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METAMORPHIC STUDIES

IN THE SOUTH-EAST TAUERN WINDOW, AUSTRIA

by

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SHORT ABSTRACT

Detailed structural and mineralogical mapping was undertaken in an area of 130 square kilometres within the Pennine Zone of the south-east Tauern Window, the Eastern Alps.

A coherent tectono-metamorphic history is established for this area. Three pre-Alpine metamorphic events are recognised. The Alpine history involves two discrete phases of intense penetrative deformation which occurred under conditions of rising temperature, a widespread phase of thermal crystallisation at 35-40 mybp representing the attainment of maximum temperatures, and a minor post-metamorphic phase of deformation. The regional tectonic significance of this sequence is discussed.

The spatial variation in Alpine peak metamorphic conditions is examined in terms of the distribution of mineral zones and isograds in metapelites. Staurolite+biotite assemblages were developed at low structural levels, and chloritoid+biotite assemblages along a thin wedge-shaped zone. Barrovian and Stonehavian facies series are both represented.

Oxygen isotope studies on whole-rocks and mineral separates suggest that widespread circulation of metamorphic fluids did not occur during Alpine metamorphism.

Thermodynamic calculations indicate that metamorphic conditions during the Alpine thermal peak were 570°C and 7 kb at the basement-cover interface. These results can be reconciled both with models in which tectonic burial alone was sufficient to cause Alpine metamorphism of the Pennine Zone and with models in which a transient high heat flow regime is invoked.

LONG ABSTRACT

In the Eastern Alps, rocks of the Pennine Zone are exposed within a large tectonic window (the Tauern Window) through a thick and extensive thrust mass of continental basement (the Austro-Alpine nappes). Two subdivisions of the Pennine domain are recognised: the Basement Complex, a series of ancient metamorphic rocks (the Inner Schieferhülle) intruded by granites and granodiorites of Permian age (the Zentralgneis), and the Peripheral Schieferhülle, a series of largely Mesozoic metasediments and metabasites.

The purpose of this thesis has been to establish a coherent structural and metamorphic history for the Pennine Zone in the south-east Tauern Window, and to estimate, as far as possible, the pressure and temperature conditions to which the rocks were subjected at each stage in their history, with particular emphasis on the later part of the Alpine orogenic cycle.

The spatial variation in metamorphic grade attained during the 35-40 mybp Alpine thermal peak was also examined.

Two adjacent areas, one of 100 sq. km centred on Mallnitz and one of 30 sq. km east of Obervellach, were mapped on a scale of 1:10000 and the results are presented at a scale of 1:15000. Within these areas, detailed structural and mineralogical investigations were carried out. Mineralogical mapping was also undertaken in various parts of a much larger area, mostly to the east of the map area.

Three pre-Alpine metamorphic events can be distinguished in the Inner Schieferhülle:

(i) a pre-Variscan eclogite-facies metamorphism (M_e) represented

by relict garnet+clinopyroxene parageneses preserved within metabasites outcropping over a very restricted area,

(ii) a Variscan amphibolite-facies regional metamorphism (M_v) represented by relict assemblages in both metapelites and metabasites,

(iii) a phase of high temperature contact metamorphism (M_z) associated with the intrusion of the Zentralgneis, the effects of which are only recognisable in certain migmatitic schists.

Although poorly constrained, the metamorphic conditions estimated for these three events are distinctly different from one another. M_v and M_e were accompanied by intense penetrative deformation.

An Alpine tectono-metamorphic history has been established; it involves two distinct major deformation events (D_a^1 and D_a^2) and a complex history of mineral growth culminating in a widespread post-D_a^2 phase of static crystallisation at ca. 35-40 mybp.

D_a^1 involved an early period of thrusting and imbrication during which slices of granitic continental basement and ultramafic oceanic basement were emplaced into the Peripheral Schieferhülle cover sequence. This developed into a phase of intense penetrative deformation confined mainly to the Peripheral Schieferhülle. D_a^1 is identified with the main phase of overthrusting of the Austro-Alpine nappes. The earliest recognisable Alpine mineral assemblages crystallised at this stage under greenschist facies conditions. The occurrence of a clinozoisite+margarite+quartz pseudomorph after lawsonite preserved within a pre-D_a^2 garnet in one rock suggests that part of the cover sequence was subjected to high-pressure low temperature metamorphism (<400°C) during or immediately after D_a^1, but other evidence of

eo-Alpine blueschist-facies metamorphism is entirely lacking. This is in direct contrast to the south-central Tauern where relict glaucophane-schists and eclogites are abundant.

