zones, a series of parautochthonous and allochthonous thrust sheets involving Jurassic to Eocene sediments which crop out along the northern margin of the Eastern Alps (see Figs. 1-1, 1-2). These are overlain by the Northern Calcareous Alps, an allochthonous unit of Mesozoic carbonate rocks, with a complex internal structure. To the south of these is a zone of phyllites of probable Palaeozoic age known as the Grauwackenzone.

In the Axial Zone, the lowest units (Pennine zone) are exposed in tectonic windows, of which the largest is the Tauern Window. The Pennine zone consists of a basement complex of igneous rocks (now gneisses, making up the Zentralgneis) intruded into a series of older metamorphic rocks (the Inner Schieferhülle), and a cover series of metamorphosed sediments and basic rocks, the Peripheral Schieferhülle. Much of the latter has recently been interpreted as a tectonic mélange (Norris et al., 1971) composed of fragments of oceanic crust, with pelagic and continental margin sediments (Bickle and Pearce 1975) which has been emplaced onto the Pennine basement.

The Pennine Zone is overlain by the so-called East Alpine thrust sheets which make up most of the rocks outcropping in the Axial Zone. The Lower East Alpine (Unterostalpin) sheet rests upon the Peripheral Schieferhülle around the Tauern Window and consists of a Mesozoic, carbonate-rich succession, Palaeozoic phyllites, and thin slivers of crystalline basement rocks. Along the southern margin of the Tauern Window it is represented by the Matrei Zone, a highly disrupted zone containing rock-types typical of both the East Alpine and Peripheral

Schieferhülle units.

Above the Lower East Alpine Sheet lie the schists and gneisses of the Altrkistallin Sheet, which in turn is overlain by scattered patches of Mesozoic sediments and Palaeozoic phyllites, some of which are apparently autochthonous upon it. The mutual sedimentary and tectonic relationships between these fragmentary sequences, the Altrkistallin, the Northern Calcareous Alps and the Grauwackenzone are still a matter for debate among Alpine geologists. The essence of the controversy is whether or not it is possible to accommodate all the Mesozoic sequences of the East Alpine units on top of the available basement (the Altkristallin).

One school of thought (e.g. Kober, 1955) maintains that all the Palaeozoic and Mesozoic sequences are part of the sedimentary cover of the Altkristallin sheet, and that the overthrust relationships (carrying e.g. Palaeozoic over Mesozoic) visible in the Axial Zone are of local significance only. The whole unit is termed the Upper East Alpine Sheet (Oberostalpin).

A different view (e.g. Tollman, 1963) is that the Northern Calcareous Alps and the Palaeozoic phyllites represent a separate tectonic unit (Upper East Alpine Sheet), comprising a basement and cover sequence originating further to the south, and thrust over the Altrkistallin and its originally more extensive autochthonous cover represented by the fragmentary Mesozoic sequences (Middle East Alpine Sheet).

To avoid confusion in the nomenclature of the East Alpine units, that part comprising the Altkristallin schists and gneisses with their largely autochthonous sequences is referred to henceforward as the Altkristallin Sheet.

B. The Altkristallin Sheet

The internal history of the Altkristallin Sheet remained controversial until radiometric age data began to be accumulated. Complex deformational and metamorphic histories had been described in many areas but no very successful regional correlations or syntheses had been made. With the aid of geochronology, a pattern is now emerging.

A possible Caledonian metamorphism has been described from the area south of the Tauern Window (Borsi et al. 1973). Widespread mineral ages of around 300 My indicate the effect of major metamorphic events connected with the Hercynian orogeny. While many of the orthogneisses in the Tauern Window are of Permian age, c. 250 My (e.g. Cliff, 1968), all orthogneisses so far investigated from the Altkristallin Sheet yield whole-rock isochron ages in the range 350-440 My (Grauert, 1966; Harre et al.,1968, Borsi et al.,1973; Hawkesworth, 1974), showing the contrasted geological histories of the two basement complexes.

In several areas, ages of around 80 My indicate an early Alpine thermal event in the Altkristallin Sheet (Oxburgh et al., 1966; Lambert, 1970; Schmidt et al., 1967, etc.), which in the area of the Schneeberger Zug (see Chapter 7, Fig. 7-1) may be linked with an episode of mineral growth which reaches up into the Mesozoic sedimentary cover, (Miller et al., 1967).

Late Alpine effects, with mineral ages in the range 16-35 My are locally encountered close to the edge of the Tauern Window (e.g. Borsi et al., 1973).

Structural work in the Ötztaler Alpen (e.g. Schmidt, 1965) indicates that the major deformations, at least in this part of the Altkristallin, are at least Hercynian in age. The nature and extent of Alpine deformation in the Altkristallin Sheet are not well-known, although some syntheses, e.g. that of Metz (1966), have been proposed.

C. The Altkristallin of the Kreuzeck Group

The rocks of the Kreuzeck Group (K on Fig. 1-1) are dominantly mica-schists and paragneisses with subsidiary leucocratic gneiss and amphibolite. To the north-west of the Kreuzeck Group the thrust contact of the Altkristallin over the Matrei zone and Pennine units is preserved (Exner, 1964). To the north-east, the boundary with Pennine zone rocks is probably a high-angle fault, marked by a strong topographic lineament, the Mölltal line. Along the southern edge ofthe Kreuzeck Group, Altkristallin rocks are faulted against unmetamorphosed clastic sediments and carbonates (Mesozoic) of the Drauzug. In the Goldeck Group, a little to the south-east, the sediments are unconformable on probable Altkristallin phyllites.

No detailed geological map of the Kreuzeck Group exists, and until recently the area has received very little attention from geologists. H. Beck (1928-1939) spent several field seasons in the area, but the results of his mapping efforts were never published. F. Angel (1930) published a petrographical study of rocks collected from the Kreuzeck Group. Holzer (1958) prepared a map based on the study of aerial photographs, supplemented by a limited amount of field observation. A few other authors have commented on the presence of such features as mineralisation and a young dyke-suite. A fuller account of previous work and further references may be found in Wright (1973).

K-Ar geochronology by Oxburgh et al. (1966) yielded mineral ages of 80 My in the area north-east of Polinik (see Plates 2), and a sharp geochronological break across the edge of the Tauern Window. A study of the structures in the same area revealed three superimposed phases of folding (Oxburgh 1966). Further K-Ar work in the Kreuzeck Group was undertaken by Brewer (1970) who found more early Alpine ages in the north, but ages of 220-300 My in the phyllitic mica-schists of the south.

The major geochronological break coincided with a high-angle fault or shear zone along the Nigglai Tal.

In 1973, R.C. Wright completed a thesis study of an area in the east central Kreuzeck Group, lying astride the Nigglai shear zone. He concluded that a low-grade metamorphic event reaching biotite grade (greenschist facies) was responsible for the early Alpine ages to the north.

Other work of relevance to this study by members of the Alpine Project, includes that within the Tauern Window summarised by Cliff et al. (1971), and that of M.J. Bickle (1973) and C.J. Hawkesworth (1974) across the edge of the Window to the west and east of the Kreuzeck Group.

D. Aims of the Present Study

The principal aim of the study was to investigate the structure and metamorphism of the Altkristallin in the Kreuzeck Group, in the hope that the early Alpine isotopic event could be further linked to events in the structural and metamorphic history. An area to the west of that of Wright (1973) where a NNE-SSW trending ridge of excellent and continuous exposure (between Polinik and Kreuzeck) connects the area of Oxburgh (1966) and Oxburgh et al. (1966) where some measure of structural and geochronological control already existed, with the area of 220 + My ages of Brewer (1970).

There were further advantages in choosing an area further west than that of Wright. The vertical movements on the Mölltal line which obscure the relationship between the areas of Wright and Cliff et al. (1971) die out towards the west, and the base of the Altkristallin may be studied in stream sections near Söbriach, and in the Sadnig Group. In addition it was hoped that the influence of the Nigglai fault on the age pattern would die out towards the west.

E. The Map Area

The map area is covered by the Österreichische Karte (1:50,000) Sheet 181 (Obervellach), and by 1:25,000 sheets 181/1 (Obervellach) and 181/3 (Greifenburg) topographic maps. Mapping was carried out on 1:10,000 scale enlargements of the 1:25,000 maps, which were found to be adequate for locating outcrops in the field, and recourse to aerial photographs was seldom necessary.

A detailed geological map of part of the northern Kreuzeck Group, covering an area of about 50 km^2, is presented at a scale of 1:10,000 (Plate 1). A smaller-scale map covering the whole width of the Altkristallin outcrop from the Tauern Window to the Drauzug Mesozoics has also been prepared at 1:50,000 (Plate 2). This map is a compilation and adaptation of maps by Exner (1962), Cliff et al. (1971) and Wright (1973) together with observations by the author outside the area of the detailed map, and elements from the unpublished map of H. Beck.

Plate 3 is a geological cross-section through the map area at a scale of 1:10,000.

Some 33 weeks were spent in the field during the years 1971 to 1973, with briefer visits in 1974 and 1975. Nearly 600 hand specimens were collected. Some of these samples were prepared for electron microprobe analysis, and others for K-Ar analysis of separated minerals. About 400 were examined in thin section, and a further 100 sections from the collections of E.R. Oxburgh and M.S. Brewer were studied.

F. Terminology and Abbreviations

The expression "early Alpine" is used somewhat loosely to indicate events probably occurring during the Cretaceous, prior to the onset of overthrusting. "Late Alpine" describes events which largely post-date overthrusting.

The terms macroscopic, mesoscopic and microscopic are used to signify map-scale, outcrop to hand-specimen scale, and thin-section scale respectively, after Turner and Weiss (1963).

Plate 1 has a coordinate system which is used to identify localities mentioned in the text: for example Polinik (C3).

Mineral abbreviations used in the text are listed in Table 1-1.

Table 1-1. Abbreviations

Mineral Abbreviations

ab	albite	hbl	hornblende
alm	almandine	ilm	ilmenite
an	anorthite	ksp	potash feldspar
and	andalusite	ky	kyanite
ann	annite	mu	muscovite
ap	apatite	or	orthoclase
as	aluminium silicate	pg	paragonite
bi	biotite	phl	phlogopite
cc	calcite	plag	plagioclase
chl	chlorite	pyr	pyrope
clzo	clinozoisite	qz	quartz
cpx	clinopyroxene	rut	rutile
dol	dolomite	ser	sericite
ep	epidote	sill	sillimanite
fe-chl	iron chlorite	sp(ess)	spessartine
fe-st	iron staurolite	sph	sphene
gross	grossularite	st	staurolite
gt	garnet	tour	tourmaline

Other Abbreviations

Kb	Kilobars (pressure)
My	Million years
P	Pressure
P_i	Partial pressure of i
X_i	Mole fraction of i

CHAPTER 2

Lithologies in the Map Area

Introduction

In this chapter the petrography and field relations of the principal lithological units in the map area are described. The rocks are dominantly metasediments comprising psammitic, semi-pelitic and pelitic schists and gneisses. Calcareous rocks are rare. Subordinate amounts of meta-igneous material are present in the form of amphibolites and leucocratic gneisses.

The metasedimentary lithologies present considerable problems of nomenclature and classification. Variation in composition in terms of mineral composition (e.g. quartz-rich, mica-rich) is fairly easily observable in outcrop and presumably represents original differences in sediment composition. Problems in the mapping of metasedimentary lithologies are caused by first, small-scale interbanding of different compositions, such that any lithological unit mapped must of necessity be defined as an association of two or more lithologies, lateral discontinuity of rock-types, probably due to the extreme disruption of the stratigraphy by multiple deformations and high strains, so that almost all rock units are lenticular in form on various scales, and cannot be traced far along the strike. A non-geological problem affecting the recognition of lithologies in the field is the luxuriant growth of lichens on outcrop surfaces (e.g Figs. 3-8, 3-30).

In the map area it was concluded that there was little to be gained by producing a detailed lithological map of the schists, which would have little bearing on the central problem in the map area, which is the variation of metamorphic conditions and structures over the whole area. Instead the metasedimentary schists have been classified

and mapped on combined lithological association and textural criteria.

One of the most prominent lithological features of the map area is the difference in texture between the phyllitic mica-schists of the south, with their frequently conspicuous fine-scale quartz-mica banding and porphyroblastic garnets, and the coarser-grained mica-schists and paragneisses of the north. These two contrasted types of textural and lithological association are given the names Kreuzeck Schist and Polinik Schist respectively. In the intervening tract between these two types, the schists display textures characteristic of both types, and are termed Intermediate Schists. These three textural types are geographically separated from each other by lithologically distinct biotite- and plagioclase-rich schists.

The schist units differentiated in the map area are thus:

A Kreuzeck Schists

B Polinik Schists (locally subdivided)

C Intermediate Schists

D Biotite-feldspathic Schists

Particularly quartzitic associations, usually with the quartzites occurring in closely-spaced bands, are distinguished from the above:

E1 Schists with abundant black quartzites

E2 Schists with abundant grey quartzites

Other metasedimentary lithologies recognised are:

E3 White quartzite

F Marble

usually in conspicuous single or multiple thin bands.

Metamorphosed igneous rocks are represented by:

G Amphibolites and eclogitic amphibolites

H Leucocratic gneisses

I Fine biotite gneiss (including biotite augen-gneiss)

Largely unmetamorphosed, cross-cutting lithologies include:

J Pegmatites

K Dykes

L Shear zones: crush rocks and mylonites

Modal analyses of some typical rocks are presented in Chapter 5, Table 5-2.

A Kreuzeck Schists

The Kreuzeck schists are the characteristic metasedimentary rock-type of the southern central Kreuzeck Group, and consist of an inter-banding of two principal lithological components, one quartzose, one micaceous, on a scale which varies from millimetres to one or two metres. The relative proportions of the two components are variable. In out-crop they are grey in colour with a silvery phyllitic sheen on foliation surfaces. Small-scale folding of the banding gives the outcrops a massive, unfoliated appearance from a distance.

The quartzose component consists dominantly of quartz, with minor muscovite, biotite, chlorite and platy-shaped ore (ilmenite). The quartz is generally equigranular and fine-grained (0.2 mm). Plagioclase is rather rare, and when found is more often associated with the mica-ceous component. Within particularly quartz-rich areas the micas and chlorite occur as separate, small (0.1 - 0.2 mm long) narrow laths, often without a strong preferred orientation, lying along quartz grain boundaries, but where micas are a little more abundant they tend to be larger (0.5 mm) and link up into trains defining a crude foliation parallel to the compositional banding.

The micaceous component consists mostly of muscovite, with variable but subordinate amounts of garnet, quartz, biotite, chlorite, plagioclase, occasional staurolite, and accessory minerals. Rocks consisting mostly of this component are dark grey, finely foliated with a phyllitic sheen,

frequently studded with garnet porphyroblasts up to 1 cm across, and often with the mica foliation deformed into millimetre-scale crenulations.

In thin sections cut perpendicular to the foliation the phyllitic sheen is seen to be due to the fabric of the muscovite, which consists of long narrow parallel laths each up to 0.5 mm long, and some ten times as long as broad, but generally so matted together in parallel orientations that it is difficult to distinguish individual laths from their neighbours. The resulting effect is like well-brushed felt (see Figs. 4-3 to 4-9). These parallel felts are very often deformed into crenulations, around which the muscovites are bent. The grey, sometimes almost black colour of the mica felts is due to graphite in dusty trails parallel to the felted fabric. Other accessories such as platy ilmenites, and tiny greenish- or bluish-yellow tourmalines occur as oriented trails of inclusions in the mica felts. Sometimes parts of the mica-felt are recrystallised to fine-grained (usually less than 0.1 mm) stubby, randomly oriented muscovite grains. Recrystallised parts of the mica felts contain appreciably less graphite (see Figs. 4-7 and 8).

Garnet occurs as large porphyroblasts up to 1 cm or more in diameter, and often showing rhombic dodecahedral crystal faces, which in hand specimen usually appear rusty-brownish or greenish due to alteration. In section they are seen to be much shattered, and the alteration, usually to chlorite or a birefringent, fine-grained micaceous material, has sheathed them and proceeded inwards along the many cracks. The trails of graphite, ore and tourmaline, together with elongate quartz grains, are found included within the garnets, often rotated into S-shaped configurations which indicate syntectonic growth.

Staurolite is found in the Kreuzeck Schists, though not commonly, and is not usually apparent in hand specimen. The porphyroblasts are

up to 4 mm across, invariably smaller than, and subordinate to, garnet in the same rock, and usually surrounded by a rim of chlorite or sericite alteration.

Biotite is not abundant in the Kreuzeck schists, only rarely making up more than 5% of the mode and often present only as an accessory mineral. It displays two modes of occurrence which probably represent two generations of mineral growth. Occasionally it is found in narrow laths within the parallel felts intergrown with and parallel to the muscovites. More usually it is found as broad flakes up to 2 mm long which cut across the parallel felts (See Fig. 4-4) and post-date at least one of the crenulation deformations. A later crenulation produces kink-bands in the porphyroblasts (Fig. 2-1). Sometimes these post-tectonic biotites are themselves chloritised by replacement of the whole crystal by a single lath of chlorite (see Fig. 4-5).

Chlorite occurs in the micaceous component either as knots and patches of radiating laths, or as an alteration product of garnet or biotite. Its birefringence varies widely, from anomalous blue colours through isotropic to greyish white, implying considerable variation in composition (Albee 1962). Chlorite pseudomorphing biotite can often be distinguished from other chlorite laths by the presence of bundles of radiating dark brown needles, which are probably of rutile. Chlorite after garnet is usually composed of disoriented aggregates of small, blue birefringence (iron-rich) laths, yellow-green in colour. The chlorite in small individual laths associated with more quartzose segregations is almost colourless, with a birefringence indicating more magnesium-rich compositions. Chlorite is of late appearance in the assemblage since it grows across all the other phases, but it is possible that more than one generation of chlorite is present in the Kreuzeck Schists.

Plagioclase occurs as untwinned porphyroblastic patches up to

1.5 mm across growing over the folded parallel felts (see Fig. 4-6). It is therefore probably of the same generation as the porphyroblastic biotite. From comparison of refractive indices with quartz and the mounting medium, and from examination of optic axial angles, its composition appears to be oligoclase in most cases.

The amount of quartz in the mica felts is variable, and most gradations from quartz-free rocks up to those quartz-rich enough to be considered with the quartzose lithological component can be found. Quartz within the felts is usually in lenticular stringers or bands, annealed into a granular mosaic around the pre-porphyroblast microfolds, but strained where affected by the later microfolding which has kinked biotite porphyroblasts and bent plagioclases. Larger (around 10 cm long by 2 cm broad) lenticles of vein quartz which can often be seen to consist of detached fold hinges (Fig. 3-23) are common in some parts of the micaceous Kreuzeck Schists.

B Polinik Schists

The Polinik schists form the bulk of the northern Kreuzeck Group north of the Teuchl shear zone, and most of the Sadnig Group. They embrace a range of lithological types varying from psammitic to pelitic composition, and are very similar in lithology and texture to the schists and paragneisses of the Schober Group (Bickle 1973, Troll and Hölzl 1974) and of the Liesertal (Hawkesworth 1974), and the Sadnig Group (Exner and Prey 1964).

The distinguishing features of the Polinik Schists are the fairly coarse, even grain-size of all the minerals, and the greater abundance of biotite and plagioclase than in the Kreuzeck Schists. Weathered exposures are rusty-brownish in colour due to the weathering of biotite, and the lithological banding often gives a flaggy appearance to the outcrop.

The minerals of the Polinik schists are quartz, plagioclase (usually oligoclase), muscovite and biotite, in varying proportions, with subsidiary garnet, chlorite, locally staurolite and kyanite, and accessory ore, tourmaline, apatite, clinozoisite and zircon. Fig. 2-3 shows a fairly typical example in thin section.

Quartz is the most abundant mineral and occurs in grains 0.2 to 1 mm in diameter, equidimensional or elongate in the plane of mineral foliation when constrained between micas. Quartz is often strained, with sutured grain boundaries, and sometimes finely recrystallised yielding mortar textures. Textures produced by strain in the quartz fabric are described in more detail in Chapter 3.

Plagioclase occurs as grains of similar size and shape to quartz, and forms from 5 to 40% of the mode, sometimes equalling quartz in abundance. More unusual feldspar textures are found in feldspar-rich schists near the margins of large amphibolite bodies (see Chapter 4) where plagioclase occurs in large poikiloblastic grains up to 5 mm diameter which include most of the other phases of the rock and appear to be replacing the mica fabric. It is often difficult to distinguish plagioclase from quartz in the Polinik Schists, since their refractive indices are close in value, but in many sections the presence of lamellar twinning, and of fine dusty alteration products around grain boundaries and in cracks, serve to distinguish the plagioclase. Measurement of refractive index, optic sign and symmetrical extinction angle of twin lamellae indicate an oligoclase composition in nearly every case.

Muscovite and red-brown biotite are found in laths about 1 mm long in micaceous lithologies, though smaller in quartzofeldspathic rocks. They tend to be broader (length/width ratios about 3) than the non-porphyroblastic micas of the Kreuzeck Schists, and though they usually define a mineral foliation, their alignment is not so strong

as in the Kreuzeck schist parallel felts. Muscovite and biotite are usually present in roughly equal proportions together making up 10 to 60% of the mode. Muscovite predominates over biotite in pelitic compositions (mica-rich schists with garnet, staurolite and/or kyanite), and biotite over muscovite in more quartzose and quartzo-feldspathic lithologies, and in certain micaceous schists in which plagioclase is also relatively abundant. These latter biotite-rich schists could be classified with the biotite-feldspathic schist lithology, but it has not been possible to map them separately within the Polinik schists. Both muscovite and biotite are frequently deformed around microfolds.

Garnets in the Polinik Schists are seldom more than 2 or 3 mm in diameter, and are not shattered, in contrast to the larger Kreuzeck schist garnets, and they are found in the more muscovite-rich, micaceous schists. Their textures are described further in Chapter 5 along with the analytical data. Locally in the Polinik Schists they are partly replaced by chlorite.

Staurolite and kyanite occur together or separately in medium to coarse-grained, usually muscovite-rich schists of pelitic composition, in crystals up to 5 mm in length, or larger still in and near synmetamorphic veins. Kyanite blades up to 6 cm long have been found in a quartz-kyanite-biotite vein in pelitic schists on Polinik. Kyanite also occurs in two localities as small (0.1 mm) grains, associated with biotite and garnet, picking out narrow bands of aluminous composition in garnet-two mica-schists. Both staurolite and kyanite may be partly or wholly retrogressed in places to aggregates of fine sericite.

Chlorite occurs either independently as sheaves and radiating knots of greenish laths with anomalous dark blue to olive-brown birefringence, or as an alteration product of garnet and biotite, usually

with blue birefringence.

Accessory tourmaline is fairly common in the Polinik Schists, forming larger crystals than in the Kreuzeck Schists, sometimes up to 4 mm in diameter, and often concentrated in particular horizons in the schists. These concordant tourmaline-rich bands possibly have a genetic connection with the tourmaline-bearing younger pegmatites. The colour of these tourmalines is golden- or slightly greenish-yellow, in contrast to the bluish shades observed in the Kreuzeck Schists.

Locally the Polinik Schists can be subdivided into the following distinctive lithologies, whose geographical distribution is shown on Plate 1:-

1 flaggy schist

2 coarse-grained schist

3 medium-grained schist

4 pelitic schist

1. Flaggy schist

This is the most typical Polinik Schist lithological association, and has not been distinguished from 'undivided' Polinik Schist on the map. The flaggy schists consist of interbanded quartzo-feldspathic or quartzose layers, and mica-rich layers, 1 cm to 2 m thick (Fig. 2-2). The micaceous layers have a strong preferred orientation of micas which produces an intense schistosity parallel to the layering. However, where the layers are much affected by tight mesoscopic folding, the flaggy appearance is lost, and the outcrops take on a massive appearance, since the mineral foliation and lithological banding are no longer parallel in the hinge zones.

2. Coarse-grained schist

This is a lithology found mainly to the south and south-east of Polinik, and consists of homogeneous, rather quartzo-feldspathic mica-schists whose principal characteristic is their large grain-size (quartz and plagioclase up to 2mm diameter, micas up to 5 mm long). The micas tend to occur in semi-continuous trains, and the alternating quartz-feldspar stringers and mica trains give the rock a crude foliation which is often deformed into tight, sharp-hinged, centimetre-scale folds (e.g. see Fig. 3-9). The minerals, however, are not bent round these folds, but have crystallised without destroying the coarse-textured folded fabric. Coarse-grained schists of more pelitic composition are found in the Blässenbühel and Geistlacke areas (F3) and south-east of Kessel See (G3). These schists have a very silvery appearance due to an abundance of coarse-grained muscovite flakes on foliation surfaces, and are often studded with large (up to 2 cm) staurolite prisms or kyanite blades. Chlorite is rare in or absent from the coarse-grained schists.

3. Medium-grained schist

This is the typical homogeneous mica-schist lithology to the north of Polinik, and it occurs either as units a metre or two thick within flaggy successions, or as larger thicknesses of homogeneous mica-schist. The grain size of the constituent minerals is around 1 mm. In outcrop this lithology is massive and grey-brown in colour, often with numerous thin creamy-white quartz stringers. Microfolds are often well-developed in this lithology and their axial surfaces can become as well-developed a planar feature as the mineral foliation. The medium-grained

schists are often garnetiferous, and when muscovite-rich grade into the pelitic schist lithology.

4. Pelitic schists

Schists of pelitic composition are not common within the Polinik Schists, and usually occur in scattered small and rather discontinuous bands. If the characteristic minerals staurolite and/or kyanite are partly retrogressed or fine-grained, these rock compositions cannot usually be recognised in the field. There is, however, a fairly large lenticular body in the Mörningtal (C3) where the assemblage of qz-plag-mu -bi -gt- st -ky (Fig. 2-4) is developed over some 35 m thickness, often in a very fresh, unretrogressed state, and this body is distinguished on the map (Plate 1).

In areas of poorer exposure or limited accessibility the coarse-grained, medium-grained and pelitic schists are not distinguished from the flaggy schists, but are mapped together as "undivided Polinik Schists".

C Intermediate Schists

The category of Intermediate schists covers mica-schists with characters intermediate between the Kreuzeck and Polinik schists, and their area of outcrop covers the tract between the outcrop of the other two schist types. In the Intermediate schists the parallel-felts of the Kreuzeck schists are modified by increasing recrystallisation of the muscovite fabric, such that in hand specimen the rocks lose their phyllitic sheen, and larger discrete laths of muscovite appear, either randomly oriented or defining a new foliation axial-planar to the micro- and mesoscopic folding of the parallel-felts. Towards the north, with

increasing recrystallisation, the new fabric becomes coarser and more even-grained, and the preferred orientation of new large micas more intense, until these rocks approach identity with the medium-grained and pelitic lithologies of the Polinik Schists.

The change in texture is accompanied by mineralogical changes. Much of the Intermediate schist in the map area is of semi-pelitic and pelitic composition, equivalent to the micaceous component of the Kreuzeck schists. These micaceous Intermediate schists are much more biotite-rich than the Kreuzeck schists and often contain abundant euhedral staurolite prisms up to 1 cm long. Garnet becomes subordinate to staurolite. The intermediate schists are characterised by an abundance of reaction textures, minerals rimming and replacing others, which are not attributable to late-stage retrogression. In thin section, large Kreuzeck-schist-type garnets are seen in various stages of breakdown, primarily to biotite, and staurolite grows from muscovite-rich areas. The parallel-felted muscovite fabric is substantially or wholly replaced by new, broader muscovite laths with a less distinct preferred orientation. The grain size of the Intermediate schists is very variable.

Other minerals which occur in the Intermediate schists are chlorite, either in large laths cutting across the fabric or as an alteration product of garnet or biotite; occasional andalusite in large porphyroblasts, spongy or packed with inclusions of most of the other phases; locally a little fibrolitic sillimanite associated with biotite; and accessory tourmaline, ilmenite, apatite, graphite and zircon.

The sequence of textural and mineralogical changes is described in more detail in Chapter 4.

D. Biotite-feldspathic schists

Biotite and plagioclase-rich schists outcrop in two texturally distinct areas, one around Schulter in the upper Teuchl valley (B8), and the other on the Tröger Alm (D5 etc.).

The plagioclase weathers opaque white or cream-coloured and the biotite to a rusty brown, and the brown and cream streaked outcrops with deep rusty-brown debris are characteristic of this lithology.

1. Schulter biotite-feldspathic schist

These schists are texturally akin to the Kreuzeck schists. Quartz and plagioclase (30-50 modal % each) and biotite (10-20%) with some muscovite, chlorite and accessory ilmenite, graphite, tourmaline, apatite and zircon make up the typical biotite feldspathic schist, but some horizons are markedly more aluminous, containing more abundant muscovite, with garnet and occasionally a little staurolite or kyanite. These garnets are large (to 1 cm or more) and often show spectacular spiral trails of quartz and ore. Where muscovite is sufficiently abundant it forms parallel felts as in the Kreuzeck schists.

Within the outcrop of this lithology there are a few small enclaves of quartz-muscovite-rich Kreuzeck schist (Plate 1, B9) and locally near the southern boundary of this rock group the biotite feldspathic and Kreuzeck schists are somewhat interbanded.

2. Tröger biotite-feldspathic schist

These are texturally similar to the Polinik schists. A typical modal analysis, sample 0474, can be found in Table 5-2 in Chapter 5, and their compositional similarity to biotite-rich elements of the Polinik schists can be seen by comparing this analysis with sample 058.

The Tröger biotite feldspathic schists have a strongly-developed planar mineral foliation and a medium grain size with quartz 0.5 to 3 mm, plagioclase 0.2 to 1 mm and micas up to 2.5 mm long (Fig. 2-5). Apart from quartz, plagioclase, biotite and minor muscovite, fibrolitic sillimanite is quite common associated and intergrown with biotite and quartz. In the centres of larger mats of fibrolite, more coarsely crystalline sillimanite can often be found. Garnet is rare, and staurolite recorded from only one occurrence near the lithological boundary with the Intermediate schists south of Schneestellkopf (C6). Accessory ilmenite, tourmaline and apatite occur, and occasional retrogressive chlorite after biotite and mats of sericite after sillimanite may be found.

E. Quartzites

Black, grey and white quartzites have been distinguished on the map, but the black and grey varieties also occur within the schist units as laterally discontinuous single layers on a scale too small to show on the map, which are particularly frequent in the Kreuzeck schists. Their grain size is very sensitive to modification by deformation (see Chapter 3), and with the exception of some parts of the white quartzite bands, they appear as fine-grained rocks (less than 0.5 mm grains).

1. Grey quartzites

These are most common in the scattered mode of occurrence, as single layers from a few cm to 1 m thick within the schists, but are locally abundant enough, with little intervening micaceous material, to be mapped separately. Such concentrations of quartzitic material occur north of Strieden Kopf summit (A7) and north of the Feldner Hütte (C-D 10). The grey quartzites are light grey in colour with tiny

muscovite flakes visible on their crude foliation surfaces and occasional partings. Muscovite, opaque ore and occasional skeletal garnet, often concentrated in narrow bands, are minor constituents, and biotite, partly chloritised, is sometimes found. The narrow muscovite laths, average length around 0.2 mm, usually have a well-defined preferred orientation.

2. Black quartzites

Black quartzites are most abundant north of the Feldner Hütte (C 10), where they occur in mappable units, traceable a certain distance along strike, interbanded with grey quartzites and mica-schists. They are well-jointed rocks with good partings every 1 to 5 cm and form bands up to 3 m thick. Single thick bands have a tendency to subdivide along strike, and a single band on the map may represent a group of bands on the ground. This concentration of quartzitic material cannot be traced round to the north side of Kreuzeck, or to the east of the Feldner Hütte.

Black quartzite occurs sporadically through the Kreuzeck Schists (e.g. south end of Dechant Schneid (D 10), H. Gerbers Hütte area (Plate 2)), in thin, 1 to 10 cm thick bands, sometimes deformed into boudins of lenticular cross-section. Black quartzites are uncommon outside the Kreuzeck schists, but have been found as thin (up to 1 m) usually solitary bands in both the Intermediate and Polinik Schists.

The black colour is due largely to graphite, both disseminated and in amorphous clumps, which rubs off on the fingers when the rock is handled. In paler varieties, without abundant graphite, the colour appears to be due to a very fine disseminated opaque dust within the quartz grains. Fine-scale grey-black banding is found in some quartzites caused by variation in the amount of opaque material. Muscovite is usually present, and biotite, chlorite and garnet found

in small amounts.

3. White quartzites

White quartzites are apparently confined to the upper Teuchl valley area (B 10, D9 etc.). They are conspicuous white rocks, sometimes with a faint greyish-black streaky banding, occurring in bands 50 cm to 2 m thick, crudely foliated and conformable with the surrounding schists. Muscovite is the only other important constituent (around 3 modal %).

F. Marbles

Carbonate rocks are uncommon in this part of the Altkristallin Sheet but several small occurrences exist within the map area. The smaller bands are discontinuous, but the more substantial ones can be followed across country. The five more important occurrences are marked on Plate 1 and described below:

1. South Wunzental (C, D2)

A creamy-coloured, grey and green-speckled band of impure marble about 2 m thick outcrops in two stream-gulleys on the West side and at one locality near the foot of the crags on the East side of Wunzental. Calcite in 0.5 mm grains is the main constituent. Dolomite was not proved optically. Other minerals present are blue-green amphibole, chlorite, plagioclase, K-feldspar, clinozoisite, quartz and sometimes a little diopside, white mica or phlogopite/biotite. Sphene is a common accessory. In one section scapolite is common, in association with plagioclase, and its high birefringence (up to second order sensitive tint) indicates a composition close to the meionite (Ca-rich) end-member.

The assemblages in many of these rocks are unlikely to be in

equilibrium, and the textures are complex. Two plagioclases are present in many sections, one anorthite-rich, usually bytownite, one more sodic (albite to andesine). Clinozoisite occurs in small 1 to 2 mm clusters with plagioclase, perhaps pseudomorphing on An-rich plagioclase. Chlorite and amphibole generally do not coexist in the same section. It would seem that a high-grade amphibole-(diopside) An-rich plagioclase assemblage is being replaced by a lower grade one containing clinozoisite-chlorite-An-poor plagioclase. Since quartz is generally present the absence of dolomite might indicate metamorphic conditions outside the stability field of dolomite + quartz for the high-grade assemblage.

2. West side of Geistlacke corrie (F3)

Here is a 30 cm thick band of calcite-dolomite marble, pure white when fresh. The outcrop is illustrated in Fig. 3-3. The marble is banded, with alternating pure calcite and dolomite + calcite layers, which are distinguishable because of the finer grain-size (less than 1 mm) and rougher weathered surfaces of the dolomitic bands compared with the up to 5 mm grains of the calcite bands (Fig. 2-6). Occasional silica-bearing veins which cut the marble contain large (up to 3 cm) crystals of zoisite, tremolite and talc. The marble is generally separated from the surrounding schists by about 1 cm of blue-green amphibole rock.

3. East side of the Raggatal (B4)

These are white, yellowish and impure greyish marbles often banded in colour and composition. In places they are strongly mineralised along veins and fissures which in former times were worked for iron ore. Their fine grain size (0.1 mm) is due to mechanical reduction by shearing, and large carbonate grains, or porphyroblasts of other minerals are found within "augen". The carbonate-rich bands

have taken up most of the strain, and bands containing more calc-silicate minerals are less deformed. Both calcite-rich and dolomite-rich lithologies are represented, and other minerals occurring include phlogopite, clinozoisite, a pale (probably tremolitic) amphibole, quartz, plagioclase and sphene.

4. South and southeast of Strieden Kopf (B8)

The Strieden Kopf marble band, varying in thickness from 2 to 25 metres, extends from the edge of the map area just southwest of Strieden Kopf southeastwards to near spot height 2378 m where the band is cut out against the Bären Kopf shear zone. The band is interfolded with the surrounding schists on a mesoscopic scale (e.g. Fig. 3-47) and locally in its thicker parts is somewhat interbanded with mica-schists. The rocks consist of buff or blue-grey coloured calcite marbles, often colour-banded, with dark grey streaks, or bands rich in impurities (mostly feldspar) which stand out on weathered surfaces (Fig. 2-7). The colour banding can be seen folded around isoclinal folds.

In thin section, the minerals are usually all deformed, the calcites being intensely lamellar-twinned, bent, and often reduced to a fine grain-size (less than 0.1 mm) by mechanical granulation. Calcite appears to be the only carbonate present: dolomite was not proved optically. The principal impurities are quartz, plagioclase and white mica. Pale brown phlogopite is found in some grey bands. Accessories are sphene and opaques which include graphite and an ore mineral (its white alteration suggesting ilmenite). The colour-banding is largely produced by the variable distribution of opaque material.

5. Above the Penker Gemeinde Alm (D8)

1000 m NNE of Dechant, and 250 m SE of the Penker Gemeinde Alm, a narrow band of impure marbles and calc-silicates about 1 m in width crops out in a small stream. The rocks are banded on the scale of a few millimetres, the usual alternation being of bands rich in calcite with minor phlogopite, quartz and clinozoisite, interleaved with bands of calc-silicates (clinozoisite and quartz, with or without plagioclase and actinolite) in which calcite is subsidiary or absent. Black, blue-black and cream colour-banding is caused by the different distribution of opaque material (ore and graphite) among the bands.

G. Amphibolites

Amphibolites occur as small lenticular, pod-like bodies from 1 to 5 m long, as larger irregularly shaped bodies, or as fairly continuous concordant bands. The amphibolites of the map area can be divided initially into three groups, those occurring within the areas of outcrop of the Polinik, Intermediate and Kreuzeck Schists. Further subclassifications are recognised and described below:

(1) Amphibolites from the Polinik Schists

  (a) Eclogitic amphibolites, containing fresh or relict clinopyroxene

  (b) Non-eclogitic amphibolites, but including some amphibolites which display textural evidence of derivation from eclogitic amphibolites

(2) Amphibolites from the Intermediate Schists

(3) Amphibolites from the Kreuzeck Schists

  (a) Coarse-grained amphibolite

  (b) Fine schistose amphibolite

On Plate 1, type (1)a is distinguished from the other, non-eclogitic types.

1. Amphibolites from the Polinik Schists

(a) Eclogitic amphibolites

These are easily recognisable in the field by the pale green colour of the omphacite and its pseudomorphs. Other amphibolitic rocks are dark green to black in colour. The "Eclogitamphibolit" of Teuchl referred to by Austrian authors (for example Angel, 1930) is the Xanitzl Alm body (FG3) of this map, which in certain bands displays conspicuous light-green pseudomorphs after clinopyroxene, but no fresh grains or relict cores were discovered in this study.

In a number of localities, displayed on Plate 1, pod-like bodies containing fresh pyroxene are found, and in three cases the fresh assemblage garnet-clinopyroxene-hornblende-quartz-rutile is preserved in the cores of the pods. This assemblage appears to be in equilibrium. Plagioclase is absent.

In hand specimen fresh rocks display flesh-pink garnet, black amphibole and pale green clinopyroxene. The garnets are generally euhedral, 0.5 to 2 mm in diameter and are often crowded with inclusions of hornblende, quartz and rutile. Clinopyroxene is colourless to pale green in thin section, in prisms usually 2 to 5 mm long but sometimes up to 4 cm, enclosing whole garnets and occasionally small amphiboles. The optic axial angle 2V is in the range 60° to 75° (positive sign) which is typical for omphacite. The amphibole ? is a very pale brown colour and occurs in stubby grains interstitial to omphacite and garnet, about 0.5 to 1 mm long, and also as small inclusions in garnet. They are optically negative with a $2V_{\alpha}$ of around 80° and

are probably 'carinthine' type hornblendes like those described from the eclogites of the Saualpe Altkristallin by Mottana et al. (1968) and other authors. Quartz is in unstrained interstitial grains, and rutile is a common accessory.

Suites of specimens collected between the cores and the margins of these pods revealed a sequence of textural and mineralogical changes indicating breakdown of this assemblage and its replacement by a non-eclogitic assemblage. Some of these textures are illustrated in Figs. 2-9 to 2-11. The omphacite is the first mineral to break down. It is replaced, starting usually at mutual interfaces with garnet and hornblende, by a fine symplectite composed of a small-scale intergrowth of green amphibole and plagioclase (Fig. 2-9). Further towards the margin the sympletcite continues to grow at the expense of garnet (Fig. 2-10) and the pale brown hornblende. A biotite-plagioclase intergrowth also appears in many specimens, corroding garnet and pale-brown hornblende. Rutile becomes rimmed by sphene. Generally, the greater the extent of the alteration, the coarser the grain-size of the symplectite. Sometimes the garnets develop a reaction rim of fuzzy-outlined clinozoisite prisms, and some opaque ore forms, apparently as a result of garnet breakdown. The final stage of alteration, at the pod rim, is a rock consisting of a rather spongy greenish amphibole, with or without biotite, straggly remnants of garnet, intersitital plagioclase and quartz, and accessory sphene and ore (Fig. 2-11).

Locally another type of alteration is apparent. Garnet is replaced by chlorite (blue birefringence), plagioclase (apparently albite) and clinozoisite in varying proportions while the omphacite is replaced isostructurally by a very pale blue-green amphibole.

(b) Non-eclogitic amphibolites

Most of the other amphibolites within the Polinik Schists are dark green or black rocks whose mineralogy is dominated by a blue-green or brownish-green hornblende, often pale in colour in thin section, making up 70 to 90% of the mode. Plagioclase feldspar is usually present in small amounts, but is not conspicuous in hand specimen. Other minerals occurring are quartz, biotite, garnet, clinozoisite/epidote, and accessory chlorite, sphene ore and calcite. The hornblende is in the form of interlocking prisms 0.5 to 3 mm long, frequently very spongy with inclusions of plagioclase, a texture which suggests that these rocks too have been derived from eclogitic types. Patches, streaks and bands of paler green rock within dominantly non-eclogitic bodies are sometimes found, and these consist of the characteristic symplectic blue-green hornblende-plagioclase intergrowth of recrystallised eclogitic rocks, studded with somewhat corroded porphyroblasts of garnet with or without some clinozoisite/epidote. The plagioclase of Polinik amphibolites is of highly variable composition, a range from albite to bytownite having been recorded, though andesine or oligoclase compositions are most common. Its habit is usually interstitial, or ophitic with respect to amphibole. Garnets are either small and straggly, or in the process of breaking down. In the Xanitzl Alm body (FG3) garnets are seen in all stages from 1 cm porphyroblasts crowded with inclusions (mostly of hornblende) through reaction rims to complete replacement by biotite and plagioclase. The large, unaltered garnets often have an internal fabric which is discordant to the alignment of hornblende in the rock.

A faint compositional banding due to alternations of garnet-rich, biotite-rich or plagioclase-rich, plagioclase-poor amphibolite on a scale of a few cm or tens of cm is noticeable in places.

Aspects of the field relations of the amphibolites within the Polinik Schists are further described and interpreted in the following chapter.

2. Amphibolites from the Intermediate Schists

Amphibolites from the rest of the map area show no textural evidence that they once contained eclogitic assemblages.

In the Intermediate Schists amphibolites occur in concordant bands 1 to 10 metres in thickness. Their dark green to black colour makes them fairly conspicuous in the field. Blue-green hornblende is the major component (some are almost monomineralic hornblende rocks), with subsidiary plagioclase, quartz, clinozoisite/epidote, chlorite or biotite. Garnet is rare (only one occurrence noted). The grain size is variable. Coarse varieties have hornblendes between 1 and 2 mm long, fine-grained rocks may be largely composed of prisms 0.1 mm diameter and 0.5 mm long. Chlorite seems to be present as a breakdown product of biotite, replacement being complete in some samples. A fine-scale dark-light banding in some of these amphibolites is due to alternating hornblende-rich and clinozoisite-plagioclase rich layers.

3. Amphibolites from the Kreuzeck Schists

Amphibolites crop out in three principal concentrations within the area of Kreuzeck Schists studied in this work; the Kreuzeck-Feldner Hütte area (C-D 10-11), the Hochtristen area, and the H. Gerbers Hütte area (Plate 2). They occur in large, somewhat irregularly shaped bodies, e.g. around the Feldner Hütte or on Hochtristen, or as continuous bands up to 50 m thick. Two lithological types can be distinguished, coarse-and fine-grained, but are too closely associated in the field to be distinguished on the map.

(a) Coarse-grained amphibolites

This type contains markedly more plagioclase than other amphibolites in the map area. Foliation surfaces display plumose clusters of black hornblende prisms up to 2 cm in length set in white plagioclase, or plagioclase and apple-green epidote. Banding on various scales is produced by variation in the proportions of these principal minerals, or by variation in grain-size from band to band, which causes the fine-grained bands to appear darker and more intensely foliated than the coarser bands (Fig. 2-12). Quartz, chlorite and/or biotite are minor constituents, with accessory sphene apatite, ore and calcite. In thin section, at least in the Kreuzeck area coarse amphibolites, the grain size is seen to be finer than it appears in hand specimen. The large prisms of hornblende show undulose extinction or have broken down into smaller subgrains of slightly different orientation. Epidote patches are not single crystals but composed of aggregates of tiny (50μ) grains (Fig. 2-13). Plagioclase (generally oligoclase) occurs in equidimensional 0.1 mm grains. Biotite forms olive-brown pleochroic laths up to 0.5 mm long, often partly or completely replaced by chlorite.

(b) Fine schistose amphibolites

The fine-grained amphibolites are even-grained and green to almost black in colour. Their usually well-developed planar schistosity often has a silky sheen. Many of these fine amphibolites contain similar proportions of the same minerals as the coarse amphibolites, and it is possible that they represent a textural modification of the same rock composition due to tectonic events at the time of crystallisation of the minerals. The fine schistose amphibolites tend to be found at the margins of amphibolite bodies, and grade into coarser amphibolites towards the centre. The silky-

looking amphibolites generally contain chlorite as an important constituent (bright grey-white birefringence, optically positive, thus probably a magnesium-rich chlorite). Variously-sized (0.1 mm to 1 mm) prisms of hornblende are set in parallel-felted mats of chlorite, with interstitial oligoclase and quartz.

H. Leucocratic gneisses

The leucocratic gneisses of the map area can be subdivided geographically into the following groups, which are also distinct in composition and lithological character:-

1. Launsberg Alm area, 2 km west of Obervellach (Plate 2)
2. Gamskar Area (D3)
3. The Kallukerl - Blass Alm gneiss body (D,E 4 and Plate 2)
4. Schistose Leucocratic gneisses in the Kreuzeck schists (C 10 etc.)

1. Launsberg Alm

These are white rocks with a coarse grey-green streaky gneissose texture, often with 2 cm feldspar augen. They form at least two bands about 20 m thick, trending WNW-ESE, north of the Launsberg Alm. One of these is quarried near the river bank 2 km west of Obervellach.

In thin section they consist of microcline perthite, albite, quartz, muscovite and chlorite with accessory sphene, apatite, epidote and ore. Usually all the minerals are somewhat deformed. Albites are "gefüllte" with microlites mostly of white mica. Occasionally 0.5 mm euhedral garnets may be found, now mostly pseudomorphed by chlorite, or chlorite + albite. The microcline perthites occur as large, 1 cm or greater grains in the augen or as smaller 0.2 to 0.5 mm grains in the groundmass. Larger grains often show evidence of two phases of perthitic exsolution (Fig. 2-14):- 0.4 mm rounded blebs of

albite, now packed with microlites of white mica, sit in a microcline host which now shows patches and stringers of exsolved sodic feldspar.

2. Gamskar area

Leucocratic gneisses occasionally occur in thin bands in the Polinik Schists. The most conspicuous example is a single band of biotite-microcline-augen-gneiss, varying from 1 to 8 metres thick, which crops out in the corrie east of Gamskar. It is a grey-white rock with white augen, with biotite usually an obvious constituent in hand specimen. The assemblage is quartz-microcline-oligoclase-biotite-muscovite with accessory zircon, apatite, epidote and chlorite. The augen consist of granular, strained microclines 1 to 4 mm in diameter, rather than single crystals, some slightly perthitic but most not obviously so. The groundmass consists of quartz, microcline and oligoclase of 0.5 mm mean grain size, with flakes of biotite and muscovite up to 1.5 mm long, generally oriented within the gneissic foliation.

3. Källukerl - Blass Alm gneiss body

This large body, 10 km long and 2 km wide, called the Blass Alm augen gneiss body by Wright (1973), consists mainly of a coarse-grained crudely foliated white augen gneiss, sometimes discoloured brown-orange due to the weathering of iron-bearing minerals filling veins and fissures. Near the margins of the body, and where the gneiss forms thin screens surrounded by schists, the foliation is much more strongly developed, and parallel to the lithological contacts. In these marginal gneisses, feldspar augen are much less abundant, or absent.

In thin section the rocks are seen to consist of microcline, albite or albite-oligoclase, quartz and muscovite with accessory ore, apatite and chlorite. Augen are of two kinds: the larger (1 or 2 cm

diameter) consist of perthitic microcline, the exsolved sodic feldspar in the form of irregular blebs and streaks, the smaller (2 to 4 mm) of gefüllte albite, with microlites of white mica (Fig. 2-15). In some rocks, generally those with more abundant large augen, microcline predominates over albite in the groundmass, but in others albite is more abundant than K-feldspar. Most rocks in which albite predominates are from the more strongly foliated marginal rocks.

4. Leucocratic gneisses in the Kreuzeck Schists

These are white, often schistose, rocks, usually without augen, more strongly foliated than other leucocratic gneisses in the map area. Their gneissic banding is frequently seen to be tightly folded (e.g. Fig. 3-21). They are frequently associated with amphibolites, as narrow 50 cm to 2 m bands either contained within or at the margins of amphibolite bands and bodies. They are mostly more sodic in composition than the other leucocratic gneisses, and consist of quartz, albite and muscovite, with or without a little biotite or chlorite. An exception is the rather unusual lithology represented by sample 0424, from part of the gneiss body associated with amphibolites 230 m northeast of Kreuzeck (C10). Multiply folded quartz stringers are set in a finer grained matrix (see Fig. 3-22) in which a Ca-rich plagioclase has broken down to epidote + albite + muscovite, and is associated with clear microcline and quartz.

The grain size of these leucocratic gneisses is rather variable, as the quartz is frequently broken down into very fine granular aggregates. Albite and ungranulated quartz are usually in the range 0.2 to 0.5 mm, with micas about 0.5 mm long.

1. Fine biotite gneiss and biotite augen gneiss

The principal occurrence of these rocks is on the upper Kleinblässen Alm, southwest of Kallukerl (C,D 4,5). They are grey or purplish-grey fine grained rocks, which contain variable amounts of cream-coloured feldspathic material as isolated prophyroblasts, elongate streaks or continuous bands. The biotite augen gneisses have the light-coloured feldspathic material in augen and stringers defining a gneissic banding which is often deformed around mesoscopic folds.

Mineral assemblages in the biotite gneisses are:- quartz-albite or oligoclase-microcline-biotite-clinozoisite, or quartz-albite or oligoclase-biotite-clinozoisite with accessory muscovite, sphene, apatite and chlorite. Microcline occurs in augen up to 2 cm across which may be made up of large single crystals or of aggregates of 1 mm grains which may have narrow rims of albite. They usually display fine thread-like perthite exsolution, and may contain larger blebs of lamellar-twinned albite. Augeniferous rocks have microcline in the groundmass as well as a plgioclase, but potassium feldspar appears to be absent from the more uniform, augen-free, fine biotite gneiss. The fine grained groundmass (quartz and feldspars less than 0.2 mm, biotite generally less than 0.5 mm) consists of quartz, feldspars, biotite and clinozoisite. Varying proportions of biotite + clinozoisite to quartz + feldspar sometimes produce a fine-scale, dark-light banding in augen-free rocks (Fig. 2-16). Clinozoisite occurs as colourless rounded grains or as needles with a small extinction angle. Large rounded grains occasionally have red-brown pleochroic, higher birefringent cores (Allanite ?).

The body 700 m north of Kreuzeck is a coarser grained streaky biotite gneiss (quartz and plagioclase 0.2 mm, micas 1 mm) with dispersed augen. Clinozoisite and K-feldspar are not present.

J. Pegmatites

Pegmatitic veins in the map area can be divided into two types, distinguished as older and younger pegmatites, according to their texture and relationship to deformations in the host rocks.

1. Older pegmatites

These are white, feldspar-rich, usually medium-grained rocks of sugary appearance, in bands from a few centimetres up to 3 metres thick, concordant with the schists in which they lie. They are deformed, and frequently show pinch and swell, boudinage and rodding structures related to the major deformations in the host rock (e.g. Fig. 3-11). They consist mostly of feldspar but grains of clear quartz and muscovite are apparent in hand specimen, and sometimes also pink garnets up to 4 mm across. In thin section, the feldspar is seen to be albite, and together with quartz forms an equigranular mosaic of 0.5 mm grains. Muscovite, garnet, sphene, epidote and apatite are accessories.

2. Younger pegmatites.

Coarse-grained pegmatite veins from a few centimetres to 25 metres thick are common in two parts of the map area, in the Polinik schists, and on the Tröger and northern Eisen Alms (B-D 5-6). They may be more or less concordant with the schists, or markedly cross-cutting on both the small (Fig. 2-17) and the large scales (e.g. large "dyke" west of Schneestellkopf, B5, 6). A slight regional variation in texture and field relations is noticeable. On the Tröger Alm they are very coarse-grained and structureless, and almost invariably discordant to the schistosity. Towards the north, in the Polinik Schists, however, many of the pegmatites are more or less concordant, and some have a weak foliation, which in most cases appears to be a tectonically produced

feature and not related to a primary orientation of minerals such as books of muscovite. Fig. 2-18 shows a pegmatite intruded into folded Polinik Schists. An arm of the pegmatite is conformable with the schists on the limb of the fold, and is crudely foliated. The irregular body in the fold hinge is structureless and cuts the folded layers.

The commonest mineral assemblages in the younger pegmatites are

quartz-albite-muscovite, and

quartz-albite-muscovite-tourmaline

A rarer but important assemblage is

quartz-albite-muscovite-Al silicate (andalusite, sillimanite or kyanite)

Biotite and rare garnet are found occasionally in the younger pegmatites. The tourmalines are frequently broken, with fractures filled with quartz. The grain size is somewhat variable, but may be up to several centimetres. Muscovite books up to 4 cm thick, black tourmaline 10 cm long, and pink andalusite (an occurrence well-known to local mineral collectors, 600 m south of Schneestellkopf, C6) up to 12 cm or more long may be found.

K. Dykes

A suite of dykes of intermediate composition (andesitic-dacitic), often garnet-bearing, cut the schists and gneisses of the Kreuzeck Group Altkristallin. Their petrography, field relations and chemistry have been described by Wright (1973) and not a great deal need be said about them here. They are 30 cm to 12 m thick, discordant, undeformed, and often extensively hydrothermally altered. The dykes which crop out within the present map area can be divided into three types, of which types 1 and 2 are broadly equivalent to Wright's (1973) types C and D respectively.

Type 1: light coloured, grey or green-grey, weathering light grey, grey-brown or cream. Phenocrysts of plagioclase, biotite, quartz, garnet and occasionally hornblende. Localities: Blässenbach dyke (F5) (not precisely located, in inaccessible stream gulley, sampled from fallen blocks); Dechant schneid dyke (D 10); Zweiseetörl dyke (Plate 2).

Type 2: dark grey, weathering brown-grey. Phenocrysts: needles of hornblende, plagioclase. Localities: small dyke 300 m NW of Dechant (C9); Scheuchen Kopf dykes (B8)

Type 3: grey or black and white speckled (according to grain size), weather brownish or cream-speckled. Coarser grained, less distinction between phenocrysts and groundmass. Assemblage plagioclase (65%) + hornblende (30%) + biotite. Phenocrysts: hornblende, or absent. Locality: 2 dykes near Gössnitztörl (A9).

Alteration products include chlorite, calcite, epidote, sphene and ore after biotite and hornblende, and white mica, albite, zeolites and calcite or clinozoisite after plagioclase.

L. Shear zones

Cataclastic rocks and mylonites occur in a number of shear zones which cut the Altkristallin of the Kreuzeck Group and trend E-W to SE-NW. The zones vary from a few metres to over 100 metres in width, and their present attitude is steep: most of them dip south at between 60^{o} and 90^{o}. Their course can be traced in the field by topographic depressions and lineaments, by fine dark flaky debris in the superficial cover, and sometimes by lines of springs (e.g. the large spring near the Penker Gemeinde Alm (D8) rises from the Bären Kopf shear zone).

Near the zone margins, and in less-deformed "screens" within zones, the nature of the wall rock is recognisable though strongly leached and cut by blackened shear surfaces. In the central parts of shear zones cataclastic rocks are developed which may be divided into two types, called here a cataclastic and a mylonitic type. The two varieties are not usually found in the same zone. Wright (1973) reviewed the classification and nomenclature of cataclastic rocks with reference to the shear zones of the Kreuzeck group, but the distinction between these two types, which is largely independent of the clast-to-matrix ratio, does not readily find a place in this classification. A third rock-type, a cross-cutting ultramylonite, is sometimes found.

1. Cataclastic type

These are brittle, black rocks which tend to break into lenticular fragments. Breaking these open reveals a black matrix set with rock and mineral fragments (Fig. 2-19). In thin section (Fig. 2-20) the matrix is brown or black, isotropic or opaque. Feldspar clasts are usually sericitised, but quartz clasts survive, though strongly strained. Fine sericite and chlorite replace micas and garnet. Some rocks, such as 067 from the Launsberg shear zone, may have been derived from amphibolites, and now consist of chlorite + calcite + quartz.

2. Mylonitic type

The mylonites of the map area are grey or grey- green rocks which in hand specimen look like fine siltstones (Fig. 2-26). They are frequently finely banded and often fissile parallel to the banding. In places the banding is folded or irregularly disturbed, and the disturbances truncated by planar banding of character identical to the deformed banding. The colour and texture of the mylonite may

depend on the parent lithology. Mylonites formed from quartzose rocks are pale grey, whereas those formed from micaceous rocks are dark grey or dark greenish grey.

In thin section (Fig. 2-22) they are fine-grained rocks with fewer and smaller clasts than type (1), or none at all. The brownish matrix of the darker varieties is not opaque or isotropic but consists of very fine white mica (sericite) and finely granulose quartz, with a little chlorite. Often the matrix tends to extinguish parallel to the fine banding under crossed polars. Pale varieties consist primarily of extremely fine grained quartz (Fig. 2-23). Assemblages recorded are:-

quartz

quartz + sericite

quartz + sericite + chlorite

quartz + sericite + chlorite + albite

plus occasional clasts of quartz, muscovite, chloritised garnet, etc. and accessory opaque material. Banding in the mylonites either by varying proportions of quartz and sericite + chlorite, or by varying concentrations of very fine dusty opaque material.

3. Ultramylonite

"Intrusive" mylonites, discordant dark veinlets of black, finely milled rock with some small rock and crystal fragments are occasionally found intruding less sheared rocks along irregular cracks (Fig. 2-24). The most important occurrences are associated with the Teuchl shear zone in the Teuchlbach near Napplach (Plate 2), and with the Eisen Alm shear zone. They occur in association with both cataclastic and mylonitic shear zones.

The existence of two types of shear zone lithology, which are generally not found together, suggests two phases of shear zone formation and movement, perhaps under different conditions. This is discussed further in the following chapter.

CHAPTER 3

Structural Geology:- Analysis, Correlation and Tectonic History

A. Introduction

The object of structural analysis in the map area was to work out the tectonic history in terms of a chronological succession of deformational events and episodes of mineral growth, and to throw light on the nature and amount of internal strain in the Altkristallin Sheet during episodes of its history. The emphasis in this work is laid on correlation of folds rather than on descriptive aspects of structural geology. The problems of correlation of deformational events in metamorphic terrains are not small, and the difficulties have been summarised in the cautionary observations of Park (1969). Particular attention has been paid to correlation in the map area by means of style and orientation of folds, and the relationship of deformation to mineral growth.

When using fold style in correlation, the factors which affect deformational style should be considered. Important factors are the nature (e.g. how homogeneous, nature and scale of inhomogeneity or anisotropy) and mechanical properties of the rock. The mechanical properties further depend on the physical conditions at the time of deformation; temperature and pressure, and on time-dependent factors such as rate of stress loading and resulting strain rate.

Over wider areas, variation in these parameters may cause the style of folds of a single deformation event to vary accordingly. Elevated temperatures have the effect of promoting rheid flow, and decreasing rock strength. While concentric folding preserves the uniform thickness of the folded layer, the generation of similar folds requires the internal deformation of the layer, largely by shear parallel to the fold axial surfaces (Ramsay, 1962). Thus a folding deformation

which shows a gradual change in style and intensity over a few kilometres from more concentric to more similar folds in comparable lithologies, may indicate a temperature gradient at the time of deformation.

In a small area, physical conditions can be regarded as uniform at any one time, and the range of styles displayed by any one deformational event will depend very largely on lithological factors alone. Under these circumstances it is relatively simple to establish a structural sequence in a restricted area.

Given reasonably continuous exposure and no large tectonic discontinuities, it is possible to trace gradual regional variations of style and orientation into neighbouring areas and effect correlations with other locally determined structural sequences. This is the technique employed in this study.

The uplift of the Hohe Tauern areas has exposed different levels of the Altkristallin sheet in the Kreuzeck Group, which therefore represent different physical conditions at earlier times. Two areas of distinct structural style are evident in the map area, with an intervening zone characterised by elements belonging to each of the other styles. Structural sequences have been established for the Polinik schists, north of the Teuchl shear zone, and for the Kreuzeck schists in the southern part of the map area, and lateral variations are traced northwards from the latter area until it becomes possible to compare the two sequences directly to each other. Later brittle deformations which affect the whole map area are treated separately after the correlation of the principal phases of folding.

The notation employed is a sequential numbering of folding deformations with a subscript number, with a superscript letter to denote which of the local sequences is being referred to. Thus F_3^N denotes the third folding deformation in the northern part of the

map area. Linear features associated with folds are denoted by L and axial planes by S, e.g. S_2 are the axial surfaces to F_2 folds. The superscripts are generally omitted once correlation throughout the map area has been demonstrated. Non-folding deformational events such as the formation of the shear zones are not included in this scheme.

Terms used to describe the tightness of folds are employed in the sense used by Fleuty (1964).

B. Structural sequence in the North

The structures in part of the area of the Polinik Schists were studied during the summer of 1964 by E.R. Oxburgh (Oxburgh, 1966) who established a sequence of three main folding deformations, which correspond to F_1^N, F_2^N and F_3^N of this work.

1. F_1^N deformation

The first folding phase on Polinik F_1^N, the 'isoclines' of Oxburgh, are visible only in scattered localities, generally as refolded single layers making interference patterns of Ramsay's type 3 (Ramsay, 1967) with folds of the following generation (see Figs. 3-1, 3.2). Little can be said about F_1 folds other than that they are almost isoclinal. It is probable that some of the small-scale isoclinal folding in quartz stringers within the mica-schists also belongs to this phase especially as quartz-rodding lineations discordant to neighbouring F_2 fold axes are locally found. These anomalous rodding directions generally trend about NW-SE where F_2 folds are E-W.

The Geistlacke corrie (F3) marble band displays asymmetric isoclinal folds (Fig. 3-3) which very likely belong to this generation.

2. F_2^N deformation: (a) Folds

The F_2^N folds, ("buckles" of Oxburgh) are generally close to tight folds defined in the paragneisses and flaggy mica-schists by quartzofeldspathic bands 10 cm. to 2 m. thick, or by multiple layers of these competent bands separated by less competent micaceous material (Figs. 3-4, 3-5). Minor folds are frequently seen to be asymmetric but except on the south side of the Hohe Leiten ridge, where the majority of F_2^N minor folds have a clockwise sense looking east, there is no consistent rotation sense. The orientations of F_2^N axial planes (see Plate 4, rows 2-7) spread over a similar range to that of the principal planar feature, i.e. subhorizontal in the northern part of the Polinik schists, south-dipping in the south.

The style of F_2^N folds approaches that of similar folding with significant thickening of the hinge in most cases. This can be illustrated by means of isogon- (Class 2 of Ramsay (1967)) and thickness-plots of F_2^N folds after the manner of Ramsay (1967, p. 361) (Figs. 3-6, 3-7). Minor asymmetric folds in quartzofeldspathic bands often show very much thickened short limbs (Fig. 3-8). Quartzitic bands tend to keep a more nearly parallel profile around folds, and folded successions of competent quartzitic and incompetent micaceous schist frequently consist of alternations of Ramsay's (1967, p. 367) classes 1C and 3 folds. This is an indication that the F_2^N folds may have formed by a buckling mechanism, though their shapes have been substantially modified by flattening strain (Ramsay, 1967, p.432-3 and references).

The distribution of mesoscopic F_2^N folds is irregular. Small areas where the folds are abundant and stacked up in "cascades" alternate with zones where folds are rare and the foliation and lithological banding are of uniform attitude. This pattern implies the existence of larger structures which are not evident because of the lack of mappable continuous marker horizons.

In more homogeneous lithologies folds of this generation are usually not seen, but in the coarse-grained mica-schists small close to tight sharp-hinged folds with limb lengths of 2 to 5 cm. occur (Fig. 3-9), defined by coarse quartz-plagioclase lamellae separated by mica-trains.

(b) Linear structures

Thin quartz stringers within the Polinik schists are frequently deformed into tight or isoclinal small-scale folds, whose thickened hinges when exposed on foliation surfaces give rise to a prominent quartz rodding (Fig. 3-10). Frequently the rods have become partly or completely detached from the fold limbs and are "floating hinges" in the micaschist matrix. Quartz rodding directions and F_2^N fold axes have been plotted together on the stereograms of Plate 4.

In amphibolites a strong mineral lineation is developed in the hinge zones of F_2^N folds.

(c) Boudinage and lenticular structures

Non-brittle boudinage and pinch-and-swell structures (see Ramsay, 1967, pp. 103-6) are fairly common in the Polinik Schists, giving rise to boudins with a lenticular cross-section. Fig. 3-11 shows an almost separated boudin from a pinch-and-swell structure in an old, pre F_2^N pegmatite vein.

There is a sense in which the whole fabric of the Polinik Schists is lenticular, in that any stratigraphy there may have been has been totally disrupted, and competent lithological units cannot be traced far along strike before pinching out.

The most obvious expression of this disruption is in the 1 to 5 metre bodies of amphibolite and eclogite-amphibolite (Fig. 3-12) which have the three-dimensional form of flattened lenses, usually flattened

in the plane of the regional foliation. It is probable that these lenses have been formed by the disruption of basic dykes or sills by high flattening strain, but their irregular, stacked-up or en echelon distribution in large outcrops suggest that their mutual relationships have been considerably modified since their formation as separate lenses - i.e. that their formation predated F_2^N and they were subsequently reshuffled during F_2^N. Disrupted quartz veins also show this style (Fig. 3-13; example shown is from the Intermediate Schists of the Eisen Alm (D6)).

One amphibolite pod (Fig. 3-14) has been folded around what is interpreted as an F_2^N fold.

At about 1500 metres altitude between the Wunzen Alm and the Polinik Hütte (D1) a deformed quartzite band consists of a series of boudins of lenticular cross-section which contain asymmetric folds with much-thickened hinges and short limbs (Fig. 3-15). The boudin axes and asymmetric fold axes trend about 170° which may be interpreted as an F_1^N orientation. The boudins have been redistributed into an en echelon arrangement, and a compositional banding between them is folded about axes trending about 120°, an orientation appropriate for F_2^N in this subarea. The mineral foliation in the outcrop is axial planar to this later generation of folds, and passes discordantly into the boudins.

This evidence further supports the suggestion that the major disruption of competent lithological units in the Polinik Schists predates F_2^N.

(d) Foliation

In the Polinik Schists the mineral foliation or schistosity is usually parallel to the lithological banding, except near the hinges of F_2^N folds, where in micaceous lithologies, the schistosity defined

by unstrained micas is seen to be axial-planar to the folds and becomes discordant to the lithological banding as the latter curves towards the hinge (see Fig. 3-8). The quartzo-feldspathic layers which define the folds contain fewer and smaller micas which usually do not show any strong preferred orientation. The principal planar feature may thus be defined by the lithological banding (rotated parallel to the S_1 axial planes of isoclincal F_1^N folds) or by the S_2 schistosity axial-planar to F_2^N folds which for the most part is also parallel to lithological banding and S_1.

Foliation/lithology relationships become complex at the margins of larger irregularly-shaped bodies such as the Xanitzl Alm amphibolite body. In places a marked discordance is seen on a large scale between the foliations internal and external to the body (Fig. 3-16), but on a smaller scale (over the last 2 metres or so) the schist foliation swings into parallelism with the amphibolite margin. In the interior of large masses, the amphibolites display a crude foliation, usually parallel to any compositional banding, which becomes more intense towards the margin (Fig. 3-16). Locally thin screens or elongate lenticular masses of amphibolite occur within the schists close to the margins. These could represent sheared-out fold noses, or detached projections from an irregular body, now rotated into parallelism with the margin. Elsewhere (such as on the West wall of the Geistlacke corrie (F3)) the entire "body" consists of alternating more-or-less concordant bands of amphibolite and schist, each a few metres thick. These features indicate that a body of complex geometry predated the F_2 deformation.

3. M_2^N Mineral Growth

The episode of post-F_2^N mineral growth which defines the S_2 schistosity is designated M_2^N. Most of the rock-forming mineral assemblages in the Polinik Schists can be assigned to this phase, though an earlier M_1 phase is suggested by the relict eclogite assemblages within some amphibolites and perhaps by certain garnets with inclusion trails markedly discordant with the external fabric.

4. F_3^N Deformation

In micaceous lithologies throughout most of the outcrop of the Polinik Schists in the map area, a microfold lineation is conspicuous (Fig. 3-17). Coaxial with the millimetre-scale microfolds and with parallel axial surfaces, are gentle, open flexures of the foliation and lithological banding on scales of 50 cm. to a few metres. Locally the minor folding deformation is intense enough to destroy the earlier foliation, though only rarely does the crenulation cleavage produced by the microfolding become the dominant planar feature of the rocks.

The relationship of F_3^N microfolds to F_2^N can be seen in Fig. 3-18, a hand-specimen cut perpendicular to F_2^N and F_3^N fold axes. A thin quartzofeldspathic band is folded in a series of flattened F_2^N folds, and the micaceous matrix displays a strong mineral foliation which is axial-planar to these folds. F_3^N microfolds with well-defined axial surfaces deform this foliation, and in thin section are seen to bend the minerals which define it.

F_3^N axial surfaces are usually approximately perpendicular to the earlier foliation, and frequently the axial directions coincide with F_2^N. Where they do not, crossed lineations are commonly displayed, particularly in micaceous schists which contain abundant F_2^N quartz-rodding in quartz stringers. In the Ebeneck structural subarea (Plate 4 row 3A) F_2^N and F_3^N axes consistently diverge by about 40^o.

5. M_3^N Mineral Growth

The F_3^N folding deforms all minerals of the M_2^N generation but the patchily-distributed regional chloritisation observed in the Polinik schists of the map area post-dates F_3^N. Knots of chlorite laths overgrow F_3^N microfolds (Fig. 3-19). Some retrogression of M_2^N minerals has accompanied this growth of chlorite: staurolite and kyanite are partly sericitised, and garnet and biotite chloritised. The pseudomorphs are not deformed by the F_3^N microfolds.

6. F_4^N Deformation

The F_4^N structures are megascopic fold structures, responsible for the mappable changes in orientation of the regional foliation in the Northern part of the map area. The principal F_4^N feature is a large antiformal structure, the Polinik antiform (see Plate 3), which has an E-W or ENE-WSW axis and passes through the summit of Polinik. Smaller, wavy open folds on a 10 - 100 metre scale are displayed at the northern end of the Mörningköpfe (Fig. 3-20) and in the Bodensee area (B2 - C2), on the northern limb of the Polinik antiform. These are distinguished as a deformational event separate from F_3^N, since the F_3^N axial surfaces are folded around these structures, preserving their angular relationships with all earlier structural elements. This can be illustrated by comparing π-pole, F_2^N and F_3^N orientation data in those subareas affected by F_4^N folding (Plate 4, rows 3-6). Late strains recorded in the mineral fabric (particularly in quartz and in M_3^N generation chlorite etc. can be related to F_4^N folding (see discussion of quartz fabric strain, later in this chapter).

The Polinik antiform is truncated by the Teuchl shear zone and thus appears to be earlier in the tectonic sequence than the cataclastic shear zones.

TABLE 3-1. Structural sequence in the north of the map area.

	DEFORMATION		MINERAL GROWTH
Time ↓	F_1^N Early isoclinal deformation(s).	?	M_1^N Early mineral assemblages - eclogites
	F_2^N Tight, flattened buckles, similar folds		M_2^N Amphibolite facies mineral growth, axial-planar foliation to F_2 folds
	F_3^N Microfolds and flexures		M_3^N Retrogression, growth of chlorite, patchily developed.
	F_4^N Large open folds (Polinik antiform)		

7. The northern tectonic sequence can be summarised as shown in Table 3-1.

C. Structural sequence in the south

The Kreuzeck schists, their structural style very different from the Polinik schists, were studied in the area round the Feldner Hütte and in the region of the Hugo Gerbers Hütte (for location see Plate 2) which exhibit directly comparable structural sequences. Orientation data are presented in Plate 4, rows 12-13.

1. F_1^S Deformation and M_1^S mineral growth

The most obvious mesoscopic folds in competent lithologies (quartzites, leucogneisses and coarse-grained amphibolites) are similar-type close to tight folds with flat-lying axial surfaces and approx. N-S axes which deform a foliation generally parallel to the compositional banding (Figs. 3-21, 22). In micaceous and well-foliated rocks, folds of this generation are evident only as intrafolial isoclines and detached 'pothook' hinges in quartz veinlets and stringers (Fig. 3-23). Quartz-rodding, the three-dimensional expression of these detached hinges, is common in mica-schists and in the more schistose leucogneisses.

On the microscopic scale, there is evidence that the Kreuzeck schists were undergoing high strains while their amphibolite-facies mineral assemblage (the M_1^S mineral growth) was growing. Rotational strain has produced garnet and occasionally staurolite porphyroblasts with S-shaped trails of inclusions, and shear strain during growth may have been responsible for the characteristic parallel-felted muscovite fabric of the Kreuzeck Schists (see Figs. 4-1, etc.). This parallel-felted texture is absent from micaceous lithologies only in rare localities where F_1^S folding is less strongly developed than usual.

The effects of variable intensity of F_1^S strain can be seen in the

leucogneisses and amphibolites. Towards the centres of bands a few metres thick, the rocks are coarse-grained, crudely foliated, and compositional banding can usually be traced continuously around close to tight F_1^S folds. Towards the contacts of the leucogneiss or amphibolite with less competent material, however, the F_1^S folds become much tighter, with sheared-out limbs, the foliation much more intense, and the grain-size of the rocks finer.

The foliation in micaceous lithologies is defined by the parallel-felted micas, and is parallel to the compositional banding at least on the small scale, the latter having been rotated into parallelism with the F_1^S axial surfaces. The formation of the foliation of these leucogneisses and amphibolites where the F_1^S folds are not sheared-out predates F_1^S.

2. Deformational events predating F_1^S

Evidence of at least one significant folding deformation earlier than F_1^S is clearly seen in one outcrop of massive quartz-rich gneisses within the Kreuzeck schists south-west of Kreuzeck (Fig. 3-24). The flat-lying F_1^S folds are seen to refold earlier tight folds defined by light-coloured quartz stringers in the quartz-rich gneiss. It is possible that some of the phenomena elsewhere attributed to F_1^S, e.g. some of the tight similar-type folding of the compositional banding in the quartzites, may be attributable to this pre-F_1^S phase. Bands of quartzose material in micaschists (see Fig. 3-32) and in leucogneisses (Fig. 3-21) frequently appear to have been deformed more than once prior to F_2^S.

3. F_2^S deformation

Open buckle-folds accompanied by strongly developed microfolds in micaceous lithologies, with axes trending from N-S to WNW-ESE and steep-dipping axial surfaces, deform the regional foliation and flat-lying F_1^S axial surfaces in the Feldner Hütte area, (Fig. 3-25). These are designated F_2^S. The buckle-folds are defined by more competent bands such as quartzites, and interlimb angles rarely exceed 90^o. They are not so abundant or strongly-developed in the H. Gerbers Hütte area.

The microfolds are best studied in thin section. They are often sharp-hinged, chevron-like in muscovite parallel-felts (Fig. 3-26), but have a more rounded profile in quartz stringers and where quartz occurs within the felts. The mica fabric is frequently deformed around and between M_1^S porphyroblasts (Fig. 3-27).

4. M_2^S mineral growth

Certain minerals occur as porphyroblasts which grow over the F_2^S microfolds, and quartz has generally been annealed after the F_2^S folding. Thus an M_2^S crystallisation episode involved growth of biotite, muscovite, chlorite, plagioclase and quartz.

5. F_3^S deformation

A second set of buckle-folds affecting competent layers, and associated with microfolds in micaceous lithologies, is locally developed in the Kreuzeck Schists of the Feldner Hütte area. The fold axes trend E-W and do not scatter so widely as the F_2^S axes. The axial surfaces dip steeply north. Where F_2^S and F_3^S can be seen affecting the same layers, both F_3^S buckle-folds and microfolds are somewhat smaller-scale than the F_2^S structures (Fig. 3-28).

Although in the Feldner Hütte area this phase is not universally developed, around the H. Gerbers Hütte it becomes the dominant mesoscopic

folding phase in the micaceous lithologies. Fairly sharp-hinged nesting chevron-style folds of 1 to 10 cm. limb-length (Fig. 3-29) characterise the mica-schists, though folds of this generation in more competent bands are rounded open buckles. The distribution of poles to foliation reflects open folding about the F_3^S axis (Plate 4, row 13).

Under the microscope the F_3^S microfolds can be seen to deform the porphyroblastic minerals grown during M_2^S. Micas develop kink-bands and plagioclases become bent, showing undulose extinction. Chlorites which have formed in some localities by subsequent replacement of porphyroblastic biotite do not appear deformed by the F_3^S microfolds, suggesting a later, post F_3^S chloritisation event which may be termed M_3^S. Certain other 'late' retrogressive phenomena within the Kreuzeck schists may also be related to this event.

The Kreuzeck structural sequence is summarised in Table 3-2.

6. Lateral variation of structural elements and correlation with the north

Although the orientations of structures in the southern part of the map area are mesoscopically very variable, statistical consideration of the collected data reveals elements of the larger-scale structures (Fig. 3-31).

North and northeast of Kreuzeck (c 10) there is a significant change in the attitude of most structural elements. The foliation, which in the Feldner Hütte area fluctuates around the horizontal (Plate 4, row 12) becomes more consistently south-dipping. The effects of F_3^S folding disappear and the mesoscopic variation in foliation attitude is largely due to F_2^S. The F_2^S lineations and fold axes change from a subhorizontal mean orientation to a southeasterly or southsoutheasterly plunge of about 45^o over this same interval. Orientations of F_2^S axial surfaces are little affected as their trace makes a high angle with the trace of this

TABLE 3-2. Structural sequence in the south of the map area.

	DEFORMATION	MINERAL GROWTH	
Time ↓	F_0^S Early isoclinal and sub-isoclinal deformation(s).		
	F_1^S Tight folds, "pothooks" in mica-schists	M_1^S	Amphibolite facies syntectonic mineral growth
	F_2^S NW-SE buckles, microfolds	M_2^S	Porphyroblast growth
	F_3^S E-W buckles, chevron folds, microfolds	M_3^S	Local chloritisation

flexure. This large-scale, west-plunging, gentle synformal structure is displayed on the cross-section (Plate 3), and is henceforward referred to as the Kreuzeck Synform.

On the northern side of Kreuzeck and Dechant, the F_2^S folds are tighter, and more homogeneously developed. Their typical appearance is shown in Figs. 3-32, 33. The fold axes and axial planes cluster more tightly as the deformation becomes more intense (Plate 4, row 11). At first the folds are approximately symmetrical, but further north they become asymmetric with a dominant clockwise rotation-sense looking down the south-east plunging fold axes. Occasionally F_1^S quartz-rodding can be seen refolded about F_2^S folds, but the increasing tightness of F_2^S generates its own quartz rods, which soon become the dominant linear feature in outcrop towards the north.

On the Eisen and Troger Alms (C 5-7) minor folds become difficult to distinguish. M_2^S mineral growth has increased in intensity such that the dominant foliation in outcrop is now axial-planar to F_2^S folds, and in more homogeneous lithologies the folds have effectively disappeared. Occasionally large, metre-scale folds can be seen where quartzitic or quartzo-feldspathic horizons provide enough lithological contrast to be distinguished in the field. Here the folds are similar-type, or much-flattened buckles, with a well-developed axial-planar mineral foliation in the micaceous incompetent matrix, and are indistinguishable in style from F_2^N in the Polinik schists, just as the intermediate schists texturally converge with the Polinik schists towards the north. The orientations of the folds are similar to, and the dominant sense of asymmetry the same as on the south side of the Hohe Leiten ridge. This is sufficient evidence to effect a direct correlation between F_2^S and F_2^N. It follows from comparison of the tectonic sequences that F_1^S and M_2^S may be correlated with F_1^N and M_2^N respectively.

Correlation of F_3^S northwards is hampered by the disappearance of

this deformation north of Kreuzeck. On the Eisen Alm, post M_2 folds, rather kink-like in style (Fig. 3-30) occur in a narrow zone just north of the Eisen Alm shear zone. These are similar in style and orientation to F_3^S (allowing for the change in orientation of structural elements around the Kreuzeck synformal bend). These folds, designated F_3^S, are apparently not related to the formation of the Eisen Alm shear zone. There is no intensification of deformation towards the zone, no progression through tight folds, and lenticular fabrics to intensely sheared material as noted by Bickle (1973) in situations where both shear zones and fold structures in the surrounding terrain are expressions of the same deformational event. F_3^E is thus presumed to predate the shear zone, and is correlated with F_3^S. Correlation with F_3^N is more difficult to establish, as in the map area they differ somewhat in style and orientation although F_3^N can be traced westward and towards the base of the Altkristallin Sheet in the Sadnig Group (Plate 2) where the similarity to F_3^S and F_3^E is much more pronounced. Fig. 3-34 shows the areal distribution of the F_3 deformations in the map area.

D. Structures in the Källukerl zone

The Källukerl zone, the kilometer-wide tract between the Teuchl and Kleinblässen shear zones, contains structural features worthy of mention. Here the fold styles and rock textures are analogous to those considerably further south in the sequence of increasing F_2 intensity and M_2 recrystallisation. The very heterogeneous lithological character of this zone, contrasted with the monotonous nature of the Tröger Alm biotite-feldspathic schists and the Polinik schists, has produced prominently exposed but complex structures.

The most obvious mesoscopic folds, correlated with F_2 on grounds style, orientation and relationship to M_2 mineral growth, are best displayed by the fine biotite-gneiss lithology (Fig. 3-35). They are

characterised by vertical, very much sheared-out limb zones some few metres wide, followed by a hinge zone of a similar width of crumpled mesoscopic folds less than a metre in amplitude, then a further steep, sheared limb zone. The difference in texture between limbs and hinge-zones is clearly displayed by the augen-bearing gneisses (Figs. 3-37 and 3-38). The manner in which the feldspathic augen have been streaked out to many times their original length indicates the amount of strain on the limbs of the folds.

The mica-schists are often almost structureless. Occasionally minor F_2 folds are visible in quartz stringers (Fig. 3-36) and a foliation axial planar to F_2 is developed.

The leucocratic augen gneiss displays a steep-dipping planar foliation in the narrow 'screens' within schists, and near the margins of larger masses, but a more shallowly-dipping, less clearly defined foliation is characteristic of the interior of the body. Rodding and mullion structures, and a crude microfold lineation parallel F_2 fold axes within the body and in cross-section the gneissic foliation is seen to be crumpled, with microcline augen squashed together, as in the augen-bearing fine biotite gneiss (Fig. 3-38). This indicates that the leucocratic augen-gneiss had a gneissic foliation before F_2.

Minor F_2 folds in the steep-dipping limb zones show a much greater scatter of orientations than those in the hinge zones. In the latter, the minor axes cluster strongly about the β of the π-pole girdle (Plate 4 row 8), whereas in the limb zones they tend to scatter around an almost complete girdle in the plane of the F_2 axial surfaces. This probably reflects modification of the orientations of the first-formed minor folds with increasing and somewhat inhomogeneous strain in the limb zones (Ramsay, 1962).

E. Kink-bands and leached joints

Kink bands are frequently developed in lithologies suitable for their formation, e.g. mica-schists with a strongly-developed planar foliation such as the micaceous elements of the Polinik schists, and the Tröger Alm biotite-feldspathic schists.

Often more than one set of kinks can be seen at any one locality but in the Polinik schists it appears that there is one dominant, very widespread set which strikes north-south and has a constant rotation-sense (clockwise, looking south) on the intersection of the kink-plane and the foliation. The bands have a width of up to 5 cm , but are usually 1 to 2 cm wide. They frequently pass laterally into joint surfaces. Occasionally kink-bands of opposite rotation sense appear, with an angular relationship to the dominant set suggesting that together they constitute a conjugate set of kink-bands.

On the Tröger Alm, there appear to be three independent sets of kink-bands. Clockwise- and anti-clockwise sense sets are fairly equally developed, but the spread of orientations is large and conjugate relationships are not demonstrable.

The kink-bands of the Källukerl area strike north-east/south-west and have a steeply-plunging intersection on the well-foliated steep limbs of the F_2 folds, with an anti-clockwise rotation sense. In more competent lithologies (e.g. fine biotite gneiss) these are represented by prominent joints, or narrow zones of en echelon tension gashes.

The kink-bands are almost always strongly leached. Biotites are chloritised within the hinge zones and often up to 2 or 3 cm. into the rock away from the kink zone. The associated joints, into which the kink zones frequently pass in more competent lithologies, show the same leaching and chloritisation features in the surrounding rocks and are often mineralised. Quartz, albite and chlorite are the most commonly found minerals filling these joints, but in the neighbourhood of the

Raggatal iron mines (84) Fe-rich minerals are common on joint-surfaces.

These kink-joint sets can be distinguished from later joint sets, which are not leached or mineralised in this manner.

F. Quartz Fabric Strain

Quartz fabrics have received a certain amount of attention in the geological literature, but mostly in order to study mechanisms of grain nucleation and recrystallisation, e.g. Hobbs (1968), Phillips (1965), rather than as a tool in structural analysis. However, Wilson (1973) has described a sequence of quartz textures produced by a deformation event synchronous with a prograde metamorphism at Mount Isa, Australia. At lower grades of metamorphism (chlorite and biotite zones) the textures observed by Wilson are very similar to those described here.

Observations in the present area suggest, however, that a similar systematic variation may be explained by increasing strain under more-or-less uniform physical conditions (perhaps equivalent to the lower greenschist facies) and that a contemporaneous or subsequent temperature increase is not essential for the production of most of these textures.

In the map area, the M_2 metamorphism has annealed all the pre-existing quartz fabrics, even in the Kreuzeck Schists (see Chapter 4), and the strains observed are due to comparatively late events in the tectonic history. The textures described below are interpreted as a sequence of increasing strain. Unstrained quartz occurs in 0.5 to 1 mm equidimensional grains with approximately straight boundaries against neighbouring quartz grains. The first indications of strain are undulose extinction and wavy grain boundaries. With increasing deformation, discrete strain lamellae are produced, and the extinction angle may fan out more than 20° across the grain (Fig. 3-39). The grain boundaries become convoluted and sutured. Further strain results in a "mortar texture" (Spry, 1969) with fine granulated quartz concentrated

along the old grain boundaries, surrounding relict, strained cores (Fig. 3-40). Eventually, in highly strained rocks, the granulation process is completed and all the quartz is reduced to fine (usually considerably less than 0.1 mm) equidimensional grains. Occasionally, for example, close to a shear zone, these textures may be observed as a small-scale sequence indicating increasing strain over a few metres towards the shear zone.

The amount of strain displayed in the quartz varies somewhat according to lithology. In micaceous lithologies the micas take up much of the strain, and quartz is less intensely deformed than in associated quartz-rich rocks. Nevertheless, a regional variation in quartz fabric strain can be observed, and Fig. 3-41 is a summary of the state of the quartz fabrics in the map area. It is interesting to note that the quartz fabric strain does not correlate well with the intensity of the post-M_2 F_3 deformation either in thin section or regionally, except possibly south of the Nigglai-Gössnitz Törl shear zone (Compare Figs. 3-41 and 3-34). Rocks strongly deformed by F_3 in the Polinik schists are often characterised by a reduction of grain size compared with undeformed rocks, but do not necessarily display a highly strained quartz fabric. This implies that temperatures may well have been high enough during the M_3 retrogressive event to anneal much of the strain produced by the F_3 deformation. In the Sadnig Group, close to the base of the Altkristallin Sheet, quartz is always relatively unstrained while the mica fabric often records high F_3 strains with close minor folding.

In the Polinik area, rocks containing coarse-gained quartz (greater than 1 mm) are largely confined to the south side of the Hohe Leiten ridge. The medium-grained schists to the north differ from the coarse-grained schists only in the grain size, particularly that of quartz, and the greater intensity of F_3 microfolding. It is possible that the

distinction between these lithologies is entirely textural and caused by recrystallisation of the quartz fabric, during or after the F_3 deformation, and increasing in intensity towards the edge of the Tauern Window.

The three zones of maximum strain on Fig. 3-41 (neglecting the northernmost zone between the Möll"tal Line and Launsberg shear zone) coincide approximately with, first, the axial zone of the F_4 Polinik antiform; second, the neighbourhood of the strongly retrogressed "green band" of Eisenriegel; and third, the axial zone of the Kreuzeck synformal bend, which may be another F_4 structure. It is therefore probable that the pattern of quartz fabric strain is a product of deformations later than F_3, and is possibly the small-scale manifestation of the large-scale flexuring of F_4.

G. Shear zones

The shear zones, whose petrography and field relations have already been described in the previous chapter, divide up the Altkristallin sheet in the Kreuzeck Group into a number of blocks (see Plate 2 and Fig. 3-45). Marked discordances in the structural and metamorphic sequence across the zones indicate that they occurred late in the tectonic history. Their present attitude is almost vertical or steeply south-dipping. The effect of rotation of the Altkristallin due to uplift of the Tauern is difficult to assess, but any which predate this uplift may have been rotated from shallower south-dipping structures.

The existence of at least two different lithological types of shear zones, which might represent two (or more) generations of shear zone movement, was suggested in the previous chapter.

The first type is essentially cataclastic, characterised in thin section by broken crystal and rock fragments in an almost isotropic, dark matrix.

The second type is mylonitic, consisting in the most strongly sheared parts of the zone of greyish, often finely-banded mylonites, which contain finely recrystallised quartz, white mica and chlorite.

The Launsberg and Teuchl shear zones belong to the first type. The Kleinblässen shear zone, together with its small SSE-trending branch which follows the Kleinblässen stream down to the Teuchlbach, is of the second type.

The shear zones of the Eisen Alm and upper Teuchl valley appear to converge, and constitute one major shear zone (the Nigglai shear zone) which passes eastwards down the Nigglaital (described by Wright 1973). It is possible that this set of shear zones has had a complex history, since the sheared rocks cannot always be consigned unambiguously to either of the above categories. The southernmost shear zone, passing through the Gnoppnitz Törl and crossing the next ridge to the west some 500 m south of Hochtristen (Plate 2), was not studied in this work, but has been previously described (Exner 1957, Wright 1973). It would appear to be of the second type.

1. Control of movement sense

There are very few minor structures associated with the shear zones. The planar black-coated movement surfaces very rarely show any lineation or folding. Usually, proximity to a shear zone is indicated only by the increasing abundance of blackened shear surfaces cutting the rock.

A study was made on the north-east margin of the Kleinblässen shear zone. Here the rocks adjacent to the shear zone have a fabric of lenticular masses some centimetres thick enclosed by intersecting shear surfaces which often display a fine striation, which may be an a-lineation parallel to the direction of movement on the surface. The lenticular bodies occasionally enclose fairly tight asymmetric folds.

The sketch (Fig. 3-42) shows a possible dynamic relationship between the movement direction of the shear zone, lineations on sheared surfaces, and asymmetric folding. The relative movement direction would lie in the girdle of lineation directions, and the sense of movement would be consistent with the rotation sense of the asymmetric folds. Fold axis directions with rotation sense, and striation lineation directions were measured, but the results proved inconclusive (Fig. 3-43). Lineations did not define a recognisable girdle, and folds did not show a consistent rotation sense. Comparison of the folds with minor F_2 folds further outside the shear zone led to the conclusion that most of the folds probably predated the shearing movements.

In the Kleinblässen stream, near its junction with the Teuchlbach (F5) the branch of the Kleinblässen shear zone cuts the surrounding biotite-feldspathic schists at a fairly high angle (unlike most of the other shear zones, which are seen to be cross-cutting only on a much larger scale). Here the foliation is seen to bend into the shear zone from each side, in a manner which suggests that the relative movement was of the west wall upwards at 45^{o} towards the south (Fig. 3-44). Asymmetric folding of the mylonitic banding, and here and there a possible a-lineation on shear-surfaces support this observation.

The best indications of movement sense arise from a comparison of the pre-existing structures and the state of metamorphic recrystallisation on either side of a shear zone. There is rapid variation in structural style and recrystallisation across the strike through much of the map area, and it can be assumed that this is a vertical variation within the Altkristallin Sheet, which uplift of the Hohe Tauern area has made visible as a north-south sequence at this level of erosion. Thus a south side downthrow is suggested for the Launsberg shear zone, since strong M_3 retrogressive effects (see Chapter 4) are restricted to the northern side, adjacent to the edge of the Tauern Window. On

these criteria, the Teuchl shear zone clearly throws down on the south side, and the Kleinblässen shear zone down on its north side. The Eisen Alm, Bärenkopf and Gössnitz Torl shear zones of the upper Teuchl valley probably all throw down on the south side, since they all create gaps rather than repetitions in the otherwise smooth sequence of structural and metamorphic variation in this area. In any case, their eastward continuation, the Nigglai shear zone, shows evidence of a very substantial south side downthrow , and causes the major geochronological break in the eastern Kreuzeck Group (Brewer 1970).

The amount of strike-slip movement on the shear zones cannot be assessed on these criteria, since the zones strike approximately parallel to the rocks and their "isograds" of metamorphic and structural grade.

Fig. 3-45 shows the distribution, nature and probable movement senses of the Kreuzeck Group shear zones.

2. Control on the age of movements

Control on the time of movements in the shear zones is difficult to obtain, particularly since it appears that at least two separate generations of shear zone movement are involved. The relationships to other structural units are often obscure .

It would seem likely that the first type were formed at shallower depths in cooler rocks, and are therefore likely to be substantially younger than the mylonitic set.