D_a^1 was followed by a static period during which temperatures in the Pennine Zone increased in response to thermal relaxation following comparatively rapid tectonic burial. The second Alpine deformation phase, D_a^2, occurred while temperatures were still increasing. The strains experienced by the cover sequence during this event were much lower than during D_a^1, but the basement was involved to a much greater extent. Large asymmetric antiforms with cores of basement, and complementary synforms, were formed at this stage, but no major thrusts developed. The orientation of D_a^2 structures and the intensity of the deformation varied considerably over the two map areas.

Maximum temperature (and entropy) mineral assemblages were developed some time after D_a^2 movements ceased. Combined experimental work (from the literature) and thermodynamic calculation indicate that conditions during the 35-40 mybp thermal crystallisation were 570°C and 7 kb at the basement-cover interface east of Obervellach. Oxygen isotope data from separated minerals confirm the temperature estimate. Calculations also indicate that the metamorphic temperatures increased with structural depth, and this is confirmed by the arrangement of metamorphic zones and isograds in pelitic rocks. Staurolite+biotite assemblages are only developed in the basement and lowest levels of the Peripheral Schieferhülle east of Mallnitz. The occurrence in this area of a thin wedge-shaped chloritoid+biotite zone is taken as evidence for an isothermal gradient in pressure from higher values in the south-east to lower values in the north-west.

The oxygen isotopic composition of H_2O, and the $X_{H_2O}:X_{CO_2}$ ratios of late

Alpine high temperature metamorphic fluids were buffered by the compositions of the coexisting solid phases on scales as small as 1 to 10 cm. Most lithologies equilibrated with relatively H_2O-rich fluids. No widespread circulation of metamorphic fluids occurred while the main Alpine mineral assemblages were crystallising, and it is unlikely that significant heat was transported by convection, at least before the Alpine thermal peak.

Estimated peak metamorphic conditions were substituted into an already existing one-dimensional conductive thermal model. The results can be reconciled both with the theory that tectonic burial alone was sufficient to cause Alpine metamorphism of the Pennine Zone and with theories involving transient high heat flow regimes.

The thermal peak was followed by a period of rapid differential uplift and erosion. Regional updoming, centred on the Tauern Window resulted in the rotation of foliation surfaces and isograds to steep attitudes near the margins of the Tauern Window, particularly adjacent to the Möll1tal Line, a steep fault which developed at this stage. A minor, possibly polyphase, deformation event (D_a^3) probably developed in response to stresses generated during updoming. The minerals in one D_a^3 vein crystallised at 525°C and 2.1 kb.

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CHAPTER 1

INTRODUCTION

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INTRODUCTION

A. INTRODUCTION TO THE GEOLOGY OF THE EASTERN ALPS

The European Alps form an arcuate mountain chain, convex towards the north and west, stretching from the Ligurian Sea to Vienna. This is part of a comparatively young, semi-continuous orogenic belt which extends westwards through the Betic Cordillera of southern Spain to the Atlas mountains in north Africa, and eastwards, through the Carpathians and the mountains of Greece, Turkey and Iran, to the Himalayas.

The Alps are bounded on all sides by Tertiary sedimentary basins: the Molasse basin to the north, the Po basin to the south, and the Hungarian basin to the east. A large area of continental basement, the Bohemian Massif, outcrops to the north of the Molasse; it consists, essentially, of ancient gneisses and Palaeozoic sedimentary rocks, and forms the northern foreland of the Alps.

The Eastern Alps extend for about 500 km in an east-west direction, and occupy much of the area of Austria. They are laterally continuous with the Swiss Alps; the upper Rhine is normally taken as the boundary between the two. They are bounded to the north and west by the Molasse and Vienna basins respectively, and are separated from the Southern Alps by the Gailtal line, a major fault which correlates with the Insubric line further west.