A sample from the mylonitic Kleinblässen shear zone yielded a whole-rock K/Ar age of 61 m.y. This single age determination might be subject to geological errors such as incomplete outgassing of the source rock or excess Ar^{40} pressures, and in the absence of further determinations cannot be regarded as good evidence for the age of formation of the mylonite. Kink bands cut the mylonitic Kleinblässen shear zone but have not been recorded in any of the cataclastic shears.

The Bärenkopf shear zone is cut by a dyke 1 km east of Scheuchenkopf (B8). The dyke is not continuous through the zone, but forms a series of discrete but undeformed lenses, which cut sheared surfaces within the zone. The dyke thus post-dates most, if not all, the movement associated with this zone. Evidence for the absolute age of the Kreuzeck Group dykes, if indeed they represent only one intrusive episode, is somewhat ambiguous (see Chapter 6). This particular dyke has not been dated.

H. Synthesis of the major deformational event F_2

Due to the absence of suitable marker horizons it has not been possible to map macroscopic structures on the ground. However, the variation in tightness, orientation and asymmetry of mesoscopic F_2 folds allows a synthesis to be made across the map area, with interesting results and implications.

Where marker bands exist it is useful to study the local relationships between them and the mesoscopic structures:

Example 1. North flank of Dechant (D9)

The marker is a 1½ to 2 metre thick white quartzite band, conformable with the foliation of the enclosing mica-schists. On the scale of the map the quartzite band is essentially straight, i.e. shows no megascopic folding. On the outcrop scale (Fig. 3-46) it is seen to be folded gently in approximately symmetrical waves, with a wavelength of 2 to 5 metres. The folds in the mica-schists above and below the quartzite are smaller-scale (5 cm. wavelength) but follow the outline of the larger folds in the marker band. The larger-scale foliation envelope (strong dashed line in the figure) parallels the outcrop of the marker.

Example 2. Strieden Kopf marble band (B8)

In the neighbourhood, F_2 folds are tight, strongly asymmetric, and less abundant than in the previous example. The marble band is occasionally seen to be tightly interfolded with the schists (Fig. 3-47). The lithological banding on the long limbs is rotated almost parallel to the axial surfaces of the folds. On the map scale, the strike of the marker, and thus the foliation envelope of the minor folds, is indistinguishable from the regional axial-planar orientation of the F_2 folds.

The transition between these two styles (similar to that observed by L.E. Weiss, in Turner and Weiss, 1963, p.167) takes place in the upper Teuchl valley where the only marker is the rather ill-defined boundary between the quartz-muscovite-rich Kreuzeck schists and the Schulter biotite-feldspathic schists. Fig. 3-48 is a sketch map of this area showing F_2 minor folds from large schist outcrops, schematically projected onto a horizontal surface to show their asymmetry, style and trend of foliation envelope. From this it is apparent that they outline a large antiformal structure, plunging to the south-east, which has been termed the Schulter antiform.

To the south of this structure, the F_2 folds become markedly less intense and exhibit a much broader spread of axial directions. To the north, as far as the Kleinblässen shear zone, the F_2 minor folds are tight and asymmetric with a remarkably consistent clockwise rotation-sense on their south-east plunging axes. Local departures from this rotation sense, in narrow zones, can usually be seen in the field to belong to the short limbs of larger clockwise-sense structures on the scale of a few metres.

In the Kallukerl area, no dominant sense of asymmetry is apparent, and strongly sheared limbs alternate with crumpled symmetrical zones, suggesting larger fold structures on a 20-30 metre scale, which are,

however, not readily identifiable in the field. It is possible that the Kallukerl area represents the core of a kilometer-scale structure similar to the Schulter antiform.

North of the Teuchl shear zone, and on the south side of the Hohe Leiten ridge (D-E3), F_2 folds are dominantly clockwise-asymmetric (looking east) once more, though local exceptions are more frequent than on the Eisen and Tröger Alms. However, on the northern side of Polinik equal numbers of either rotation sense can be found, and there is no regular distribution consistent with the observed folds being parasitic on the limbs of larger structures. F_2 folds of 0.5 to 5 metre scale occur in cascades stacked-up to 50 metres or more high, and these zones are interspersed by regions where F_2 folds are infrequent. It is not uncommon to find folds of both rotation senses in the same outcrop (see Oxburgh, 1966).

Collation of this evidence has allowed a tentative synthesis of major F_2 structure through the whole map area. Fig. 3-49 shows F_2 structures in the lower parts of the Altkristallin sheet restored to their possible orientation at the time of their formation, in the form of a NNE-SSW section through the map area (The regional foliation south of Kreuzeck is assumed to be more-or-less unaffected and unrotated by deformational events from F_2 onwards, and is taken as approximately horizontal). The effects of deformations later than F_2, such as the Polinik antiform, the Kreuzeck synformal bend, and the shear zones, have been removed.

1. Relationship between F_2 and M_2 - the axial-planar foliation

It is important to the understanding of the tectonic history of the Altkristallin Sheet to establish the relationship between the F_2 deformation and M_2, the most prominent episode of mineral growth in the map area.

All the M_2 mineral growth appears to be post-kinematic to F_2. Garnets in the Polinik schists occasionally show inclusion trails discordant to the external foliation, but the trails are straight, make a sharp angle with the external foliation at the margin, and most probably record an earlier fabric element. The F_2 folds, however, display a very prominent axial planar foliation, defined by the orientation of unstrained micas, in more micaceous lithologies throughout the map area north of the Eisen Alm shear zone, and a strong mineral lineation parallel to F_2 fold axes is observed in amphibolites.

There are two likely explanations for the widespread development of this M_2 foliation:-

(i) The M_2 foliation-defining minerals (muscovite, biotite, etc.) are mimetic on an earlier axial-planar foliation.

(ii) The M_2 minerals began to grow while the stress field responsible for the F_2 folding was still in existence, though later growth of the minerals took place post-kinematically.

The most significant evidence in favour of the second explanation is the spatial correlation between the intensity and increasing ductile nature of the F_2 deformation, and the grade of M_2 metamorphism. The tightest, and most nearly similar-type folds are found in the Tröger Alm sillimanite zone and in the Polinik schists, and parallel-sided buckled layers in the biotite zone Kreuzeck Schists, while styles of intermediate character occur in between.

It may be adopted, as a working hypothesis, that the F_2 deformation just predates the M_2 metamorphic episode, and is part of the same

tectono-metamorphic event. While the linking of F_2 and M_2 cannot yet be considered conclusively proved, the hypothesis has interesting implications (see Chapter 7).

J. Conclusions

The sequence of deformational and crystallisation events established by correlation through the map area may be summarised as shown in Table 3-4.

The pre-F_1 history of the Altkristallin in the map area is obscure, but probably involved at least one major folding event.

F_1 and M_1 together constitute a major thermal and tectonic event in this part of the Altkristallin sheet. The F_1 deformation was characterised by high finite strain and the disruption of lithological units (unless this had already been accomplished prior to F_1). M_1 mineral assemblages grew syntectonically to F_1, and parallel-felted mica-fabrics and rotated porphyroblasts are typical. However, over most of the map area evidence for the nature of the F_1/M_1 event is obliterated by later features, and large-scale F_1 structures can no longer be recognised.

The style and intensity of the F_2 deformation vary significantly across the map area, corresponding to a variation with depth in the Altkristallin sheet, and suggesting a temperature gradient across the unit at the time of deformation. The upper part of the sheet was little deformed by F_2, but the amount of strain involved in this deformation increases greatly downwards in the sheet, and substantial rotation and transposition of pre-existing structures has clearly occurred at deeper levels. Towards the north, deeper in the Altkristallin, the more concentric style of folding of the Kreuzeck schists is replaced by similar and isoclinal styles. In the map area the internal structure of the Altkristallin after F_2 folding has been tentatively

TABLE 3-3. Summary of the tectonic history of the map area.

	DEFORMATION		MINERAL GROWTH, etc.
F_0	Early deformation(s)		
F_1	Tight mesoscopic folds (KS) Refolded isoclines (PS)	M_1	KS assemblages - gt-mu-schists ?eclogites in PS
F_2	Buckles, passing northwards into tight and flattened folds.	M_2	increasing in grade northwards, mineral fabric axial-planar to F_2 folds.
F_3	Open buckles, chevron folds, microripples		
-	? Early shear zone movements		
		M_3	Patchy retrogression, greenschist facies recrystallisation of basal 1.5 Km of Altkristallin.
		-	? intrusion of dykes
F_4	Large scale flexures		
-	Kinks and leached joints		
-	Later shear zone movements		
-	Other joints (not leached)		

Time ↓

KS = Kreuzeck schists, PS = Polinik schists.

reconstructed (Fig. 3-49). The parallel increase in metamorphic grade of the M_2 recrystallisation suggests a possible link between F_2 and M_2. Micas of the M_2 generation define a foliation axial-planar to F_2 folds from the Eisen Alm northwards, but most mineral growth appears to be post-kinematic to F_2. This metamorphic episode is treated in detail in the following chapters.

The sequence of events from F_3 onwards is less clear, although some order of elements is suggested by field relations. For example: dyke intrusion post-dates at least some shear zone movement; F_4 mineral strain bends M_3 chlorites, but not joint-filling minerals; kink-band and leached joint orientations are discontinuous across the cataclastic Teuchl shear zone. The following sequences can be established, but the evidence required to set up a complete sequence is not available.

(i)	(ii)	(iii)
F_3^E	early shears	F_3^N
early shears	kinks/leached joints	M_3
dykes	late shears	F_4
		kinks/leached joints
		late shears

The F_3 deformations, although geographically separated, are tentatively correlated with each other on grounds of orientation, style (when in comparable lithology), and relationship to other events in the sequence. The total strain involved in F_3 deformations in this part of the Altkristallin Sheet is small, except possibly close to the base of the sheet exposed in the Sadnig Group.

At some stage after M_2, but otherwise of indeterminate relative age, a first set of shear zones (mylonitic set) appeared in the Altkristallin sheet, dividing it into separate blocks. At least one

of these early shears had a reverse-fault sense of movement.

F_4 deformation produced very few recognisable minor structures apart from microscopic mineral strain, but was responsible for large-scale changes in orientation of the regional foliation in the Polinik area (C3) (the Polinik antiform and associated structures) and possibly also in the Kreuzeck Schists in the neighbourhood of Kreuzeck and Dechant (C10). These structures may have been related to movements at depth in units below the Altkristallin Sheet and F_4 might reflect the rotation of the central part of the map area to its present steep-dipping attitude during the uplift of the Hohe Tauern area, the F_4 major structures being now displayed at either end of the steep-dipping zone.

The last major structural event to affect the map area is the formation of the late cataclastic shear zones, which are apparently large normal faults, and are possibly connected with movements along the Mölltal line. It is likely that these represent later stages of differential uplift of the northern parts of the map area close to the area of maximum Alpine uplift in the Central Hohe Tauern.

Consideration of the absolute ages of the tectonic events in the map area is deferred until the geochronological evidence has been considered (Chapter 6) and discussion of the tectonic sequence in relation to the history of the Eastern Alps will be found in Chapter 7.

CHAPTER 4

Metamorphism I - Textural evidence

Introduction

The tectonic history of the map area has been established in the previous chapter. The relationship of deformational events to episodes of mineral growth makes it apparent that there are two distinct major metamorphic episodes in the map area, separated in time by the deformational event F_2. The metamorphic recrystallisation of the second, post-F_2 episode M_2, increases in intensity northwards, gradually obliterating minerals and textures characteristic of the metamorphic episode M_1.

A. Textural characteristics of the metamorphic episode M_1

Though the mineral assemblages are similar to those developed in M_2, there are textural criteria by which minerals belonging to M_1 may be distinguished from those grown during M_2. The muscovite- and quartz-rich Kreuzeck Schists of the southernmost part of the map area are the least affected by M_2, and best display the textures characteristic of M_1. Many of these textures have already been described in Chapter 2.

M_1 minerals frequently show evidence of syntectonic growth during F_1. The large garnets often show spiral inclusion trails of quartz, opaque ores, graphite and tourmaline (Fig. 4-3). The parallel-felted muscovite fabric, though infrequently described from other metamorphic belts, may indicate syntectonic growth of micas during deformation (see Figs. 4-3 to 4-9).

Quartz in the Kreuzeck Schists has usually been annealed after the F_2 microfolding, and plagioclase and biotite most commonly occur

as porphyroblasts which grow over the folded mica parallel-felts. These are thus of M_2 rather than M_1 generation.

Occasionally rocks can be found which preserve the earlier assemblage almost intact. One such is sample 0342 (Fig. 4-1 and analysis in Table 5-2) from just east of Hochtristen (Plate 2), in which the biotite occurs in oriented narrow laths within parallel-felts of muscovite and biotite, and the plagioclase occurs with quartz in a more-or-less equigranular mosaic between the felts. The rock contains accessory staurolite and kyanite in small rounded grains, and the total assemblage is qz-plag-mu -bi - gt-st - ky-chl-ore. The chlorite is most probably secondary.

The eclogitic assemblages occurring in amphibolites within the Polinik Schists are probably relict M_1 assemblages (see later discussion).

The distribution map of M_1 index minerals (Fig. 4-2) indicates that the grade of this metamorphism does not vary much in the central Kreuzeck Group, though the absence of staurolite, and more chloritic nature of the garnet phyllites (Kreuzeck Schists) of the H. Gerbers Hütte area may indicate that the metamorphic grade falls off to the west. Elsewhere the widespread though uncommon occurrence of staurolite and occasionally of kyanite implies middle amphibolite facies conditions in an intermediate to high pressure (Barrovian) facies series. Platy ilmenite is the characteristic ore mineral, and the abundant graphite indicates low oxygen fugacity for this event.

The absence of substantial variation in metamorphic grade suggests either that the M_1 isogradic surfaces were gently inclined to the present erosion level, or possibly that thermal gradients were low during this metamorphism. The abundance of garnet and muscovite, as compared to M_2 assemblages in rocks of equivalent composition, and the absence of any low-pressure metamorphic minerals, may indicate rather high pressures.

B M_2 Progressive Metamorphism

1. Zonation criteria

The Kreuzeck schists frequently contain minerals which cut the prominent M_1 mineral fabric, and whose growth post-dates the F_2 deformation. As the rocks are traced northwards, the M_1 mineral fabric is gradually obliterated and replaced by assemblages and textures characteristic of M_2 as developed in the northern part of the map area. Different stages of this replacement can be recognised as a progressive sequence through the Kreuzeck and Intermediate Schists.

The three most sensitive indices of the extent of the M_2 overprint are:

i. the amount of recrystallisation of the M_1 felted mica fabric

ii. textural phenomena involving garnets

iii. textural phenomena involving staurolite

A crude zonation can be established for progressive M_2 metamorphism on the basis of these criteria. Many of the recrystallisation phenomena are patchily and irregularly developed, a fact which perhaps reflects the distribution and availability of fluids.

2. Recrystallisation of the mica fabric

Early stages of mica-felt recrystallisation in the Kreuzeck Schist are characterised by the growth of fairly large (1-2 mm) porphyroblasts of biotite, and sometimes of chlorite and muscovite. These minerals cut the parallel felts (Fig. 4-4) and grow over F_2 microfolds. Mg-rich chlorites (optically positive, grey to brownish birefringence, (Albee, 1962)), occur as single cross-cutting laths, or as knots of laths which are occasionally seen to be pseudomorphing staurolite, and are probably coeval with the biotite growth. More Fe-rich chlorite porphyroblasts (anomalous blue birefringence), often containing bundles

of radiating rutile needles, appear to be of secondary formation, replacing M_2 biotite (Fig. 4-5). Plagioclase porphyroblasts also grow at the expense of the mica felts, (Fig. 4-6).

The muscovite of the parallel felts initially recrystallises to small, stubby, randomly oriented laths (Fig. 4-7, 4-8). The amount of recrystallisation is patchily distributed in the rock, and often proceeds selectively along certain layers in the felts. Recrystallisation is often accompanied by the disappearance of a great deal of the graphite occurring within the felts, but the fact that some graphite almost always remains in the recrystallised felts suggests that f_{O_2} in the fluid phase may still have been buffered at a fairly low level.

The grain size of the recrystallised material is variable, from 10 - 20μ up to 0.5 mm, but there is a tendency for the grain size to increase with increasing M_2 metamorphic grade, and for the new grains to become oriented in the axial surfaces of F_2 folds. In the Intermediate Schists it can be hard to evaluate the amount of recrystallisation when the grain size is coarse and there are few F_2 fold hinges to allow differentation between narrow, bent M_1 and broader, undeformed M_2 muscovite laths. It is also possible that mimetic recrystallisation of new laths following the trend of the older felts has occurred in some rocks.

Fragments of parallel-felted mica persist up into quite high-grade M_2 rocks. Fig. 4-9 shows one such fragment from sample 0125 from the Intermediate Schists of Kallukerl (D4), where the felts form isolated patches some 3 mm across, comprising 4% of the total micas.

The zonation of mica-fabric recrystallisation (Fig. 4-10) has been plotted by studying thin sections and estimating by eye the area of recrystallised material as a percentage of the total mica-fabric area. This is not a very precise measure, because a thin section is too small

an area to be representative of the state of recrystallisation in larger volumes of rock, particularly where recrystallisation is of a patchy nature. However, a regional trend of northwards-increasing intensity of recrystallisation does emerge from the distribution.

3. Garnet textures

Textural phenomena involving garnets in the map area can be subdivided into five types, whose systematic regional distribution can be attributed to increasing temperatures towards the north during M_2 metamorphism.

(i) Textural type 1 consists of unaltered M_1 generation garnets, usually large (up to 1 cm), very much shattered, and with an internal, often spiral, fabric of ore, quartz and tourmaline inclusions.

(ii) Type 2. The large shattered garnets break down to an Fe-rich, blue birefringent chlorite around their rims (Fig. 4-11) and along fractures (Fig. 4-1), and locally this replacement proceeds to completion. There is some doubt about the age of chloritisation in the Kreuzeck schists, and it is possible that some of this chloritisation is an effect of events later than M_2.

(iii) Type 3. The breakdown products include biotite and chlorite, which are usually coarse-grained laths surrounding and penetrating garnet, distinct from the generally fine-grained Type 2 alteration which proceeds along narrow cracks. Exceptions to this are the two recorded instances within the Kreuzeck Schist (samples K6 and K7), where the fine-grained musc + bi + chl pseudomorphs may be later than M_2, since M_2 biotite porphyroblasts are also somewhat affected by alteration.

Breakdown assemblages include bi + chl + mu and bi + chl + plag as well are purely bi + chl. In some cases it is not certain whether the biotite and chlorite are coeval or whether the chlorite has formed from biotite at a later stage.

(iv) Type 4. Garnet breaks down to biotite (Fig. 4-12), or biotite + muscovite (Fig. 4-13). The products of this breakdown are coarse flakes up to 1.5 mm across, which surround and invade the garnets. Invasion by muscovite and biotite often leads to a garnet with a "lace-doily" texture with mica-filled cavities. Biotites occupying the cores of old garnet grains often have abundant pleochroic haloes around tiny radioactive indusions, probably inherited from the garnet. Sometimes the garnets are completely pseudomorphed.

(v) Type 5. Growth of new garnet. Newly formed M_2 garnets are not often seen in the Intermediate Schists, probably because the M_1 assemblage usually contained more garnet than would be stable under M_2 conditions. Where they do occur, they are texturally distinctive, small, inclusion-free, unbroken and often euhedral grains about 0.5 mm in diameter. In the Intermediate Schists of Källukerl (C4 etc.) these new grains can be found in the same thin section as M_1 relics rimmed with broad biotite flakes. Another texture evident under higher grades of M_2 metamorphism on the northern part of the Eisen Alm (C6), is the annealing of the relict garnet cores within the biotite or biotite-muscovite partial pseudomorphs. The cracks heal and the garnets tend to assume more unbroken outlines and compact shapes. This presumably reflects the final adjustment of the garnet towards equilibrium with the M_2 meta-

morphic assemblage.

Fig. 4-14 shows the regional distribution of the textures involving garnet. A garnet isograd cannot be constructed for the M_2 metamorphism since the dominant M_2 feature is the breakdown of pre-existing garnet, although the rocks of Kallukerl could be said to belong to an M_2 "garnet zone", since they display new garnet growth without the development of staurolite.

4. Staurolite textures

Staurolite is not a particularly common mineral in M_1 assemblages. It reacts to incipient M_2 metamorphism by altering to sericite or chlorite at the margins or along fractures. Towards the north, the extent of the alteration increases until the staurolite is completely pseudomorphed, generally by white mica. Large pseudomorphs are common in the Kallukerl Intermediate Schists, and are also found along strike from Strieden Kopf (B7) and near the Penker Gemeinde Alm (D8).

Staurolite of the M_2 generation is found as clear, euhedral porphyroblasts which grow across F_2 folds in the mica fabric. They virtually always grow in muscovite-rich parts of the rock, and sometimes within muscovite-rich pseudomorphs probably after earlier staurolite (Fig. 4-15). On occasions, both generations, shattered and sericitised remnants of M_1 and clear euhedral M_2 staurolite, are seen in the same thin section, and thus the regional distribution of M_1 and M_2 staurolite may overlap. In some places M_1 staurolite is completely pseudomorphed before the appearance of the new generation.

Fig. 4-18 shows the distribution of M_1 and M_2 staurolite. A staurolite isograd has been constructed where new staurolite growth becomes abundant in rocks of the appropriate composition. This is an isograd in the sense of Tilley (1924), marking the first appearance

of staurolite rather than one based on a specific reaction.

5. Textural evidence for the mechanism of a staurolite-producing reaction

The figures 4-14 and 4-18 show the coincidence of the zones of M_2 staurolite growth and zone 4 of the garnet textural phenomena, where garnets break down to biotite or biotite + muscovite. Reactions can be written to describe these local textural phenomena in the manner of Carmichael (1969). Each local reaction takes place in a small system of a very few millimetres scale. The small system may be considered closed to components such as aluminium, which is believed to be the least mobile element in rock systems, but open to more mobile species such as H_2O, H^+ (possibly as H_3O^+), alkali cations, Fe^{2+} etc., which can therefore move from one local system to another on a centimetre or greater scale.

Consider the way in which staurolite first appears in muscovite-rich areas of the rock (Fig. 4-15). The constant-aluminium reaction

$$\underset{\text{muscovite}}{4\ K\ Al_3Si_3O_{10}\ (OH)_2} + 3(Mg,Fe)^{++} = \underset{\text{staurolite}}{3(Mg,Fe)Al_4Si_2O_{10}\ (OH)_2} + \underset{\text{quartz}}{6\ SiO_2} + 4\ K^+ + 2\ H^+$$

describes this texture, and involves the exchange of $(Fe,Mg)^{++}$, K^+, H^+ and possibly SiO_2 with the larger system. The simplified staurolite formula used by Carmichael has been adopted here for convenience, to keep the reaction coefficients manageably small.

To represent the breakdown of garnet to muscovite and biotite, the following reaction can be written

$$2(Fe,Mg)_3Al_2Si_3O_{12} + 2K^+ + 4H^+ = K(Mg,Fe)_3AlSi_3O_{10}(OH)_2 + KAl_3Si_3O_{10}(OH)_2 + 3(Fe,Mg)^{++}$$

garnet — biotite — muscovite

while garnet being replaced by biotite alone can be represented by

$$(Fe,Mg)_3Al_2Si_3O_{12} + 3SiO_2 + 2K^+ + 3(Fe,Mg)^{++} + 6H_2O = 2K(Fe,Mg)_3AlSi_3O_{10}(OH)_2 + 8H^+$$

garnet — quartz — biotite

These two reactions may be regarded as occurring simultaneously in garnets which are being replaced by biotite and a smaller amount of muscovite. Very little quartz occurs in the neighbourhood of garnets replaced in this way, confirming the likelihood of this mechanism. In practice, the biotite produced in these reactions will be more aluminous than the annite/phlogopite represented here, which will result in the production of less muscovite by the first reaction, and the consumption of less quartz by the second.

These three local reactions can now be combined into a closed system as shown in Fig. 4-16, with the mobile components exchanged between the local systems as indicated. The net reaction is

$$7 \text{ garnet} + 5 \text{ muscovite} + 6H_2O = 6 \text{ Staurolite} + 5 \text{ biotite} + 9 \text{ Quartz}$$

It should be noted that the prograde reaction consumes H_2O, and thus the formation of staurolite will be strongly dependant on the availability of water-bearing fluids.

These phase relationships can also be considered on a Thompson AFM projection (Fig. 4-17). The breakdown of a more Mg-rich M_1 garnet (by analogy with the >25 mol.% pyrope type 3 garnets in the Polinik Schists, Chapter 5) would yield staurolite + biotite + a smaller amount of more almandine-rich garnet. Increased solid solution of pyrope in garnet at higher pressures (Yoder and Chinner,

1960) would restrict the range of rock compositions in which staurolite and kyanite occur in the M_1 assemblages, accounting for their less frequent occurrence in the Kreuzeck Schists.

It is probably necessary to postulate some exchange of Mg^{2+} and Fe^{2+} between the partly pseudomorphed garnet and phases elsewhere in the rock, since it is unlikely that the reacting garnet and its biotite product have the same Mg/Fe ratios.

This reaction cannot be satisfactorily used to define a staurolite-in isograd because of the problem of disequilibrium, the bivariance of the reaction when Mg/Fe ratios are allowed to vary, and the strong H_2O-buffering capacity of the reaction (see Chapter 5).

6. Mobility of ions

While Carmichael's approach has been successfully applied to reactions in the staurolite zone, at higher grades it becomes more difficult to regard microscopic systems as closed to even relatively immobile components. In the Tröger Alm area and in the Polinik Schists there is ample evidence of the mobility of aluminium.

On Polinik, post F_2 veins contain Al-rich minerals such as kyanite and muscovite. On the Tröger Alm, the fairly abundant pegmatites are aluminous, containing muscovite, tourmaline, sillimanite and in one spectacular case, abundant andalusite, which indicates that large quantities of aluminium were moving about in solution. Pegmatitic veinlets penetrate the rocks on a small scale too, and it is very probable that Al was free to move through the rock.

At the lower grades prevailing in the southern Eisen Alm and upper Teuchl valley, disequilibrium textures indicate limited mobility of components: relict folded parallel-felts can be found in the same thin section as M_2 staurolite; the new assemblages can show a very variable development over the scale of a single outcrop; and comp-

ositional variation of phases within the probed assemblage 0462 (see Chapter 5) from the staurolite zone is much greater than in the Tröger Alm and Polinik assemblages.

7. Al-silicates associated with M_2: the andalusite-sillimanite isograd

(a) Andalusite: Andalusite is found in the Intermediate Schists. Although not common, its occurrences can be spectacular. Sample 0462 from 550 m ENE of Strieden Kopf (B7) contains large porphyroblasts 2 cm or more in diameter, which enclose all other phases including euhedral garnet and staurolite. In sample 0452, 400 m SW of Schneestellkopf (B6), the andalusite is not evident in hand specimen, but in section is seen to constitute some 25% of the mode in very spongy, optically continuous porphyroblasts probably up to 5 cm across, dusty with graphitic trails. Sample 0222, from the Wöllatal road 1.5 km ENE of Stall (Plate 2), along strike from the aforementioned localities, contains common 1 - 2 mm prismatic andalusites of square cross-section (Fig. 4-19). In a synmetamorphic younger pegmatite 400 m NW of Eisenriegel (C6) prisms of crimson-pink andalusite up to 15 cm long are associated with quartz, feldspar, muscovite and biotite.

(b) Sillimanite: In one part of the section from sample 0452 a mat of fibrous sillimanite grows in association with a knot of biotite laths set in quartz (Fig. 4-20). To the north and northeast of this point, largely within the Tröger Alm biotite-feldspathic schists, sillimanite becomes the characteristic Al-silicate of the mica-schists, and is commonly found as matted fibrolite along with biotite and quartz. A continuous range of grain size, from almost submicroscopic sillimanite fibres up to prisms .05 mm diameter, and occasionally greater, can be observed, very often within the same thin section.

Several authors express doubt about the stability field of the fibrolitic variety of sillimanite compared to that of sillimanite as determined by experiments. Holdaway (1971) is of the opinion that fibrolite is a metastable mineral with an excess molar free energy of some 150 cal/mole which forms as a result of significant overstepping of the andalusite-sillimanite boundary. From a study of natural fibrolite, Cameron and Ashworth (1972) concluded that it does not differ from prismatic sillimanite either in ordering scheme or composition, and that significant bivariance of the andalusite-sillimanite boundary due to minor element partition seems to be ruled out (see also Albee and Chodos, 1969). They further suggested that the excess free energy of fibrolite proposed by Holdaway, if it exists, is more likely to be caused by the higher surface energy of the fibrous habit. Bell and Nord (1974) investigated partly fibrolitic sillimanite from Brandywine Springs by microscopic and electron diffraction techniques and concluded that the fibrolite consisted purely of narrow sillimanite crystals intergrown by simultaneous crystallisation with a small amount of quartz.

On balance there is no reason to suppose that the largely fibrolitic sillimanite of the map area did not form in the stability field of sillimanite, and since the grain size and hence surface free energy of the sillimanite is variable in almost every sample from the sillimanite zone, the amount of overstepping of the andalusite-sillimanite boundary, if any, is not thought to be significant.

In a few places in the zone, later retrogressive effects, probably connected with the M_3 event, have resulted in the replacement of the sillimanite by very fine-grained white micaceous material, frequently preserving the fibrous habit of the former.

The more general and widespread occurrence of Al-silicate in the biotite-feldspathic schists may be a function of two factors; either

the more Mg-rich biotite feldspathic schist compositions tend to fall on Al-silicate-biotite tie lines (see Thompson AFM projections, Fig. 4-17) or with the disappearance of staurolite, a wider range of rock compositions develop Al-silicate than within the staurolite zone.

On Fig. 4-18, an andalusite-sillimanite isograd has been constructed to separate occurrences of the two polymorphs, and to contain all the sillimanite occurrences. Because of the low free-energy change of the andalusite-sillimanite transition, the kinetics of the isograd reaction are likely to be sluggish, and the coexistence of both minerals close to the isograd (as in sample 0452) is to be expected.

(c) Kyanite: one rock from near the sillimanite isograd contains kyanite, rimming staurolite and undoubtedly of M_2 generation, in addition to sillimanite associated with biotite. This possibly indicates M_2 conditions very close to the triple point.

The characteristic Al-silicate of the Polinik Schists is kyanite implying that somewhat different conditions prevailed during M_2 in this northern region.

8. The garnet-Al silicate-biotite isograd

An isograd corresponding to the breakdown of staurolite in muscovite- and quartz-bearing assemblages, and based on the reaction

$$\text{Staurolite} + \text{muscovite} + \text{quartz} = \text{garnet} + \text{Al-silicate} + \text{biotite} + H_2O$$

was suggested by Carmichael (1969, 1970) and termed by him the Al-silicate-garnet-biotite isograd. This reaction is univariant in the system K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O but may be rendered divariant by the presence of other components or if the vapour composition is

buffered by the assemblage. Fig. 4-17 shows the change in topology of the phase relationships on a Thompson AFM projection.

On Fig. 4-21 reactant and product assemblages are distinguished by open and closed circles, respectively, and rocks containing all six phases are represented by crosses. For the Eisen and Tröger Alm section an isograd can be drawn which is roughly parallel to the andalusite-sillimanite isograd, and approximately follows the strike of the lithological boundary between the Intermediate and Biotite-feldspathic schists. Although the biotite-feldspathic schists may have higher Mg/Fe ratios than the Intermediate Schists, and to be somewhat less aluminous, this should not affect the position of the isograd. Al-silicate-garnet-biotite assemblages are not restricted to the biotite-feldspathic schists. In Sample 0452, a staurolite-free Intermediate Schist, these three minerals make up more than 35% of the mode. Above the isograd, whenever sillimanite-bearing biotite-feldspathic schists are Fe-rich enough to contain appreciable garnet, staurolite is absent. It can be concluded that the isograd does represent the upper limit of staurolite stability, and is not directly controlled by the change of lithology.

No isograd can be drawn for the Polinik Schists. The possible reasons for this are discussed in Chapter 5.

9. Pegmatites associated with M_2

There is strong evidence to link the younger pegmatites of the map area (see Chapter 2) with the M_2 event.

They are usually discordant, and not deformed by F_2. They occur only in the higher-grade zones of M_2 crystallisation, and the distribution of Al-silicate polymorphs in the pegmatites is the same as that in the country rocks (i.e. sillimanite-bearing pegmatites occur in the sillimanite zone, etc.). In sample T6 (Fig. 4-22)

individual muscovite laths in the wall-rock pass continuously out into the pegmatite, increasing vastly in size towards the interior of the vein. A K/Ar age on a pegmatitic muscovite (Oxburgh et al. 1966) is indistinguishable (but for a presumed grain-size effect) from those obtained from country-rock micas and interpreted as cooling ages from the M_2 metamorphism (see Chapter 6). This evidence strongly suggests that the younger pegmatites are directly associated with the climax of the M_2 metamorphism.

All gradations of pegmatite vein size exist from large, mappable intrusive bodies to centimetre-scale semi-concordant mineral veins and segregations. However, their variable, often very aluminous and tourmaline-rich compositions are nothing like those which would be produced by the partial melting of schists, and although temperatures high enough to initiate melting may have existed not far below the levels presently exposed, the origin of these fluids is somewhat problematical.

In particular, it would be interesting to know whether they have formed by "sweating out", i.e. by the segregation of rock fluids into veins, with crystallisation more or less in situ, or whether they represent the intrusion of fluids from deeper levels. In the former case, the pegmatites and mineral veins are merely a symptom of high temperatures and fluid pressures in the rocks, while in the latter case, their presence might suggest that fluids played a significant part in transporting heat for the metamorphism from deeper levels.

10. The fluid phase during M_2 metamorphism

Experimental investigations of mineral equilibria are generally carried out under hydrostatic ($p_{H_2O} = p_{total}$) conditions, and in order for these results to be applied to natural assemblages, it is necessary to consider among other things, the nature and composition of the metam orphic fluid phase. If the p_{H_2O} were considerably lower than the total pressure, natural assemblages could exist at temperatures lower than those indicated by the experimental work. However, at the critical localities there is good evidence for the presence of a free H_2O-rich fluid phase during metamorphism:

(i) In the zone of pelitic rocks on Polinik there are abundant quartz-rich veins which either follow or gently cut across the metamorphic foliation, and contain minerals typical of the surrounding schists. Assemblages recorded in these veins include quartz + feldspar, quartz + musc. + feldspar, quartz + kyanite, quartz + biotite + kyanite, muscovite + kyanite. It is very unlikely that these phenomena could have been produced with fluid pressures less than the total pressure, or that they were formed at any time other than close to the metamorphic maximum. One of the younger pegmatites in the Polinik Schists carries kyanite.

(ii) In the Tröger Alm biotite-feldspathic schists, and the northern parts of the Intermediate Schist around Eisenriegel (C6) a good measure of textural equilibrium has been achieved during the M_2 event, no further relics of M_1 minerals and fabrics appear, and the grain size is coarse. The synmetamorphic younger pegmatites are common. These points are taken as evidence to indicate that here, too, the fluid pressure during M_2 metamorphism was high.

For these two critical areas, it is probably reasonable to

assume that P_{fluids} approximately equalled P_{total}. It remains to be considered what was the partial pressure of water within this fluid phase. Studies in C-H-O fluid systems by French (1966) indicate that in the presence of free carbon (as graphite) only one additional constraint is required to buffer the P_{H_2O} at about 0.5, methane and carbon dioxide being the principal other components of the fluid phase under these metamorphic conditions.

Outside the areas considered above, i.e. further south in the map area, textural disequilibrium and patchy development of the new metamorphic assemblages suggest a less abundant fluid phase. Here the water-consuming reactions effecting the transformation to M_2 assemblages (such as the probable staurolite-forming reaction discussed earlier) are probably directly responsible for the removal of the fluid phase.

There is also the question of the origin of the fluid phase, abundantly developed at lower levels (i.e. pegmatites and veins, in the Polinik and Tröger Alm schists), but substantially decreasing in the staurolite zone. Assuming that the present M_2 assemblages in the higher-grade zones (Polinik Schists and Tröger Alm Schists) were produced from water-poor M_1 assemblages, as is the case for the Intermediate Schists, these higher-grade levels of the Altkristallin are highly unlikely to have been the source of large quantities of metamorphic fluid. Two principal alternatives remain , that the fluids were derived from still higher-grade crustal levels deeper than the present base of the Altkristallin Sheet, by dehydration reactions such as the breakdown of muscovite, or that the fluids are at least in part magmatic, and derived from below the crust.

C. Feldspathisation at Amphibolite Margins

Near the margins of larger bodies of amphibolite in the Polinik Schists (and also to a lesser extent in the Intermediate Schists of Eisenriegel) the schists are especially rich in plagioclase. This plagioclase is usually oligoclase or albite-oligoclase, forms up to 60% of the mode, and occurs in large irregular porphyroblasts up to 1 cm or more across, poikilitically enclosing narrow flakes of muscovite or biotite and small grains of opaque material or garnet (Fig. 4-23). In the case of the Xanitzl Alm body (FG3) the parent lithology is usually the silvery, muscovite-rich coarse-grained schist. Adjacent to the amphibolites, or in screens within them, the rock is markedly depleted in mica, and in thin section the plagioclase appears to be forming at the expense of muscovite, replacing the mica-rich parts of the rock.

An explanation of this metasomatism may be sought in the study of solid-fluid equilibria. Such a study has been made in the system $K\,Al\,Si_3O_8$-$Na\,Al\,Si_3O_8$ - Al_2SiO_5-SiO_2 - H_2O - HCl by Wintsch (1975), which delimits the stability fields of Na and K feldspars and micas with respect to the activity of Na^+, K^+ and H^+ in the fluid phase over a range of temperatures and pressures. The phase diagram for 600^oC and 6^{kb} relevant to the present study is reproduced in Fig. 4-24.

In a closed system, because of the preponderance of the mass of solid phases over the fluid phase, the composition of the fluid phase becomes buffered by the assemblage, i.e. in the case of coexisting muscovite and Na-rich feldspar, to the muscovite-albite boundary on Fig. 4-24.

Within the amphibolite body, the abundant retrogressive reaction textures, involving the breakdown of clinopyroxene and garnet to hydrous assemblages, bearing amphibole, biotite and clinozoisite, and the concentric disposition of the reaction zones with the least-hydrated

assemblages in the centre of the body, indicate that water-bearing fluids were passing into the amphibolite from the surrounding schists. Also, the growth of biotite within the amphibolites, where no K-rich phase is found in eclogitic assemblages, probably implies an influx of K^+.

The effect of these reactions on the fluid phase in and around the amphibolite would be to increase the $K^+/_{H+}$ and Na^+/H^+ ratios by removal of water, and to increase the Na^+/K^+ ratio by the formation of biotite. In either case the fluid phase composition in the schists is displaced into the stability field of albite. As long as this displacement from equilibrium is maintained, by the continued removal of components from the fluid phase by the amphibolite, and the migration of fluids through the marginal schists, Na-rich plagioclase will form at the expense of muscovite.

The width of this feldspathised layer, about 2 m next to an amphibolite body a few tens of metres thick, is such that the relative amounts of material to be exchanged and transported are by no means unreasonable.

D. M_3 Retrogressive Metamorphism

1. M_3 near the base of the Altkristallin

Through most of the Altkristallin sheet M_3 phenomena are limited to patchily distributed retrogression of higher-grade porphyroblastic phases, e.g. garnet and biotite (Fig. 4-25) replaced by chlorite, kyanite and staurolite by sericite.

North of the Launsberg shear zone much more extensive retrogressive metamorphism is evident. Here the relationships between the Altkristallin and the Pennine zone rocks within the Tauern Window have been somewhat modified by vertical movements later than the M_3

metamorphism. For this reason, a short time was spent in a study of the lower parts of the Altkristallin in the Sadnig Group, further to the West (Plate 2), where the base is a fairly low-angled (40-45°) thrust plane. The following remarks on the more intense M_3 metamorphic effects draw on evidence from both the Sadnig and Söbriach-Launsberg areas. Retrogressed rocks near the base of the Altkristallin sheet are also described by Bickle (1973) and Hawkesworth (1974).

This retrogressed zone of Polinik schists is characterised by the breakdown of plagioclase to albite, white mica and clinozoisite. The products may be fine-grained, or may occur as gefüllte albite porphyroblasts (Fig. 4-26). Garnet and biotite are generally chloritised, at least in part, and a muscovite fabric deformed by F_3 microfolding is cut by flakes of newly-grown white mica. No biotite remains within 500 m of the base of the Altkristallin sheet in the Sadnig Group, though occasional garnets may survive. The characteristic lithology near the base is a well-foliated muscovite-chlorite schist with gefüllte albite feldspars and undeformed pseudomorphs of chlorite after garnet. The strongly-oriented mineral growth is post-tectonic, and quartz is largely unstrained.

The zone of vertically-dipping rocks between Söbriach and the Launsberg shear zone shows similar textures to the Sadnig Group rocks. Here in addition there are amphibolites and leucocratic gneisses. The plagioclase of the amphibolites has broken down to albite, clinozoisite and white mica, and hornblende is often partly rimmed by chlorite. Microcline porphyroblasts in the leucocratic gneisses may show two generations of perthitic exsolution (see Fig. 2-14).

Exposure is not good enough in the Mölltal to demonstrate an increase in the intensity of retrogression towards the window edge. At Söbriach, schists with biotite, hornblende and garnet, but retrograded feldspar, may be found within 200 m of the base of the Altkristallin.

These rocks also show more evidence of post-M_3 shearing (see quartz strain map, Fig. 3-41) than those of the Sadnig Group, and have been rotated and possibly attenuated near the base by later movements, probably connected with those along the Mölltal line.

2. The Eisenriegel "green band"

In a narrow zone some 30 m broad, which runs approximately parallel to the strike east and west of Eisenriegel (C6), the Intermediate Schists together with intercalated amphibolites are strongly retrogressed. In the field the colour change, from brown-grey to a paler green, is striking. In thin section biotite and garnet are chloritised, staurolite and Al-silicates sericitised, and plagioclases inverted to albite, white mica and clinozoisite. The reaction products are all very fine grained. The band of alteration is not associated with any obvious geological feature such as a dyke or shear zone, but its coincidence with a narrow zone of highly deformed quartz fabrics suggests that there may be a zone of weakness, perhaps a strike fault of small displacement. Such a structure could have acted as a channel for hot, water-rich fluids through a cool and relatively impervious Altkristallin. The fluids may have been a part of those expelled from the underlying Peripheral Schieferhülle during the M_3 metamorphism.

The metamorphic evidence is summarised at the end of the following chapter (Chapter 5).

CHAPTER 5

Metamorphism II : Conditions

A. Introduction

1. Metamorphic Conditions

This chapter contains a discussion of the analytical results obtained on certain mineral assemblages in the map area, and is largely devoted to establishing the conditions of metamorphism of the M_2 event.

From information on the distribution of metamorphic minerals and assemblages presented in the previous chapter, certain limits may already be placed on the conditions of metamorphism by comparison with experimentally determined equilibria (some of the equilibria relevant to pelitic assemblages are shown in Fig. 5-1). For instance, the schists of the Eisen and Tröger Alms involve an andalusite-sillimanite transition in rocks also characterised, at least in the andalusite zone, by staurolite + quartz. The pressure-temperature profile that this succession represents must then pass below the Al-silicate triple point, but within the stability field of staurolite + quartz. The latter are apparently unstable over most of the Tröger Alm sillimanite zone. The Polinik schists on the other hand lie within the stability fields of kyanite and of staurolite + quartz.

What these observations mean in terms of absolute temperatures and pressures is not so simply derived from experimental work, since the presence of other chemical components in the minerals, and the composition of the vapour phase, must be taken into account. For the Al-silicates one must decide which of the published phase equilibria to employ.

In an effort to define more closely the pressure and temperature during M_2, some analytical work, using the Geoscan electron microprobe

in the Department of Mineralogy and Petrology, Cambridge, was undertaken on a few carefully chosen samples from parts of the map area.

2. Selection of rocks for analysis

For investigation of phase equilibria, the chosen rocks should be the product of one metamorphic episode (i.e. M_2) and rocks with relict minerals or products of retrogression should be avoided; they should contain as many phases as possible without violating the Phase Rule (i.e. an assemblage of low variance) so that the compositions of most of the phases should be controlled by chemical equilibria between them; and mutual contacts of all the mineral species with one another should be observable. Certain of the lithological groups in the map area do not lend themselves well to this treatment. The calc-silicate rocks have complex textures, probably containing minerals of several generations. The basic rocks are chemically rather complex, and once again their textures make the recognition of equilibrium assemblages difficult. The pelitic schists were thought to be the best rocks for the purpose.

Rocks from four localities were selected for study:

(i) the central part of the body of pelitic schists north of Polinik (C3) contains the fresh maximum phase assemblage qz + plag+ mu + bi + gt + st + ky. The texture is coarse, unstrained and there is no evidence of retrogression except for a flake or two of chlorite present in some thin sections. Three probe sections, from samples 033, 034 and 0524 were examined.

(ii) The calcite-dolomite marble from near Geistlacke (E3), sample 0174, was analysed in order to provide an estimate of the temperature using the calcite-dolomite geothermometer. Although the horizontal distance from the Polinik locality is 2500 m, the

difference in structural level, i.e. perpendicular to the foliation, is estimated at about 700 m. (That the distance perpendicular to the foliation is a measure of structural depth in the map area is suggested by the isograds in the Intermediate Schists, which are parallel to the strike of the foliation)

(iii) A sillimanite-rich rock, 0479, from the Tröger Alm (C6), containing the assemblage qz + plag + mu + bi + sill + ore was chosen to represent the sillimanite zone.

(iv) Sample 0462, from the ridge-top just south of the Eisen Alm shear zone, 100 m north of spot ht. 2639 m (B7), contains the assemblage qz + plag + mu + bi + gt + st + and with later chlorite and tourmaline. The specimen is perhaps not ideal because of its rather inhomogeneous texture, but was chosen as being a low-variance assemblage from the staurolite zone.

Thin section descriptions of the analysed pelitic rocks can be found in Appendix 1, and modal analyses of samples 034, 0524 and 0479 are presented in Table 5-2. 033 and 0462 are too heterogeneous for modal analysis on the scale of a thin section to be meaningful.

First the chemical mineralogy of the analysed phases is examined, and a chemographic treatment of the assemblages is briefly presented. Certain equilibria are chosen for more detailed study, and their bearing on the conditions of the M_2 metamorphism is discussed.

B. Chemical Mineralogy

1. Biotite

It is believed by many workers that the composition of biotite does not vary greatly with grade, but rather reflects the composition of the host rock (Schwander et al. 1968, Butler 1967).

Metamorphic biotites are generally more aluminous than the annite-phlogopite end-member series $K_2(Mg,Fe)_6Al_2Si_6O_{20}(OH)_4$, the compositions being closer to eastonite-siderophyllite $K_2(Mg,Fe)_5Al.Al_3Si_5O_{20}(OH)_4$. Eastonite is related to phlogopite by the coupled substitution $Al^{VI}\ Al^{IV} \rightleftharpoons Mg\ Si$. Butler (1967) suggests that Al^{IV} may vary with grade, but that Al^{VI} is not sensitive. However, Wenk (1970) reported a grade-related variation of Al^{VI} in biotites, and a consequent variation of the distribution of Al^{VI} between coexisting muscovite and biotite.

Fe:Mg ratios of biotite are believed to be controlled by the composition of the parent rock and the distribution of Fe and Mg between biotite and other ferromagnesian minerals in the rock (Butler 1965).

The data from the present study do not cover a sufficiently broad range of metamorphic grade to test any of these observations.

A convenient method of plotting chemical variation of micas within the sheet structure (i.e. excluding interlayer alkali cations) is to use the S-A-F plot devised by Lambert(1959) where

$$S = \frac{X_{Si}}{\Sigma} \times 100 \quad \text{per cent}$$

$$A = \frac{X_{Al} + X_{Fe}^{3+}}{\Sigma} \times 100$$

$$F = \frac{X_{Fe}^{2+} + X_{Mg} + X_{Mn}}{\Sigma} \times 100$$

where X_i = no. of atoms of i per formula unit, and

$$\Sigma = X(Si + Al + Fe^{3+} + Fe^{2+} + Mg + Mn)$$

Since the analyses in this work were performed on the electron probe, which does not distinguish between Fe^{2+} and Fe^{3+}, all the Fe has been taken as Fe^{2+} (as in the S-A-F plots of Ashworth (1975)). Assignment of some Fe to Fe^{3+} has the effect of shifting the plotted points of Figs. 5-3 etc. a little to the left, parallel to the base of the diagram.

Fig. 5-2 shows the S-A-F system with some mica end-members plotted. Figs. 5-3 and 5-4 show the compositions of the analysed biotites. With the exception of one analysis from the staurolite zone 0462, they all fall in the narrow field enclosed by dashed lines, lying on the Si-rich and Mg-Fe-poor side of the phlogopite-eastonite join. The exceptional analysis 0462-bi5, from a biotite in contact with muscovite, falls on a tie-line between the mean biotite and the mean muscovite compositions for 0462. Biotites from non-calcareous metamorphic rocks of the Lepontine region of the Central Alps, from analyses presented in Schwander et al. (1968), are plotted in Fig. 5-5. They form a broad spread of data, extending beyond the Kreuzeck group analyses (enclosed field) on all sides, although the grades of metamorphism represented are similar. The tighter grouping of the Kreuzeck group analyses may well reflect the narrow range of rock compositions and assemblages represented here.

The alkali contents of the biotites are discussed together with the muscovites.

2. Muscovite

Substitutions within the sheet structures, e.g. the phengite substitution, may be represented on the S-A-F plot described above.

It is generally agreed (Butler 1967, Schwander et al. 1968, Cipriani et al. 1971) that the extent of the phengite substitution Mg.Si $\rightleftarrows$ Al.Al decreases with increasing metamorphic grade. Fig. 5-6 shows fields of muscovite compositions from different grades of metamorphism. Data and assignment of the data to "petrogenetic fields" was taken from Cipriani et al. The amount of phengite substitution may vary widely under any particular set of metamorphic conditions, but in certain limiting assemblages the maximum equilibrium phengite content possible under those conditions should be achieved. According to Velde (1965) and Ashworth (1975) the limiting assemblage is mu + bi + ksp (+ qz), but assemblages such as mu + qz + bi + gt or chl should also approach the maximum phengite content in muscovite. Fig. 5-7 shows the compositions of Kreuzeck Group muscovites displayed on a S-A-F plot. The broken line surrounds analyses of Evans and Guidotti (1966) from the sillimanite zone (upper- and lower-) of Western Maine, and the thin continuous line bounds those of Ashworth (1975) from the sillimanite zone of the Dalradian. It can be seen that the amount of phengite in Kreuzeck Group muscovites is very low, and is even somewhat lower than that of higher grade micas (sill-ksp zone) from other areas. The effect of pressure on the phengite content of white micas was studied experimentally by Velde (1965, 1967) whose results indicate that phengite content is greatest at high pressures and low temperatures. Velde presents pressure-temperature plots contoured in Si content of stable phengitic micas. This does not resolve the above discrepancy, however, since the pressures during the M_2 and Maine metam orphisms were probably quite similar.

Guidotti (1973) presented muscovite data from Maine representing a range of metamorphic grade from the upper staurolite to the upper sillimanite zone, which can usefully be compared with the data from the Kreuzeck Group, and the comparisons are presented in Table 5-1. The values quoted from Guidotti are mean values from high-alumina assemblages (containing Al-silicate or staurolite). These show much more consistent trends with increasing grade than micas from low-alumina assemblages.

The mean Si contents of Kreuzeck Group micas are all lower than those from Maine (Table 5-1). Ti content apparently increases with grade in Guidotti's data, and the Kreuzeck Group relationships, staurolite zone less than kyanite zone less than sillimanite zone, is consistent with the probable temperature distribution between the zones here. Total Al is somewhat higher in all Kreuzeck Group rocks. Octahedral Al is comparable in the staurolite and sillimanite zones, but is higher than any of Guidotti's data for the kyanite zone. Total Mg + Fe for the Maine samples increases with grade, and this is apparently not a phengite substitution, since Si tends to decrease. The Kreuzeck Group data have (Mg + Fe) consistently higher than the Maine values, although the phengite contents, based on Si content, are lower.

The ratio Na:Na+K decreases markedly through Guidotti's sequence with increasing grade. This ratio is not so meaningful for the Kreuzeck Group micas because of the rather large alkali deficiency of the Polinik kyanite zone micas. The ratio of Na to total interlayer sites is 23.1 for the staurolite-andalusite zone, 22.9 for the kyanite zone and 17.4 for the sillimanite zone; these are consistent with the metamorphic grade variation in the map area.

The solid solution between muscovite and paragonite has been investigated by Eugster and Yoder (1955) and more recently by

Table 5-1. Comparison of muscovite compositions from the Kreuzeck Group and N.W. Maine, U.S.A.

	Atomic proportions					
	Si	Ti	Al^{vi}	Total Al	$Mg+Fe_{tot}$	Na/Na+K %
MAINE U.S.A.						
Upper staur. zone	6.144	.036	3.822	5.68	.158	24.3
Transition zone	6.142	.043	3.80	5.66	.179	21.4
Lower sill zone I	6.123	.051	3.79	5.67	.184	18.9
Lower sill zone II	6.135	.055	3.78	5.64	.193	19.6
Upper sill zone	6.118	.068	3.755	5.64	.206	14.2
KREUZECK GROUP						
0462 staur. zone	6.082	.033	3.80	5.72	.225	23.7 (23.1)
0479 sill zone	6.035	.081	3.73	5.70	.247	18.0 (17.4)
033,034 ky zone	6.104	.055	3.85	5.74	.221	27.1 (22.9)

Maine data are mean values for muscovites from high-Al assemblages in each metamorphic grouping, taken from Guidotti, (1973).

Values in parentheses for Kreuzeck Group muscovites in the Na/Na+K column are calculated as $\frac{\text{Na x 100}}{\text{total sites}}$ %

Eugster et al. (1972). The relationship may be summarised in a pseudobinary phase diagram muscovite-Ca-paragonite (Fig. 5-8) taken from Evans and Guidotti (1966, Fig. 10). Cipriani et al. (1971) present a curve for Na content of muscovites calibrated approximately for temperature, whose shape conforms to the left-hand-side of the binary phase diagram, with a maximum of around 23.5% Na:total sites at about 640°C. The temperature of the maximum Na content of Eugster and Yoder's curve was somewhat higher, around 670°C, but Cipriani et al. do not give a specific reason for lowering this temperature other than "adopting ..... reasonably acceptable temperature values on the basis of present knowledge". The mean Na content of their collected muscovite data, which will include non-limiting as well as limiting assemblages, peaks at about 17% Na:total sites in minerals of the "kyanite-almandine-muscovite subfacies" to which they allocate temperatures of around 600°C. It should be borne in mind, however, that paragonite breaks down at lower temperatures in the presence of quartz (Chatterjee 1972) and in addition the breakdown temperature, and that of the corresponding maximum paragonite content of muscovite, will depend on the X_{H_2O} in the vapour phase, decreasing with the dilution of H_2O in the vapour. Chatterjee's experimental curve for the breakdown of paragonite + quartz passes through 590-600°C at 5kb at $P_{H_2O} = P_{total}$.

Guidotti (1973, Fig. 2) indicates that the maximum paragonite content of muscovite should be achieved in the limiting high-alumina assemblage mu + Na-plag + Al silicate + qz, which is present in each of the analysed Kreuzeck Group rocks.

The data for the kyanite and staurolite zones of the map area, with their high paragonite contents, suggest temperatures of formation of muscovite in the region of the maximum solubility of paragonite, which probably lies at about 600°C. No free paragonite was

detected during the electron probe analysis of white micas.

Use of the Na content of muscovite as a geothermometer (e.g. in Lambert 1959) has not been very successful, and the following reasons may have contributed:

1. In low-alumina, paragonite-free assemblages, the maximum Na-content of muscovite is not attained.
2. High-alumina assemblages, especially at low to medium grades of metamorphism, may contain free paragonite which would contaminate wet-chemically analysed muscovites, yielding too high a Na content. Instances of finely interlayered paragonite and muscovite are known to occur at low grades of metamorphism (e.g. Höck 1974).

It remains possible that with the correct choice of mineral assemblage and analytical method, the Na content of muscovite may be a useful indicator of metamorphic temperatures.

Fig. 5-9 shows the distribution of Na and K in white micas, in numbers of atoms per formula unit. Micas from staurolite-Al silicate bearing assemblages 033, 034 and 0462 are considerably more paragonite-rich than the analyses of Butler (1967) from the biotite up to staurolite zones, and of Evans and Guidotti (1966) from the sillimanite zone. The diagonal line represents the condition Na + K 2 atoms and it can be seen that both Kreuzeck Group and other data are alkali-deficient and fall to the left of the line. Particularly for samples 033, 034, the discrepancy is larger than can be accounted for by the presence of unanalysed rare alkalis. This deficiency was noted in both muscovite and biotite by Butler (1967), who demonstrated an apparently related distribution of alkali deficiency between the two minerals, with biotites more deficient than coexisting muscovites.

Both mean values for muscovite and biotite in each rock, and values for spot analyses of mica pairs in contact, have been plotted

on Fig. 5-10. In samples 0462 and 0479 both mean and individual values lie on the trend defined by data from Ardnamurchan and Morar, whereas for the Polinik assemblage, the mean values lie close to the 1:1 distribution line, and only two individual pairs approach the field occupied by the Scottish micas. Coexisting pairs from Schwander et al. (1968) scatter widely and unsystematically over the diagram and have not been plotted.

A number of ways of compensating for alkali deficiency in the mica structure have been suggested, the most obvious being the replacement of interlayer K^+ and Na^+ by charged water molecules (H_3O^+), a hydromuscovite substitution, which may have occurred at the crystallisation of the mineral, or may be a result of weathering processes. Since water cannot be estimated on the electron probe, the existence of this substitution should lead to low analysis totals, and a positive correlation between alkali content and analysis total. Such a correlation could also be produced by loss of alkalis during analysis (see Appendix 1) or by other substitutions involving species which were not analysed. Fig. 5-11 is a plot of total alkalis against analysis total for micas from the Kreuzeck Group. No correlation is evident for any of the mineral groups. The analysis total is presumably too much a function of the accumulated uncertainties of the other elements to give an adequate assessment of H_2O and other undetermined species.

On the basis of a statistical analysis of a large quantity of muscovite analytical data, Cipriani et al. (1971) suggested the existence of a coupled substitution involving vacancies in the interlayer site, of the type

$$Si^{IV} + \text{vacancy} \rightleftharpoons Al^{IV} + K$$

On Fig. 5-12, total alkalis are plotted against Si for the muscovites from the Polinik assemblage (Samples 033,034). There

appears to be a weak correlation.

3. Garnet

(a) Zoning in garnet : models

Garnets from metamorphic rocks frequently display compositional zoning, and a number of models have been put forward to explain the most commonly observed zoning patterns. These different models may be summarised as follows:

(i) Equilibrium growth during prograde metamorphism, e.g. Miyashiro (1953). The composition at a point within a garnet thus represents a composition in equilibrium with the rest of the assemblage at the time that point cyrstallised at the garnet rim, and the garnet will preserve within itself a record of the prograde metamorphic conditions. Miyashiro explained MnO zoning on the grounds that the first garnet stable in pelitic rocks is a Mn-rich one (c.f. Hsu, 1968), and that at higher grades more Fe and Mg are able to enter the structure. Due to the changing partition of Fe and Mg with temperature and pressure between garnet and other ferromagnesian minerals, notably biotite (e.g. Albee, 1965) the garnet may be expected to be zoned with respect to these elements.

(ii) Depletion or segregation models, e.g. Hollister (1966), Atherton (1968), which seek to explain the bell-shaped profiles frequently shown by Mn and occasionally by Ca (see Fig. 5-14d, 0462-gt5). Elements such as Mn, of low abundance in the rock as a whole, are preferentially incorporated in garnets, which progressively deplete their surroundings in these elements. The shapes of Mn zoning curves can be fitted to curves generated by models

assuming simple Rayleigh fractionation, or more complex models involving a changing distribution coefficient between garnet and reservoir, or an increasing volume of diffusion in the reservoir with grade.

(iii) Diffusion models. The previous types of model assume that diffusion of the constituent elements is negligible within the garnet. Models which take account of diffusion in the garnet lattice may be considered under two groupings:- first, the models of Anderson and Buckley (1973) who seek to explain all common garnet zoning trends as diffusion phenomena, and second, the models of Grant and Weiblen (1971), Dallmeyer (1974) de Béthune et al. (1975), who invoke a limited diffusion of components to explain unusual zonations, which are often restricted to the rims of garnets, and believed to be the effect of retrogression.

Evidence so far available indicates that garnets from low and medium grades of metamorphism tend to be strongly zoned, whereas at higher grades, garnets are more homogeneous (Butler 1967, Evans and Guidotti 1966, Blackburn 1969), implying that recrystallisation occurs, or that diffusion within garnet becomes fast enough at high temperatures to smooth out zoning patterns. Evidence for the unreactiveness and slow re-equilibration of garnet at moderate temperatures is common. In the map area, garnet is the last M_1 mineral to readjust to the M_2 conditions.

Clearly some of the above models are generally applicable, some relevant to specific cases; the models are not necessarily mutually exclusive, in that in any one garnet the content of each different element may be controlled by a different set of factors.

(b) Optical character of Kreuzeck Group garnets

Garnets occur in all the probed pelite samples except 0479.

(i) Eisen Alm locality. The garnets of sample 0462 are relatively small (c. 1 mm) and evenly distributed, and are distinct from the larger, shattered and biotite-rimmed relict M_1 garnets of neighbouring rocks. They are assumed to have nucleated and grown during the metamorphic episode M_2.

(ii) Polinik locality. 033, 034 and 0524. On optical criteria, the garnets may be divided into three main types.

Type 1 are 2 mm diameter rounded garnets, with cores often noticeably pink in thin section and crowded with tiny highly birefringent or larger squarish lower birefringence (quartz ?) inclusions, surrounded by a clearer rim (Fig. 5-13). Sometimes, more particularly in sample 0524, the garnets are hollow-cored, and filled with quartz and micas.

Type 2 garnet forms incomplete rims to some Type 1 garnets. They are more spongy, and contain elongate rutile inclusions, often oriented in one or more parallel sets, and the contact with Type 1 garnets is often marked by a band of small opaque inclusions.

Type 3 garnets are inclusion-free and colourless in thin section, and seem to be particularly associated with the staurolite-kyanite-rich knots in sample 033.

(c) Zoning patterns in Kreuzeck Group garnets

Electron probe traverses across garnets from samples 0524 and 033 (Polinik) and 0462 (Eisen Alm) are displayed in Fig. 5-14 abcd, and the zoning paths are plotted on triangular diagrams Figs. 5-15,16.

A number of different zonation patterns can be distinguished,

which broadly conform to the optical distinctions described above.

(i) 0462-gt 5 (Fig. 14d)

Roughly bell-shaped patterns appear for Mn and Ca; Mg and Fe increase towards the rim. This is the characteristic pattern found by Atherton and other workers in garnets from medium-grade rocks, e.g. Atherton (1968). The Mn curve shows a slight marginal enrichment, apparently at the expense of Fe (see later discussion of Mn-enriched rims).

(ii) Type 1 garnets from Polinik locality (remaining profiles except 033-gt 2, -gt 5 and rims of 0524-gt 4).

These show many of the characteristics of 'typical' zoning, i.e. rather flattened bell-shaped Mn and Ca, though there are interesting modifications. Sometimes the Mn profile shows a depression in the centre of the grain. Mg profiles often show symmetrical maxima near the garnet rims (e.g. 0524-gt 4).

This distribution of Mg is responsible for the strongly curved zoning paths on Fig. 5-16. The Mg/Mg + Fe ratios change markedly across the garnet. This is not immediately explicable in terms of segregation or diffusion models, and may well reflect changing equilibrium conditions between ferromagnesian minerals with increasing grade. When staurolite and kyanite join the Polinik assemblage, the chemical system (ignoring minor components) becomes fully defined according to the Phase Rule (see second part of this chapter) and the garnet zoning paths should then converge on the same rim composition. Fig. 5-16 suggests that this is indeed the case. It is therefore possible that the change in direction of the zoning path marks the entry of a new phase into the assemblage, and the resulting equilibria require the garnet to become less Mg-rich.

The compositions of garnet cores spread out along the zoning paths pertaining to the rock, the reason being either that the profile investigated did not pass through the exact centre of the garnet, or possibly that the garnets nucleated at different times during the evolution of conditions in the rock. The former alternative is assumed (see also Atherton 1968), and the "true" core composition is taken to be that of the beginning of the longest zoning path for each rock. The core compositions from samples 0524 and 033 differ: garnets from the former developed from a core composition of approximately $Alm_{73}\ Pyr_{12}\ Gross_{7.5}\ Sp_{7.5}$ whereas 033 cores are around $Alm_{68}\ Pyr_{16}\ Gross_{5}\ Sp_{11}$. This difference probably reflects differences in host rock composition. Sample 033 has lower Ca than 0524: all the garnets are poorer in grossular, and the plagioclase is both less abundant and lower in anorthite content. The host rocks probably have different Mg:Fe ratios. However it is notable that the different garnet core compositions converge towards very similar rim com positions, implying that though the garnet composition may initially have been affected by that of the host rock, at a later stage it was controlled by equilibria within the maximum phase assemblage.

Hollow garnets (0524-gt 2 and -gt 3) do not appear to be strongly zoned, except for a compensating Mg and Fe zonation in 0524-gt 2. Inner margins are similar in composition to external rims.

(iii) Type 2 garnets. 0524-gt 4, Fig. 5-14b.

Narrow type 2 rims were investigated in 0524-gt4 (Fig. 5-13) and were found to be significantly more grossular-rich. The amount of the Ca-enrichment seems to vary inversely with the width of the rim. It is not known whether the type 2 garnets of the Ca-poorer rocks 033 and 034 are similarly grossular-

enriched. That the type 2 rims are a late addition is strongly suggested by the fact that the average garnet rim-composition, towards which both type 1 and type 3 garnets converge, is achieved inside the type 2 rims. The type 2 garnet has probably not achieved equilibrium with the rest of the assemblage, and may have grown during waning metamorphism, exchanging Ca with neighbouring plagioclase, and perhaps thus accounting for local variations in plagioclase composition in these rocks.

(iv) Type 3 garnets. 033-gt 2, -gt 5, Fig. 5-14c.

Two clear, colourless, inclusion-free garnets from sample 033 display an entirely different zonation pattern. The core is pyrope-rich and very low in Ca and Mn, approximately Alm_{70} Pyr_{27} $Gross_{1.5}$ $Sp_{1.5}$. By analogy with other prograde zoning patterns and garnet composition variations, this should represent a higher-grade, and probably a higher pressure environment of formation than other garnets analysed here. The zoning path approaches the same average rim composition as type 1 garnet. A clue to the nature of the zoning is given by 033-gt 2 (Fig. 5-17) which contains a broad internal fissure, filled with minerals of the external assemblage (quartz and micas). The core composition with maximum Mg content is achieved close to the fissure, in the geometrical centre of the garnet, and there is then a very steep compositional gradient towards the fissure, the average rim composition being reached at the very edge. This pattern suggests that the type 3 zoning is a product of diffusion into and out of a pre-existing, cracked, high-pyrope garnet of the components necessary to bring the garnet into equilibrium

with the M_2 assemblage. Cracked M_1 garnets, the deformation having occurred before M_2, are common in the Kreuzeck and Intermediate Schists.

These type 3 garnets, with their fundamentally different, pyrope-rich core composition, are believed to be relics of the M_1 assemblage, which is partly preserved in basic rocks in the Polinik area by eclogites in the cores of amphibolite lenses. Sample 033 contains scattered knots, 1-2 cm in diameter, of staurolite- and kyanite-rich material associated with type 3 garnets which include those analysed. This texture may well be a relic of a coarse-grained M_1 fabric.

(v) Mn-enriched rims (e.g. Fig. 5-14d)

Certain of the Mn profiles show an increase in spessartine content over the last 100μ at the rim, which appears to be at the expense sometimes of Fe (0462-gt 5), sometimes of Mg (033-gt 9). Both type 1 and type 2 garnet rims seem to be affected, and possibly also type 3, but the effect would be indistinguishable from the characteristic type 3 zoning pattern. This indicates that if all such phenomena may be ascribed to the same event, it occurred late in the metamorphic history.

Similar Mn-enrichment has been described by Grant and Weiblen (1971), Dallmeyer (1974) and de Béthune et al. (1975). The main problem is the source of the manganese. Two possibilities present themselves: either the Mn is derived from the breakdown of a Mn-bearing mineral elsewhere in the rock, or it is derived from the garnet itself, by breakdown of the outermost rim, and selective removal of Fe, Mg and Ca. The second alternative is favoured by the above-mentioned authors. De Béthune et al. present excellent textural and analytical evidence to support their model of garnet resorbtion by retrograde reactions

such as chloritisation, with incorporation of Mn in the garnet rims by migration along grain boundaries and limited diffusion into the garnet lattice.

In the analysed rocks there is little direct evidence of resorbtion or retrogression at Mn-enriched garnet rims except in the case of 0462-gt 5, where a thin zone of chlorite and sericite of late, retrograde origin intervenes between the garnet and the adjacent staurolite. However, samples 033, 034 and 0524 do contain small quantities of chlorite, its texture indicating its late, retrograde origin, and its presence revealing retrogressive reactions which may have involved the redistribution of manganese in garnet rims.

In summary, the analysed garnets display a number of different zoning patterns:-

1. Bell-shaped profiles for Mn and Ca in some garnets may be ascribed to segregation or fractionation during growth.
2. Mg and Fe contents of garnets in the maximum phase assemblages are believed to be controlled by equilibria between garnet and other ferromagnesian minerals, which may have changed with time during garnet growth.
3. Certain relict pyrope-rich M_1 garnets display a zonation caused by diffusion into and out of them of the components required to bring them into equilibrium with the M_2 assemblage.
4. A late period of garnet growth, possibly during waning metamorphic temperatures, produced partial rims around M_2 garnets which appear to be significantly more grossular-rich.
5. A later event, possibly involving retrograde chlorite-forming reactions and some garnet resorbtion, has produced some Mn enrichment at garnet rims.

(e) Garnet rim compositions.

The compositions of garnet rims are displayed on triangular diagrams in Figs. 5-18 to 5-20. The solid symbols in Figs. 5-18 and -19 represent the tightly grouped points at the end of zoning paths. This is the garnet composition believed to be in equilibrium with the rest of the assemblage. Open symbols are "anomalous" points which, by comparison with Figs. 5-15 and -16, can be seen to lie on the zoning paths relevant to the type of garnet to which they belong. These points are believed in many cases to represent spot analyses insufficiently close to the garnet edge. Note that 033 and 034 garnet rims contain less Ca and slightly more Mn than those of 0524, although their Mg:Fe ratios are the same.

4. Staurolite

The formula of staurolite has not yet been established beyond argument. The ideal structural formula given by Deer et al. (1962) and Ganguly (1972) is $(Fe,Mg)_2\ Al_9\ Si_4\ O_{23}\ (OH)$ but analyses generally show more (OH), apparently balanced by deficiencies in the Si and Fe sites. Richardson (1967) argued that a more realistic formula would be $(Fe,Mg)_2\ Al_9\ Si_{3.75}\ O_{22}\ (OH)_2$ since synthesis experiments using starting mixtures of the ideal composition with excess H_2O almost invariably yielded staurolite plus quartz.

The analyses in this work are calculated on the basis of 23 oxygen equivalents, conforming to the Richardson formula. They may be summarised as follows:-

033:- $(Fe,Mg,Mn,Zn)_{1.92}\ Al_{8.86}\ Si_{3.84}\ O_{22}\ (OH)_2$

034:- $(Fe,Mg,Mn,Zn)_{1.82}\ Al_{8.85}\ Si_{3.89}\ O_{22}\ (OH)_2$

0462:- $(Fe,Mg,Mn,Zn)_{1.81}\ Al_{9.04}\ Si_{3.77}\ O_{22}\ (OH)_2$

Ti has not been allocated because of some doubt about which site it enters. Smith (1968) believed it to occupy Fe sites. Eisen Alm

staurolites (0462) show more Al and less Si than Polinik analyses, and are close to the Richardson formula except for an Fe site deficiency. In other respects the differences between samples 033 and 034, rocks from the same locality no more than 50 metres from each other, are as great as the differences between Polinik and the Eisen Alm; for example the Mg/Fe ratios:

	033	034	0462
$X_{Fe}^{Fe\ site}$	.766	.740	.738
$X_{Mg}^{Fe\ site}$	.168	.146	.1405
Mg/Fe	.219	..197	.190

Zinc and cobalt are elements preferentially concentrated in staurolite. Zinc was detected to the extent of 0.25 wt.% ZnO in 034 and 0.49% in 0462, and a trace of cobalt in 034. The zinc content of staurolite will depend on both the Zn content of the rock and the modal content of staurolite.

5. Plagioclase

Some plagioclases from samples 033 and 0524 were examined for compositional zoning. While most proved fairly homogeneous, some showed a slight zoning from more albite-rich cores to more anorthite-rich rims (see Fig. 5-21). There is, however, little evidence to suggest whether these reflect growth under changing conditions, or later, possibly retrograde, exchange of Na and Ca. Equilibria which may have controlled the composition of plagioclase in these rocks include:

grossular + 2 Al-silicate + quartz = 3 anorthite
(in garnet)

(see discussion of this equilibrium later in the chapter), and

paragonite + quartz = albite + Al-silicate + H_2O
(in muscovite)

Increasing temperature displaces both these equilibria in favour of the right hand side, so it is clear that the An content of plagioclase will not be a simple function of temperature in these rocks.

Sample 0462 staurolite-zone plagioclase compositions are very variable, from An_4 to An_{15}, with a mean of $An_{9.5}$. Compositional zonation was not investigated in this sample, but is quite likely to be present. A slight and gradual optical zonation was noted in some grains.

Polinik samples show less variation in composition within a probe section, but there is significant difference between sample 0524 (mean An_{15}) and samples 033/034 (mean An_{11}). This probably reflects the host rock composition, since the garnets are also significantly different in Ca content.

Plagioclases from the sillimanite-grade Tröger Alm (sample 0479) fall in the range An_{23} to An_{27}. In the absence of garnet, plagioclase is the only major Ca-bearing phase in the rock.

6. Aluminium Silicates

The Al-silicates analysed are close to Al_2SiO_5 in composition, with traces of Fe and Mg. Fe contents of andalusite and sillimanite are around 0.5 to 0.6 wt.% Fe_2O_3. That of kyanites in 033 and 034 (from partial probe analyses not presented in Appendix 1) is of the same order, suggesting that no significant differences between Al-silicates in their affinity for Fe occur. Unequal partition of Fe has been suggested as a possible reason for the experimental difficulty of precisely determining Al silicate phase boundaries, particularly andalusite-sillimanite (e.g. Holdaway, 1971). However,

it would seem that in natural systems the distribution of Fe between andalusite and sillimanite is approximately equal (see also Albee and Chodos 1969).

7. Chlorite

Two late, cross-cutting, probably retrograde chlorites from sample 0462 were analysed, one associated with garnet, the other with biotite. They are almost identical in composition and Mg/Fe ratio. On the classification of Hey (1954) they fall near the ripidolite-pseudothuringite boundary with a $Fe_{total}/Fe + Mg$ ratio of 0.56. The formula calculation in Appendix 1 has allocated an arbitrary fraction (0.13) of the total Fe to Fe^{3+}.

8. Ilmenite

With the exception of a few unidentified grains associated with sillimanite in sample 0479, the opaque phase in all rocks investigated proved to be ilmenite. The atomic totals of the full analyses presented in Appendix 1 show marginally less Fe than Ti, suggesting that hematite solid solution is negligible, possibly reflecting conditions of low oxygen fugacity.

9. Distribution of Fe and Mg between minerals

(a) Garnet-biotite

The distribution of Fe and Mg between garnet and biotite has been studied by a number of workers, and in particular by Albee (1965), who came to the conclusion that the distribution coefficient K_D which may be expressed as

$$\frac{(Mg/Fe)_{Gt}}{(Mg/Fe)_{Bi}}$$

depends on metamorphic grade and on the Mn content of the garnet. By plotting Mn/Mn + Mg + Fe against Mg/Mg + Fe, a value of Mg/Mg + Fe yielded by extrapolation to zero Mn may be obtained and used to calculate a K_D. This correction for Mn content was found by Albee to amount to a reduction in K_D of about .007 per 1 atom % increase in Mn/Mn + Mg + Fe. Albee obtained K_D's characteristic of metamorphic zones as follows: garnet zone 0.20; staurolite zone 0.215; kyanite zone 0.23; sillimanite - K feldspar zone 0.30 - 0.37.

Plots of Mn/Mn + Mg + Fe against Mg/Mg + Fe are shown for Polinik samples 033, 034, 0524 (Fig. 5-22), and for the Eisen Alm sample 0462 (Fig. 5-23). Only those points plotted as solid circles on Figs. 5-18 and 5-19 (i.e. the rim compositions corresponding to the end of the zoning paths) have been used in the Polinik diagram. A line has been fitted by eye through the trend of the data points to intersect the abscissa at about 16% Mg/Mg + Fe. This figure yields a K_D of 0.215, with a sensitivity of about 0.0035 per 1% Mn/Mn + Mg + Fe.

All but two of sample 0462 garnets lie close to a line which yields a corrected Mg/Mg + Fe of 12.25 at zero Mn, and a K_D of 0.157 with a sensitivity of .004 per 1% Mn/Mn + Mg + Fe. The remaining two garnets appear to lie on a parallel trend.

The K_D values of 0.215 and 0.157, respectively from the kyanite and staurolite zones, are lower than the equivalent values of Albee. This may be due to other factors such as pressure. Omitting the Mn-correction, or making allowance for the presence of Fe^{3+} in the Kreuzeck Group biotites only reduces the K_D's even further.

(b) Staurolite - biotite

Hounslow and Moore (1967) found the distribution of Fe and Mg between staurolite and biotites to be regular but not ideal over a range of $(Mg/Fe)_{St}$ values from 15 to 35%, K_D varying with the Mg/Fe ratio of the staurolite. The K_D values from the present study are 0.238 (sample 033), 0.226 (sample 034) and 0.217 (sample 0462). These values plot close to the distribution curve of Hounslow and Moore for $(Mg/Fe)_{Bi}$ against $(Mg/Fe)_{St}$, Fig. 5-24, in a range where there is a gap in their data, and would appear to confirm their conclusion. No variation of K_D with temperature was noted by Hounslow and Moore.

10. Graphical Presentation of Pelite Assemblages

(a) Estimation of whole-rock compositions

Rock compositions were calculated from the modal proportions, mean analysed compositions, and densities of their constituent minerals. Modal proportions were determined by point-counting the area of the probe section. Samples 033 (with 1 cm knots of staurolite and kyanite) and 0462 (a large andalusite porphyroblast constitutes half the section) were judged not homogeneous enough to yield a meaningful whole-rock composition by this method and were not counted. The method is probably a good approximation for the remaining three probed assemblages 034, 0479 and 0524 which are homogeneous, and the sampled area representative of larger volumes of rock.

Some other rocks were point counted for comparison, 1000-2000 points per section, and whole rock compositions calculated for these, selecting mineral compositions from the probed assemblage closest in character to the rock in question. Clearly the possibilities of error in these unanalysed cases are much greater than for the

probed rocks, particularly 0342 which is an M_1 assemblage about whose minerals no precise compositional information is available. The results of modal analysis and whole-rock composition calculation are shown in Table 5-2, and the whole-rock compositions are plotted on a Thompson AFM diagram, projected from muscovite, in Fig. 5-25.

(b) Mineral compatibilities

Compositions of minerals from the probed assemblages have also been plotted on Thompson AFM projections (Fig. 5-26). Only garnet may contain significant components outside the system AFKM. The projections here were plotted with a computer program, primarily designed for dealing with rock compositions, which does not subtract Al equivalent to grossular content of garnets. However, in this case the positions of garnets are not significantly affected because of the low grossular contents (about 2 mol % in those plotted). Mn is considered together with Fe.

The properties of this projection are dealt with extensively by Thompson (1957). In the presence of excess quartz, muscovite and H_2O, an assemblage should be capable of representation on this projection by a single mineral, or a tie-line between two minerals, or as a three-phase field. The bulk composition of the rock should be contained by the field of the assemblage.

The maximum phase assemblages from Polinik and the Eisen Alm contain four phases that project on the AFM plane, implying that the conditions of the projection have been violated. However, there is no indication that the assemblage is out of equilibrium, and indeed ky-st-gt-2 mica-qz schists are by no means uncommon in the Altkristallin and elsewhere. Similar situations involving an extra phase are often encountered in metamorphic rocks, for example the frequently documented persistence of muscovite into the sillimanite-

Table 5-2. Modal analyses and computed whole-rock compositions of mica-schists from the Kreuzeck Group.

	Pelitic Polinik schist	Pelitic Polinik schist	Polinik schist (biotite-rich)	Coarse grained Polinik schist	Tröger Alm biotite feldspathic schist	Tröger Alm Al-rich bi-fsp schist	Pelitic Intermediate schist, Eisenriegel.	Pelitic Intermediate schist, Källuker1.	Kreuzeck schist
	0524*	034*	058	0121	0474	0479*	0477	0152	0342
Qz	26.8	27.2	42.3	61.8	42.2	32.0	3.5	18.6	42.5
Plag	11.6	2.6	26.4	18.6	33.1	19.6	21.7	.9	4.8
Mu	29.3	38.7	8.1	8.3	2.3	11.9	41.2	62.6	36.7
Bi	16.7	18.4	22.3	8.7	22.0	25.9	20.7	7.4	13.2
Chl	tr	tr	tr	.6	-	-	1.7	2.9	.4
Gt	10.4	7.2	tr	1.2	-	-	1.2	7.7	1.6
St	2.6	4.9	-	-	-	-	9.3	-	tr
Ky	2.4	1.1	-	-	-	-	-	-	tr
Sill	-	-	-	-	tr	10.5	-	-	-
Ilm	tr	tr	.8	.9	.3	tr	.7	tr	.8
	*	*				*			
SiO_2	56.9	55.7	69.4	79.3	70.7	61.4	45.7	52.9	66.2
Al_2O_3	22.6	24.0	13.5	9.8	13.1	21.7	30.4	27.1	18.0
TiO_2	.5	.6	1.1	1.0	.8	.6	1.3	.5	1.2
Fe_2O_3	.4	.5	.4	.3	.3	.4	.8	.6	.5
FeO	8.5	7.6	5.0	3.1	4.1	4.4	6.8	6.0	4.1
MnO	.3	.2	.0	.1	.0	.0	.1	.2	.1
MgO	2.4	2.5	2.4	1.1	2.5	2.9	2.7	1.9	1.7
CaO	.6	.3	.8	.9	1.6	.9	.6	.2	.3
K_2O	3.4	4.2	2.6	1.4	2.3	3.3	4.8	5.0	3.9
Na_2O	1.6	1.0	2.8	1.8	3.0	2.0	2.8	1.2	1.1
H_2O	2.3	2.9	1.6	1.0	1.3	2.0	3.5	3.8	2.6
Total	99.5	99.5	99.5	99.8	99.7	99.6	99.4	99.4	99.7
	0524*	034*	058	0121	0474	0479*	0477	0152	0342

* indicates samples for which mineral composition data are available.

K feldspar zone, where the phases present are represented by the equilibrium mu + qz = ksp + sill + H_2O. (Evans and Guidotti, 1966; Ashworth 1975). There are two principal alternative explanations for the persistence of an extra phase:

(i) First, the existence of chemical components outside the ideal system in one of the phases may be sufficient to stabilise it. This is the interpretation given by Hounslow and Moore (1967) to explain the presence of garnets, which contain appreciable Mn, in ky-st-gt-2 mica-qz assemblages from the Grenville schists. However, the Mn or Ca contents of Kreuzeck Group garnets (all in the range 90-97% Mg + Fe/total R^{2+} at equilibrium) are not sufficient to explain the widespread occurrence of the maximum phase assemblage (see Chapter 4, Fig. 4-21) in the Polinik Schists, nor is it possible to explain either the staurolite or biotite as excess phases on similar grounds (see also Chakraborty and Sen, 1967). Garnet clearly cannot be regarded as an excess phase in sample 0524 where it makes up over 10% of the mode, and causes the whole-rock composition to fall in the gt-st-bi field. This argument is not circular, since whatever the Mg/Fe ratios chosen for biotite etc., the preponderance of garnet still causes the rock composition to fall to the left of st-bi tie-lines.

(ii) The second possibility is that H_2O cannot be regarded as a perfectly mobile component, and that the solid assemblage is capable of buffering the fluid composition. Given that a certain proportion of the fluid phase consists of diluting components such as CO_2, CH_4, it would be possible for the X_{H_2O} to be controlled by the release of H_2O into, or its removal from the fluid by the adjustment of an equilibrium between

the solid phases to changing pressure and temperature. Such an equilibrium would be the reaction

$$st + mu + qz = gt + as + bi + H_2O$$

relating the phases of the maximum pelite assemblage. The fact that the maximum phase assemblage is widespread in the Polinik Schists but rarer on the Eisen and Tröger alms, where the equilibrium does appear to define an isograd, suggests that in the Polinik Schists the vapour composition was more effectively buffered than in the latter area. The greater abundance of pegmatitic veins (i.e. veins of distinctly different composition from the host rock, as opposed to mineral segregations of similar composition) on the Tröger Alm may be connected with this as evidence of greater fluid mobility during M_2 metamorphism. The buffering capability of this reaction is calculated later in the chapter.

Fig. 5-26 summarises the Thompson projection information.

The sample 0462 biotites vary considerably in Mg/Fe ratio, which might be thought due to local equilibration on a scale smaller than that of a probe section, except the tie-lines drawn from minerals in contact, cross each other (Fig. 5-26). The chlorite composition falls inside the area enclosed by the other phases and is clearly of retrograde origin.

The assemblage and biotite compositions of sample 0479 indicate that the three-phase field of gt-bi-as is to be found at lower Mg/Fe ratios. This is consistent with the deployment of bi—as tie-lines in the other rocks.

C. Calcite - Dolomite Geothermometry

The determination of temperatures using the $MgCO_3$ content of calcite coexisting with dolomite utilises the temperature-dependence of the maximum solubility of $MgCO_3$ in $CaCO_3$. The geothermometer is calibrated by the experimental work of Goldsmith and Newton (1969), who also investigated the comparatively slight pressure-dependence of the solvus.

The sample chosen, 0174, is from the narrow horizon of banded calcite-dolomite marble from Geistlacke corrie (E3) pictured in Fig. 3-3. The dolomitic bands consist only of dolomite (80%) and calcite (20%), and the rock has an even-grained equilibrium texture produced during the M_2 metamorphism. The tendency for dolomite grains to be convex against calcite is often noted in carbonate rocks and is probably due to the greater surface energy of dolomite. Some of the calcites were somewhat dusty in appearance. During analysis, some of the more inclusion-rich areas were analysed to determine whether these consisted of exsolved dolomite. With the exception of one inhomogeneous large calcite grain, these test analysis did not show abnormally high Mg contents.

The presence of other components such as Fe and Mn introduces uncertainties into the calibration of the geothermometer (Bickle 1973, and in prep. 1975) but fortunately these are not a problem in sample 0174 where the mean FeO in calcite is 0.05% by weight, with Mn below the detection limit.

40 calcites and 17 dolomites were analysed, the calcite analysis spots generally being chosen close to adjacent dolomite grains.

The distribution of $MgCO_3$ contents of these calcites is shown in the histogram, Fig. 5-27. The range of values is broad, and requires some interpretation. Carbonate systems are thought to equilibrate more readily than silicates. Re-equilibration of dolomite

solid solutions in calcite is believed to occur rapidly at temperatures above around 500°-550°C (Bickle 1973, Sheppard and Schwarcz 1970) so that calcites raised to higher temperatures in the presence of dolomite, and then cooled, will tend to exsolve dolomite until the temperature falls to about this value. The maximum concentration of data should then reflect the temperature at which most of the calcites ceased to re-equilibrate; in fact the histogram peak of 5 mol % $MgCO_3$ represents a temperature of about 520°C.

The scatter towards higher values may then be interpreted as those calcites which ceased to equilibrate at temperatures closer to the maximum. The sharp cut-off above 8% should then represent a minimum estimate for the highest temperature reached. Appropriate pressures, in the lower part of the kyanite stability field will be in the range 4-7 Kb, depending on the triple point adopted. The maximum homogeneous magnesian calcite corresponds to a temperature of about 620°C in this pressure range, with a sensitivity of -3°C per Kb increase, suggesting that temperatures reached at least this value during M_2.

D. Pressure Control from the System $CaO-Al_2O_3-SiO_2$

The pelite samples analysed contain the subsystem $CaO-Al_2O_3-SiO_2$ in which the following univariant equilibrium may be written

3 Anorthite = Grossular + 2 kyanite + quartz

The possible usefulness of this equilibrium for determining pressure limits was first suggested by Kretz (1964). Experimental investigations have been made by Hays (1966) and Hariya and Kennedy (1968) at high temperatures and pressures. Because of its relatively low slope, it was hoped that some absolute pressure limits might be placed on conditions in the map area.

Although the best-fit curve presented by Hariya and Kennedy does not coincide with that of Hays, in fact an equilibrium curve compatible with both sets of data may be constructed, which is parallel to that of Hays, and passes through temperatures 10°C lower at equivalent pressures. This best-fit curve has been projected back into the pressure range of interest (Fig. 5-28a).

The equilibrium can be calculated using the computer program THERMOCALC written by R. Powell (see Powell, 1973). THERMOCALC calculates the equilibrium temperatures of reactions by summing heats of formation, heat contents, entropies and molar volumes of the participating phases. The thermochemical data for pure phases are taken from compilations such as Robie and Waldbaum (1968). To a large extent the data have been made self-consistent, and data for additional phases have been retrieved from reversed experimental equilibria. However, inconsistencies remain, particularly for aluminous phases. Activities of pure phase end-members in solid solutions may be input in order to calculate the displacement of an equilibrium involving impure phases.

The calculated curve for

$$3\ an = gross + 2\ ky + qz$$

with pure phases has the same slope as the experimentally determined equilibrium curve but passes through temperatures some 70°C higher. This implies errors in the thermochemical data equivalent to almost 3 Kcal/mole in the calculated ΔH of reaction, but it is difficult to know where to attribute the error. As a compromise solution, all subsequently calculated curves have been shifted down-temperature such that the calculated pure-phase curve coincides with the experimentally determined one. The pattern and mutual spacing of the curves will not then be significantly affected if most of the error

is contained in imprecise heats of formation (ΔHf) for the phases concerned.

Equilibrium curves for the analysed assemblages have been calculated by inserting into THERMOCALC appropriate values for the activities of grossular and anorthite end-members in garnet and plagioclase. Grossular is a dilute solution in garnet (2-8 mol %), and ideal ionic mixing is assumed. The small positive excess free energy of mixing in grossular-almandine solid solutions (about 40 cal/mole for sample 0524 using estimates of Ganguly and Kennedy 1974) has an insignificant effect on these equilibria. To calculate the activities of anorthite in plagioclase, the data of Orville (1972), determined at 700°C and 2 Kb, were employed. Since temperatures in excess of 600°C were indicated by the calcite-dolomite geothermometry, the effect of any peristerite solvus at lower temperatures (Crawford 1966) is assumed to be negligible.

Fig. 5-28b shows the set of calculated curves, adjusted so that the pure-phase curve coincides with the experimental one (Curve 1). In addition, three Al-silicate phase diagrams, those of Althaus (1967), Richardson et al. (1969) and Holdaway (1971) are shown for comparison.

Curve 2 is that calculated for the more Ca-rich pelite 0524 from the Polinik locality, using the mean plagioclase composition, and the mean garnet rim composition. The curve lies within the kyanite fields of Richardson et al. and of Holdaway at reasonable temperatures, but well outside that of Althaus. In addition it passes through the temperature indicated by the calcite-dolomite assemblage at around 6 Kb.

Curve 3 represents mean values from samples 033 and 034. These rocks are poorer in Ca than 0524, and 033 at least contains relict very Ca-poor (type 3) garnets. Curve 2, for sample 0524, in which plagioclase and garnet are both more abundant and more Ca-rich, is believed to be a more reliable indication of the garnet-plagioclase equilibrium at

the Polinik locality, especially since curve 3 passes below all the Al-silicate triple points on Fig. 5-28.

The mean composition of the Ca-enriched rims on 0524-gt 4 has been employed in the calculation of curve 4. There is some doubt as to the plagioclase composition in equilibrium with these, or indeed whether they truly represent an equilibrium involving plagioclase, kyanite and quartz, but at least it appears that they probably reflect a lower temperature of formation than the maximum assemblage.

Curve 5 has been calculated for sample 0462 using andalusite as the Al silicate. Since the garnets are strongly zoned, with Ca decreasing markedly in 0462-gt 5 towards the rim, the garnet composition represented by the three low-Ca garnets (2.2 mol %) has been chosen as representing the end of the zoning paths and thus the true grossular content in equilibrium at the maximum metamorphic conditions. The plagioclase contents also scatter widely, but the mean value of 10 mol % anorthite has been used here, since no information on plagioclase zonation is available. The resulting curve is compatible with all the Al-silicate phase diagrams.

The possible error on these curves is hard to assess, but is probably of the same order as the discrepancy between the two curves (2 and 3) for the Polinik locality, i.e. ± 1.5 Kb. However, if the Richardson et al. triple point is accepted, and curve 3 attributed to disequilibrium, the error may be reduced to c. ± 1 Kb. The accuracy is probably sufficient to allow the following conclusions to be drawn:

1. The analytical data favour a low-pressure, high-temperature position for the Al-silicate triple point, since the Polinik curves are inconsistent with the kyanite field of Althaus.

2. Pressures of 6 Kb (± 1 Kb) are indicated for the Polinik assemblage at the temperatures deduced from calcite-dolomite geothermometry.

E. Phase Equilibria in the system K_2O-FeO-(MgO)-Al_2O_3-SiO_2-H_2O

1. The analytical results indicate that the above six-component system accounts for over 95% of the composition of the investigated pelitic rocks. In the Polinik and Eisen Alm assemblages, these components are distributed among six solid phases: quartz, Al silicate, muscovite, biotite, garnet and staurolite.

Since Mg and Fe are interchangeable in the three phases (biotite, garnet and staurolite) in which they occur, it is permiss ble to consider a simplified system K_2O-FeO-Al_2O_3-SiO_2-H_2O (or KFASH) consisting of the Fe-end-members annite, almandine and Fe-staurolite (subsequently abbreviated as ann, alm and fe-st). The six-phase assemblage qz + ky + mu + ann + alm + fe-st is invariant in a pure H_2O vapour, and five univariant reactions radiate from this invariant point. These are

(1) (Mu)(Bi) fe-st + qz = alm + ky + H_2O

(2) (fe-st) alm + mu = ann + ky + qz

(3) (alm) fe-st + mu + qz = ann + ky + H_2O

(4) (ky) fe-st + ann + qz = alm + mu + H_2O

(5) (qz) fe-st + alm + mu = ann + ky + H_2O

Two extra degrees of freedom are introduced by allowing:

(i) Variation of the vapour phase composition by dilution of H_2O with other components.