The geology of the Eastern Alps has been summarised in English by Oxburgh (1968) whose nomenclature and tectonic classification are used in this study. A sketch map and cross-section are presented in Figs. 1.1 and 1.2.

The Eastern Alps can be conveniently subdivided into (a) a synformal Northern Zone, composed mainly of unmetamorphosed sedimentary rocks, which is an area of net basement depression, and (b) an antiformal Axial Zone, consisting mainly of metamorphic rocks, which is an area of net basement elevation.

The Northern Zone consists of a series of east-west trending fault-bounded slices of Mesozoic and Tertiary rock which appear to have been thrust northwards over the Molasse. Four major thrust-sheets are recognised, each composed of numerous subsidiary slices. Immediately overlying the Molasse, and outcropping to the south of it, is the Helvetic Zone, a sequence of lower Jurassic to Eocene marine sandstones and marls with an internal structure complicated by both folding and thrusting. The Helvetic Zone is, in turn, overlain by the Flysch Zone, a southward dipping unit composed of lower Cretaceous to Eocene pelagic sediments, predominantly calcareous. These two units are thought to be essentially parautochthonous, i.e. the Flysch Zone basin of deposition probably lay immediately to the south of the Helvetic Zone. The Northern Calcareous Alps form a thick allochthonous unit, consisting mainly of Triassic carbonate rocks from a variety of depositional facies, which overlies and outcrops to the south of the Flysch Zone. Rocks of the Flysch Zone are exposed in tectonic windows up to 20 km from the northern margin of the Northern Calcareous Alps, indicating that the latter must have been transported over distances greater than this. At their southern margin, the Northern Calcareous Alps rest, apparently unconformably, on the Grauwackenzone, a monotonous series of Palaeozoic quartz-phyllites and chloritic schists.

The Axial Zone is separated from the Northern Zone by a major fault, the North Tauern Line. Not only has the Axial Zone been arched on an east-west axis, but differential uplift has given rise to a series of culminations and depressions along its length; greater erosion in the culmination zones has resulted in the formation of tectonic windows through the uppermost units. Of the

three main tectonic windows (Engadine, Tauern, Wechsel), the Tauern is the largest and most studied.

Two major tectonic units are recognised in the Axial Zone:

(i) the Pennine Zone,

(ii) the East Alpine thrust sheets, or Austro-Alpine Zone.

The Pennine Zone is the structurally lower of the two, and, in the Eastern Alps, is only exposed in the tectonic windows mentioned above. In the Central and Western Alps, where the overlying Austro-Alpine units have largely been removed by erosion, the Pennine Zone is much better represented. Part of the Pennine Zone consists of a basement complex of polymetamorphic Variscan and older metamorphic rocks (the Inner Schieferhülle) intruded by large volumes of granite, granodiorite and tonalite of Permian age (the Zentralgneis). This is thought to be essentially autochthonous and to represent the southern continuation of the Bohemian Massif. Structurally overlying the basement complex is a predominantly allochthonous cover sequence of highly deformed and metamorphosed sedimentary and basic volcanic rocks of Mesozoic age (the Peripheral Schieferhülle). A wide variety of sedimentary compositions and depositional facies are represented in the cover sequence, ranging from pelagic types (e.g. calc-schists, which are compositionally similar to much of the Flysch and are commonly associated with serpentinites and metabasites) to epicontinental types (calcareous quartzites, dolomites, marbles and evaporites). Norris et al. (1971a) interpret the Peripheral Schieferhülle as a tectonic mélange.

The East Alpine thrust-sheets overlie the rocks of the Pennine Zone and occupy the greater part of the Axial Zone. The Lower East Alpine sheet, or Unterostalpin, rests directly on the rocks of the Pennine Zone and is exposed round the margin of the Tauern Window. It consists mainly of a highly disrupted series of fossiliferous sediments of low metamorphic grade, including carbonate rocks and black shales, together with some slices of crystalline schists and

gneisses. The quartz-phyllites which form the Lower East Alpine sheet round the eastern margin of the Tauern Window are thought to be of Palaeozoic age (Hawkesworth, 1976). Around the southern margin, however, this sheet is represented by a comparatively narrow zone (the Matrei Zone) containing slices and lenses of highly sheared metasedimentary rocks resembling low-grade equivalents of the Peripheral Schieferhülle.