(ii) Variation of the solid phase compositions by solid solution, and hence variation of the activity of end-members in the phases. Strictly speaking, the variance of the assemblage is increased by one for each extra component, but MgO is the only significant one in this case, even in the garnets.

The addition of MgO to the system causes the reaction

(6) $st + mu + qz = gt + bi + as + H_2O$

(which is of course Carmichael's (1969) isograd reacion, see Chapter 4) to be theoretically univariant, although it can be rendered divariant by variation of the X_{H_2O} in the vapour, or the significant presence of other components.

In the six-phase assemblage in a more complex system with solid solutions and a variable vapour phase the activities of the model system end-members in the phases must be such that all the equilibria 1 to 5 are satisfied. The P-T-X_{H_2O} curves for these equilibria can be calculated, given adequate experimental and thermochemical data. Analysis of the phases of variable composition allows activities of the KFASH end-members to be calculated. Then calculation of the five equilibria defines a line in P-T-X_{H_2O} space which is the locus of the KFASH invariant point for the six-phase assemblage. Fixing one other variable, P,T or X_{H_2O} yields unique solutions for the other two.

In practice, this approach is complicated by the fact that the thermochemical data for phases such as fe-st, alm and ann are poorly known.

Experimental work in this part of the model system is limited. Reaction (1) has been investigated by Richardson (1968) using sillimanite as the Al-silicate, and by Ganguly (1972), using kyanite. Reaction (3) was attempted by Hoschek (1969) but not conclusively reversed.

2. Calculation of Equilibria

The five equilibria were calculated using the method and computer program THERMOCALC of R. Powell (1973) as for the anorthite-grossular equilibrium described in the previous section.

Mean composition of phases from the analysed Polinik assemblage were used to calculate the activities of end-members. These were derived on the basis of ideal ionic mixing on sites, which is believed to be a good approximation for garnet, mica and a number of other minerals (Powell 1973). The activity of, say, almandine in a garnet would then be

$$(X_{Fe}^{Fe\ site})^3 \cdot (X_{Al}^{Al\ site})^2 \cdot (X_{Si}^{Si\ site})^3$$

where $X_i^{i\ site}$ is the fraction of the possible i sites occupied by i .

The reactions (1) to (5) were balanced for the following formulae:

Quartz	SiO_2
ky	Al_2SiO_5
mu	$K\ Al_3Si_3\ O_{10}\cdot (OH)_2$
ann	$K\ Fe_3\ Al\ Si_3\ O_{10}\ (OH)_2$
alm	$Fe_3\ Al_2\ Si_3\ O_{12}$
fe-st	$Fe_4\ Al_{18}\ Si_{7.5}\ O_{44}\ (OH)_4$

The balanced reactions are:

(1) 6 fe-st + 25 qz = 8 alm + 46 ky + 12 H_2O

(2) alm + mu = ann + 2 ky + qz

(3) 6 fe-st + 8 mu + 17 qz = 8 ann + 62 ky + 12 H_2O

(4) 6 fe-st + 23 ann + 48 qz = 31 alm + 23 mu + 12 H_2O

(5) 6 fe-st + 17 alm + 25 mu = 25 ann + 96 ky + 12 H_2O

The thermochemical data contained in THERMOCALC mostly come from fairly reliable compilations such as Robie and Waldbaum (1968),

cross-checked against experimentally determined equilibria. However, data for many Fe-bearing phases are not available, and their enthalpies of formation have been extracted from a chain of experimental equilibria starting from hercynite and magnetite (data of Zen, 1973).

Data for almandine were extracted by Powell from the reactions

(7) $Fe_5Al_2Si_3O_{10}(OH)_8 + \frac{2}{3}O_2 = \frac{4}{3}Fe_3O_4 + Fe\,Al_2O_4 + 3\,SiO_2 + 4\,H_2O$

Fe-chlorite — magnetite — hercynite

and

(8) $Fe_5Al_2Si_3O_{10}(OH)_8 + \frac{1}{3}O_2 = Fe_3Al_2Si_3O_{12} + \frac{2}{3}Fe_3O_4 + 4\,H_2O$

almandine

both reversed by Hsu (1968). Fe-chlorite and almandine data may thus be assumed to be concordant, although they can both be affected by any error in the thermochemical or activity coefficient data for hercynite and magnetite, or in the composition assumed for the chlorite run products. Staurolite data were then extracted by Powell from the Richardson (1968) investigation of reaction (1), staurolite data thus being linked to that for almandine. The annite data contained in THERMOCALC are extracted from experimental work of Wones et al. (1971) and are relevant for annites formed under low f_{O_2} conditions.

Entropies for these minerals were approximated by a summation method from oxides, which Powell believes to be sufficiently accurate.

The Al-silicate data were found to be not quite compatible with the Richardson et al. triple point, the one preferred for reasons to be outlined later, although the differences are only of a few tens of calories. The reactions were calculated using sillimanite since the staurolite data were extracted from a reaction involving this polymorph, although the Al-silicate in the assemblage is kyanite. Since the estimated conditions for the Polinik assemblage lie close to the kyanite-sillimanite boundary of Richardson et al. and of Holdaway, and indeed lie within the sillimanite field of the THERMOCALC data,

the error introduced by using sillimanite is very small.

At the calculated activities of Fe end-members, the reactions should all coincide at the P, T and X_{H_2O} of the Polinik assemblage. However, the calculated equilibrium temperatures are widely discordant and require a correction to the ΔH of reaction of more than 20 Kcal per mole annite, almandine or muscovite to bring them into the same temperature range. Therefore this approach cannot be used to place further limits on the metamorphic conditions.

Possible reasons for a discrepancy include the following:

(i) Disregard for the Fe^{3+} content of biotite and staurolite. However, the apparently low oxidation ratio of the rocks (see Chapter 4) should correspond fairly well to the biotite for which the properties were determined. No great changes in oxidation state are necessarily expected in equilibria (1) to (5). Any oxidation state effect is difficult to quantify, but is not expected to be great in this case.

(ii) Errors in the entropies of minerals. The summation method is usually a good approximation, but the slopes of some calculated reactions, e.g. reaction (1) are anomalous. However, deviations from the assumed quantities are extremely unlikely to total as much as 20 cal/mole.

(iii) Errors in the calculated activities due to departures from ideality in solid solutions are possible, but unlikely to exceed an order of magnitude in any phase. The total discrepancy is equivalent to five orders of magnitude on the activities. It may be concluded that non-idealities in solid solutions are unlikely to be a major cause of the discrepancy.

(iv) Incorrect enthalpies for the phases, totalling -21 Kcal per mole annite. The enthalpies of formation (ΔH_f) are the most likely quantities to be in error to the extent required to satisfy equilibrium

in the Polinik assemblage. Quartz, Al-silicates and muscovite data are fairly well-defined, but Fe-bearing phases may be subject to much greater errors due to the method of their derivation. The error on alm, fe-chl and fe-st will be systematically related by the reactions (8) and (1), but if these are inconsistent with the annite data, a large error in the ΔH of reaction could result.

3. Calculation of best-fit enthalpy corrections

Assuming the discrepancy to be due to incorrect ΔH_f values, it was decided to use the data from the Polinik assemblage to set up a self-consistent set of thermochemical data for the phases in the model system. The P, T and X_{H_2O} of the Polinik assemblage can be estimated from evidence already available, and thus the position of the KFASH invariant point approximately defined.

Evidence for temperature: the calcite-dolomite geothermometry indicated a temperature of at least 620°C at a locality not far from the Polinik assemblage in rocks of uniform metamorphic grade.

Evidence for pressure: garnet-plagioclase equilibria suggest a pressure of about 6 Kb at 620-630°C, which is in the lowest-pressure part of the Richardson et al. kyanite field, close to the triple point. This is consistent with the existence of an andalusite-sillimanite progression a few kilometers to the south.

Evidence for X_{H_2O}: the presence of small amounts of graphite in the Polinik pelites may have buffered the vapour phase, or at least until it became buffered by the six-phase pelite assemblage. The work of French (1966) implies X_{H_2O} of around 0.5 in graphite-bearing rocks over most metamorphic temperatures. The X_{H_2O} is taken to be 0.5 for the purposes of the calculation.

Recalculation of the experimentally determined reaction (1) at an X_{H_2O} of 0.5 yielded an equilibrium temperature of 654°C at 6 Kb,

which is in fact just inside the sillimanite field of Richardson et al. but is otherwise broadly consistent with the evidence outlined above. These conditions also lie within the rather vague bracket from Hoschek's determination of reaction (3).

THERMOCALC was then employed to calculate the enthalpy corrections necessary to cause the reactions (1) to (5) to coincide at the estimated conditions of 654°C, 6 kb and $X_{H_2O} = 0.5$. The total correction required is equivalent to a correction to the enthalpy of formation of annite of -21.2 Kcal which may be considered to be distributed between the most suspect phases annite, almandine, Fe-chlorite and Fe-staurolite. Since almandine, Fe-chlorite and Fe-staurolite are related by reactions (1) and (8) there are two unknown quantities. Another experimentally determined equilibrium, independent of reactions (1) to (5), may now be used to set up and solve simultaneous equations whose roots are the "best-fit" enthalpy corrections for annite and almandine/Fe-staurolite. Such a reaction is:

(9) 31 fe-chl + 41 mu = 8 fe-st + 41 ann + 33 qz + 108 H_2O

This was studied by Hoschek (1969) who obtained a well-defined reversal bracket at 7 kb of 550-570°C. Hoschek used natural minerals as his starting materials but gives enough analytical data to enable activities of the Fe-end members to be calculated. An enthalpy correction equivalent to -7.80 $\pm$ 0.35 Kcal per mole annite is necessary for the calculated curve to pass through the experimental brackets.

The best enthalpy corrections, those which satisfy the experimental equilibria (1), (8) and (9) and cause reactions (1) to (5) to converge on the estimated conditions can now be calculated. They are:

Almandine)		(new ΔH_f = -1282.9 Kcal
) + 14.6 Kcal	(	
Fe-chlorite)		(" = -1717.4 Kcal
Fe-staurolite	+ 19.5 Kcal	" = -5697.5 Kcal
Annite	- 6.6 Kcal	" = -1241.6 Kcal

While this work was in progress, M. and R. Powell (personal communication, December 1974) suggested that as a result of their recent investigation into the thermochemistry of biotite, the ΔH_f of annite should be about 6 Kcal more negative than the value in THERMOCALC. The agreement between this and the suggested "best-fit" correction is encouraging.

The new value for the ΔH_f of almandine is intermediate between that of Zen (1973) and that initially assumed by Powell.

The possible error on these best-fit corrections may be estimated. Misplacing the conditions of the Polinik assemblage by $\pm$ 30°C or $\pm$ 2 kb has an effect on the total ΔH correction of $\pm$ 1 Kcal. The X_{H_2O} chosen has no greater effect than this unless the true X_{H_2O} is much less than 0.1. Other possible sources of error, such as solid solution non-ideality, or variable oxidation state, have been discussed earlier. The "best-fit" corrections are estimated to be reliable to $\pm$ 2 Kcal on annite, or $\pm$ 4 Kcal on almandine and Fe-staurolite.

4. Geological Implications of Equilibria in the system KF(M)ASH

The new consistent set of thermochemical data may now be used to calculate the effect of changing certain of the variables, the most important in this case being the variation of equilibrium temperature with X_{H_2O}. Fig. 5-29 shows the result of calculating the T-X_{H_2O} curves of reactions (1) to (5) at 6 Kb for the Polinik assemblage. It can be seen that the equilibrium temperatures change slowly over most of the range of vapour compositions, and low values of X_{H_2O} are

required to reduce the temperatures substantially, for example reaction (3) requires an X_{H_2O} of less than 0.05 to bring its equilibrium temperature below 600°C.

The curve for the Carmichael isograd reaction

(6) st + mu + qz = gt + bi + as + H_2O may also be calculated, and it falls between the curves for reactions (1) and (3) (Fig. 5-30). There is evidence from the map area that this is the reaction controlling the disappearance of staurolite in the Eisen and Tröger Alm (see Chapter 4). This staurolite-out isograd would appear to correspond to temperatures in the range 620-665°C, for X_{H_2O} values in the range 0.1 to 0.9 which are reasonable when $P_{fluids} = P_{total}$.

In Chapter 4 the approximate geographical coincidence of the staurolite-out and andalusite-sillimanite isograds on the Eisen and Tröger Alms was noted. The presence of kyanite in one assemblage near the isograds, and the existence of extensive kyanite-zone rocks a short distance to the north, implies a P-T gradient through the Eisen and Tröger Alm succession which passes close to the Al-silicate triple point. According to Holdaway's (1971) phase diagram the andalusite-sillimanite isograd should correspond to the polymorphic inversion at 500-540°C at around 3.5 kb. To depress the equilibrium temperature of the staurolite-out isograd reaction as low as this requires X_{H_2O} values which are unreasonably low (of the order of 0.005) for an area where there is evidence for high fluid pressures. The above estimate of 630-660°C is closely in accordance with the andalusite-sillimanite boundary of Richardson et al. at around 5 Kb. The field evidence, supported by calculated equilibria, is thus strongly in favour of a high temperature triple point for the Al-silicates, such as that of Richardson et al.

A corollary of the T-X_{H_2O} curve of reaction (6) is the indication of the buffering capacity of this reaction. The six-phase assemblage

can remain in equilibrium with vapour phases of X_{H_2O} from 0.1 to 0.9 over a temperature range of about 50°C, which is quite adequate to ensure a broad outcrop of the maximum phase assemblage in the field, if the buffering mechanism operates. An equilibrium such as (4), the staurolite-producing reaction of the Eisen Alm, has an even greater buffering potential, the tendency being in Al-silicate-free assemblages for the X_{H_2O} in the fluid to become reduced with increasing temperature.

F. Summary of Metamorphic Evidence

The first metamorphic event recognised in the map area, M_1, is widely preserved in the southern Kreuzeck Group, where it is of amphibolite facies with staurolite and occasional kyanite. The mineralogy of pelitic compositions is dominated by muscovite and garnet. M_1 mineral growth is responsible for the characteristic texture of micaceous Kreuzeck schists: porphyroblasts of garnet are set in fine-grained parallel-felted muscovite with a phyllitic sheen in hand specimen. Relict eclogitic assemblages in basic rocks within the Polinik schists are also attributed to the M_1 event.

Petrographic mapping revealed the progressive overprint of M_1 by an M_2 event in the Kreuzeck and Intermediate Schists, with the following features in the approximate order of appearance, south to north:

(i) growth of biotite and plagioclase porphyroblasts;

(ii) gradual increase in muscovite fabric recrystallisation;

(iii) growth of staurolite and breakdown of old garnet;

(iv) breakdown of M_2 staurolite and growth of Al-silicates (sillimanite, andalusite, and kyanite).

Three isograds may be constructed:

(i) a staurolite isograd, marking the first appearance of M_2

staurolite. Textural evidence suggests that staurolite appears by the reaction

$$gt + mu + H_2O = st + bi + qz$$

but this is not a very suitable reaction on which to base an isograd, because of its bivariance in the pelite model system KFMASH, its large H_2O-buffering capacity, and the abundance of disequilibrium textures at this grade of M_2 metamorphism.

(ii) an andalusite-sillimanite isograd. Because of the comparative rarity of andalusite this cannot be very precisely defined. One occurrence of kyanite near the isograd suggests that conditions passed close to the Al-silicate triple point.

(iii) a garnet-biotite-Al-silicate isograd, based on the reaction

$$st + mu + qz = gt + bi + as + H_2O$$

may be drawn for the Eisen and Tröger Alm succession, and corresponds to the breakdown of staurolite (staurolite-bearing assemblages without muscovite or quartz were not recorded). It approximately coincides with the andalusite-sillimanite isograd. In the summer of 1975 these isograds (ii) and (iii) were traced westwards as far as the mouth of the Wöllatal, 2.5 km ENE of Stall (Plate 2) giving them an outcrop length of more than 8 km, over which they approximately follow the strike of the foliation and lithological banding.

The Polinik schists, further north and separated from the Intermediate Schists, by shear zones, are all in the kyanite zone, and display no gt-bi-as isograd, the six-phase assemblage of gt-bi-ky-st-mu-qz being quite common.

Evidence for a free fluid phase during M_2 metamorphism may be found in the Polinik schists, and the Tröger Alm and Eisenriegel areas, in the form of mineral veins and segregations containing similar

assemblages to the host rocks, and common pegmatitic veins on various scales from narrow tourmaline or muscovite-rich veinlets to mappable discordant bodies. In the staurolite and lower zones, however, disequilibrium textures are common, and small-scale compositional variations are apparent in the probed sample 0462. This is consistent with the reduced importance of the fluid phase, depleted by hydration reactions forming biotite and staurolite largely at the expense of M_1 garnet.

The origin of the fluids, both metamorphic and pegmatitic (if indeed there should be any real distinction) is somewhat problematical. The tentative conclusion is that they were not generated in the presently exposed levels of the Altkristallin, but came from below, either from below the crust or by dehydration reactions in levels of the Altkristallin now lost due to overthrusting. The fluids may have assisted in the transfer of heat to higher levels in the crust.

Analytical work using an electron microprobe yielded data on mineral compositions from five pelitic rocks, and one calcite-dolomite marble. This latter indicated maximum temperatures of at least 620°C during M_2 metamorphism in the Polinik schists, deduced from the $MgCO_3$ contents of calcites.

The high paragonite contents of muscovites suggest that the ky and st zone minerals formed at around 600°C, near the breakdown temperature of paragonite + quartz. The Na content of muscovite should be a reliable geothermometer, with the correct choice of mineral assemblage, and in situ analysis by electron probe.

Garnets exhibit compositional zonations, of which most may be related to growth during prograde metamorphism, possibly accompanied by depletion or segregation effects involving Mn and Ca. Some more unusual patterns can be attributed to diffusion zoning of pre-M_2 garnets

of substantially different composition; or later compositional modifications involving some garnet rims.

Representation of mineral assemblages on Thompson AFM projections from muscovite reveals that the maximum phase assemblages from the Polinik schists and the Eisen Alm contain an extra phase, indicating violation of the conditions of the projection. Buffering of the vapour composition by the assemblage is thought to be the explanation, rather than stabilisation of one or more phases by extra components outside the model system KFMASH.

Calculation of the equilibrium

$$3\ an = gross + 2\ ky + qz$$

indicated pressure of 6 ± 1 Kb for the temperature range $600\text{-}650^{o}C$ in the Polinik schists.

Recalculation of the gt-bi-as isograd reaction with self-consistent thermochemical data yielded equilibrium temperatures in the range $620\text{-}665^{o}C$ for reasonable ranges of fluid composition.

The Al-silicate triple point of Richardson et al. is preferred on the following grounds:

(i) the coincidence of the and-sill and gt-bi-as isograds in the field, together with the calculated range of equilibrium temperatures for the latter isograd reaction.

(ii) the relatively low pressures and high temperatures indicated by the plagioclase-garnet equilibrium compositions.

Metamorphic conditions may now be assessed for M_2 in the map area:

(i) In the Polinik Schists, temperatures were in the range $620\text{-}665^{o}C$ depending on the X_{H_2O}, which was most probably buffered by the assemblage, with a pressure of 6 ± 1 Kb.

(ii) In the Eisen and Tröger Alm areas, the presence of a gt-bi-as isograd indicates that X_{H_2O} was not buffered by the assemblage.

If instead the graphite present in the assemblages at the isograd succeeded in buffering the vapour phase composition (at $X_{H_2O} \simeq 0.5$), the isograd should represent temperatures of around 640°C and 5 Kb. This is consistent with both the Richardson et al. and-sill boundary and the plagioclase-garnet equilibrium curve for the staurolite-andalusite zone.

If a mean density of 2.75 g/cc is assumed for the overlying rocks (these probably included Kreuzeck-type garnet-muscovite schists, quartz phyllites and carbonates) a depth of burial of c. 22 Km is indicated for the Polinik schists, and c. 18 Km for the Eisen and Tröger Alms. Average geothermal gradients (assuming a surface temperature of 0°C) are 28-30°C/Km for the former area, and about 35.5°C/Km for the latter.

The third metamorphic event M_3, which is prograde and of green-schist to amphibolite facies within the Tauern Window, is retrograde in the Altkristallin Sheet, and has significantly affected only the basal 1.5 Km of the Altkristallin in the Sadnig Group and north of the Launsberg shear zone. Elsewhere its effects are limited to patchily distributed chloritisation of biotite and garnet.

CHAPTER 6

K-Ar Geochronology

A. Purpose of study, and sample selection

The original aim of the geochronological investigation in this work was to attempt to limit the area of c. 80 My ages (see Chapter 1), and in particular to link the age data obtained, to the metamorphic and structural history of the rocks. It was also expected that excess argon phenomena, perhaps similar to those described by Brewer (1969, 1970) might be found in the map area, and that good exposure and detailed geological observations might enable these to be related to petrographic and tectonic features of the rocks.

To define the southern limit, material was collected from a traverse stretching southwards from the area of Oxburgh et al. (1966), through the area mapped by the writer to connect with the area of ages in excess of 200 My described by Brewer (1970), (see Plate 5).

To investigate the northern boundary, and to study age relations across the edge of the Tauern Window, samples were collected from the northern Schober, southern Glockner and Granatspitz Groups by M.J. Bickle. In addition a few samples were collected from rocks near the window edge at Sȯbriach near Obervellach.

Most of the minerals separated for K-Ar age determination were micas from mica-schists and gneisses, and wherever possible both muscovite and biotite from the same sample were determined. In addition, three chlorites from mica-schists (to investigate the concentration and composition of argon in K-poor minerals), four amphiboles from metamorphic rocks of basic composition, one mylonite (whole-rock), and a biotite and two amphiboles from late dykes in the

Kreuzeck Group were analysed.

Samples K6, K7 and L2-c were selected from the collection of M.S. Brewer, and 366RCW, 624RCW, 697RCW and 731RCW from that of R.C. Wright.

B. Analytical Methods

Minerals were separated from crushed and sieved rock samples by conventional techniques including a water panning device to concentrate micas, and a Franz magnetic separator. Heavy liquids were necessary only for the amphiboles from dyke samples. Either the 40-80 mesh fraction or 60-120 mesh fraction size ranges were used, according to the grain size of the original sample.

Argon was extracted on a glass line in the Mineralogisches-Petrographisches Institut, University of Bern, Switzerland, and analysed on a Varian Mat GD 150 mass spectrometer, as described by Purdy (1972). A 99.98% ^{38}Ar spike was used, and was calibrated against Bern muscovite 4M using a value of 6.31×10^{-6} STP cc/g ^{40}Ar rad. The spike was checked against U.S.G.S. P207 muscovite. The mean of six measurements yielded the value $27.81 \pm 0.33 \times 10^{-6}$ STP cc/g ^{40}Ar rad.

A double furnace was used, enabling two samples to be analysed per day without letting any part of the system up to air. The line was baked overnight at about 150°C, with the samples in, to ensure a low argon blank and a good starting vacuum.

After fusion of mineral samples in hydrofluoric and perchloric acids, their potassium contents were measured on a Baird flame photometer which employs a lithium internal standard. The operating technique used was the "null method" recommended by the manufacturers. Most samples were analysed in duplicate, with some triplicated.

Constants used in the age calculation were:

$\lambda_{\epsilon} = 0.585 \times 10^{-10}\ y^{-1}$

$\lambda_{\beta} = 4.72 \times 10^{-10}\ y^{-1}$

atomic abundance of $^{40}K = 1.19 \times 10^{-4}$ moles ^{40}K/mole K

The 2σ error for the whole age determination is estimated at $\pm$ 4% for the Bern laboratory (Hunziker, 1974) broken down as follows:

± 1% on the K determination

± 1% on the spike calibration

± 0.6% on the air argon ratio

and a drift of ± 1% against international standards.

Reproducibility on the Baird flame photometer was ± 2% (on replicate analyses, and against USGS standards and local standards supplied by R.K. O'Nions) and thus the overall 2σ error on the determinations in this study is estimated at ± 5%.

C. Compilation and Presentation of Data

The analytical data are listed in Tables 6-1 to 6-3.

Petrographic descriptions and localities of samples from the Kreuzeck Group (except for dyke samples of R.C. Wright) may be found in Appendix 2. Descriptions and localities of Schober and Glockner Group samples are contained in Bickle (1973).

The regional distribution of the samples and their ages, together with the data of Oxburgh et al.(1966), Brewer (1970), and Lambert (1970) are shown on Plate 5.

Table 6-1. K-Ar results on micas from the central Kreuzeck Group.

Specimen No. and mineral		Rock type	Ar rad. STPcc/g x10^{-6}	%rad	%K	$^{40}Ar/^{36}Ar$	$^{40}K/^{36}Ar$ x 10^3	Age, My
0201	M	2 mica Polinik schist	20.77	92.7	6.31	4007	752	82.6
"	B	"	20.94	91.0	6.78	3109	610	77.6
0204	M	chloritised Intermed. schist	18.14	94.8	5.01	5688	992	93.9
0206	M	2 mica qz-rich Intermed. schist	13.00	94.0	3.31	4914	784	98.1
"	B	"	30.84	97.2	6.03	10610	1345	126.9
0125	M	gt-2 mica Intermed. schist	18.11	92.9	4.40	4173	628	102.8
"	B	"	19.25	93.7	5.39	4688	820	89.4
0255	M	st-gt-2 mica Intermed. schist	49.47	93.8	6.62	4780	400	183
"	B	"	42.87	96.6	5.99	8594	773	175
0282	M	st-gt-2 mica Intermed. schist	40.77	97.5	5.04	11966	962	197
"	B	"	54.75	97.9	5.85	14202	990	226
0280	M	st-gt-mu Intermediate schist	42.98	97.1	4.54	10262	701	229
366RCW	M	2 mica Kreuzeck schist	45.13	97.0	4.87	9971	697	224
"	B	"	44.00	97.3	5.08	10811	809	210
K 7	M	gt-mu Kreuzeck schist	36.43	97.5	5.18	11866	1096	172
K 6	M	gt-2 mica Kreuzeck schist	34.47	98.1	6.03	15260	1744	141
"	B	"	7.353	83.0	4.17	1732	543	44.7

Mineral abbreviations: M - muscovite, B - biotite, C - chlorite, H - hornblende, W - whole rock.

The 2σ error on model ages is estimated at $\pm$ 5%

Table 6-2. K-Ar results from the southern Glockner and northern Schober Groups.

Specimen No. and mineral	Rock type	Ar rad. STPcc/g x10^{-6}	%rad	%K	$^{40}Ar/^{36}Ar$	$^{40}K/^{36}Ar$ x 10^3	Age, My
SP 614 M	2 mica gneiss, Granatsp. Kern	13.73	91.7	7.58	3561	1201	45.9
" B	"	10.86	93.4	6.66	4453	1699	41.4
SP 622 M	2 mica schist, Riffldecke	7.32	88.2	4.88	2502	981	38.1
" B	"	4.99	81.2	4.47	1569	759	28.5
SW 42 M	ctd-mu schist, Periph. Schief.	9.17	85.6	8.03	2046	1021	29.1
SP 210 M	calc-phyllite, Periph. Schief.	4.93	85.8	4.04	2068	970	31.0
SP 506 M	augen gneiss	8.91	81.1	5.30	1561	502	42.6
" B	"	5.24	80.6	4.22	1521	657	31.6
SP 596 M	graphitic, pyritous phyllonite	33.50	96.8	4.67	9450	850	175.5
SP 15 M-C	mu-chl schist	14.00	90.8	2.76	3205	382	126.0
SP 203 M	white quartz schist	19.48	89.6	7.47	2831	648	65.7
SP 89 M	leucocratic gneiss	23.73	91.8	7.83	3610	729	76.2
SP 719 M	2 mica schist	18.99	95.6	5.88	6626	1308	81.0
" B	"	29.83	92.0	4.61	3669	348	159.0
SP 708 M	2 mica schist	14.70	92.0	5.51	3698	851	67.2
" B	"	15.80	80.5	3.95	1508	202	99.9

Table 6-3. K-Ar results from micas, whole rock, chlorites and amphiboles from metamorphic rocks, a shear zone, and dykes from the Kreuzeck Group.

Specimen No. and mineral		Rock type	Ar rad. STPcc/g x10^{-6}	%rad	%K	$^{40}Ar/^{36}Ar$	$^{40}K/^{36}Ar$ x 10^{3}	Age, My
Base of Altkristallin, and shear zone								
0218	M	phyllonite	23.74	96.7	6.17	8798	1473	96.2
0217	M	phyllonite	26.45	97.0	5.88	9871	1417	112.1
0215	M	hbl-gt-2 mica Polinik schist	20.64	90.2	4.26	3006	373	120.3
"	B	"	12.22	82.9	3.78	1720	294	81.1
0197	W	mylonite, Kleinblässen shear z.	6.74	93.4	2.80	4480	1154	61.0
Chlorite analyses								
0217	C	phyllonite	2.298	63.2	-	801	-	-
0204	C	chloritised Intermed. schist	3.464	80.8	-	1531	-	-
L 2	C	pelitic schist (see Brewer,1970)	21.62	92.9	-	4167	-	-
Hornblende analyses from amphibolites								
0213	H	amphibolite	2.279	37.4	0.38	471	19.5	147.
07	H	amphibolite	6.084	88.9	0.85	2657	219	175.
028	H	eclogitic amphibolite	7.288	92.8	0.42	4089	146	397.
0260	H	amphibolite	6.453	84.9	0.32	1949	55.3	453.
Analyses from dykes								
624RCW	H	andesite	2.564	73.7	0.81	1120	173	80.0
697RCW	H	andesite	1.203	45.2	0.81	537	109	37.6
731RCW	B	dacite	5.187	67.2	4.12	897	319	32.0

D. The Interpretation of Radiometric Ages from Metamorphic Rocks

Age determinations, both Rb-Sr and K-Ar, on metamorphic minerals have been interpreted as cooling ages, i.e. the time elapsed since the mineral cooled below temperatures at which radiogenic daughter products could escape by diffusion (e.g. Armstrong et al., 1966; Harper, 1967).

The rate of diffusion of argon (or strontium) in a mineral lattice depends strongly on temperature, such that above a critical temperature (usually known as the blocking temperature) argon may diffuse out relatively rapidly, while below that temperature it is retained and the mineral accumulates argon. The blocking temperature varies with both the mineral and the diffusing species, and a grain-size effect, with coarse-grained minerals proving more retentive, is also observed.

From studies on thermal aureoles (e.g. Hart, 1964) and regional metamorphic terrains (Wanless et al., 1970, expressed in terms of the resistance of minerals to rejuvenation) the generally accepted relative order of blocking temperatures is, highest to lowest: Rb-Sr in muscovite; K-Ar in muscovite; Rb-Sr and K-Ar in biotite. Attempts have been made by calculation (Dodson, 1973) and by geological example (Jäger et al., 1967) to give absolute values to the blocking temperatures of minerals. Values suggested by Purdy and Jäger, quoted in Hunziker (1974), are 500°C for Rb-Sr in muscovite, 380°C for K-Ar in muscovite, and 300°C for both Rb-Sr and K-Ar in biotite. The last value is in good agreement with the calculations of Dodson.

Thus cooling ages on metamorphic minerals should yield a minimum age for the last thermal event sufficiently intense to exceed the blocking temperatures and reset the K-Ar and/or Rb-Sr mineral systems. Anomalous K-Ar ages, which cannot be referred to a time of cooling after a thermal event, may be produced in a number of ways, of which

the most important are:

(i) by partial retention of radiogenic argon from an earlier event, termed inherited argon.

(ii) by radiogenic argon incorporated during the formation of a mineral, or introduced by diffusion at a later stage, usually termed extraneous or excess argon, since the apparent age is then greater than the true formation age of the mineral.

E. Cooling age patterns in the Study Area.

1. The area under consideration in this chapter is extended to include the Glockner and Schober Groups, and eastwards through the Kreuzeck and Reisseck Groups as far as the area studied in detail by Brewer (1969) (see Plate 5). In this area the interpretation and linking of radiometric ages with thermalevents is simplified by the abundance of geological and petrological data.

Although anomalous K-Ar ages are of quite common occurrence in the polymetamorphic rocks of the south-east Tauern (e.g. Brewer 1969, 1970, Lambert 1970), normal cooling age patterns, related to specific metamorphic events, may be recognised on the following criteria:

(i) Statistical concentrations of data-points in time.

(ii) Concordance of ages over an appreciable area on the ground.

(iii) Textural homogeneity of samples within this area, for example the existence of equilibrium assemblages of one mineral generation, and evidence for abundant fluids during metamorphism to flush out argon inherited from earlier events.

For the purposes of the following discussions, the data of Oxburgh et al. (1966), Lambert (1970), and those of Brewer from the Kreuzeck and Goldeck Groups (Brewer 1970) are considered together with the new determinations.

The simplest method of displaying statistical concentrations of data-points in time is by the histogram. Three maxima are evident from Fig. 6-1, at 20-30 My, around 80 My, and 260-320 My. From the regional distribution of the data (Plate 5), three age-provinces may be defined:

(i) The ages of 260-320 My, were mostly determined by Brewer on whole-rock phyllites and phyllitic garnet-mica schists from the Goldeck and Southern Kreuzeck Groups.

(ii) The 70-90 My ages occur along the northern edge of the Alt-kristallin Sheet, south of the Tauern Window.

(iii) Ages of the youngest group, 20-30 My, occur within the Tauern Window.

The third criterion, that of the textural nature and homogeneity of the samples, is the most important in relating the radiometric data to the metamorphic events.

(i) The oldest group of ages in the Kreuzeck Group come from the Kreuzeck schists with their dominantly M_1 mineral fabrics. Two M_1 parallel-felted muscovite separates yielded ages close to 300 My. The Kreuzeck Group whole-rocks, some of which contain a younger generation of biotite, scatter towards ages younger than 300 My. Rocks from the Goldeck, particularly those close to the Permo-Triassic unconformity, appear to be less affected by later rejuvenation. Brewer calculated mineral ages from the whole-rock age and muscovite: biotite abundance ratio in the sample (determined by X-ray diffraction) on the assumption that the muscovite preserves an age of about 300 My, while the biotite is progressively rejuvenated, and does not itself lose argon until the biotite is fully rejuvenated at about 100 My. The whole-rock ages then fall into two groups: firstly those with 300 My muscovites and partly rejuvenated biotite, mostly

from the southern part of the Kreuzeck Schists, and the Goldeck Group, and secondly those with partly rejuvenated muscovites and 100 My biotites, mostly from more northern parts. This correlates reasonably well with textural observations on both the dated samples and neighbouring rocks: those in the first group contain biotites which are mostly of the M_1 generation, while in the second group almost all the biotite may be assigned to the cross-cutting M_2 generation.

(ii) The concordant 70-90 My ages occur in Polinik schists of the northern Kreuzeck Group and the westward continuation of the same rocks along strike south of Döllach and into the Schober Group. Ages in this range are also characteristic of rocks from the Millstätter See area to the east (Brewer, 1969) when not affected by large quantities of excess argon. Here the metamorphic grade, textures and fabrics are similar to the Polinik schists. The rocks which yield 70-90 My ages are with few exceptions M_2 assemblages in textural equilibrium, with evidence (see Chapter 4) for the presence of an abundant fluid phase during the M_2 metamorphism.

(iii) Ages of the 20-30 My group are most uniformly developed in the Peripheral Schieferhülle, which is mostly of uniform greenschist to lower amphibolite facies, though parts show evidence of an earlier eclogitic or blueschist phase. Studies by Bickle (1973) and Hawkesworth (1974) have shown this greenschist event to cross into the base of the Altkristallin Sheet, and to be responsible for the approximately 1500 m thick zone of retrograde metamorphism which is M_3 of this work.

2. Age of metamorphism

On the above criteria the mineral age maxima may be correlated with the three texturally defined metamorphic events M_1, M_2 and M_3. Interpreted as cooling ages, the three peaks of the age distribution correspond to times of maximum uplift rate and cooling, and are minimum ages for the respective metamorphic events. It would be useful to know the approximate duration of the events, and the time at which temperatures were greatest. On uplift, the upper parts of a metamorphosed pile will be the first to cool through the mineral blocking temperatures. The closest approximation to the time of maximum temperatures should then be given by minerals from the uppermost parts, and the mineral ages should decrease with increasing structural depth. Another possibility which may arise is that a mineral may grow below its blocking temperature, and subsequently never exceed it, in which case the age recorded would be a formation age, close in time to the metamorphic maximum (see Jäger, 1970). In practice, cases such as these are difficult to substantiate, though some Rb-Sr muscovite ages of 110-120 My from East Alpine rocks west of the Tauern Window are believed to be formation ages (M. Satir, personal communication, 1975).

Unfortunately, in the lower-grade upper levels of metamorphism, conditions for complete resetting of K-Ar clocks are less than ideal, particularly in dry polymetamorphic or igneous rocks where severe textural disequilibrium is usually the case at lower grades of overprinting. Partial outgassing or excess argon phenomena may be displayed in these circumstances and the age pattern must be interpreted with caution.

Little further evidence for the dates of metamorphic maxima is contributed by this study, but the evidence so far available may be summarised:

(i) M_3 event. The most useful piece of evidence concerning M_3 comes from sample SP89, an Altkristallin sheet leucocratic gneiss with an M_3 mineral fabric from about 500 m. below the upper limit of muscovite and feldspar recrystallisation in the retrogressed zone (Bickle, 1973). An Rb-Sr muscovite-albite mineral isochron (Hawkesworth, 1974) yielded an age for the muscovite of 36 ± 1 My which may be regarded as a close approximation to the metamorphic climax here. The K-Ar ages from this and neighbouring samples are anomalous and are discussed later. K-Ar ages in excess of 36 My exist within the Tauern Window, but are all probably explicable in terms of inherited argon. SP506 is a leucocratic gneiss of the Matrei Zone which clearly displays an older generation of muscovite, while some samples from the Granatspitz area (SP614, SP622 of this work), and a biotite and several amphiboles from the Reisseck Group (Oxburgh et al., 1966, Cliff et al., 1971) are all from reworked igneous or metamorphic pre-Alpine basement and may not have been fully outgassed during the Alpine metamorphism (see later discussion).

The highest K-Ar age from the essentially monometamorphic Peripheral Schieferhülle is 34 My from a muscovite near Döllach. Cliff et al. 1971 placed upper and lower limits of 55 and 35 My for the time of Alpine metamorphism in the south-east Tauern Window on the basis of this K-Ar date and the partially outgassed hornblendes from the Reisseck Group.

(ii) M_2 event. There is a tendency for both K-Ar (Brewer, 1970) and Rb-Sr (Hawkesworth, 1976) cooling ages from the M_2 event to increase towards higher levels of the Altkristallin Sheet. K-Ar ages reach 99 My in the Kreuzeck Group (two whole rock, effectively muscovite, ages from near the Nigglai shear zone) and a maximum of 103 My from a biotite in the Goldeck Group. From these, Brewer

deduced the metamorphic maximum to have occurred at about 105 My. Hawkesworth's Rb-Sr mineral ages from Altkristallin sheet rocks east and south-east of the Tauern Window led him to confirm the conclusion of Brewer.

In the map area the pattern in the Intermediate schists is complicated by anomalous ages involving inherited or excess argon (see later discussion).

(iii) M_1 event. Brewer (1970) suggested that the climax occurred between 380 My (on the basis of a Rb-Sr whole-rock determination by Cliff, 1968, later confirmed by an isochron of Hawkesworth (1974), on pre-M_1 leucocratic gneisses) and 320 My (the highest muscovite and whole-rock cooling ages).

F. Interpretation of Anomalous Ages

The interpretation of cooling age patterns rests on the assumption that temperature is the most important factor in determining whether mineral or rock systems are open or closed to the diffusion of radiogenic argon. However, anomalously high K-Ar ages are often found in polymetamorphic rocks where the temperature, deduced independently, should have been considerably above the usually observed mineral blocking temperatures. An example from the literature is a study of mantled gneiss domes from southern Idaho, U.S.A. (Armstrong and Hills, 1967). A polymetamorphic Precambrian basement with a Palaeozoic cover sequence was metamorphosed in late Mesozoic time. While ages in the cover sequence fit a cooling pattern, decreasing systematically with stratigraphic depth, ages in the basement are variable, up to four times as great as ages in the lower parts of the cover. The grade of metamorphism here is in the amphibolite facies. The authors point out the importance of factors other than temperature in this

case, and suggest the presence of a low P_{H_2O} in the basement as a possible reason for the retention of radiogenic argon.

Now the resetting of K-Ar mineral systems in polymetamorphic rocks requires that all radiogenic argon accumulated since previous events is expelled. Among other things this means that the diffusion of argon through the rock (along grain boundaries, or in rock voids if such exist) should be rapid compared with volume diffusion out of minerals, otherwise outgassing of the minerals will be restricted. Since movement of argon through the rock along grain boundaries will be subject to different controls from volume diffusion in minerals, this condition may not always be fulfilled, and anomalous ages will result if not all the pre-existing radiogenic argon succeeds in escaping from the system.

1. Argon as a component of the rock fluid phase: a qualitative model for the treatment of anomalous K-Ar ages

Once out of the mineral, argon can be considered as a component of the intergranular fluid phase, and independent evidence for the behaviour of the fluid phase is likely to be relevant to any treatment of argon movement through rocks. In the example described by Armstrong and Hills (1967), the retention of variable quantities of radiogenic argon, despite the high temperatures, was tentatively ascribed to an insufficiently developed fluid phase. Textural evidence, in the form of local, partial preservation of Precambrian mineral fabrics, suggests that the mobility of many other species besides argon was restricted.

The treatment of anomalous argon phenomena which follows requires some assumptions to be made:

(i) Under certain circumstances the rate of escape of argon from the whole-rock system may be less than that for individual phases, and may determine the argon retention characteristics of the system as a whole.

Equations and data given in Dodson (1973) indicate that at amphibolite facies temperatures the time constant for volume diffusion of radiogenic daughter species becomes geologically significant for systems on the scale of metres or even centimetres. The size of mineral systems is of course a few orders smaller, and the interpretation of K-Ar mineral ages assumes that the argon is lost rapidly from the whole-rock system. Inhibition of diffusion through the rock along grain boundaries could make the escape of argon from large whole-rock systems slow compared to escape by volume diffusion from minerals. This leads to the second principal assumption:

(ii) An abundant fluid phase, with P_{fluids} close to P_{total}, or at least in the form of a continuous intergranular film, is necessary to allow uninhibited diffusion of argon and other species out of the whole-rock system, even when temperatures exceed mineral blocking conditions.

There should be no shortage of fluids in sedimentary rocks undergoing their first prograde metamorphism, but polymetam orphic and igneous rocks are liable to be fluid deficient during subsequent metamorphism unless large quantities of fluid can be supplied from outside. It is in these latter rock types that almost all the major occurrences of anomalous argon distribution occur. Hunziker (1974) observed that the amount of excess argon in polymetamorphic basement rocks in the Western Alps was strongly dependent on the degree of schistosity of the rocks, with schists showing lower ages than massive rocks.

Inhibited diffusion of argon out of rocks in which mineral systems

are open carries the important consequence that the argon concentrations in minerals will tend to equilibrate with each other and with the ambient P_{Ar} in the fluid. This may involve:

(i) retention of argon by minerals which would otherwise outgas,
(ii) diffusion of argon into minerals of low potassium content above their blocking temperatures, and
(iii) occlusion of argon by newly crystallised minerals.

This equilibration can cause the apparent ages to be significantly greater than the true formation ages in cases (ii) and (iii).

If the whole-rock system may be considered closed, the argon will be redistributed between the minerals and the fluid, and will tend towards an equilibrium partitioning. In a partly open system, where argon slowly escapes from the whole rock during the thermal event, different minerals show disequilibrium argon contents because of their different blocking temperatures. In such situations, as the metamorphism of a 'dry' homogeneous basement surmounted by a fluid-rich cover sequence, or a body of igneous rock enclosed in wet sediments in which argon escape is rapid, approximately simultaneous cooling through the blocking temperature of a particular mineral may preserve a record of argon concentration gradients through the rock body, which may then be recognisable in the age pattern. Access of fluids from neighbouring "wet" rocks, or deformation, may promote the diffusion of argon. If these effects are not distributed homogeneously, irregular patterns of concentration gradients will be set up in the rock.

In general, partly open whole-rock systems with an argon concentration gradient are likely to be more common than closed, equilibrium systems.

If equilibrium between minerals has been achieved, a K-Ar isochron will reveal the true age of the system. The construction

and properties of K-Ar isochrons have been dealt with by a number of authors (e.g. Harper, 1970; Roddick and Farrar, 1971) and are further discussed, together with some possible isochron relationships from the study area, later in this chapter. However, demonstrable K-Ar isochrons are rather rare, and disequilibrium situations are more commonly found (see Hunziker, 1974). Since mineral systems close to argon diffusion at different temperatures, and the P_{Ar} and its isotopic composition are likely to vary with time as the rock cools, this is not particularly surprising.

2. Discordance of Rb-Sr and K-Ar mineral ages

Because of the obvious differences in systematics between model K-Ar and Rb-Sr isotopic systems it is possible for the relative blocking temperature relationships to be reversed in cases of restricted diffusion. Strong discordance, involving much greater K-Ar than Rb-Sr mineral ages on the same sample are reported by Besang et al. (1968), Harre et al. (1968), Brewer (1969), and the present study (sample SP89), and are evidently of quite common occurrence in polymetamorphic and meta-igneous rocks.

3. Excess Argon: mechanisms for creating high argon pressures

Large quantities of excess argon have been reported in metamorphic micas in an area adjacent to the present study area (Brewer 1969, 1970).

Excess or extraneous argon may be defined as ^{40}Ar present in a mineral in excess of that which could have formed by radioactive decay of ^{40}K since the crystallisation of the mineral. Two reasons are suggested by Dalrymple and Lanphere (1969) for its incorporation in minerals: either argon was occluded during mineral growth, or it diffused in at a later stage.

In terms of the model discussed above, excess argon may be incorporated in minerals of low potassium content by redistribution in a closed or partly-closed whole-rock system. In minerals of higher K content, however, some special mechanism for concentrating radiogenic argon must usually be invoked, because of the larger quantities involved. The pattern of age data reported by Brewer requires that a very high ^{40}Ar pressure appeared suddenly in the rocks in the middle of their cooling history. Brewer suggested two possible special mechanisms for providing a rapid influx of ^{40}Ar at this relatively late stage. First, the overthrusting of the Altkristallin sheet onto a very old, cold basement shortly before 80 My ago (an explanation somewhat improbable on geological and geophysical grounds), or second, the concentration of ^{40}Ar outgassed from deeper levels in a magmatic fluid, with sudden release of the argon into the rocks as the residual magma crystallised (the alternative preferred by Brewer himself, personal communication, 1974).

The latter is an attractive hypothesis in the area concerned, but since such phenomena are likely to be of uncommon and localised occurrence, a more generally applicable model is needed for the more widespread and varied anomalous age patterns reported here and elsewhere. A general model may be derived by treating argon as a part of the rock fluid phase. In a pile of rock undergoing metamorphism, fluids, together with outgassed radiogenic argon, tend to move upwards under the effect of gravity. If the rocks through which the fluids pass were once dry polymetamorphic or igneous rocks such as the Altkristallin, two significant observations may be made about the cooler, upper levels:

(i) the cooler rocks above may constitute a barrier to diffusion (see also Brewer, 1969).

(ii) the upper levels are characterised by retrograde reactions

yielding assemblages more hydrous than the pre-existing ones.

The second point may be elaborated: removal of H_2O from the fluid phase has the effect of relatively concentrating argon in the remaining fluid. Any buffering of the fluid composition by solid phases (for example, the graphite buffer of French, 1966) will accentuate this effect by removing other important components (such as CO_2, CH_4) as well. The resulting high partial pressure of argon, in a reduced vapour phase whose pressure is probably now considerably less than the total lithostatic pressure, may be expected to act analogously to the inhibited diffusion model discussed above, with similar consequences. Argon may be incorporated in minerals which are above their blocking temperatures, or are undergoing recrystallisation, or the outgassing of older minerals may be restricted.

4. Discussion of Anomalous K-Ar data

A large body of circumstantial evidence can be called in favour of the treatment outlined in sections 1 and 3 above, both from this work and other sources.

Both normal and abnormal age patterns may be conveniently displayed on the profiles (Figs. 6-2 to 6-4) which represent sections through the structural units of the study area at three locations. Locally, height above or below main structural marker horizons (for example the base of the Altkristallin Sheet) may be fairly accurately assessed. This is true for the area of the south Glockner and north Schober groups mapped by Bickle (1973), but for the Döllach area the perpendicular distance NNE or SSW of the window edge has been employed as a measure of the structural height in the uniformly rotated, SSW-dipping units. The section in the map area, Fig. 6-4, is cut by shear zones and affected by F_4 large-scale flexuring,

so that here structural height is not linearly related to distance south of the window edge.

(a) Anomalous ages in the Pennine basement

The lower part of Fig. 6-2 consists of two mineral pairs from the Pennine basement of the Granatspitz Kern and the Riffl Decke. Both muscovites and biotites appear to show a trend of increasing apparent age with depth over the normal Schieferhülle value of around 30 My. This is not consistent with the usually observed trend of decreasing age with depth, nor is there any evidence for an inversion of the thermal gradient, which suggests that the Pennine basement has retained some radiogenic argon from earlier events. The basement consists mainly of igneous rocks recrystallised during the Alpine metamorphism and it may be expected that their fluid content prior to this was low. Under these conditions, argon may not have been able to leave the system rapidly. The two mineral pair ages are consistent with the existence of argon concentration gradients out of the basement, with the biotites recording the argon distribution at a later time than the muscovites. Additional factors, such as the effect of fluids from the Schieferhülle, and the penetrative Alpine deformation, both of which die out downwards in the basement, have probably affected the distribution of argon and the shape of concentration gradients.

Textural evidence indicates that the minerals belong to an Alpine generation, in which case the ^{40}Ar they contain in excess of their true age must have been contributed by the whole-rock system. Four mineral ages are of course too few to preclude a fortuitous pattern, and this situation would repay further investigation.

Hornblende ages from the Reisseck and Hochalm Groups (Cliff and Oxburgh, 1971) show ages which increase with depth in the basement,

Fig. 6-5. A regular distribution of ages with depth is less likely to be achieved in this case since the hornblendes are from meta-basic rocks which are not the most abundant rock type. They may have had different radiogenic argon contents at the time of Alpine heating, or the ambient argon pressure may have been influenced by neighbouring rocks.

Sample A346 is a biotite-amphibolite. Argon from pre-Alpine biotite in the rock may have contributed to the high ^{40}Ar content of the hornblendes by redistribution. The K-Ar isochron age of A346 is 24 $\pm$ 3 My, which is concordant with nearby K-Ar ages on biotites, suggesting that argon ceased to be redistributed at about this time.

The "fossil" concentration gradient hypothesis cannot be tested with the Armstrong and Hills (1967) example because of the lack of detail concerning depth of samples in the basement. The apparently irregular distribution of anomalous ages with distance inside basement outcrop, estimated from their locality map, may reflect either basement topography, or other localised factors such as the influx of fluids, or heterogeneous deformation.

(b) Excess argon associated with the base of the Altkristallin Sheet

The next feature of interest in the age profiles is the striking pulse of high apparent ages associated with the base of the Altkristallin sheet, and visible on all three profiles (Figs. 6-2 to 6-4). Anomalous ages occur in the basal 750 m or so of Altkristallin, and at one locality in the Matrei zone (Lambert, 1970).

The temperature at the base of the Altkristallin Sheet in the Schober Group during the Alpine metamorphism is estimated by calcite-dolomite geothermometry to be $435^{o} \pm 20^{o}C$ (Bickle, 1973) which is fairly close to the probable blocking temperature of muscovite. The

polymetamorphic Altkristallin could then act as an impermeable, cool cap-rock to contain the fluids expelled from the Pennine units during the Alpine metamorphism. These fluids contained argon outgassed from the largely Variscan Pennine basement. More important, however, the basal zone of the Altkristallin exhibits textures indicating strong retrogression of M_2 mica-schists and gneisses to more hydrous assemblages, in particular the chloritisation of biotite and garnet and the breakdown of plagioclase to albite and white mica. Around the time of the peak of Alpine metamorphism these reactions could have removed large quantities of H_2O from the fluid phase, and concentrated argon in the remaining fluid. It may be noted that the highest anomalous age in the northern Schober Group is yielded by SP596, a very graphitic and pyritous phyllonite very close to the base of the Altkristallin sheet. In such a rock the buffering effect of the graphite may have removed carbon-bearing components (CO_2, CO, CH_4) from the fluid in addition to H_2O, allowing the maximum concentration of argon to be achieved.

The muscovites in this basal zone in the N. Schober Group are predominantly of M_3 generation, so they are all affected to some extent by excess argon. For example, SP89 muscovite, which is almost all M_3, over half of it produced by the breakdown of feldspar, yields a K-Ar age of 76 My, although its Rb-Sr age is 36 ± 1.

SP 15 muscovite contains an appreciable amount of chlorite which may have contributed ^{40}Ar to the analysis. To investigate the argon content of other minerals, three chlorites from Altkristallin rocks were analysed (Table 6-3) including one from sample 0217 near the base of the Altkristallin at Söbriach. This last contained 2.3×10^{-6} STPcc/g of radiogenic argon.

In the terms of the model, the argon concentration would reach a maximum at around the time of the metamorphic peak when the retro-

grade reactions were proceeding most rapidly, and may be expected to disperse slowly after this. The absence of biotite from the lower parts of the Altkristallin Sheet makes it impossible here to trace the effect of the high ^{40}Ar pressure with time (as could Brewer, 1970). However, the M_2 generation biotites from higher levels in the Schober Group yield anomalously high ages (of 100 My and 159 My) compared with their coexisting muscovites. It is possible that these biotite ages record the slow upward migration of the high ^{40}Ar pressure through the Altkristallin, argon being trapped in biotites as they cooled through their blocking temperatures some 2 km above the zone of anomalous muscovite ages, although excess argon associated with an earlier event cannot be ruled out.

(c) Anomalous ages in the Central Kreuzeck Group

The third zone of anomalous ages to be explained involves those rocks from the staurolite zone of M_2 metamorphism in the central Kreuzeck Group, where the analysed minerals are of M_2 generation but yield apparent ages in the range 175 - 229 My (i.e. 0255M, B; 0282M, B). Staurolite zone temperatures should have been sufficient to exceed the blocking temperatures of both micas especially sample 0255 which lies close to the sillimanite isograd. The evidence for the nature and distribution of the fluid phase discussed in Chapter 4 is relevant here, in particular that the staurolite-forming reaction, for which there is good textural evidence, consumes H_2O, and that there is ample evidence for the restricted mobility of H_2O and other components from staurolite zone levels upwards during M_2. An argon concentration could then have developed in a similar manner to that proposed for the base of the Altkristallin sheet. Retrograde and water-consuming reactions are spread over a wider zone in the central Kreuzeck Group than at that base of the Altkristallin, and the

argon concentration effect appears to have been correspondingly less pronounced, since no apparent ages in excess of the probable M_1 values of 300 My occur.

(d) Anomalously low ages from the study area

The only mineral age from this study which is lower than indicated by its mineral generation is 45 My on an M_2 biotite from sample K6 in the northern Kreuzeck schists. This mineral is strongly kinked by microfolds of probable F_3 generation (Fig. 2-1) and is marginally somewhat altered to sericitic material. It is possible that it has been affected by proximity to a late dyke, of which there are a number in this neighbourhood.

5. K-Ar Isochrons

(a) Types of isochron

Two types of graphical presentation of K-Ar data have been employed in the literature.

The first involves the plotting of radiogenic ^{40}Ar against ^{40}K or K %. Hayatsu and Carmichael (1970) call these initial argon diagrams. An isochron defined by minerals with no initial argon content passes through the origin, whereas a cogenetic set of minerals with equal concentrations of excess ^{40}Ar but discordant model ages will fall on an isochron whose positive intercept is the initial concentration of excess argon. Good isochrons composed of the same mineral species are seldom achieved, however, because of the low spread of K contents usually found.

The second plot, analogous to the Rb-Sr isochron diagram, employs the ratios $^{40}Ar/^{36}Ar$ against $^{40}K/^{36}Ar$. On this plot, minerals of the same model age define a line of slope proportional to the age

and which intersects the $^{40}K/^{36}Ar$ axis at the value for atmospheric argon. To make use of this graphical method in cases of excess argon it is important that the extraction technique should minimise the atmospheric contamination of the sample. In Bern the line is baked overnight with the sample inside to ensure a low argon blank, and all the ^{36}Ar may then be assumed to have come from the sample (see also Hayatsu and Carmichael, 1970). The calculation of model K-Ar ages assumes that any argon contamination, or initial argon present in the minerals, has present-day atmospheric isotopic ratios, and in igneous or metamorphic rocks this is not necessarily true (Hayatsu and Carmichael, 1970; Roddick and Farrar, 1971). Of course if the contribution of initial argon is small compared with that liberated by subsequent radioactive decay within the mineral, the effect on the model age will be negligible.

The isochron plot may be useful in the case of minerals which have equilibrated with a fluid phase with a high P_{Ar}. At any particular time, the $^{40}Ar/^{36}Ar$ ratio of the fluid phase argon may be constant over a considerable volume of rock. Minerals which become closed systems at approximately the same time would then fall on an isochron whose intercept defines the $^{40}Ar/^{36}Ar$ ratio of the initial argon atmosphere. To satisfy the isochron it is not necessary that each mineral take up the same concentration of argon. A better spread of data points from minerals of similar K content may also be obtained by this method of plotting.

The most important criteria to be satisfied in order to obtain isochrons from investigated rock samples are therefore proximity of rock samples (in a volume of rock which may have contained an argon atmosphere of constant isotopic composition) and approximate synchroneity of closure of radiometric systems (in practice this would mean either a situation of rapid cooling, or minerals of the

same type). Mineral pairs from the same rock satisfy the first criterion, but the time taken for the rock to cool slowly through the interval between the mineral blocking temperatures, where these differ appreciably, may well have allowed the isotopic composition of argon in the rock fluid to change substantially.

(b) Interpretation of data

The analytical data from this work have been plotted on both initial argon and isochron diagrams. No isochron relationships due to equal concentrations of excess argon are evident from the initial argon diagrams (Figs. 6-6, 6-7) but some features of the isochron plots (Figs. 6-8, 6-9) merit further discussion.

Two-point isochrons from coexisting mineral pairs should be interpreted with caution, for the reasons outlined above. However, the muscovite and biotite of sample SP614 yield an isochron age of 30 ± 4 My with an initial $^{40}Ar/^{36}Ar$ of 1450 ± 250, which could reflect equilibration with an argon-bearing fluid of this composition at around 30 My ago, during cooling from the Alpine metamorphism. This is not necessarily inconsistent with the "fossil" concentration gradient hypothesis for the Granatspitz samples, provided the isotopic composition of the argon remained constant.

On the other hand, mica pairs from samples SP708 and SP719 yielded isochron ages of 57 and 52 My, with initial argon ratios of 840 and 2550 respectively. There is no independent evidence for an equilibration at these times. Moreover at the apparent isochron age, the biotites of each pair would have contained three times as much excess argon as the muscovites. Since there is no evidence to suggest that the equilibrium partition of argon between the two micas should be as unequal as this, it is not very likely that these "isochrons" represent real events. Many of the other mineral pairs

from the study area yield "isochrons" of negative slope or negative intercept indicating some kind of disequilibrium.

Of more interest are the isochrons made up mostly of minerals of the same species, from the same geographical area. Two such isochrons may be constructed through data from the Kreuzeck Group.

The first (Fig. 6-10) involves the four southernmost samples, from the area around Dechant (D9) i.e. K6, K7, 366RCW, 0280. Except for K6B which has clearly been affected by a much later event, all the analysed minerals, i.e. four muscovites and a biotite, may be fitted to an isochron of 80.8 ± 3.4 My with an initial $^{40}Ar/^{36}Ar$ of 6750. This suggests that at around 81 My ago the argon atmosphere with which the minerals equilibrated was of approximately constant isotopic composition over an area some 3 km or more broad. The proximity of sample 0282M to the isochron suggests that at this locality the composition of the ambient P_{Ar} was similar when this mineral became closed to argon diffusion. The 81 My age is of course consistent with the normal cooling ages from M_2-metamorphosed rocks.

The second example (Fig. 6-11) comes from the Källukerl zone (D4). The three Källukerl zone muscovites 0204M, 0206M and 0125M may be fitted on an isochron together with 0204 chlorite, (assuming a negligible K content for the latter) which gives an age of 71 ± 3 My with an initial argon ratio of 1531. The presence of a small amount of K in the chlorite would raise the age a little without destroying the isochron. 0125 biotite also falls close to the isochron, but not within error. Once again the isochron age is broadly consistent with the M_2 cooling ages.

G. Amphibole Ages

Four hornblende mineral ages were determined from the Kreuzeck Group, three from amphibolites (0213 from north of the Launsberg shear zone, 07 from Ebeneck (D1), 0260 from Eisenriegel (C6)), texturally of M_2 generation, and one from an eclogitic amphibolite, of pre-M_2 generation, sample 028. The apparent ages scatter widely, and are not concordant with any of the mineral age maxima.

However, if the presence of relatively small amounts of initial argon is assumed, with concentrations and isotopic ratios of the same order as those contained in the analysed chlorites, and suggested by the isochrons, then the hornblende ages (with the exception of 028H) may be made concordant with the 80 My ages obtained from other M_2 minerals.

The eclogitic amphibole 028H would require the concentration and $^{40}Ar/^{36}Ar$ ratio of initial argon to be unreasonably high compared with other rocks from the Polinik Schists if its true age were also around 80 My. It is therefore probable that 028H records a pre-M_2 event. Since the rock is texturally unaffected by M_2, the argon systematics may also have remained intact; but the meaning of the 397My apparent age is uncertain.

H Age of the Kleinblässen shear zone

A whole-rock K-Ar age of 61 My is recorded on a mylonite from the Kleinblässen shear zone. The rock consists of very fine-grained, predominantly recrystallised white mica and quartz, with a little chlorite, and the well-oriented fine-grained and finely banded texture of this assemblage suggests that it was produced at the time of the shearing. There is no reason why reliable dates should not be obtained on material of this kind, but the validity of the 61 My age could only be confirmed by the analysis of further samples from this

shear zone, in the hope that they should be concordant, or fall on an isochron.

1 Ages from late dykes

An attempt was made to determine the age of the suite of late dykes (see Chapter 2 and Wright, 1973; Exner, 1955; Holzer and Scharbert, 1958) by the K-Ar method, on minerals separated from material collected by R.C. Wright. A biotite from sample 731RCW (from a dyke on the ridge-top 600 m NW of Stawipfel, Plate 2) yielded anapparent age of 32 My, which compares well with K-Ar ages of 31, 31, 31.5 and 44 My determined on biotites from the Wöllaner Kopf granodiorite stock (unpublished results of Oxburgh). However, a Rb-Sr age of 50.5 ± 3.3 for this biotite was obtained by Hawkesworth (1974) from a two-point biotite-feldspar isochron.

Hornblendes from two other dykes in the Grakofel Area (Plate 2), samples 624RCW and 697RCW, yielded apparent ages of 80 and 38 My respectively. The possibility that the late magmatic fluids responsible for the widespread hydrothermal alteration of the dykes may have upset the argon systematics, by introducing argon of abnormal isotopic composition, makes these apparent ages difficult to interpret. It is difficult to suggest a mechanism, other than a subsequent outgassing of a mineral, to allow the true age to be greater than the lowest apparent age, so 32 My is probably a good indication of the age of at least the hydrothermal alteration of dyke 731RCW.

The possibility that more than one period of late dyke intrusion has occurred in the Kreuzeck Group cannot be ruled out.

J. Summary of Geochronological Evidence

On the basis of both new and old age determinations, the three groups of cooling ages at c. 300 My, 80 My and 30 My are correlated with the three metamorphic events, M_1, M_2 and M_3 respectively.

The age of each metamorphic maximum will be somewhat earlier than the age distribution maximum.

Ages which do not fall into the 3-phase cooling age pattern are regarded as anomalous, and containing inherited or excess radiogenic argon. These are principally found along the margin of the Tauern Window and in the central Kreuzeck Group.

Anomalous ages may be qualitatively understood if it is assumed:

(i) that in certain circumstances the rate of escape of argon from the whole-rock system (on the scale of metres or larger) may be slow compared with escape from millimetre-scale mineral systems.

(ii) that rapid diffusion in whole-rock systems requires the presence of abundant fluids, and that assumption (i) is applicable to fluid-deficient rocks.

There are corollaries to this approach which can be tested, with evidence both from the literature and from this study:-

(i) redistribution, and possibly equilibration, of argon will occur between minerals which are open to diffusion within a larger rock system which is closed, or partly closed.

(ii) Growth of new minerals in a partly-closed whole-rock system leads to occlusion of argon, yielding an apparent age greater than the true formation age.

(iii) The closure of mineral systems in rock systems which were partially open may allow the recognition of "fossilized" concentration gradients of argon out of bodies of fluid-deficient rock. Patterns suggestive of preserved concentration gradients occur in the Granatspitz

basement within the Tauern Window.

The occurrence of large quantities of excess argon in K-rich minerals requires a special concentration mechanism. The hypothesis of Brewer (1969, 1970) for the Millstätter See area (to the East of the area covered by Plate 5) is not challenged, but an alternative explanation is suggested for the excess argon associated with the base of the Altkristallin Sheet along the southern margin of the Tauern Window. Here, the fluids, generated by dehydration and decarbonation reactions in the Peripheral Schieferhülle during the Alpine (M_3) metamorphism, and containing radiogenic argon expelled from the Pennine basement, passed up into the base of the Altkristallin, where they were rapidly diminished in quantity by retrograde reactions in the basal 1.5 km of the Altkristallin sheet. Argon was thus relatively concentrated in the fluid phase, and passed into, or restricted diffusion out of, muscovites which were still above their blocking temperature.

Hydration reactions are also a characteristic feature of the staurolite and lower zones of M_2 metamorphism in the central Kreuzeck Group, and a similar concentration mechanism, affecting argon outgassed from lower levels of the Altkristallin during M_2, and transported upwards with the metamorphic fluids, is believed to have operated here, giving rise to the second principal zone of anomalous ages.

Andesitic dykes which cut the Altkristallin of the Kreuzeck Group were intruded at least 32 My ago. The age pattern of the dykes is complicated by their hydrothermal alteration and the discordance of K-Ar and Rb-Sr ages.

CHAPTER 7

Geological Synthesis

The main purposes of this chapter is to compare the evidence for early Alpine events in the map area with that from elsewhere in the Alps, and to attempt a synthesis of early Alpine events in the Eastern Alps. The problem of the detachment of the Altkristallin thrust sheet from its lower crust and mantle is also discussed.

A. Generalised conclusions from the map area

The structural, metamorphic and geochronological conclusions have been summarised at the end of Chapters 3, 5 and 6 respectively, and need not be repeated in full. The principal, general conclusion is the recognition of a metamorphic event, M_2, characterised by high temperatures and relatively low pressures, which affects the lower parts of the Altkristallin sheet in the map area, and which occurred during the Cretaceous, yielding late-Cretaceous cooling ages of 70-100 My. This event occurred before the overthrusting of the Altkristallin sheet over the Pennine units within the Tauern Window (see e.g. Brewer, 1970). A deformation event, F_2, which is possibly associated with the M_2 metamorphism, increases in intensity downwards in the Altkristallin, and displays more ductile deformation with increasing M_2 metamorphic grade.

B. Early Alpine effects elsewhere in the Altkristallin sheet

Well-developed early Alpine metamorphic effects occur in the "Schneeberger Zug", south of the Ötztal and Stubai Alps to the west of the Tauern Window. (For geology, see Schmidt, 1965; for geochronology, Schmidt et al., 1967; Miller et al., 1967; and Harre et

al., 1968). Here the metamorphic grade reaches amphibolite facies, with temperatures probably in excess of 600°C. (Oxygen isotope fractionations of H. Friedrichsen, personal communication, 1975; breakdown of paragonite in the presence of quartz, F. Purtscheller, personal communication, and Chatterjee, 1972). Cooling ages of around 80 My on micas are typical of the Schneeberger Zug and some neighbouring areas, including a biotite age of 77 My from the Brenner Mesozoics. An unconformity of Permian age is preserved in the garnet grade of early alpine metamorphism at Egeten Joch (Purtscheller, personal communication). The Schneeberger Zug has an ESE-WSW-trending synformal structure which involves Mesozoic rocks, at least at the Eastern end, but the effect of the early alpine metamorphism dies out northwards and southwards into the surrounding Altkristallin. Very little deformation is associated with the recrystallisation. At the Egeten Joch the unconformable Permo-Trias has an intermittently developed foliation, and some strained pebbles, but is essentially unaffected by folding. The new mineral growth in the intensely folded Altkristallin below the unconformity (these structures are thus at least pre-Permian) is randomly oriented. This is clearly different from the map area, where at lower levels the early Alpine minerals are oriented axial-planar to a major post-M_1 (Hercynian) deformation. The zone of highest temperatures approximately coincides with the Schneeberger Zug outcrop, but is displaced slightly towards the southern side of the synformal axis. Thus it appears that an antiformal distribution of early Alpine isotherms was superimposed on a synformal structure.

The preservation of this synformal structure of the Schneeberger Zug is not consistent with uplift of the area, after the early Alpine recrystallisation, relative to the Altkristallin to north and south, making it unlikely that the antiformal thermal structure is produced

by differential vertical movements. It is more probable that the Schneeberg recrystallisation occurred in an E-W or ENE-WSW trending zone of high early Alpine thermal gradients, with zones of lower thermal gradient to north and south.

South of the western end of the Tauern Window, the Altkristallin Sheet has a narrow outcrop and is disturbed by E-W trending tectonic lines, and by intrusions of the Periadriatic igneous suite (Borsi et al., 1973). Although some minerals, mostly muscovites, record early Alpine dates, biotites often show late Alpine ages so that the zone of c. 300 My ages abuts directly onto the zone of 20-30 My ages. Apparently the effects of the late Alpine metamorphism in the lower parts of the Altkristallin are more strongly developed here than they are in the map area.

In the Deferegger Alpen, further to the East (described by Senarclens-Grancy, 1965), no mineral age determinations are currently available. A visit to the Lasörling area (L on Fig. 7-1) by the present writer in 1975 revealed rocks which are garnet-grade schists texturally and petrologically similar to the Kreuzeck schists, with a weak, later overprint of porphyroblastic albite and chlorite, down to within 2.5 km of the base of the Altkristallin.

In the Schober Group, lithologies and textures are comparable with the Polinik schists, being coarser-grained, with staurolite and kyanite in pelitic compositions (Troll and Hölzl, 1974; Bickle, 1973). Preliminary results from a Rb-Sr geochronological study in the central Schober Group (Troll, personal communication, 1975) indicate early Alpine ages for micas.

Immediately to the east of the Tauern Window, Hawkesworth (in press, 1976) reports Rb-Sr muscovite and biotite ages of 90-100 My less than 1 km below the base of the Mesozoic sediments of Stangalm, at Innerkrems (I on Fig. 7-1).

Concerning the Altkristallin rocks of Steirmark and East Kärnten, there is an abundance of conflicting interpretation, which need not be discussed in detail here; (for one viewpoint, see Metz, 1966). Detailed geochronol ogical evidence is not yet published. However, the Glein-, Kor-and Sau-alpe areas are characterised by early Alpine mineral ages (S. Scharbert and W. Morauf, personal communication). The gneisses of the Seckauer Tauern to the north, show early Alpine biotite but muscovites retain pre-Alpine ages.

The distribution of dated early Alpine mineral assemblages in the Altkristallin sheet is summarised on Fig. 7-1.

C. Regional character of the early Alpine metamorphism in the Altkristallin Sheet

Although some features of the distribution of early Alpine metamorphism (Fig. 7-1) are due to variable erosion level in the Altkristallin, evidence such as that from the Schneeberger Zug, and the variation in thermal gradient in the map area, suggest that the early Alpine metamorphism may be concentrated in an east-west trending belt in the central zone of the Eastern Alps, the effects of which locally reached up into the Mesozoic sedimentary cover.

Overthrusting has destroyed much useful evidence, by removing the lower levels of the Altkristallin sheet, and in the map area we are indeed fortunate to be able to see so deep into an area of high early Alpine geothermal gradient, unaffected by later Alpine structures or metamorphism. The high thermal gradients necessary for an andalusite-sillimanite succession require the input of heat from below (Oxburgh and Turcotte, 1971). This probably happens by a convective process, involving either intrusion of igneous material at high levels, or circulating fluids.

In the map area, the late pegmatites, though themselves volu-

metrically insignificant as a heat source, are evidence that circulating fluids may have been responsible for the transport of heat. Partly in contradiction to this, however, is the evidence, particularly from parts of the Polinik schists, that the rock fluid composition was locally buffered by the mineral assemblage. Possibly fluids did not flow uniformly through the rocks, but the movement was concentrated into "channels", with "stagnant volumes" in between, which allowed local buffering in appropriate assemblages. It is notable that possible fluid-buffering assemblages are much less common in the Eisen and Tröger Alm area which is characterised by a higher temperature : pressure ratio and more abundant pegmatites than the Polinik Schists.

The somewhat patchy distribution of early Alpine metamorphism in the Altkristallin, with evidence for horizontal temperature gradients, suggests a metamorphic belt which is dependent on the convective transfer of heat.

D. The high-pressure phase of early Alpine metamorphism

Early Alpine events in the Western Alps are of a different character. A metamorphism earlier than the main (Lepontine) event, and characterised by minerals indicating high pressures, was recognised by Bearth (1962), mostly within Pennine units of the Western Alps. In the Sesia-Lanzo zone (Dal Piaz et al., 1972) this early high-pressure phase has been dated, with 60-90 My ages on phengites, with both K-Ar and Rb-Sr methods. Glaucophanes from the Mesozoic Piemont ophiolites give K-Ar ages of 80-100 My (Hunziker, 1974; Bocquet et al., 1974).

In the Central Alps, to the east of the Lepontine metamorphic culmination, there is further evidence for an early, high-pressure metamorphic event, with mineral ages in the range 60-80 My (Frey et

al., 1974, Jäger, 1973).

In the Tauern Window, a zone of relict eclogitic and glaucophane-bearing rocks within the Peripheral Schieferhülle, best displayed south of the Grossvenediger, indicate an early growth of minerals under high-pressure conditions (e.g. Miller, 1974). The probable coexistence here of eclogitic assemblages with a free fluid phase implies very high pressures (T.J.B. Holland, personal communication). These rocks are undated, but probably belong to the same early Alpine phase, which, in various segments of the Alpine chain (Dal Piaz et al., 1972; Ernst, 1973; Hunziker, 1974; Frey et al., 1974) is interpreted as being due to metamorphism in a subduction zone.

E. Possible paired metamorphic belts of Cretaceous age

The occurrences described in the previous section appear to constitute a belt of high-pressure metamorphism of similar age to the high-temperature M_2 event. The geograhical relationship of the two belts has been altered by the overthrusting of East Alpine units over Pennine units, and much of the evidence to connect the two metamorphic events has been obliterated by subsequent metamorphism, uplift and erosion. Minimum allowances for the amount of overthrusting place the andalusite-sillimanite progression at least 55 km south of the eclogite and blueschist occurrences within the Tauern Window, (55 km is the width of the Window at its Eastern end), although the position of the latter relative to the Pennine basement which they now overlie was probably also changed by overthrusting.

The contrasted metamorphic character of the two belts, the similarity in age, and their probable original spatial relationship suggest that these are paired metamorphic belts similar to those described by Miyashiro (1961) and interpreted as forming in an island arc and trench environment (Oxburgh and Turcotte 1971).

The principal difference between the metamorphism in the Altkristallin Sheet and more typical high-temperature low-pressure belts described by Miyashiro is the scarcity of igneous rocks in the former. There is no evidence for large-scale igneous intrusions directly associated with the M_2 event, except for that postulated at greater depth by Brewer (1970) to explain the distribution of excess argon south-east of the Tauern Window.

Very little is known about the factors affecting magma generation above subduction zones. If the generation of large volumes of intermediate and acid magma, and its subsequent emplacement at higher levels, requires a significant length of time from the onset of subduction (say, of the order of 30-40 My), interpretation of M_2 metamorphism as the high-temperature member of a paired association is valid, provided that subduction beneath the Altkristallin sheet was relatively short-lived. The maximum possible duration of subduction in the Eastern Alps might be estimated if the range of formation ages of ophiolitic fragments in the Pennine Zone (which are believed to represent pieces of subducted oceanic crust, Bickle and Pearce, 1975) were known. Although ages up to 180 My have been recorded on ophiolites in the French Alps (Frey et al., 1974, and references), no data are available for the Eastern Alps.

Another difference between the Alps and paired metamorphic belts of Miyashiro type is that in the former, the high- and low-temperature belts are separated by a low-angle overthrust fault, in contrast to the major, high-angle fault which characterises the better-preserved examples of the circum-Pacific region.

A plate-tectonic model to describe the Mesozoic and Tertiary history of the Eastern Alps in terms of a paired metamorphic belt with subsequent continental collision, taking into account the sedimentary, structural and metamorphic evidence from the Eastern

and Central Alps, has been presented elsewhere (Hawkesworth et al., 1975). Fig. 7-2 shows the suggested reconstruction of the central Eastern Alps from the Cretaceous to the present day.

F. Development of the Altkristallin thrust sheet

The presently exposed base of the Altkristallin sheet at the southern margin of the Tauern Window represents a crustal level approximately 22 km deep at the time of M_2 metamorphism. This thickness is derived from metamorphic evidence (see Chapter 5). This represents only about two-thirds of continental crust of average thickness, (a "flake" as defined by Oxburgh, 1972), and a mechanism for removing the remaining 10-15 km, together with the lower lithosphere, must be incorporated in any model for the development of the Eastern Alps.

A possible mechanism is outlined by Armstrong and Dick (1974) who suggest that in regions of elevated geothermal gradient, such as may be found behind volcanic arcs, sheets of crystalline rock several kilometers thick may become detached from underlying material which has become thermally softened. The dislocation will occur approximately parallel to an isotherm, which probably marks the transition between brittle and ductile behaviour of the crystalline rocks.

With regard to the map area, two principal alternative hypotheses for the time and manner of removing the base of the Altkristallin present themselves.

(i) Detachment occurred in the manner suggested by Armstrong and Dick, at an early stage, when temperatures at this level in the Altkristallin were high. At high temperatures, and low rock strengths, one might expect the zone of deformation in the lower Altkristallin to be broad, and partly preserved today in the detached sheet. If the suggested link between the F_2 deformation and M_2 metamorphism is

accepted, then it is possible that the F_2 deformation is connected with the removal of the lower Altkristallin. Deformations correlated with F_2 are recognised both in the Schober Group (Bickle, 1973) and east of the Tauern Window (Hawkesworth, 1974). Later localisation of the deformation along a narrow zone, as suggested by Armstrong and Dick, may explain why the F_2 deformation higher in the Altkristallin did not outlast the M_2 metamorphism. However, the origin of the tectonic stresses required to produce F_2 deformation at this early, pre-M_2 stage is unknown.

(ii) Detachment occurred at a later stage, after the M_2 metamorphism, and perhaps at the onset of continental collision, immediately before overthrusting over the Pennine domain. The absence of a penetrative deformation in the lower parts of the exposed Altkristallin sheet in this interval (F_3 is correlated, through the Sadnig and Schober Groups, with deformations accompanying overthrusting, Bickle, 1973) requires that the deformation was restricted to a narrow, subhorizontal shear zone. This alternative is mechanically similar to the "flake" of Oxburgh (1972), although the subduction polarity, according to the interpretation of M_2 as a behind-trench high temperature metamorphism, is reversed.

The first explanation has the advantage of accounting for the occurrence and distribution of the F_2 deformation in the map area, but fails to provide a compelling reason for the existence of a stress field at that time. It is possible that a combination of the two hypotheses provides the most plausible explanation: that while most F_2 deformation was penetrative, later stages produced a more localised dislocation which either allowed slow and continuous movement until, or was reactivated at, the onset of continental collision. The lower lithosphere was then displaced towards the south by the underthrust, northern continental mass. The displaced material may still have been

sufficiently hot to have deformed in a plastic manner.

The continental material removed (it is assumed that any mantle material also displaced can be accommodated by flow within the mantle) must be accommodated below the Southern Alps. Geophysical data from Angenheister et al. (1972) indicate that the Mohorovicic discontinuity reaches a maximum depth under the central Eastern Alps of c. 52 km at a position just south of the Tauern Window, and that the depth to the Moho under the Southern Alps is c. 40 km in contrast to 30 km to the north of the Alps. The present extra thickness of continental crust in the south may be enough to have accommodated that removed from below the Altkristallin sheet, if the original thickness of the crust south of the Alps was similar to that to the north.

The ideas discussed above are points raised by considering the implications of a high-temperature metamorphic event in the East Alpine units. Clearly the difficulties of interpretation are great, but the evaluation of the significance of the early Alpine metamorphism will probably prove critical to our understanding of the Alpine orogeny.

APPENDIX 1

Mineral Analyses

A. Techniques

The analyses were performed on the Geoscan MK.11 electron microprobe in the Department of Mineralogy and Petrology, Cambridge, under the supervision of Dr. J.V.P. Long and Mr. G.G.Price. Specimen preparation and analysis correction techniques are those described by Sweatman and Long (1969).

Standards used were:-

Fe - pure iron

Mg - periclase

Al - corundum

Si and Ca - wollastonite

Mn - pure manganese

Ti - pure titanium

K - orthoclase

Na - jadeite

Zn - pure zinc

Co - pure cobalt

All analyses were performed at 20 Kv. The speciment current was varied, since some minerals, e.g. particularly carbonates, and to a lesser extent plagioclase and micas, tend to decompose in the electron beam, which particularly affects the determination of alkali elements. A 'safe' specimen current was selected which showed no fall-off in the count-rate for K or Na on continuous analysis over 2-3 minutes. However, some of the micas in samples 033 and 034 analysed with the high specimen current (see below) showed an approximately linear fall-off of K and Na peaks. The count-rate for these samples was corrected by extrapolating

back to zero time. There is then no systematic difference between the corrected analyses and those which displayed no peak fall-off.

Specimen currents used were:

4×10^{-8} amps for most of sample 033 and 034 minerals

3×10^{-8} amps for sample 0462, 0479 minerals

2.5×10^{-8} amps for plagioclases

2×10^{-8} amps for carbonates

4 x 10 second counting periods were used for most silicates and 8 x 10 seconds for Mg in carbonates.

Apparent concentrations were calculated from raw count data using a computer program written by R.Powell, which also calculates the standard error on the apparent concentration from the counting statistics for each spot analysis. These errors can be used to estimate the precision of the analyses, which varies with both the atomic number and the concentration of the element in question. In Fig. 8-1 the maximum 2 error has been plotted against apparent concentration for several elements.

With the exception of the ilmenites and carbonates, the analyses have been recalculated using the computer program RECALC, written by R. Powell, to which I have added further mineral subroutines and end-member calculations.

B. Descriptions of analysed samples

033. Pelitic Polinik Schist

Locality: on Polinik path, at 2450 m altitude on N side of Mörningtal, 575 m NE of Polinik (C3).

Assemblage: ky - st - gt - mu - bi - qz - plag

The rock is an even-grained ky - st - gt - pelitic schist (qz and plag 0.33 m diameter, mica laths 1 mm long) with fairly strong preferred orientation and with scattered "knots" of

st + ky c. 1.5 cm across.

034. Pelitic Polinik Schist

Locality: same as 033

Assemblage: ky - st - gt - mu - bi - qz - plag

Medium-grained pelitic schist with common euhedral staurolite (3 mm), garnet (2 mm), and kyanite set in 1 mm long mu and bi laths with strong preferred orientation, qz and plag. (0.3 mm). See Fig. 2-4.

0174. Dolomitic marble

Locality: 2300 m alt. on E side of Geistlacke corrie, 350 mm E of Geistlacke (F3).

Assemblage: dol - cc

Mean grain size c. 0.7 mm. Dolomite (c. 80%) in rounded grains, often convex against neighbouring calcite. Calcite (20%) sometimes has dusty appearance.

0462. Pelitic Intermediate Schist

Locality: c. 2600 m alt. on ridge ENE of Strieden Kopf, 100 m N of spot height 2639 m. (B7).

Assemblage: mu - bi - qz - and - gt - st - plag - chl

Gt small, 0.55 mm; st euhedral, to 4 mm; and in large (3 cm or more in hand specimen) porphyroblasts; sequence of growth indicated by gt included in st included in and; mu + bi in broad flakes, 0.5 x 0.2 mm. Rock cut by tourrich veins. Accessory ilm.

0479. Sillimanite-biotite Schist

Locality: 150 m ENE of spot height 2309 m, on Penker Tröger Alm (C5).

Assemblage: qz - plag - bi - mu - sill - ore

0.5 mm grains of qz and lamellar-twinned oligoclase, with broad 0.5 mm flakes of bi and mu, weak preferred orientation; sill occurs in knots up to 1 cm across associated with bi and qz, and often also some ore. Coarser prismatic sill occurs, usually near centres of knots.

0524. Pelitic Polinik Schist

Locality: same as 033

Assemblage: ky - st - gt - mu - bi - qz - plag

Qz and plag 0.33 mm diam., micas c. 1 mm long, strongly oriented in the foliation; gt plentiful, to 3 mm, often hollow cored; euhedral st, up to 3 mm; some large ky blades up to 6 mm across, many more smaller ones. Accessory ilm and tour.

TABLE 8-1

Mineral analyses from pelitic schists

Note: Fe has been calculated as FeO for muscovite, biotite, and staurolite; and as Fe_2O_3 for Al silicates. For plagioclase and chlorite, an arbitrary, fixed fraction of Fe_{total} (0.5 and 0.13 respectively) has been calculated as Fe_2O_3. For garnets, Fe^{3+} (for andradite molecule) has been apportioned on a site-occupancy basis (Si + Al + Fe^{3+} = 5).

Garnet and plagioclase analyses are rim compositions except where expressly stated.

034 GARNET 1	
ØXIDE	WT %
SIØ2	37.83
AL2Ø3	20.94
TIØ2	0.00
FE2Ø3	0.14
FEØ	36.16
MGØ	3.21
MNØ	2.15
CAØ	0.56
NA2Ø	0.00
K2Ø	0.00
TØTAL	101.01

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.020
AL	
AL	1.971
TI	
FE3+	0.008
FE2+	2.414
MG	0.382
MN	0.146
CA	0.048
NA	
K	

ENDMEMBERS	
% PYRØPE	12.66
% ALM	79.90
% SPESS	4.82
% ANDR	0.42
% GRØSS	1.18
% M/M+F	13.68

034 GARNET 2	
ØXIDE	WT %
SIØ2	37.85
AL2Ø3	20.96
TIØ2	0.00
FE2Ø3	0.00
FEØ	36.36
MGØ	3.46
MNØ	1.55
CAØ	0.43
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.61

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.026
AL	
AL	1.975
TI	
FE3+	
FE2+	2.431
MG	0.412
MN	0.105
CA	0.037
NA	
K	

ENDMEMBERS	
% PYRØPE	13.63
% ALM	80.44
% SPESS	3.47
% ANDR	0.00
% GRØSS	1.22
% M/M+F	14.48

034 GARNET 3	
ØXIDE	WT %
SIØ2	37.42
AL2Ø3	21.19
TIØ2	0.00
FE2Ø3	0.00
FEØ	34.60
MGØ	3.15
MNØ	2.76
CAØ	0.77
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.89

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.012
AL	
AL	2.010
TI	
FE3+	
FE2+	2.329
MG	0.378
MN	0.188
CA	0.066
NA	
K	

ENDMEMBERS	
% PYRØPE	12.60
% ALM	77.62
% SPESS	6.28
% ANDR	0.00
% GRØSS	2.22
% M/M+F	13.96

034 GARNET 4	
ØXIDE	WT %
SIØ2	37.37
AL2Ø3	20.94
TIØ2	0.00
FE2Ø3	0.36
FEØ	35.69
MGØ	3.10
MNØ	2.55
CAØ	0.76
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.78

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.997
AL	0.003
AL	1.977
TI	
FE3+	0.022
FE2+	2.394
MG	0.371
MN	0.173
CA	0.065
NA	
K	

ENDMEMBERS	
% PYRØPE	12.33
% ALM	79.64
% SPESS	5.75
% ANDR	1.10
% GRØSS	1.08
% M/M+F	13.41

034 GARNET 5	
ØXIDE	WT %
SIØ2	36.94
AL2Ø3	21.13
TIØ2	0.00
FE2Ø3	0.30
FEØ	36.10
MGØ	3.41
MNØ	1.91
CAØ	0.49
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.28

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.975
AL	0.025
AL	1.981
TI	
FE3+	0.018
FE2+	2.431
MG	0.410
MN	0.131
CA	0.042
NA	
K	

ENDMEMBERS	
% PYRØPE	13.48
% ALM	79.99
% SPESS	4.30
% ANDR	0.90
% GRØSS	0.50
% M/M+F	14.42

034 GARNET 6	
ØXIDE	WT %
SIØ2	37.45
AL2Ø3	20.80
TIØ2	0.00
FE2Ø3	0.17
FEØ	35.49
MGØ	3.40
MNØ	1.89
CAØ	0.77
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.98

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.015
AL	
AL	1.974
TI	
FE3+	0.010
FE2+	2.389
MG	0.408
MN	0.129
CA	0.066
NA	
K	

ENDMEMBERS	
% PYRØPE	13.54
% ALM	79.20
% SPESS	4.28
% ANDR	0.50
% GRØSS	1.70
% M/M+F	14.60

034 GARNET 7	
ØXIDE	WT %
SIØ2	37.60
AL2Ø3	20.88
TIØ2	0.00
FE2Ø3	0.03
FEØ	36.13
MGØ	3.39
MNØ	1.62
CAØ	0.54
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.20

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.021
AL	
AL	1.977
TI	
FE3+	0.002
FE2+	2.427
MG	0.406
MN	0.110
CA	0.047
NA	
K	

ENDMEMBERS	
% PYRØPE	13.44
% ALM	80.33
% SPESS	3.65
% ANDR	0.10
% GRØSS	1.44
% M/M+F	14.33

034 GARNET 8	
ØXIDE	WT %
SIØ2	36.63
AL2Ø3	20.47
TIØ2	0.00
FE2Ø3	0.77
FEØ	35.10
MGØ	4.24
MNØ	0.88
CAØ	0.50
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.60

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.984
AL	0.016
AL	1.951
TI	
FE3+	0.047
FE2+	2.391
MG	0.515
MN	0.060
CA	0.044
NA	
K	

ENDMEMBERS	
% PYRØPE	17.01
% ALM	79.01
% SPESS	2.00
% ANDR	1.46
% GRØSS	0.00
% M/M+F	17.71

034 GARNET 9	
ØXIDE	WT %
SIØ2	37.06
AL2Ø3	20.59
TIØ2	0.00
FE2Ø3	0.78
FEØ	35.26
MGØ	3.99
MNØ	1.46
CAØ	0.56
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.69

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.991
AL	0.009
AL	1.950
TI	
FE3+	0.048
FE2+	2.380
MG	0.480
MN	0.100
CA	0.048
NA	
K	

ENDMEMBERS	
% PYRØPE	15.90
% ALM	78.91
% SPESS	3.30
% ANDR	1.60
% GRØSS	0.00
% M/M+F	16.77

033 GARNET 1	
ØXIDE	WT %
SIØ2	36.91
AL2Ø3	20.69
TIØ2	0.00
FE2Ø3	0.61
FEØ	35.26
MGØ	3.61
MNØ	2.05
CAØ	0.53
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.67

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.987
AL	0.013
AL	1.961
TI	
FE3+	0.037
FE2+	2.387
MG	0.435
MN	0.141
CA	0.046
NA	
K	

ENDMEMBERS	
% PYRØPE	14.41
% ALM	78.99
% SPESS	4.66
% ANDR	1.52
% GRØSS	0.00
% M/M+F	15.43

033 GARNET 2A	
ØXIDE	WT %
SIØ2	36.77
AL2Ø3	20.39
TIØ2	0.00
FE2Ø3	0.44
FEØ	34.38
MGØ	4.03
MNØ	1.64
CAØ	0.49
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.14

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.006
AL	
AL	1.965
TI	
FE3+	0.027
FE2+	2.351
MG	0.491
MN	0.114
CA	0.043
NA	
K	

ENDMEMBERS	
% PYRØPE	16.32
% ALM	78.11
% SPESS	3.78
% ANDR	1.35
% GRØSS	0.07
% M/M+F	17.28

033 GARNET 2H	
ØXIDE	WT %
SIØ2	36.44
AL2Ø3	20.61
TIØ2	0.00
FE2Ø3	0.59
FEØ	35.00
MGØ	3.19
MNØ	2.71
CAØ	0.53
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.08

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.977
AL	0.023
AL	1.962
TI	
FE3+	0.036
FE2+	2.391
MG	0.389
MN	0.188
CA	0.046
NA	
K	

ENDMEMBERS	
% PYRØPE	12.80
% ALM	78.73
% SPESS	6.18
% ANDR	1.52
% GRØSS	0.00
% M/M+F	13.99

033 GARNET 3	
ØXIDE	WT %
SIØ2	36.78
AL2Ø3	20.76
TIØ2	0.00
FE2Ø3	0.45
FEØ	35.28
MGØ	3.06
MNØ	2.86
CAØ	0.55
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.74

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.985
AL	0.015
AL	1.972
TI	
FE3+	0.027
FE2+	2.395
MG	0.370
MN	0.196
CA	0.048
NA	
K	

ENDMEMBERS	
% PYRØPE	12.24
% ALM	79.18
% SPESS	6.49
% ANDR	1.35
% GRØSS	0.23
% M/M+F	13.39

033 GARNET 4	
ØXIDE	WT %
SIØ2	36.68
AL2Ø3	20.79
TIØ2	0.00
FE2Ø3	0.43
FEØ	35.44
MGØ	3.36
MNØ	2.25
CAØ	0.46
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.42

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.981
AL	0.019
AL	1.972
TI	
FE3+	0.026
FE2+	2.409
MG	0.408
MN	0.155
CA	0.040
NA	
K	

ENDMEMBERS	
% PYRØPE	13.45
% ALM	79.49
% SPESS	5.10
% ANDR	1.31
% GRØSS	0.02
% M/M+F	14.47

033 GARNET 5A	
ØXIDE	WT %
SIØ2	37.04
AL2Ø3	20.99
TIØ2	0.00
FE2Ø3	0.00
FEØ	34.00
MGØ	4.38
MNØ	0.88
CAØ	0.56
NA2Ø	0.00
K2Ø	0.00
TØTAL	97.85

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	3.013
AL	
AL	2.013
TI	
FE3+	
FE2+	2.313
MG	0.532
MN	0.061
CA	0.048
NA	
K	

ENDMEMBERS	
% PYRØPE	17.72
% ALM	77.10
% SPESS	2.03
% ANDR	0.00
% GRØSS	1.62
% M/M+F	18.69

033 GARNET 5B	
ØXIDE	WT %
SIØ2	36.52
AL2Ø3	21.36
TIØ2	0.00
FE2Ø3	0.29
FEØ	35.34
MGØ	3.32
MNØ	2.77
CAØ	0.60
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.21
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.948
AL	0.052
AL	1.981
TI	
FE3+	0.018
FE2+	2.386
MG	0.400
MN	0.189
CA	0.052
NA	
K	
ENDMEMBERS	
% PYRØPE	12.97
% ALM	77.47
% SPESS	6.15
% ANDR	0.87
% GRØSS	0.81
% M/M+F	14.34

033 GARNET 6	
ØXIDE	WT %
SIØ2	36.71
AL2Ø3	20.86
TIØ2	0.00
FE2Ø3	0.32
FEØ	35.57
MGØ	3.34
MNØ	2.14
CAØ	0.49
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.43
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.982
AL	0.018
AL	1.980
TI	
FE3+	0.020
FE2+	2.417
MG	0.404
MN	0.147
CA	0.042
NA	
K	
ENDMEMBERS	
% PYRØPE	13.34
% ALM	79.80
% SPESS	4.86
% ANDR	0.97
% GRØSS	0.43
% M/M+F	14.32

033 GARNET 7	
ØXIDE	WT %
SIØ2	37.04
AL2Ø3	20.75
TIØ2	0.00
FE2Ø3	0.31
FEØ	35.06
MGØ	4.18
MNØ	1.20
CAØ	0.46
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.01
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.999
AL	0.001
AL	1.980
TI	
FE3+	0.019
FE2+	2.375
MG	0.504
MN	0.082
CA	0.040
NA	
K	
ENDMEMBERS	
% PYRØPE	16.80
% ALM	79.08
% SPESS	2.74
% ANDR	0.95
% GRØSS	0.39
% M/M+F	17.52

033 GARNET 8	
ØXIDE	WT %
SIØ2	36.51
AL2Ø3	20.43
TIØ2	0.00
FE2Ø3	1.02
FEØ	34.31
MGØ	3.77
MNØ	2.52
CAØ	0.56
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.12

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.974
AL	0.026
AL	1.935
TI	
FE3+	0.062
FE2+	2.337
MG	0.458
MN	0.174
CA	0.049
NA	
K	

ENDMEMBERS	
% PYRØPE	15.04
% ALM	76.78
% SPESS	5.70
% ANDR	1.60
% GRØSS	0.00
% M/M+F	16.38

033 GARNET 9A	
ØXIDE	WT %
SIØ2	36.73
AL2Ø3	21.01
TIØ2	0.00
FE2Ø3	0.38
FEØ	35.30
MGØ	3.07
MNØ	3.02
CAØ	0.58
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.10

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.972
AL	0.028
AL	1.976
TI	
FE3+	0.023
FE2+	2.388
MG	0.370
MN	0.207
CA	0.051
NA	
K	

ENDMEMBERS	
% PYRØPE	12.14
% ALM	78.45
% SPESS	6.81
% ANDR	1.15
% GRØSS	0.51
% M/M+F	13.40

033 GARNET 10A	
ØXIDE	WT %
SIØ2	36.39
AL2Ø3	20.95
TIØ2	0.00
FE2Ø3	0.36
FEØ	34.48
MGØ	3.07
MNØ	3.26
CAØ	0.82
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.33

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.965
AL	0.035
AL	1.977
TI	
FE3+	0.022
FE2+	2.350
MG	0.372
MN	0.225
CA	0.072
NA	
K	

ENDMEMBERS	
% PYRØPE	12.19
% ALM	76.93
% SPESS	7.36
% ANDR	1.07
% GRØSS	1.28
% M/M+F	13.68

0524 GARNET 1A	
ØXIDE	WT %
SIØ2	36.81
AL2Ø3	20.07
TIØ2	0.00
FE2Ø3	1.20
FEØ	35.24
MGØ	3.34
MNØ	1.36
CAØ	1.24
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.27

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.997
AL	0.003
AL	1.923
TI	
FE3+	0.074
FE2+	2.399
MG	0.405
MN	0.094
CA	0.108
NA	
K	

ENDMEMBERS	
% PYRØPE	13.45
% ALM	79.68
% SPESS	3.12
% ANDR	3.59
% GRØSS	0.00
% M/M+F	14.44

0524 GARNET 1M	
ØXIDE	WT %
SIØ2	35.96
AL2Ø3	20.41
TIØ2	0.00
FE2Ø3	0.72
FEØ	34.82
MGØ	3.06
MNØ	2.02
CAØ	1.10
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.09

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.968
AL	0.032
AL	1.954
TI	
FE3+	0.045
FE2+	2.403
MG	0.377
MN	0.142
CA	0.097
NA	
K	

ENDMEMBERS	
% PYRØPE	12.34
% ALM	78.76
% SPESS	4.64
% ANDR	2.19
% GRØSS	1.00
% M/M+F	13.55

0524 GARNET 1N	
ØXIDE	WT %
SIØ2	36.51
AL2Ø3	20.54
TIØ2	0.00
FE2Ø3	0.74
FEØ	35.18
MGØ	3.40
MNØ	1.65
CAØ	1.00
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.02

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.978
AL	0.022
AL	1.953
TI	
FE3+	0.045
FE2+	2.400
MG	0.413
MN	0.114
CA	0.087
NA	
K	

ENDMEMBERS	
% PYRØPE	13.62
% ALM	79.03
% SPESS	3.75
% ANDR	2.24
% GRØSS	0.63
% M/M+F	14.70

0524 GARNET 1Ø	
ØXIDE	WT %
SIØ2	36.17
AL2Ø3	20.54
TIØ2	0.00
FE2Ø3	0.78
FEØ	34.92
MGØ	3.24
MNØ	1.96
CAØ	1.08
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.70

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.965
AL	0.035
AL	1.950
TI	
FE3+	0.048
FE2+	2.394
MG	0.396
MN	0.136
CA	0.095
NA	
K	

ENDMEMBERS	
% PYRØPE	12.94
% ALM	78.31
% SPESS	4.46
% ANDR	2.38
% GRØSS	0.74
% M/M+F	14.18

0524 GARNET 1P	
ØXIDE	WT %
SIØ2	36.33
AL2Ø3	20.04
TIØ2	0.00
FE2Ø3	1.32
FEØ	33.58
MGØ	3.51
MNØ	2.23
CAØ	1.48
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.48

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.978
AL	0.022
AL	1.915
TI	
FE3+	0.081
FE2+	2.302
MG	0.429
MN	0.155
CA	0.130
NA	
K	

ENDMEMBERS	
% PYRØPE	14.13
% ALM	75.79
% SPESS	5.09
% ANDR	4.01
% GRØSS	0.26
% M/M+F	15.71

0524 GARNET 1Q	
ØXIDE	WT %
SIØ2	36.00
AL2Ø3	20.36
TIØ2	0.00
FE2Ø3	0.85
FEØ	34.87
MGØ	3.23
MNØ	1.69
CAØ	1.16
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.15

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.967
AL	0.033
AL	1.945
TI	
FE3+	0.053
FE2+	2.403
MG	0.397
MN	0.118
CA	0.102
NA	
K	

ENDMEMBERS	
% PYRØPE	12.99
% ALM	78.70
% SPESS	3.85
% ANDR	2.58
% GRØSS	0.76
% M/M+F	14.17

0524 GARNET 2A	
ØXIDE	WT %
SIØ2	36.74
AL2Ø3	20.41
TIØ2	0.00
FE2Ø3	0.99
FEØ	33.55
MGØ	4.30
MNØ	1.42
CAØ	1.33
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.73

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.983
AL	0.017
AL	1.937
TI	
FE3+	0.060
FE2+	2.278
MG	0.521
MN	0.098
CA	0.115
NA	
K	

ENDMEMBERS	
% PYRØPE	17.19
% ALM	75.22
% SPESS	3.22
% ANDR	2.99
% GRØSS	0.82
% M/M+F	18.60

0524 GARNET 2F	
ØXIDE	WT %
SIØ2	36.80
AL2Ø3	20.61
TIØ2	0.00
FE2Ø3	0.86
FEØ	34.97
MGØ	3.59
MNØ	1.65
CAØ	1.16
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.66

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.978
AL	0.022
AL	1.945
TI	
FE3+	0.053
FE2+	2.367
MG	0.433
MN	0.113
CA	0.101
NA	
K	

ENDMEMBERS	
% PYRØPE	14.28
% ALM	77.96
% SPESS	3.72
% ANDR	2.60
% GRØSS	0.72
% M/M+F	15.48

0524 GARNET 3A	
ØXIDE	WT %
SIØ2	36.79
AL2Ø3	20.51
TIØ2	0.00
FE2Ø3	1.03
FEØ	33.35
MGØ	3.87
MNØ	2.43
CAØ	1.43
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.41

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.978
AL	0.022
AL	1.935
TI	
FE3+	0.063
FE2+	2.258
MG	0.467
MN	0.167
CA	0.124
NA	
K	

ENDMEMBERS	
% PYRØPE	15.37
% ALM	74.33
% SPESS	5.49
% ANDR	3.09
% GRØSS	0.99
% M/M+F	17.14

0524 GARNET 3E

ØXIDE	WT %
SIØ2	36.17
AL2Ø3	20.15
TIØ2	0.00
FE2Ø3	1.23
FEØ	33.79
MGØ	3.44
MNØ	2.49
CAØ	1.20
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.47

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.970
AL	0.030
AL	1.921
TI	
FE3+	0.076
FE2+	2.320
MG	0.421
MN	0.173
CA	0.106
NA	
K	

ENDMEMBERS

% PYRØPE	13.79
% ALM	76.09
% SPESS	5.67
% ANDR	3.46
% GRØSS	0.00
% M/M+F	15.34

0524 GARNET 4A

ØXIDE	WT %
SIØ2	36.77
AL2Ø3	21.05
TIØ2	0.00
FE2Ø3	0.37
FEØ	35.09
MGØ	3.63
MNØ	1.60
CAØ	1.12
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.62

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.971
AL	0.029
AL	1.977
TI	
FE3+	0.022
FE2+	2.372
MG	0.437
MN	0.110
CA	0.097
NA	
K	

ENDMEMBERS

% PYRØPE	14.36
% ALM	77.90
% SPESS	3.60
% ANDR	1.10
% GRØSS	2.09
% M/M+F	15.57

0524 GARNET 4Ø

ØXIDE	WT %
SIØ2	36.71
AL2Ø3	21.60
TIØ2	0.00
FE2Ø3	0.18
FEØ	34.15
MGØ	2.81
MNØ	1.76
CAØ	3.28
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.49

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.945
AL	0.055
AL	1.988
TI	
FE3+	0.011
FE2+	2.291
MG	0.336
MN	0.119
CA	0.282
NA	
K	

ENDMEMBERS

% PYRØPE	10.88
% ALM	74.28
% SPESS	3.87
% ANDR	0.54
% GRØSS	8.60
% M/M+F	12.77

0524 GARNET 4P

ØXIDE	WT %
SIØ2	37.04
AL2Ø3	21.96
TIØ2	0.00
FE2Ø3	0.22
FEØ	35.50
MGØ	4.02
MNØ	1.49
CAØ	1.24
NA2Ø	0.00
K2Ø	0.00
TØTAL	101.49

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.935
AL	0.065
AL	1.986
TI	
FE3+	0.013
FE2+	2.353
MG	0.475
MN	0.100
CA	0.106
NA	
K	

ENDMEMBERS

% PYRØPE	15.32
% ALM	75.87
% SPESS	3.23
% ANDR	0.64
% GRØSS	2.76
% M/M+F	16.80

0524 GARNET 4V

ØXIDE	WT %
SIØ2	36.86
AL2Ø3	21.68
TIØ2	0.00
FE2Ø3	0.53
FEØ	34.80
MGØ	4.45
MNØ	1.30
CAØ	1.23
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.85

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.933
AL	0.067
AL	1.967
TI	
FE3+	0.032
FE2+	2.316
MG	0.527
MN	0.087
CA	0.105
NA	
K	

ENDMEMBERS

% PYRØPE	16.98
% ALM	74.61
% SPESS	2.82
% ANDR	1.52
% GRØSS	1.84
% M/M+F	18.54

0524 GARNET 4X

ØXIDE	WT %
SIØ2	36.38
AL2Ø3	21.04
TIØ2	0.00
FE2Ø3	0.43
FEØ	34.43
MGØ	2.88
MNØ	2.00
CAØ	2.25
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.41

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.957
AL	0.043
AL	1.973
TI	
FE3+	0.026
FE2+	2.340
MG	0.349
MN	0.137
CA	0.196
NA	
K	

ENDMEMBERS

% PYRØPE	11.39
% ALM	76.30
% SPESS	4.46
% ANDR	1.27
% GRØSS	5.12
% M/M+F	12.98

0524 GARNET 4Z	
ØXIDE	WT %
SIØ2	36.82
AL2Ø3	21.86
TIØ2	0.00
FE2Ø3	0.12
FEØ	34.50
MGØ	3.22
MNØ	1.47
CAØ	2.91
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.90

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.937
AL	0.063
AL	1.992
TI	
FE3+	0.007
FE2+	2.302
MG	0.383
MN	0.099
CA	0.248
NA	
K	

ENDMEMBERS	
% PYRØPE	12.36
% ALM	74.30
% SPESS	3.21
% ANDR	0.36
% GRØSS	7.66
% M/M+F	14.26

Ø524 GARNET 4CC	
ØXIDE	WT %
SIØ2	36.73
AL2Ø3	21.91
TIØ2	0.00
FE2Ø3	0.00
FEØ	35.03
MGØ	3.01
MNØ	1.52
CAØ	2.55
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.76

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.938
AL	0.062
AL	2.004
TI	
FE3+	
FE2+	2.344
MG	0.359
MN	0.103
CA	0.218
NA	
K	

ENDMEMBERS	
% PYRØPE	11.63
% ALM	75.89
% SPESS	3.34
% ANDR	0.00
% GRØSS	7.07
% M/M+F	13.29

0524 GARNET 5A	
ØXIDE	WT %
SIØ2	36.88
AL2Ø3	21.74
TIØ2	0.00
FE2Ø3	0.00
FEØ	35.97
MGØ	3.42
MNØ	1.91
CAØ	0.94
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.87

RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.950
AL	0.050
AL	2.000
TI	
FE3+	
FE2+	2.406
MG	0.408
MN	0.130
CA	0.081
NA	
K	

ENDMEMBERS	
% PYRØPE	13.27
% ALM	78.23
% SPESS	4.21
% ANDR	0.00
% GRØSS	2.62
% M/M+F	14.51

034 BIOTITE 1	
OXIDE	WT %
SIO2	36.55
AL2O3	18.98
TIO2	1.47
FE2O3	0.00
FEO	19.08
MGO	9.23
MNO	0.08
CAO	0.03
NA2O	0.39
K2O	7.90
TOTAL	93.70

RECALCULATED ON BASIS OF 22.0 (O)	
SI	5.580
AL	2.420
AL	0.996
TI	0.169
FE3+	
FE2+	2.437
MG	2.100
MN	0.010
CA	0.004
NA	0.114
K	1.539

ENDMEMBERS	
% PHLOG	34.99
% ANNITE	40.62
% M/M+F	46.28

034 BIOTITE 2	
OXIDE	WT %
SIO2	35.89
AL2O3	19.85
TIO2	1.27
FE2O3	0.00
FEO	19.24
MGO	8.84
MNO	0.08
CAO	0.02
NA2O	0.51
K2O	8.21
TOTAL	93.91

RECALCULATED ON BASIS OF 22.0 (O)	
SI	5.487
AL	2.513
AL	1.066
TI	0.146
FE3+	
FE2+	2.461
MG	2.015
MN	0.010
CA	0.003
NA	0.152
K	1.602

ENDMEMBERS	
% PHLOG	33.58
% ANNITE	41.02
% M/M+F	45.02

034 BIOTITE 3	
OXIDE	WT %
SIO2	34.94
AL2O3	18.83
TIO2	1.39
FE2O3	0.00
FEO	20.92
MGO	10.28
MNO	0.07
CAO	0.01
NA2O	0.37
K2O	8.31
TOTAL	95.12

RECALCULATED ON BASIS OF 22.0 (O)	
SI	5.345
AL	2.655
AL	0.741
TI	0.160
FE3+	
FE2+	2.676
MG	2.342
MN	0.009
CA	0.002
NA	0.110
K	1.621

ENDMEMBERS	
% PHLOG	37.84
% ANNITE	43.22
% M/M+F	46.68

034 BIØTITE 4	
ØXIDE	WT %
SIØ2	34.98
AL2Ø3	19.25
TIØ2	1.64
FE2Ø3	0.00
FEØ	20.33
MGØ	9.70
MNØ	0.08
CAØ	0.02
NA2Ø	0.47
K2Ø	8.16
TØTAL	94.63

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.357
AL	2.643
AL	0.832
TI	0.189
FE3+	
FE2+	2.603
MG	2.213
MN	0.010
CA	0.003
NA	0.140
K	1.594

ENDMEMBERS	
% PHLØG	36.61
% ANNITE	43.07
% M/M+F	45.95

034 BIØTITE 5	
ØXIDE	WT %
SIØ2	35.59
AL2Ø3	18.99
TIØ2	1.34
FE2Ø3	0.00
FEØ	18.82
MGØ	9.68
MNØ	0.07
CAØ	0.02
NA2Ø	0.62
K2Ø	8.10
TØTAL	93.22

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.484
AL	2.516
AL	0.933
TI	0.155
FE3+	
FE2+	2.425
MG	2.224
MN	0.009
CA	0.003
NA	0.184
K	1.592

ENDMEMBERS	
% PHLØG	37.06
% ANNITE	40.42
% M/M+F	47.83

034 BIØTITE 6	
ØXIDE	WT %
SIØ2	35.23
AL2Ø3	18.67
TIØ2	1.41
FE2Ø3	0.00
FEØ	19.39
MGØ	9.65
MNØ	0.08
CAØ	0.02
NA2Ø	0.60
K2Ø	7.94
IØTAL	92.98

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.460
AL	2.540
AL	0.872
TI	0.165
FE3+	
FE2+	2.514
MG	2.230
MN	0.010
CA	0.003
NA	0.180
K	1.570

ENDMEMBERS	
% PHLØG	36.90
% ANNITE	41.60
% M/M+F	47.01

033 BIØTITE 1	
ØXIDE	WT %
SIØ2	35.40
AL2Ø3	19.32
TIØ2	1.08
FE2Ø3	0.00
FEØ	18.63
MGØ	9.76
MNØ	0.08
CAØ	0.06
NA2Ø	0.52
K2Ø	8.29
TØTAL	93.13

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.460
AL	2.540
AL	0.973
TI	0.126
FE3+	
FE2+	2.403
MG	2.244
MN	0.010
CA	0.010
NA	0.155
K	1.632

ENDMEMBERS	
% PHLØG	37.39
% ANNITE	40.05
% M/M+F	48.28

033 BIØTITE 2	
ØXIDE	WT %
SIØ2	35.30
AL2Ø3	19.40
TIØ2	1.48
FE2Ø3	0.00
FEØ	18.42
MGØ	9.56
MNØ	0.07
CAØ	0.01
NA2Ø	0.53
K2Ø	8.23
TØTAL	93.00

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.445
AL	2.555
AL	0.973
TI	0.172
FE3+	
FE2+	2.377
MG	2.197
MN	0.009
CA	0.002
NA	0.159
K	1.621

ENDMEMBERS	
% PHLØG	36.62
% ANNITE	39.61
% M/M+F	48.04

033 BIØTITE 3	
ØXIDE	WT %
SIØ2	35.43
AL2Ø3	19.15
TIØ2	1.58
FE2Ø3	0.00
FEØ	19.27
MGØ	9.86
MNØ	0.09
CAØ	0.01
NA2Ø	0.42
K2Ø	8.48
TØTAL	94.29

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.419
AL	2.581
AL	0.872
TI	0.182
FE3+	
FE2+	2.466
MG	2.248
MN	0.012
CA	0.002
NA	0.124
K	1.656

ENDMEMBERS	
% PHLØG	37.47
% ANNITE	41.09
% M/M+F	47.69

033 BIØTITE 4	
ØXIDE	WT %
SIØ2	35.75
AL2Ø3	19.58
TIØ2	1.35
FE2Ø3	0.00
FEØ	18.80
MGØ	9.55
MNØ	0.09
CAØ	0.02
NA2Ø	0.52
K2Ø	8.47
TØTAL	94.12

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.457
AL	2.543
AL	0.981
TI	0.155
FE3+	
FE2+	2.401
MG	2.173
MN	0.011
CA	0.003
NA	0.154
K	1.650

ENDMEMBERS	
% PHLØG	36.22
% ANNITE	40.01
% M/M+F	47.51

033 BIØTITE 5	
ØXIDE	WT %
SIØ2	34.81
AL2Ø3	18.57
TIØ2	1.48
FE2Ø3	0.00
FEØ	19.33
MGØ	9.59
MNØ	0.07
CAØ	0.02
NA2Ø	0.33
K2Ø	8.70
TØTAL	92.90

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.428
AL	2.572
AL	0.842
TI	0.174
FE3+	
FE2+	2.521
MG	2.228
MN	0.010
CA	0.003
NA	0.099
K	1.730

ENDMEMBERS	
% PHLØG	37.14
% ANNITE	42.01
% M/M+F	46.92

033 BIØTITE 6	
ØXIDE	WT %
SIØ2	35.87
AL2Ø3	19.34
TIØ2	1.51
FE2Ø3	0.00
FEØ	18.54
MGØ	9.91
MNØ	0.08
CAØ	0.01
NA2Ø	0.62
K2Ø	8.29
TØTAL	94.17

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.463
AL	2.537
AL	0.936
TI	0.173
FE3+	
FE2+	2.361
MG	2.249
MN	0.010
CA	0.002
NA	0.184
K	1.610

ENDMEMBERS	
% PHLØG	37.48
% ANNITE	39.35
% M/M+F	48.78

034 MUSCØVITE 1	
ØXIDE	WT %
SIØ2	46.96
AL2Ø3	36.85
TIØ2	0.54
FE2Ø3	0.00
FEØ	0.90
MGØ	0.62
MNØ	0.02
CAØ	0.02
NA2Ø	2.09
K2Ø	7.81
TØTAL	95.80

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.127
AL	1.873
AL	3.795
TI	0.053
FE3+	
FE2+	0.098
MG	0.121
MN	0.002
CA	0.002
NA	0.529
K	1.300

ENDMEMBERS	
% MUSC	64.99
% PARAG	26.45
% CELAD	5.13
% M/M+F	55.17

034 MUSCØVITE 2	
ØXIDE	WT %
SIØ2	46.90
AL2Ø3	37.05
TIØ2	0.50
FE2Ø3	0.00
FEØ	0.76
MGØ	0.56
MNØ	0.01
CAØ	0.02
NA2Ø	1.75
K2Ø	7.38
TØTAL	94.94

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.142
AL	1.858
AL	3.862
TI	0.049
FE3+	
FE2+	0.084
MG	0.110
MN	0.002
CA	0.003
NA	0.444
K	1.233

ENDMEMBERS	
% MUSC	61.66
% PARAG	22.22
% CELAD	3.44
% M/M+F	56.73

034 MUSCØVITE 3	
ØXIDE	WT %
SIØ2	46.59
AL2Ø3	35.72
TIØ2	0.63
FE2Ø3	0.00
FEØ	1.03
MGØ	0.89
MNØ	0.02
CAØ	0.02
NA2Ø	1.60
K2Ø	7.02
TØTAL	93.51

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.191
AL	1.809
AL	3.788
TI	0.063
FE3+	
FE2+	0.115
MG	0.176
MN	0.002
CA	0.002
NA	0.412
K	1.190

ENDMEMBERS	
% MUSC	59.51
% PARAG	20.59
% CELAD	5.30
% M/M+F	60.60

034 MUSCØVITE 4	
ØXIDE	WT %
SIØ2	45.67
AL2Ø3	37.71
TIØ2	0.57
FE2Ø3	0.00
FEØ	0.96
MGØ	0.63
MNØ	0.03
CAØ	0.02
NA2Ø	2.07
K2Ø	7.87
TØTAL	95.52

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	5.992
AL	2.008
AL	3.824
TI	0.056
FE3+	
FE2+	0.105
MG	0.124
MN	0.003
CA	0.003
NA	0.526
K	1.317

ENDMEMBERS	
% MUSC	65.86
% PARAG	26.28
% CELAD	0.00
% M/M+F	54.11

034 MUSCØVITE 5	
ØXIDE	WT %
SIØ2	46.52
AL2Ø3	36.26
TIØ2	0.59
FE2Ø3	0.00
FEØ	1.07
MGØ	0.92
MNØ	0.02
CAØ	0.02
NA2Ø	2.21
K2Ø	7.54
TØTAL	95.16

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.117
AL	1.883
AL	3.738
TI	0.058
FE3+	
FE2+	0.117
MG	0.181
MN	0.002
CA	0.003
NA	0.564
K	1.265

ENDMEMBERS	
% MUSC	63.24
% PARAG	28.22
% CELAD	5.84
% M/M+F	60.70

034 MUSCØVITE 6	
ØXIDE	WT %
SIØ2	46.18
AL2Ø3	36.86
TIØ2	0.47
FE2Ø3	0.00
FEØ	0.71
MGØ	0.56
MNØ	0.01
CAØ	0.02
NA2Ø	1.81
K2Ø	7.89
TØTAL	94.50

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.102
AL	1.898
AL	3.845
TI	0.047
FE3+	
FE2+	0.079
MG	0.109
MN	0.001
CA	0.003
NA	0.464
K	1.330

ENDMEMBERS	
% MUSC	66.51
% PARAG	23.20
% CELAD	3.88
% M/M+F	58.14

034 MUSCØVITE 7	
ØXIDE	WT %
SIØ2	46.98
AL2Ø3	37.47
TIØ2	0.54
FE2Ø3	0.00
FEØ	0.77
MGØ	0.55
MNØ	0.01
CAØ	0.01
NA2Ø	1.95
K2Ø	7.72
TØTAL	96.01

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.104
AL	1.896
AL	3.843
TI	0.053
FE3+	
FE2+	0.084
MG	0.106
MN	0.002
CA	0.001
NA	0.492
K	1.279

ENDMEMBERS	
% MUSC	63.94
% PARAG	24.61
% CELAD	3.94
% M/M+F	55.84

034 MUSCØVITE 8	
ØXIDE	WT %
SIØ2	45.40
AL2Ø3	36.32
TIØ2	0.54
FE2Ø3	0.00
FEØ	1.06
MGØ	0.60
MNØ	0.03
CAØ	0.01
NA2Ø	1.55
K2Ø	7.73
TØTAL	93.25

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.087
AL	1.913
AL	3.828
TI	0.055
FE3+	
FE2+	0.119
MG	0.119
MN	0.003
CA	0.002
NA	0.402
K	1.322

ENDMEMBERS	
% MUSC	66.10
% PARAG	20.09
% CELAD	4.31
% M/M+F	50.00

033 MUSCØVITE 1	
ØXIDE	WT %
SIØ2	45.94
AL2Ø3	36.50
TIØ2	0.55
FE2Ø3	0.00
FEØ	0.83
MGØ	0.56
MNØ	0.01
CAØ	0.02
NA2Ø	1.86
K2Ø	8.21
TØTAL	94.48

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.094
AL	1.906
AL	3.802
TI	0.054
FE3+	
FE2+	0.092
MG	0.111
MN	0.002
CA	0.003
NA	0.479
K	1.389

ENDMEMBERS	
% MUSC	69.47
% PARAG	23.97
% CELAD	4.70
% M/M+F	54.65

033 MUSCØVITE 2	
ØXIDE	WT %
SIØ2	45.99
AL2Ø3	37.62
TIØ2	0.54
FE2Ø3	0.00
FEØ	0.95
MGØ	0.50
MNØ	0.03
CAØ	0.01
NA2Ø	1.48
K2Ø	8.02
TØTAL	95.15

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.042
AL	1.958
AL	3.869
TI	0.054
FE3+	
FE2+	0.104
MG	0.098
MN	0.004
CA	0.002
NA	0.377
K	1.344

ENDMEMBERS	
% MUSC	67.21
% PARAG	18.85
% CELAD	2.09
% M/M+F	48.50

033 MUSCØVITE 3	
ØXIDE	WT %
SIØ2	45.95
AL2Ø3	37.00
TIØ2	0.54
FE2Ø3	0.00
FEØ	0.89
MGØ	0.54
MNØ	0.02
CAØ	0.02
NA2Ø	1.73
K2Ø	7.77
TØTAL	94.48

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.076
AL	1.924
AL	3.844
TI	0.054
FE3+	
FE2+	0.099
MG	0.106
MN	0.002
CA	0.003
NA	0.444
K	1.311

ENDMEMBERS	
% MUSC	65.54
% PARAG	22.21
% CELAD	3.78
% M/M+F	51.92

033 MUSCØVITE 4	
ØXIDE	WT %
SIØ2	46.23
AL2Ø3	37.75
TIØ2	0.54
FE2Ø3	0.00
FEØ	0.94
MGØ	0.40
MNØ	0.02
CAØ	0.01
NA2Ø	1.87
K2Ø	7.96
TØTAL	95.73

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.041
AL	1.959
AL	3.857
TI	0.054
FE3+	
FE2+	0.102
MG	0.079
MN	0.002
CA	0.002
NA	0.475
K	1.327

ENDMEMBERS	
% MUSC	66.36
% PARAG	23.74
% CELAD	2.07
% M/M+F	43.45

034 STAURØLITE 1

ØXIDE	WT %
SIØ2	28.15
AL2Ø3	54.38
TIØ2	0.53
FE2Ø3	0.00
FEØ	12.43
MGØ	1.48
MNØ	0.16
CAØ	0.01
ZNØ	0.23
CØØ	0.07
TØTAL	97.44

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.895
AL	
AL	8.870
TI	0.055
FE3+	
FE2+	1.439
MG	0.305
MN	0.018
CA	0.002
ZN	0.024
CØ	0.008

ENDMEMBERS

% FE ST	71.93
% MG ST	15.25
% M/M+F	17.49

034 STAURØLITE 2

ØXIDE	WT %
SIØ2	28.58
AL2Ø3	54.45
TIØ2	0.66
FE2Ø3	0.00
FEØ	12.91
MGØ	1.45
MNØ	0.17
CAØ	0.02
ZNØ	0.26
CØØ	0.05
TØTAL	98.56

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.917
AL	
AL	8.799
TI	0.068
FE3+	
FE2+	1.480
MG	0.296
MN	0.020
CA	0.003
ZN	0.027
CØ	0.005

ENDMEMBERS

% FE ST	74.01
% MG ST	14.81
% M/M+F	16.67

034 STAURØLITE 3

ØXIDE	WT %
SIØ2	28.42
AL2Ø3	53.59
TIØ2	0.68
FE2Ø3	0.00
FEØ	13.31
MGØ	1.60
MNØ	0.17
CAØ	0.02
ZNØ	0.24
CØØ	0.06
TØTAL	98.10

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.925
AL	
AL	8.725
TI	0.071
FE3+	
FE2+	1.538
MG	0.330
MN	0.020
CA	0.003
ZN	0.024
CØ	0.006

ENDMEMBERS

% FE ST	76.88
% MG ST	16.51
% M/M+F	17.68

034 STAURØLITE 4	
ØXIDE	WT %
SIØ2	28.19
AL2Ø3	54.20
TIØ2	0.68
FE2Ø3	0.00
FEØ	12.70
MGØ	1.38
MNØ	0.18
CAØ	0.02
ZNØ	0.00
CØØ	0.00
TØTAL	97.35

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.903
AL	
AL	8.848
TI	0.071
FE3+	
FE2+	1.471
MG	0.285
MN	0.022
CA	0.002
ZN	
CØ	

ENDMEMBERS	
% FE ST	73.54
% MG ST	14.26
% M/M+F	16.24

034 STAURØLITE 5	
ØXIDE	WT %
SIØ2	28.18
AL2Ø3	54.69
TIØ2	0.69
FE2Ø3	0.00
FEØ	12.38
MGØ	1.16
MNØ	0.18
CAØ	0.01
ZNØ	0.26
CØØ	0.06
TØTAL	97.61

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.890
AL	
AL	8.901
TI	0.071
FE3+	
FE2+	1.429
MG	0.239
MN	0.021
CA	0.002
ZN	0.027
CØ	0.006

ENDMEMBERS	
% FE ST	71.45
% MG ST	11.97
% M/M+F	14.34

034 STAURØLITE 6	
ØXIDE	WT %
SIØ2	27.93
AL2Ø3	55.23
TIØ2	0.68
FE2Ø3	0.00
FEØ	12.88
MGØ	1.34
MNØ	0.18
CAØ	0.02
ZNØ	0.00
CØØ	0.00
TØTAL	98.24

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.835
AL	
AL	8.942
TI	0.070
FE3+	
FE2+	1.479
MG	0.274
MN	0.021
CA	0.002
ZN	
CØ	

ENDMEMBERS	
% FE ST	73.95
% MG ST	13.69
% M/M+F	15.62

034 STAURØLITE 7	
ØXIDE	WT %
SIØ2	28.14
AL2Ø3	53.77
TIØ2	0.68
FE2Ø3	0.00
FEØ	13.58
MGØ	1.84
MNØ	0.17
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	98.21

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.883
AL	
AL	8.749
TI	0.070
FE3+	
FE2+	1.568
MG	0.378
MN	0.020
CA	0.003
NA	
K	

ENDMEMBERS	
% FE ST	78.32
% MG ST	18.90
% M/M+F	19.44

034 STAURØLITE 8	
ØXIDE	WT %
SIØ2	27.67
AL2Ø3	55.16
TIØ2	0.68
FE2Ø3	0.00
FEØ	12.47
MGØ	1.12
MNØ	0.16
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	97.29

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.830
AL	
AL	9.001
TI	0.070
FE3+	
FE2+	1.444
MG	0.232
MN	0.019
CA	0.002
NA	
K	

ENDMEMBERS	
% FE ST	72.19
% MG ST	11.58
% M/M+F	13.83

033 STAURØLITE 1	
ØXIDE	WT %
SIØ2	27.27
AL2Ø3	53.44
TIØ2	0.68
FE2Ø3	0.00
FEØ	13.14
MGØ	1.64
MNØ	0.20
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.38

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.832
AL	
AL	8.853
TI	0.071
FE3+	
FE2+	1.545
MG	0.343
MN	0.024
CA	0.002
NA	
K	

ENDMEMBERS	
% FE ST	77.23
% MG ST	17.16
% M/M+F	18.18

033 STAURØLITE 2

ØXIDE	WT %
SIØ2	27.41
AL2Ø3	54.58
TIØ2	0.68
FE2Ø3	0.00
FEØ	12.81
MGØ	1.40
MNØ	0.21
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	97.11

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.813
AL	
AL	8.949
TI	0.071
FE3+	
FE2+	1.491
MG	0.290
MN	0.025
CA	0.002
NA	
K	

ENDMEMBERS

% FE ST	74.54
% MG ST	14.52
% M/M+F	16.31

033 STAURØLITE 3

ØXIDE	WT %
SIØ2	26.93
AL2Ø3	53.31
TIØ2	0.68
FE2Ø3	0.00
FEØ	13.12
MGØ	1.56
MNØ	0.20
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	95.81

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.808
AL	
AL	8.888
TI	0.072
FE3+	
FE2+	1.552
MG	0.329
MN	0.024
CA	0.002
NA	
K	

ENDMEMBERS

% FE ST	77.58
% MG ST	16.43
% M/M+F	17.48

033 STAURØLITE 4

ØXIDE	WT %
SIØ2	27.69
AL2Ø3	53.23
TIØ2	0.68
FE2Ø3	0.00
FEØ	12.97
MGØ	1.76
MNØ	0.19
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.54

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.879
AL	
AL	8.791
TI	0.071
FE3+	
FE2+	1.519
MG	0.368
MN	0.023
CA	0.003
NA	
K	

ENDMEMBERS

% FE ST	75.96
% MG ST	18.41
% M/M+F	19.51

033 STAURØLITE 5

ØXIDE	WT %
SIØ2	27.25
AL2Ø3	53.16
TIØ2	0.68
FE2Ø3	0.00
FEØ	13.17
MGØ	1.66
MNØ	0.19
CAØ	0.02
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.12

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.841
AL	
AL	8.833
TI	0.072
FE3+	
FE2+	1.552
MG	0.348
MN	0.023
CA	0.003
NA	
K	

ENDMEMBERS

% FE ST	77.61
% MG ST	17.40
% M/M+F	18.32

034 PLAGIØCLASE 1

ØXIDE	WT %
SIØ2	65.28
AL2Ø3	21.51
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.08
MNØ	0.00
CAØ	2.11
NA2Ø	10.64
K2Ø	0.10
TØTAL	99.72

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.880
AL	1.118
AL	
TI	
FE3+	
FE2+	
MG	0.005
MN	
CA	0.100
NA	0.910
K	0.006

ENDMEMBERS

% AN	9.74
% AB	88.95
% ØR	0.55

034 PLAGIØCLASE 2

ØXIDE	WT %
SIØ2	65.75
AL2Ø3	21.35
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.13
MNØ	0.00
CAØ	2.03
NA2Ø	10.49
K2Ø	0.08
TØTAL	99.83

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.893
AL	1.107
AL	
TI	
FE3+	
FE2+	
MG	0.009
MN	
CA	0.096
NA	0.895
K	0.004

ENDMEMBERS

% AN	9.55
% AB	89.52
% ØR	0.44

034 PLAGIØCLASE 3

ØXIDE	WT %
SIØ2	65.50
AL2Ø3	21.42
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.09
MNØ	0.00
CAØ	1.98
NA2Ø	10.36
K2Ø	0.07
TØTAL	99.41

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.892
AL	1.115
AL	
TI	
FE3+	
FE2+	
MG	0.006
MN	
CA	0.094
NA	0.887
K	0.004

ENDMEMBERS

% AN	9.37
% AB	88.65
% ØR	0.39

034 PLAGIØCLASE 4

ØXIDE	WT %
SIØ2	65.36
AL2Ø3	21.71
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.09
MNØ	0.00
CAØ	2.19
NA2Ø	9.86
K2Ø	0.08
TØTAL	99.29

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.886
AL	1.130
AL	
TI	
FE3+	
FE2+	
MG	0.006
MN	
CA	0.104
NA	0.845
K	0.005

ENDMEMBERS

% AN	10.37
% AB	84.45
% ØR	0.45

033 PLAGIØCLASE 1

ØXIDE	WT %
SIØ2	65.50
AL2Ø3	21.96
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.10
MNØ	0.00
CAØ	2.29
NA2Ø	10.13
K2Ø	0.10
TØTAL	100.08

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.874
AL	1.136
AL	
TI	
FE3+	
FE2+	
MG	0.007
MN	
CA	0.108
NA	0.862
K	0.005

ENDMEMBERS

% AN	10.77
% AB	86.19
% ØR	0.54

033 PLAGIØCLASE 2	
ØXIDE	WT %
SIØ2	65.41
AL2Ø3	21.93
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.11
MNØ	0.00
CAØ	2.14
NA2Ø	10.50
K2Ø	0.06
TØTAL	100.15

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.871
AL	1.135
AL	
TI	
FE3+	
FE2+	
MG	0.007
MN	
CA	0.101
NA	0.893
K	0.003

ENDMEMBERS	
% AN	9.99
% AB	88.70
% ØR	0.33

033 PLAGIØCLASE 3	
ØXIDE	WT %
SIØ2	65.19
AL2Ø3	21.47
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.11
MNØ	0.00
CAØ	2.05
NA2Ø	10.04
K2Ø	0.09
TØTAL	98.95

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.889
AL	1.122
AL	
TI	
FE3+	
FE2+	
MG	0.007
MN	
CA	0.097
NA	0.862
K	0.005

ENDMEMBERS	
% AN	9.74
% AB	86.24
% ØR	0.52

033 PLAGIØCLASE 4	
ØXIDE	WT %
SIØ2	65.29
AL2Ø3	21.74
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.10
MNØ	0.00
CAØ	2.28
NA2Ø	10.06
K2Ø	0.09
TØTAL	99.55

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.879
AL	1.130
AL	
TI	
FE3+	
FE2+	
MG	0.007
MN	
CA	0.108
NA	0.860
K	0.005

ENDMEMBERS	
% AN	10.76
% AB	86.00
% ØR	0.51

033 PLAGIØCLASE 5A

ØXIDE	WT %
SIØ2	64.73
AL2Ø3	21.11
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.29
NA2Ø	9.93
K2Ø	0.00
TØTAL	98.07

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.895
AL	1.113
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.110
NA	0.861
K	

ENDMEMBERS

% AN	10.97
% AB	86.14
% ØR	0.00

033 PLAGIØCLASE 5D

ØXIDE	WT %
SIØ2	65.34
AL2Ø3	21.07
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.67
NA2Ø	9.97
K2Ø	0.00
TØTAL	99.04

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.897
AL	1.101
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.127
NA	0.857
K	

ENDMEMBERS

% AN	12.57
% AB	84.98
% ØR	0.00

033 PLAGIØCLASE 7A

ØXIDE	WT %
SIØ2	64.23
AL2Ø3	21.92
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.24
NA2Ø	10.21
K2Ø	0.00
TØTAL	98.61

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.862
AL	1.151
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.107
NA	0.883
K	

ENDMEMBERS

% AN	10.71
% AB	88.26
% ØR	0.00

033 PLAGIØCLASE 7C	
ØXIDE	WT %
SIØ2	63.75
AL2Ø3	22.16
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.50
NA2Ø	9.99
K2Ø	0.00
TØTAL	98.40

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.848
AL	1.167
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.119
NA	0.865
K	

ENDMEMBERS	
% AN	11.95
% AB	86.51
% ØR	0.00

0524 PLAGIØCLASE 1A	
ØXIDE	WT %
SIØ2	63.86
AL2Ø3	22.03
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.21
NA2Ø	10.42
K2Ø	0.00
TØTAL	98.52

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.852
AL	1.160
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.106
NA	0.902
K	

ENDMEMBERS	
% AN	10.55
% AB	88.16
% ØR	0.00

0524 PLAGIØCLASE 1D	
ØXIDE	WT %
SIØ2	62.31
AL2Ø3	23.04
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.52
NA2Ø	9.37
K2Ø	0.00
TØTAL	98.24

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.797
AL	1.219
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.169
NA	0.815
K	

ENDMEMBERS	
% AN	16.94
% AB	81.52
% ØR	0.00

0524 PLAGIØCLASE 2A

ØXIDE	WT %
SIØ2	63.74
AL2Ø3	22.76
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.24
NA2Ø	9.57
K2Ø	0.00
TØTAL	99.32

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.825
AL	1.189
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.154
NA	0.823
K	

ENDMEMBERS

% AN	15.41
% AB	82.28
% ØR	0.00

0524 PLAGIØCLASE 3A

ØXIDE	WT %
SIØ2	62.69
AL2Ø3	22.08
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.25
NA2Ø	9.46
K2Ø	0.00
TØTAL	97.48

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.832
AL	1.176
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.157
NA	0.829
K	

ENDMEMBERS

% AN	15.73
% AB	82.87
% ØR	0.00

0524 PLAGIØCLASE 3C

ØXIDE	WT %
SIØ2	62.60
AL2Ø3	22.68
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.24
NA2Ø	9.83
K2Ø	0.00
TØTAL	98.34

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.809
AL	1.200
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.156
NA	0.855
K	

ENDMEMBERS

% AN	15.22
% AB	83.48
% ØR	0.00

0524 PLAGIØCLASE 3D	
ØXIDE	WT %
SIØ2	63.47
AL2Ø3	21.89
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	2.72
NA2Ø	9.80
K2Ø	0.00
TØTAL	97.88

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.851
AL	1.160
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.131
NA	0.853
K	

ENDMEMBERS	
% AN	13.11
% AB	85.35
% ØR	0.00

0524 PLAGIØCLASE 4	
ØXIDE	WT %
SIØ2	62.83
AL2Ø3	22.54
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.17
NA2Ø	9.92
K2Ø	0.00
TØTAL	98.45

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.815
AL	1.191
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.152
NA	0.862
K	

ENDMEMBERS	
% AN	14.83
% AB	83.96
% ØR	0.00

0524 PLAGIØCLASE 5	
ØXIDE	WT %
SIØ2	63.06
AL2Ø3	22.56
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.16
NA2Ø	9.87
K2Ø	0.00
TØTAL	98.66

RECALCULATED ØN BASIS ØF 8.0 (Ø)	
SI	2.819
AL	1.189
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.151
NA	0.856
K	

ENDMEMBERS	
% AN	14.87
% AB	84.04
% ØR	0.00

0524 PLAGIØCLASE 6

ØXIDE	WT %
SIØ2	62.88
AL2Ø3	22.65
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.28
NA2Ø	9.61
K2Ø	0.00
TØTAL	98.41

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.816
AL	1.196
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.157
NA	0.834
K	

ENDMEMBERS

% AN	15.74
% AB	83.37
% ØR	0.00

0524 PLAGIØCLASE 7

ØXIDE	WT %
SIØ2	61.97
AL2Ø3	22.66
TIØ2	0.00
FE2Ø3	0.00
FEØ	0.00
MGØ	0.00
MNØ	0.00
CAØ	3.23
NA2Ø	10.05
K2Ø	0.00
TØTAL	97.91

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.798
AL	1.206
AL	
TI	
FE3+	
FE2+	
MG	
MN	
CA	0.156
NA	0.879
K	

ENDMEMBERS

% AN	14.81
% AB	83.38
% ØR	0.00

Ø479 SILLIMANITE

ØXIDE	WT %
SIØ2	35.82
AL2Ø3	60.33
TIØ2	0.02
FE2Ø3	0.61
FEØ	0.00
MGØ	0.35
MNØ	0.00
CAØ	0.00
NA2Ø	0.17
K2Ø	0.18
TØTAL	97.47

RECALCULATED ØN
BASIS ØF 5.0 (Ø)

SI	0.996
AL	
AL	1.977
TI	
FE3+	0.013
FE2+	
MG	0.014
MN	
CA	
NA	0.009
K	0.006

ENDMEMBERS

% ALSIL	98.87

0479 BIØTITE 1

ØXIDE	WT %
SIØ2	35.46
AL2Ø3	18.86
TIØ2	1.90
FE2Ø3	0.00
FEØ	17.44
MGØ	10.37
MNØ	0.17
CAØ	0.00
NA2Ø	0.41
K2Ø	9.10
TØTAL	93.70

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.436
AL	2.564
AL	0.844
TI	0.219
FE3+	
FE2+	2.236
MG	2.369
MN	0.022
CA	
NA	0.121
K	1.780

ENDMEMBERS

% PHLØG	39.49
% ANNITE	37.26
% M/M+F	51.45

0479 BIØTITE 2

ØXIDE	WT %
SIØ2	35.24
AL2Ø3	18.71
TIØ2	2.03
FE2Ø3	0.00
FEØ	17.01
MGØ	10.39
MNØ	0.15
CAØ	0.00
NA2Ø	0.37
K2Ø	9.12
TØTAL	93.01

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.435
AL	2.565
AL	0.837
TI	0.236
FE3+	
FE2+	2.194
MG	2.387
MN	0.019
CA	
NA	0.112
K	1.796

ENDMEMBERS

% PHLØG	39.79
% ANNITE	36.57
% M/M+F	52.11

0479 BIØTITE 3

ØXIDE	WT %
SIØ2	35.35
AL2Ø3	18.93
TIØ2	1.89
FE2Ø3	0.00
FEØ	17.32
MGØ	10.56
MNØ	0.15
CAØ	0.00
NA2Ø	0.36
K2Ø	8.97
TØTAL	93.54

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.422
AL	2.578
AL	0.844
TI	0.218
FE3+	
FE2+	2.221
MG	2.414
MN	0.019
CA	
NA	0.106
K	1.756

ENDMEMBERS

% PHLØG	40.24
% ANNITE	37.02
% M/M+F	52.08

0479 BIØTITE 4

ØXIDE	WT %
SIØ2	35.29
AL2Ø3	19.32
TIØ2	1.76
FE2Ø3	0.00
FEØ	16.91
MGØ	10.22
MNØ	0.14
CAØ	0.00
NA2Ø	0.36
K2Ø	8.95
TØTAL	92.94

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.432
AL	2.568
AL	0.938
TI	0.203
FE3+	
FE2+	2.176
MG	2.344
MN	0.019
CA	
NA	0.107
K	1.757

ENDMEMBERS

% PHLØG	39.07
% ANNITE	36.27
% M/M+F	51.85

0479 BIØTITE 5

ØXIDE	WT %
SIØ2	35.62
AL2Ø3	19.66
TIØ2	1.80
FE2Ø3	0.00
FEØ	16.89
MGØ	10.15
MNØ	0.15
CAØ	0.00
NA2Ø	0.35
K2Ø	8.89
TØTAL	93.52

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.437
AL	2.563
AL	0.975
TI	0.207
FE3+	
FE2+	2.157
MG	2.310
MN	0.020
CA	
NA	0.104
K	1.732

ENDMEMBERS

% PHLØG	38.50
% ANNITE	35.94
% M/M+F	51.72

0479 BIØTITE 6

ØXIDE	WT %
SIØ2	34.21
AL2Ø3	19.37
TIØ2	1.61
FE2Ø3	0.00
FEØ	17.19
MGØ	10.79
MNØ	0.15
CAØ	0.00
NA2Ø	0.35
K2Ø	8.56
TØTAL	92.22

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.320
AL	2.680
AL	0.871
TI	0.189
FE3+	
FE2+	2.235
MG	2.502
MN	0.019
CA	
NA	0.104
K	1.698

ENDMEMBERS

% PHLØG	41.69
% ANNITE	37.26
% M/M+F	52.81

0479 MUSCØVITE 1	
ØXIDE	WT %
SIØ2	45.16
AL2Ø3	36.08
TIØ2	0.92
FE2Ø3	0.00
FEØ	0.96
MGØ	0.78
MNØ	0.03
CAØ	0.00
NA2Ø	1.37
K2Ø	9.37
TØTAL	94.68

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.029
AL	1.971
AL	3.707
TI	0.092
FE3+	
FE2+	0.107
MG	0.155
MN	0.004
CA	
NA	0.355
K	1.596

ENDMEMBERS	
% MUSC	75.82
% PARAG	16.85
% CELAD	1.43
% M/M+F	59.20

0479 MUSCØVITE 2	
ØXIDE	WT %
SIØ2	45.21
AL2Ø3	36.31
TIØ2	0.78
FE2Ø3	0.00
FEØ	0.80
MGØ	0.71
MNØ	0.04
CAØ	0.00
NA2Ø	1.42
K2Ø	9.27
TØTAL	94.54

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.033
AL	1.967
AL	3.746
TI	0.078
FE3+	
FE2+	0.090
MG	0.141
MN	0.005
CA	
NA	0.366
K	1.579

ENDMEMBERS	
% MUSC	76.01
% PARAG	17.64
% CELAD	1.66
% M/M+F	61.10

0479 MUSCØVITE 3	
ØXIDE	WT %
SIØ2	44.93
AL2Ø3	36.32
TIØ2	0.72
FE2Ø3	0.00
FEØ	0.91
MGØ	0.76
MNØ	0.04
CAØ	0.00
NA2Ø	1.33
K2Ø	9.29
TØTAL	94.29

RECALCULATED ØN BASIS ØF 22.0 (Ø)	
SI	6.016
AL	1.984
AL	3.749
TI	0.073
FE3+	
FE2+	0.102
MG	0.152
MN	0.005
CA	
NA	0.344
K	1.586

ENDMEMBERS	
% MUSC	77.01
% PARAG	16.71
% CELAD	0.79
% M/M+F	59.84

0479 MUSCØVITE 4	
ØXIDE	WT %
SIØ2	45.33
AL2Ø3	35.97
TIØ2	0.78
FE2Ø3	0.00
FEØ	0.81
MGØ	0.77
MNØ	0.03
CAØ	0.00
NA2Ø	1.26
K2Ø	9.36
TØTAL	94.31

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.063
AL	1.937
AL	3.735
TI	0.079
FE3+	
FE2+	0.090
MG	0.153
MN	0.004
CA	
NA	0.326
K	1.597

ENDMEMBERS

% MUSC	77.52
% PARAG	15.85
% CELAD	3.15
% M/M+F	62.94

0479 PLAGIØCLASE 1	
ØXIDE	WT %
SIØ2	61.05
AL2Ø3	24.13
TIØ2	0.00
FE2Ø3	0.11
FEØ	0.10
MGØ	0.19
MNØ	0.03
CAØ	5.82
NA2Ø	8.49
K2Ø	0.10
TØTAL	100.01

RECALCULATED ØN BASIS ØF 8.0 (Ø)

SI	2.716
AL	1.266
AL	
TI	
FE3+	0.004
FE2+	0.004
MG	0.013
MN	0.001
CA	0.278
NA	0.732
K	0.006

ENDMEMBERS

% AN	27.18
% AB	71.66
% ØR	0.54

0479 PLAGIØCLASE 2	
ØXIDE	WT %
SIØ2	61.34
AL2Ø3	23.32
TIØ2	0.00
FE2Ø3	0.08
FEØ	0.07
MGØ	0.22
MNØ	0.05
CAØ	5.01
NA2Ø	8.83
K2Ø	0.07
TØTAL	98.98

RECALCULATED ØN BASIS ØF 8.0 (Ø)

SI	2.751
AL	1.233
AL	
TI	
FE3+	0.003
FE2+	0.003
MG	0.015
MN	0.002
CA	0.241
NA	0.768
K	0.004

ENDMEMBERS

% AN	23.66
% AB	75.54
% ØR	0.38

0479 PLAGIØCLASE 3

ØXIDE	WT %
SIØ2	61.83
AL2Ø3	23.64
TIØ2	0.00
FE2Ø3	0.05
FEØ	0.04
MGØ	0.20
MNØ	0.02
CAØ	4.97
NA2Ø	8.97
K2Ø	0.07
TØTAL	99.79

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.749
AL	1.239
AL	
TI	
FE3+	0.002
FE2+	0.002
MG	0.013
MN	0.001
CA	0.237
NA	0.773
K	0.004

ENDMEMBERS

% AN	23.19
% AB	75.76
% ØR	0.41

0479 PLAGIØCLASE 4

ØXIDE	WT %
SIØ2	61.26
AL2Ø3	23.50
TIØ2	0.00
FE2Ø3	0.04
FEØ	0.03
MGØ	0.13
MNØ	0.03
CAØ	5.02
NA2Ø	8.98
K2Ø	0.05
TØTAL	99.03

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.746
AL	1.242
AL	
TI	
FE3+	0.001
FE2+	0.001
MG	0.008
MN	0.001
CA	0.241
NA	0.781
K	0.003

ENDMEMBERS

% AN	23.35
% AB	75.68
% ØR	0.29

0479 PLAGIØCLASE 5

ØXIDE	WT %
SIØ2	61.16
AL2Ø3	24.16
TIØ2	0.00
FE2Ø3	0.12
FEØ	0.11
MGØ	0.21
MNØ	0.01
CAØ	5.39
NA2Ø	8.65
K2Ø	0.08
TØTAL	99.91

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.721
AL	1.267
AL	
TI	
FE3+	0.004
FE2+	0.004
MG	0.014
MN	0.001
CA	0.257
NA	0.746
K	0.005

ENDMEMBERS

% AN	25.26
% AB	73.39
% ØR	0.46

0462 GARNET 1

ØXIDE	WT %
SIØ2	37.32
AL2Ø3	20.48
TIØ2	0.00
FE2Ø3	1.14
FEØ	36.69
MGØ	2.55
MNØ	1.47
CAØ	1.67
NA2Ø	0.00
K2Ø	0.00
TØTAL	101.32

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.992
AL	0.008
AL	1.928
TI	
FE3+	0.069
FE2+	2.461
MG	0.304
MN	0.100
CA	0.143
NA	
K	

ENDMEMBERS

% PYRØPE	10.09
% ALM	81.60
% SPESS	3.31
% ANDR	3.42
% GRØSS	1.33
% M/M+F	11.00

0462 GARNET 2

ØXIDE	WT %
SIØ2	36.91
AL2Ø3	20.16
TIØ2	0.00
FE2Ø3	1.06
FEØ	36.92
MGØ	2.37
MNØ	1.98
CAØ	0.80
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.21

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	3.000
AL	
AL	1.932
TI	
FE3+	0.065
FE2+	2.510
MG	0.287
MN	0.137
CA	0.070
NA	
K	

ENDMEMBERS

% PYRØPE	9.55
% ALM	83.43
% SPESS	4.54
% ANDR	2.32
% GRØSS	0.00
% M/M+F	10.27

0462 GARNET 3

ØXIDE	WT %
SIØ2	37.17
AL2Ø3	20.31
TIØ2	0.00
FE2Ø3	1.11
FEØ	36.19
MGØ	2.43
MNØ	1.68
CAØ	1.81
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.71

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.998
AL	0.002
AL	1.930
TI	
FE3+	0.067
FE2+	2.442
MG	0.293
MN	0.115
CA	0.156
NA	
K	

ENDMEMBERS

% PYRØPE	9.72
% ALM	81.13
% SPESS	3.81
% ANDR	3.36
% GRØSS	1.83
% M/M+F	10.70

0462 GARNET 4

ØXIDE	WT %
SIØ2	36.59
AL2Ø3	20.44
TIØ2	0.00
FE2Ø3	0.86
FEØ	37.40
MGØ	2.26
MNØ	1.77
CAØ	0.78
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.09

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.982
AL	0.018
AL	1.945
TI	
FE3+	0.053
FE2+	2.549
MG	0.274
MN	0.122
CA	0.068
NA	
K	

ENDMEMBERS

% PYRØPE	9.04
% ALM	84.09
% SPESS	4.02
% ANDR	2.24
% GRØSS	0.00
% M/M+F	9.71

0462 GARNET 5A

ØXIDE	WT %
SIØ2	36.93
AL2Ø3	20.26
TIØ2	0.00
FE2Ø3	1.44
FEØ	35.71
MGØ	2.29
MNØ	2.31
CAØ	1.98
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.92

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.981
AL	0.019
AL	1.909
TI	
FE3+	0.087
FE2+	2.411
MG	0.275
MN	0.158
CA	0.171
NA	
K	

ENDMEMBERS

% PYRØPE	9.06
% ALM	79.44
% SPESS	5.21
% ANDR	4.31
% GRØSS	1.34
% M/M+F	10.24

0462 GARNET 5F CØRE

ØXIDE	WT %
SIØ2	36.63
AL2Ø3	20.25
TIØ2	0.00
FE2Ø3	1.15
FEØ	31.24
MGØ	1.41
MNØ	5.39
CAØ	3.94
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.01

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.983
AL	0.017
AL	1.927
TI	
FE3+	0.071
FE2+	2.127
MG	0.171
MN	0.372
CA	0.343
NA	
K	

ENDMEMBERS

% PYRØPE	5.63
% ALM	70.20
% SPESS	12.26
% ANDR	3.49
% GRØSS	7.84
% M/M+F	7.42

0462 GARNET 6

ØXIDE	WT %
SIØ2	36.47
AL2Ø3	20.34
TIØ2	0.00
FE2Ø3	1.15
FEØ	36.26
MGØ	2.35
MNØ	1.98
CAØ	1.43
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.98

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.972
AL	0.028
AL	1.927
TI	
FE3+	0.070
FE2+	2.471
MG	0.285
MN	0.137
CA	0.125
NA	
K	

ENDMEMBERS

% PYRØPE	9.36
% ALM	81.11
% SPESS	4.49
% ANDR	3.46
% GRØSS	0.65
% M/M+F	10.34

0462 GARNET 7

ØXIDE	WT %
SIØ2	36.23
AL2Ø3	20.15
TIØ2	0.00
FE2Ø3	1.49
FEØ	36.70
MGØ	2.29
MNØ	2.43
CAØ	0.77
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.05

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.962
AL	0.038
AL	1.905
TI	
FE3+	0.091
FE2+	2.510
MG	0.280
MN	0.168
CA	0.067
NA	
K	

ENDMEMBERS

% PYRØPE	9.13
% ALM	81.94
% SPESS	5.49
% ANDR	2.19
% GRØSS	0.00
% M/M+F	10.02

0462 GARNET 8

ØXIDE	WT %
SIØ2	36.44
AL2Ø3	20.28
TIØ2	0.00
FE2Ø3	0.81
FEØ	34.93
MGØ	2.02
MNØ	2.07
CAØ	2.54
NA2Ø	0.00
K2Ø	0.00
TØTAL	99.08

RECALCULATED ØN
BASIS ØF 12.0 (Ø)

SI	2.988
AL	0.012
AL	1.948
TI	
FE3+	0.050
FE2+	2.396
MG	0.247
MN	0.144
CA	0.223
NA	
K	

ENDMEMBERS

% PYRØPE	8.17
% ALM	79.29
% SPESS	4.76
% ANDR	2.48
% GRØSS	4.90
% M/M+F	9.34

0462 GARNET 9	
ØXIDE	WT %
SIØ2	36.40
AL2Ø3	20.25
TIØ2	0.00
FE2Ø3	1.27
FEØ	36.27
MGØ	2.27
MNØ	2.41
CAØ	1.16
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.02
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.971
AL	0.029
AL	1.919
TI	
FE3+	0.078
FE2+	2.475
MG	0.276
MN	0.167
CA	0.101
NA	
K	
ENDMEMBERS	
% PYRØPE	9.07
% ALM	81.18
% SPESS	5.46
% ANDR	3.32
% GRØSS	0.00
% M/M+F	10.05

0462 GARNET 10A	
ØXIDE	WT %
SIØ2	36.69
AL2Ø3	20.33
TIØ2	0.00
FE2Ø3	0.96
FEØ	36.26
MGØ	2.33
MNØ	2.22
CAØ	1.20
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.00
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.987
AL	0.013
AL	1.939
TI	
FE3+	0.059
FE2+	2.469
MG	0.283
MN	0.153
CA	0.105
NA	
K	
ENDMEMBERS	
% PYRØPE	9.37
% ALM	81.67
% SPESS	5.07
% ANDR	2.93
% GRØSS	0.54
% M/M+F	10.29

0462 GARNET 10B CØRE	
ØXIDE	WT %
SIØ2	36.83
AL2Ø3	20.37
TIØ2	0.00
FE2Ø3	1.28
FEØ	33.37
MGØ	2.71
MNØ	2.91
CAØ	2.70
NA2Ø	0.00
K2Ø	0.00
TØTAL	100.17
RECALCULATED ØN BASIS ØF 12.0 (Ø)	
SI	2.977
AL	0.023
AL	1.919
TI	
FE3+	0.078
FE2+	2.256
MG	0.327
MN	0.199
CA	0.234
NA	
K	
ENDMEMBERS	
% PYRØPE	10.75
% ALM	74.23
% SPESS	6.55
% ANDR	3.85
% GRØSS	3.86
% M/M+F	12.65

0462 STAURØLITE 1	
ØXIDE	WT %
SIØ2	26.67
AL2Ø3	54.32
TIØ2	0.52
FE2Ø3	0.00
FEØ	13.18
MGØ	1.33
MNØ	0.16
ZNØ	0.51
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.68

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.747
AL	
AL	8.998
TI	0.055
FE3+	
FE2+	1.548
MG	0.278
MN	0.020
ZN	0.053
NA	
K	

ENDMEMBERS	
% FE ST	77.42
% MG ST	13.88
% M/M+F	15.21

0462 STAURØLITE 2	
ØXIDE	WT %
SIØ2	26.99
AL2Ø3	55.72
TIØ2	0.43
FE2Ø3	0.00
FEØ	12.13
MGØ	1.32
MNØ	0.12
ZNØ	0.45
NA2Ø	0.00
K2Ø	0.00
TØTAL	97.16

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.746
AL	
AL	9.119
TI	0.045
FE3+	
FE2+	1.408
MG	0.272
MN	0.014
ZN	0.046
NA	
K	

ENDMEMBERS	
% FE ST	70.42
% MG ST	13.60
% M/M+F	16.19

0462 STAURØLITE 3	
ØXIDE	WT %
SIØ2	26.56
AL2Ø3	54.52
TIØ2	0.44
FE2Ø3	0.00
FEØ	13.43
MGØ	1.39
MNØ	0.12
ZNØ	0.43
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.89

RECALCULATED ØN BASIS ØF 23.0 (Ø)	
SI	3.726
AL	
AL	9.020
TI	0.046
FE3+	
FE2+	1.576
MG	0.290
MN	0.014
ZN	0.045
NA	
K	

ENDMEMBERS	
% FE ST	78.80
% MG ST	14.49
% M/M+F	15.53

0462 STAURØLITE 4

ØXIDE	WT %
SIØ2	26.93
AL2Ø3	55.23
TIØ2	0.45
FE2Ø3	0.00
FEØ	11.82
MGØ	1.40
MNØ	0.12
ZNØ	0.55
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.50

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.761
AL	
AL	9.094
TI	0.047
FE3+	
FE2+	1.381
MG	0.291
MN	0.014
ZN	0.056
NA	
K	

ENDMEMBERS

% FE ST	69.06
% MG ST	14.54
% M/M+F	17.39

0462 STAURØLITE 5 CØRE

ØXIDE	WT %
SIØ2	27.44
AL2Ø3	55.45
TIØ2	0.40
FE2Ø3	0.00
FEØ	11.66
MGØ	1.38
MNØ	0.11
ZNØ	0.49
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.93

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.806
AL	
AL	9.069
TI	0.042
FE3+	
FE2+	1.352
MG	0.285
MN	0.013
ZN	0.050
NA	
K	

ENDMEMBERS

% FE ST	67.61
% MG ST	14.27
% M/M+F	17.42

0462 STAURØLITE 6

ØXIDE	WT %
SIØ2	27.47
AL2Ø3	54.92
TIØ2	0.50
FE2Ø3	0.00
FEØ	12.13
MGØ	1.46
MNØ	0.13
ZNØ	0.58
NA2Ø	0.00
K2Ø	0.00
TØTAL	97.19

RECALCULATED ØN BASIS ØF 23.0 (Ø)

SI	3.813
AL	
AL	8.989
TI	0.052
FE3+	
FE2+	1.408
MG	0.303
MN	0.015
ZN	0.060
NA	
K	

ENDMEMBERS

% FE ST	70.42
% MG ST	15.14
% M/M+F	17.69

0462 STAURØLITE 7

ØXIDE	WT %
SIØ2	26.83
AL2Ø3	54.47
TIØ2	0.50
FE2Ø3	0.00
FEØ	13.19
MGØ	1.27
MNØ	0.12
ZNØ	0.48
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.86

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.761
AL	
AL	8.999
TI	0.052
FE3+	
FE2+	1.546
MG	0.265
MN	0.014
ZN	0.049
NA	
K	

ENDMEMBERS

% FE ST	77.31
% MG ST	13.27
% M/M+F	14.65

0462 STAURØLITE 8

ØXIDE	WT %
SIØ2	27.09
AL2Ø3	54.60
TIØ2	0.47
FE2Ø3	0.00
FEØ	12.52
MGØ	1.28
MNØ	0.12
ZNØ	0.46
NA2Ø	0.00
K2Ø	0.00
TØTAL	96.54

RECALCULATED ØN
BASIS ØF 23.0 (Ø)

SI	3.793
AL	
AL	9.013
TI	0.049
FE3+	
FE2+	1.466
MG	0.268
MN	0.014
ZN	0.047
NA	
K	

ENDMEMBERS

% FE ST	73.32
% MG ST	13.38
% M/M+F	15.43

0462 BIØTITE 1

ØXIDE	WT %
SIØ2	35.33
AL2Ø3	19.35
TIØ2	1.52
FE2Ø3	0.00
FEØ	19.10
MGØ	9.12
MNØ	0.04
ZNØ	0.00
NA2Ø	0.44
K2Ø	8.85
TØTAL	93.75

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.441
AL	2.559
AL	0.955
TI	0.176
FE3+	
FE2+	2.461
MG	2.093
MN	0.005
ZN	
NA	0.133
K	1.738

ENDMEMBERS

% PHLØG	34.88
% ANNITE	41.01
% M/M+F	45.96

0462 BIØTITE 2

ØXIDE	WT %
SIØ2	33.46
AL2Ø3	18.95
TIØ2	1.26
FE2Ø3	0.00
FEØ	20.58
MGØ	9.05
MNØ	0.06
CAØ	0.00
NA2Ø	0.33
K2Ø	8.50
TØTAL	92.19

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.303
AL	2.697
AL	0.843
TI	0.151
FE3+	
FE2+	2.728
MG	2.137
MN	0.008
CA	
NA	0.103
K	1.719

ENDMEMBERS

% PHLØG	35.61
% ANNITE	45.47
% M/M+F	43.92

0462 BIØTITE 3

ØXIDE	WT %
SIØ2	35.52
AL2Ø3	20.15
TIØ2	1.41
FE2Ø3	0.00
FEØ	16.59
MGØ	10.63
MNØ	0.04
CAØ	0.00
NA2Ø	0.40
K2Ø	8.76
TØTAL	93.50

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.407
AL	2.593
AL	1.023
TI	0.161
FE3+	
FE2+	2.112
MG	2.412
MN	0.005
CA	
NA	0.119
K	1.702

ENDMEMBERS

% PHLØG	40.21
% ANNITE	35.19
% M/M+F	53.33

0462 BIØTITE 4

ØXIDE	WT %
SIØ2	34.90
AL2Ø3	20.13
TIØ2	1.47
FE2Ø3	0.00
FEØ	17.47
MGØ	9.87
MNØ	0.05
CAØ	0.00
NA2Ø	0.36
K2Ø	8.86
TØTAL	93.11

RECALCULATED ØN
BASIS ØF 22.0 (Ø)

SI	5.370
AL	2.630
AL	1.022
TI	0.170
FE3+	
FE2+	2.248
MG	2.263
MN	0.007
CA	
NA	0.108
K	1.740

ENDMEMBERS

% PHLØG	37.71
% ANNITE	37.47
% M/M+F	50.16

0462 BIØTITE 5

ØXIDE	WT %
SIØ2	37.50
AL2Ø3	22.36
TIØ2	1.50
FE2Ø3	0.00
FEØ	16.63
MGØ	8.36
MNØ	0.05
CAØ	0.00
NA2Ø	0.33
K2Ø	8.76
TØTAL	95.48

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	5.532
AL	2.468
AL	1.421
TI	0.167
FE3+	
FE2+	2.052
MG	1.839
MN	0.007
CA	
NA	0.094
K	1.649

ENDMEMBERS

% PHLØG	30.64
% ANNITE	34.19
% M/M+F	47.26

0462 BIØTITE 6

ØXIDE	WT %
SIØ2	34.41
AL2Ø3	18.53
TIØ2	1.46
FE2Ø3	0.00
FEØ	20.64
MGØ	8.48
MNØ	0.06
CAØ	0.00
NA2Ø	0.40
K2Ø	8.61
TØTAL	92.59

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	5.423
AL	2.577
AL	0.865
TI	0.173
FE3+	
FE2+	2.720
MG	1.992
MN	0.007
CA	
NA	0.122
K	1.730

ENDMEMBERS

% PHLØG	33.20
% ANNITE	45.34
% M/M+F	42.27

0462 MUSCØVITE 1

ØXIDE	WT %
SIØ2	45.99
AL2Ø3	36.20
TIØ2	0.28
FE2Ø3	0.00
FEØ	0.90
MGØ	0.62
MNØ	0.03
CAØ	0.00
NA2Ø	1.95
K2Ø	8.59
TØTAL	94.57

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.114
AL	1.886
AL	3.787
TI	0.028
FE3+	
FE2+	0.101
MG	0.123
MN	0.003
CA	
NA	0.502
K	1.456

ENDMEMBERS

% MUSC	70.14
% PARAG	24.18
% CELAD	5.32
% M/M+F	55.04

0462 MUSCØVITE 2

ØXIDE	WT %
SIØ2	45.73
AL2Ø3	36.85
TIØ2	0.32
FE2Ø3	0.00
FEØ	0.66
MGØ	0.53
MNØ	0.03
CAØ	0.00
NA2Ø	1.74
K2Ø	9.03
TØTAL	94.89

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.063
AL	1.937
AL	3.824
TI	0.032
FE3+	
FE2+	0.073
MG	0.105
MN	0.004
CA	
NA	0.446
K	1.527

ENDMEMBERS

% MUSC	73.99
% PARAG	21.61
% CELAD	3.17
% M/M+F	58.75

0462 MUSCØVITE 3

ØXIDE	WT %
SIØ2	46.52
AL2Ø3	35.94
TIØ2	0.35
FE2Ø3	0.00
FEØ	1.01
MGØ	0.90
MNØ	0.03
CAØ	0.00
NA2Ø	1.57
K2Ø	9.03
TØTAL	95.34

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.143
AL	1.857
AL	3.737
TI	0.034
FE3+	
FE2+	0.112
MG	0.178
MN	0.003
CA	
NA	0.402
K	1.521

ENDMEMBERS

% MUSC	73.47
% PARAG	19.40
% CELAD	6.57
% M/M+F	61.36

0462 MUSCØVITE 4

ØXIDE	WT %
SIØ2	45.52
AL2Ø3	37.20
TIØ2	0.24
FE2Ø3	0.00
FEØ	0.63
MGØ	0.54
MNØ	0.03
CAØ	0.00
NA2Ø	1.85
K2Ø	8.64
TØTAL	94.65

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.038
AL	1.962
AL	3.856
TI	0.024
FE3+	
FE2+	0.070
MG	0.107
MN	0.003
CA	
NA	0.477
K	1.462

ENDMEMBERS

% MUSC	72.70
% PARAG	23.70
% CELAD	1.92
% M/M+F	60.47

0462 MUSCØVITE 5

ØXIDE	WT %
SIØ2	45.69
AL2Ø3	37.28
TIØ2	0.31
FE2Ø3	0.00
FEØ	0.82
MGØ	0.61
MNØ	0.01
CAØ	0.00
NA2Ø	1.88
K2Ø	8.52
TØTAL	95.13

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.032
AL	1.968
AL	3.834
TI	0.031
FE3+	
FE2+	0.091
MG	0.121
MN	0.001
CA	
NA	0.481
K	1.434

ENDMEMBERS

% MUSC	71.70
% PARAG	24.03
% CELAD	1.59
% M/M+F	57.14

0462 MUSCØVITE 6

ØXIDE	WT %
SIØ2	45.84
AL2Ø3	36.04
TIØ2	0.43
FE2Ø3	0.00
FEØ	0.90
MGØ	0.84
MNØ	0.03
CAØ	0.00
NA2Ø	1.70
K2Ø	8.81
TØTAL	94.58

RECALCULATED ØN BASIS ØF 22.0 (Ø)

SI	6.099
AL	1.901
AL	3.753
TI	0.043
FE3+	
FE2+	0.100
MG	0.166
MN	0.004
CA	
NA	0.438
K	1.495

ENDMEMBERS

% MUSC	72.57
% PARAG	21.24
% CELAD	4.96
% M/M+F	62.45

0462 PLAGIØCLASE 1

ØXIDE	WT %
SIØ2	67.22
AL2Ø3	20.17
TIØ2	0.00
FE2Ø3	0.12
FEØ	0.11
MGØ	0.21
MNØ	0.02
CAØ	0.94
NA2Ø	11.27
K2Ø	0.06
TØTAL	100.12

RECALCULATED ØN BASIS ØF 8.0 (Ø)

SI	2.945
AL	1.041
AL	
TI	
FE3+	0.004
FE2+	0.004
MG	0.014
MN	0.001
CA	0.044
NA	0.957
K	0.003

ENDMEMBERS

% AN	4.37
% AB	94.93
% ØR	0.31

0462 PLAGIØCLASE 2

ØXIDE	WT %
SIØ2	66.61
AL2Ø3	20.09
TIØ2	0.00
FE2Ø3	0.08
FEØ	0.07
MGØ	0.23
MNØ	0.01
CAØ	1.45
NA2Ø	11.32
K2Ø	0.05
TØTAL	99.91

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.931
AL	1.042
AL	
TI	
FE3+	0.003
FE2+	0.003
MG	0.015
MN	
CA	0.068
NA	0.966
K	0.003

ENDMEMBERS

% AN	6.43
% AB	91.09
% ØR	0.29

0462 PLAGIØCLASE 3

ØXIDE	WT %
SIØ2	63.98
AL2Ø3	21.80
TIØ2	0.00
FE2Ø3	0.09
FEØ	0.08
MGØ	0.18
MNØ	0.02
CAØ	3.23
NA2Ø	10.01
K2Ø	0.06
TØTAL	99.43

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.841
AL	1.141
AL	
TI	
FE3+	0.003
FE2+	0.003
MG	0.012
MN	0.001
CA	0.154
NA	0.862
K	0.004

ENDMEMBERS

% AN	14.96
% AB	83.87
% ØR	0.34

0462 PLAGIØCLASE 4

ØXIDE	WT %
SIØ2	65.93
AL2Ø3	20.79
TIØ2	0.00
FE2Ø3	0.10
FEØ	0.09
MGØ	0.19
MNØ	0.03
CAØ	1.75
NA2Ø	10.69
K2Ø	0.06
TØTAL	99.62

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.908
AL	1.081
AL	
TI	
FE3+	0.003
FE2+	0.003
MG	0.012
MN	0.001
CA	0.083
NA	0.914
K	0.003

ENDMEMBERS

% AN	8.25
% AB	91.10
% ØR	0.33

0462 PLAGIØCLASE 5

ØXIDE	WT %
SIØ2	65.07
AL2Ø3	21.18
TIØ2	0.00
FE2Ø3	0.07
FEØ	0.07
MGØ	0.18
MNØ	0.02
CAØ	1.87
NA2Ø	10.85
K2Ø	0.07
TØTAL	99.39

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.883
AL	1.106
AL	
TI	
FE3+	0.002
FE2+	0.002
MG	0.012
MN	0.001
CA	0.089
NA	0.932
K	0.004

ENDMEMBERS

% AN	8.56
% AB	89.98
% ØR	0.39

0462 PLAGIØCLASE 6

ØXIDE	WT %
SIØ2	64.11
AL2Ø3	21.49
TIØ2	0.00
FE2Ø3	0.07
FEØ	0.06
MGØ	0.22
MNØ	0.01
CAØ	2.77
NA2Ø	10.04
K2Ø	0.06
TØTAL	98.82

RECALCULATED ØN
BASIS ØF 8.0 (Ø)

SI	2.858
AL	1.130
AL	
TI	
FE3+	0.002
FE2+	0.002
MG	0.015
MN	
CA	0.133
NA	0.868
K	0.003

ENDMEMBERS

% AN	13.16
% AB	86.16
% ØR	0.34

0462 ANDALUSITE 1

ØXIDE	WT %
SIØ2	36.27
AL2Ø3	63.18
TIØ2	0.03
FE2Ø3	0.55
FEØ	0.00
MGØ	0.21
MNØ	0.03
CAØ	0.00
NA2Ø	0.10
K2Ø	0.02
TØTAL	100.39

RECALCULATED ØN
BASIS ØF 5.0 (Ø)

SI	0.978
AL	
AL	2.009
TI	0.001
FE3+	0.011
FE2+	
MG	0.008
MN	0.001
CA	
NA	0.005
K	0.001

ENDMEMBERS

% ALSIL	97.83

0462 ANDALUSITE 2

ØXIDE	WT %
SIØ2	36.53
AL2Ø3	63.55
TIØ2	0.03
FE2Ø3	0.47
FEØ	0.00
MGØ	0.23
MNØ	0.01
CAØ	0.00
NA2Ø	0.10
K2Ø	0.03
TØTAL	100.95

RECALCULATED ØN
BASIS ØF 5.0 (Ø)

SI	0.979
AL	
AL	2.009
TI	0.001
FE3+	0.010
FE2+	
MG	0.009
MN	
CA	
NA	0.005
K	0.001

ENDMEMBERS

% ALSIL	97.95

0462 CHLØRITE 1

ØXIDE	WT %
SIØ2	23.08
AL2Ø3	22.60
TIØ2	0.11
FE2Ø3	3.94
FEØ	23.70
MGØ	12.32
MNØ	0.10
CAØ	0.03
NA2Ø	0.10
K2Ø	0.04
TØTAL	86.03

RECALCULATED ØN
BASIS ØF 14.0 (Ø)

SI	2.502
AL	1.498
AL	1.390
TI	0.009
FE3+	0.321
FE2+	2.149
MG	1.990
MN	0.009
CA	0.004
NA	0.020
K	0.006

ENDMEMBERS

% MGCLIN	39.80
% FECLIN	42.97
% M/M+F	48.08

0462 CHLØRITE 2

ØXIDE	WT %
SIØ2	22.71
AL2Ø3	23.19
TIØ2	0.10
FE2Ø3	3.96
FEØ	23.88
MGØ	12.17
MNØ	0.10
CAØ	0.03
NA2Ø	0.11
K2Ø	0.05
TØTAL	86.29

RECALCULATED ØN
BASIS ØF 14.0 (Ø)

SI	2.457
AL	1.543
AL	1.415
TI	0.008
FE3+	0.323
FE2+	2.160
MG	1.962
MN	0.009
CA	0.004
NA	0.022
K	0.006

ENDMEMBERS

% MGCLIN	38.98
% FECLIN	42.92
% M/M+F	47.59

Table 8-2

Probe analyses of ilmenites from the Kreuzeck Group.

Wt% oxides	0462-ilm1	0462-ilm2	0479-ilm1
SiO_2	0.25	0.26	0.63
Al_2O_3	0.42	0.12	0.92
TiO_2	49.92	51.66	49.93
FeO	41.94	44.06	43.92
MgO	0.31	0.37	0.35
MnO	2.06	2.36	2.79
Total	94.90	98.83	98.54
Recalculated on basis of 3 (O) to atomic proportions			
Si	.007	.007	.017
Al	.014	.004	.028
Ti	.992	.990	.958
Fe	.927	.939	.937
Mg	.013	.015	.014
Mn	.047	.051	.061
% end-members, recalculated after Rumble (1973), C.M.P., 42, 181-95			
$FeTiO_3$	90.6	92.8	89.2
$MnTiO_3$	4.7	5.1	6.1
Fe_2O_3	0.0	0.0	0.0
$FeAlO_3$	1.4	0.4	2.8
$FeSiO_3$	0.7	0.7	1.7

TABLE 8-3: SAMPLE 0174, CARBONATE ANALYSES

	Weight %				Atomic Proportions**			
Analysis No.	CaO	MgO	FeO	Total*	Ca	Mg	Fe	C
Calcite 1	53.07	2.39	0.05	99.49	.945	.060	.001	.998
2	54.89	0.48	0.05	99.37	.983	.012	.001	1.003
3	54.97	1.08	0.04	100.04	.979	.027	.001	.998
4	52.70	2.15	0.05	98.88	.942	.054	.001	1.003
5	53.06	2.13	0.06	99.24	.947	.053	.001	1.000
6	53.57	1.95	0.06	99.56	.955	.049	.001	.999
7	52.98	1.95	0.08	98.99	.947	.049	.002	1.002
8	51.90	2.63	0.07	98.62	.929	.066	.002	1.003
9	54.06	0.65	0.04	98.72	.971	.017	.001	1.007
10	53.33	1.94	0.10	99.35	.952	.049	.002	1.000
11	51.94	1.93	0.08	97.95	.934	.049	.002	1.009
12	51.57	2.65	0.07	98.30	.924	.066	.001	1.005
13	50.65	3.02	0.05	97.83	.915	.078	.001	1.006
14	53.29	1.72	0.08	99.07	.953	.043	.002	1.002
15	53.02	1.64	0.05	98.71	.950	.041	.001	1.005

*includes theoretical carbon of 11.99 wt.% for calcite, 12.99 wt.% for dolomite

**recalculated on the basis of 3 oxygens.

TABLE 8-3 cont.

Analysis No	Weight %				Atomic Proportions**			
	CaO	MgO	FeO	Total*	Ca	Mg	Fe	C
Calcite 16	53.61	1.06	0.06	98.70	.962	.027	.001	1.006
17	52.24	1.88	0.05	98.17	.939	.047	.001	1.008
18	53.64	1.20	0.04	98.86	.962	.030	.001	1.005
19	53.93	0.95	0.04	98.88	.967	.024	.001	1.005
20	53.48	2.05	0.04	99.55	.953	.051	.001	.999
21	52.00	2.78	0.05	98.83	.929	.069	.001	1.002
22	53.81	1.38	0.03	99.19	.962	.035	.001	1.002
23	53.43	1.97	0.03	99.41	.953	.049	.001	1.000
24	54.07	0.25	0.04	98.32	.975	.007	.001	1.010
25	54.73	1.12	0.02	99.83	.976	.028	.001	.999
26	52.68	2.20	0.05	98.93	.942	.055	.001	1.002
27	51.91	2.77	0.04	98.72	.928	.069	.001	1.002
28	51.14	3.15	0.03	98.33	.915	.079	.001	1.004
29	51.79	2.20	0.01	98.01	.931	.055	.001	1.008
30	51.39	3.23	0.03	98.67	.918	.081	.001	1.002
31	53.37	1.66	0.03	99.04	.955	.042	.001	1.003
32	52.24	2.24	0.04	98.51	.936	.056	.001	1.005
33	52.32	2.30	0.05	98.66	.937	.058	.001	1.004
34	55.46	0.36	0.05	99.81	.991	.009	.001	1.001
35	54.56	0.65	0.04	99.21	.978	.017	.001	1.004
36	51.32	2.79	0.04	98.16	.920	.070	.001	1.006

TABLE 8-3 cont.

Analysis No.		Weight %				Atomic Proportions**			
		CaO	MgO	FeO	Total*	Ca	Mg	Fe	C
Calcite	37	53.53	1.87	0.07	99.44	.955	.047	.001	1.000
	38	52.44	2.62	0.07	99.12	.936	.065	.001	1.000
	39	53.04	1.75	0.03	98.81	.950	.044	.001	1.004
	40	53.99	1.89	0.04	99.90	.961	.047	.001	.997
Dolomite	1	30.17	19.65	0.15	97.57	.506	.459	.002	1.018
	2	29.92	19.65	0.17	97.36	.503	.459	.003	1.019
	3	29.92	19.99	0.14	97.67	.501	.466	.002	1.016
	4	29.88	19.96	0.13	97.59	.501	.466	.002	1.017
	5	30.03	19.63	0.19	97.47	.504	.459	.003	1.018
	6	29.95	19.42	0.17	97.16	.504	.455	.003	1.021
	7	30.60	19.71	0.09	98.01	.512	.459	.002	1.015
	8	29.84	19.04	0.33	96.82	.504	.447	.005	1.024
	9	30.29	19.95	0.10	97.97	.507	.464	.002	1.015
	10	29.80	19.60	0.08	97.10	.501	.459	.002	1.020
	11	30.21	19.86	0.04	97.74	.506	.463	.001	1.016
	12	30.25	19.85	0.06	97.78	.507	.463	.001	1.016
	13	29.96	19.50	0.06	97.15	.504	.456	.001	1.020
	14	30.39	19.80	0.01	97.82	.509	.461	.001	1.016
	15	30.13	19.67	0.17	97.59	.505	.459	.003	1.018
	16	30.27	19.62	0.15	97.65	.508	.458	.002	1.017
	17	30.33	19.60	0.04	97.59	.509	.458	.001	1.018

APPENDIX 2

Localities and petrographic descriptions of specimens used for K-Ar age determinations

07. Amphibolite

Locality: from irregular-shaped small body of amphibolite, 5 m above Polinik path, 300 m SSW of Polinik Hütte (D1).

Assemblage: hbl - chl - plag - clzo- white mica - gt

Interlocking prisms of hbl c. 1 mm long make up c. 85%, with small quantities of chl, interstitial plag, tiny flakes of white mica, and clzo rimming relict gt. Accessory rutile, rimmed by sphene.

028. Eclogitic amphibolite

Locality: from centre of 4 m diameter pod. c. 2280 alt. at N end of Mörningköpfe ridge, 150 m ESE of spot height 2447 m (C2).

Assemblage: hbl - gt - cpx

Matrix of stubby, 0.5 mm pale brown hbl ("carinthine") and 1.5 mm gt, packed with hbl inclusions; large (c. 1 cm) prismatic, poikilitic pale green cpx (omphacite) encloses hbl and gt. Rutile is a common accessory.

0125. Garnet - 2 mica Intermediate Schist

Locality: 2220 m alt. on Kleinblässen Alm, 750 m SE of Källukerl (D4).

Assemblage: gt - mu - bi - qz - chl

Most mu is in stubby, randomly oriented 0.2 mm laths, with occasional relict patches of parallel-felted texutre; bi, broad 0.5 mm laths; qz, 0.2 mm slightly strained, both interstitial to micas and in discrete stringers. Chl is probably

retrograde. Tour and ore are common accessories.

0197. Mylonite

Locality: from Kleinblässen shear zone, 220 m ESE of spot height 2467 m, 460 m SSW of Källukerl (C5).

Assemblage: ser - qz - chl

Composed of very fine sericite and chl with rounded and elongate qz clasts (up to 1 x 0.2 mm). Extinguishes parallel to mylonitic banding, reflecting strong preferred orientation of submicroscopic mica flakes. Accessory opaque material and deformed tourmalines.

0201. 2-mica Polinik Schist

Locality: by Polinik path 120 m ENE of Teuchlscharte, N of Tuechl shear zone (C4).

Assemblage: qz - plag - mu - bi

Mean grain size 0.4 mm, mica flakes 1.5 mm long. Strained qz (55%) and oligoclase (35%) with narrow flakes of mu and bi (5% each), mostly mimetic on a foliation folded around tight cm-scale folds. Accessory gt, ap and ore.

0204. Deformed chlorite-muscovite Intermediate Schist

Locality: from screen of less-deformed rocks in Teuchl shear zone, 60 m S of Teuchlscharte, 400 m NNE of Källukerl (C4).

Assemblage: qz - chl - mu - plag

Strongly strained qz, in 0.5 mm grains, and less common plag. 0.5 mm laths of mu and chl (both bent). Some chl probably replaces bi.

0206. Quartz-rich Intermediate Schist

Locality: at 2580 m alt., 100 m NNE of Kallukerl (C4).

Assemblage: qz - plag - mu - bi - clzo - gt

Rock is a 2-mica quartzite of 0.33 mm grain size with narrow, coarser-grained, mu-bi-plag-gt-rich bands, without preferred orientation of minerals. Small grains of clzo are quite common. Accessory tour, ap, rut.

0213. Amphibolite

Locality: outcrop in trees 50 m SE of Moll bridge (678 m alt.), SE of Sobriach.

Assemblage: hbl - plag - qz - chl - white mica.

A fine-grained amphibolite with narrow hbl prisms up to 1 mm long, with a strong preferred orientation. Minor interstitial plag + qz, occasional flakes of chl and white mica, and accessory rutile.

0215. Hbl - gt - 2 mica Polinik Schist

Locality: c. 780 m alt. in Sobriacher Giessbach, 310 m NNW of Sobriach church.

Assemblage: qz - ab - mu - bi - hbl - gt - clzo ! chl

Rock is a crudely foliated paragneiss; strained qz 0.2 mm drain; original plag now broken down to ab + mu + clzo; 0.5 mm garnets, prisms of blue-green hbl, and broad flakes of chl, mu and bi.

0217. Phyllonite

Locality: c. 815 m alt. in Söbriacher Giessbach, 385 m NNW of Söbriach church.

Assemblage: qz - mu - chl - ab

Qz strained, 0.15 mm grains; mu fabric deformed and partly recrystallised; 1 mm unstrained ab porphyroblasts grow over strongly oriented mica foliation. Chl in patches, after gt?, and in parallel oriented laths.

0218. Phyllonite

Locality: c. 820 m alt. in Söbriacher Giessbach, 400 m NNW of Söbriach church.

Assemblage: qz - mu - dol - chl

Bands composed of deformed, fine-grained, parallel-felted mu, partly recrystallised to random laths, alternate with bands of strongly-strained 0.3 mu qz, and dol. Accessory ore, ap and tour.

0255. Pelitic Intermediate Schist

Locality: 350 m SE of spot height 2363 m on Eisenriegel (C6).

Assemblage: qz - plag - mu - bi - st - gt

St sits in 3 mm diam. pools of sericite; gt is spongy, invaded and rimmed by bi and a little mu; qz is strongly strained and sutured; oligoclase is common in 0.5 mm grains. Accessory chl, tour and ore. Fairly strong mineral foliation makes high angle to crude compositional banding.

0260. Amphibolite

Locality: outcrop by lower path, 400 m SE of spot height 2363 m on Eisenriegel (C6).

Assemblage: hbl - plag - clzo

Rock is fine-grained, mostly c. 0.2 mm but some hbl prisms up to 1.5 x 0.5 mm. Blue-green hbl (80%), interstitial plag (15%), minor amounts of fuzzy clzo, accessory ore, white mica.

0280. Intermediate Schist with staurolite and garnet

Locality: near path, 150 m WNW of Bärenkopf, at 1940 m alt. in upper Teuchltal (C8).

Assemblage: qz - mu - gt - st - bi

Probably two generations of mu: folded parallel-felts and broader laths; bi uncommon; micas bent; gt and st up to 2 mm, st partly sericitised; qz c. 0.5 mm, strongly strained and sutured. Fabric deformed, rock cut by shear surfaces associated with Bärenkopf shear zone.

0282. Pelitic Intermediate Schist

Locality: 320 m WNW oc Plotschtrattenhütte, at 1820 m alt., above upper Teuchltal path.

Assemblage: qz - plag - mu - bi - gt - st - chl

Qz, 0.4 mm, strained; plag unstrained patches up to 1 mm across; gt to 6 mm, invaded by bi and chl; st in two generations (i) cracked and rimmed by chl + ser, (ii) euhedral, growing over folds in mu parallel-felts; two generations of mu, (i) folded parallel-felts,

(ii) broader recrystallised laths; bi

in broad, 1 mm laths.

K6. Kreuzeck Schist

Locality: on Heinrich Hecht Weg, 350 m WSW of Nassfeld Törl.

Assemblage: mu - bi - qz - gt - plag

Microfolded mu parallel-felts, overgrown by plag porphyroblasts; gt up to 5 mm, shattered, altered along cracks; bi in kinked, sometimes partly recrystallised 2 mm porphyroblasts. Platy ore is a common accessory.

K7. Kreuzeck Schist

Locality: by Heinrich Hecht Weg on Bratleiten Alm, 500 m SW of Seebach Törl.

Assemblage: mu - qz - chl - plag

Qz is in strained 0.5 mm grains; mu in parallel-felts, bent around microfolds. Chl, blue birefringence, occurs in the felts; dusty plag grows over felts; gt is pseudomorphed by chl and ? pale brown mica. Accessory ore and graphite trails in the mica felts.

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