The Lower East Alpine sheet is overlain by the Altkristallin sheet, which is the thickest and most extensive allochthonous unit in the Eastern Alps. The greater part of it consists of pre-Mesozoic schists and gneisses (the Altkristallin, sensu stricto) which are overlain by patches of low grade or unmetamorphosed Palaeozoic and Mesozoic sedimentary rocks, showing a variety of structural relationships both towards each other and towards the underlying metamorphic rocks.

Two of these sediment patches are of particular interest from the point of view of reconstruction of pre-tectonic geographical relationships: (a) East of the Tauern Window, the Altkristallin is overlain unconformably by a series of sedimentary rocks known as the Stangalm Mesozoic. These, in turn, underlie the Gurktal sheet, a thrust slice of Palaeozoic quartz-phyllites resembling those of the Grauwackenzone. (b) Immediately north of the Gailtal line, similar quartz-phyllites rest directly on the Altkristallin but are separated from it by a thrust-fault. These are overlain, apparently unconformably, by a thick sequence of Triassic carbonate rocks (the Drauzug) extremely similar to the Northern Calcareous Alps.

Opinion is divided as to the identity of the original sedimentary cover of the Altkristallin. One view is that all the sedimentary rocks now resting on the Altkristallin are essentially autochthonous and, together with the Northern Calcareous Alps, represent a once-widespread cover sequence (e.g. Kober, 1955). In this interpretation, tectonic intercalation of Mesozoic and Palaeozoic sedi-

ments is of local significance only. The other principal view is that the Central Alpine Mesozoic (e.g. the Stangalm Mesozoic) are the only autochthonous sediments, while the Drauzug, Northern Calcareous Alps and Gurktal sheet were formed to the south of the Altkristallin sheet and have since overridden it (e.g. Tollmann, 1963).

Several metamorphic events have been recognised in the rocks of the Axial Zone; these are summarised in Fig. 1.2. Of particular significance is the conclusion that the Pennine Zone and Altkristallin had very different metamorphic histories prior to the Tertiary (35 - 40 my) Alpine amphibolite facies event, which affected them both. Although both the Inner Schieferhülle and the Altkristallin show evidence of Palaeozoic metamorphic events, the latter have not experienced any of the widespread intrusion of Permian granites characteristic of the Pennine basement in the Eastern, Central and Western Alps.

The Pennine Zone and Altkristallin sheet both consist essentially of a pre-Mesozoic continental basement overlain by a Mesozoic sedimentary cover sequence. The fact that the Altkristallin sheet now lies structurally above the Pennine Zone indicated that there has been extensive stratal shortening, possibly also crustal shortening, during the evolution of the Alps. In recent tectonic syntheses (e.g. Oxburgh, 1972; Dewey and Bird, 1970) the Alps have been interpreted in terms of inter-continental collision between a Eurasian plate, now represented by the Bohemian Massif and Pennine basement, and an African plate, now represented by the Altkristallin. The presence of deep-water sediments and serpentinites in the Helvetic Zone, Flysch Zone and Peripheral Schieferhülle is taken as confirmatory evidence that an ocean basin once existed between the two. In terms of the plate theory, the lithosphere beneath such an ocean would have had to have been largely destroyed by subduction for continental collision to have taken place. There is no general concensus of opinion on the orientation of the Benioff Zone. Oxburgh (1972)

postulates that subduction occurred down a northerly dipping zone beneath the Pennine basement, and that the Altkristallin sheet formed as a flake cut from the top part of the African plate. Hawkesworth et al. (1975), on the other hand, argue that the questionably contemporaneous eo-Alpine blueschist event in the Peripheral Schieferhülle and the 80-100 my low pressure amphibolite facies event in the Altkristallin (B and C respectively in Fig. 1.3) represent the two halves of a paired metamorphic belt; this would be more consistent with subduction towards the south. In this interpretation, the Altkristallin is identified with the attenuated leading edge of the African plate.

Most rocks within the Tauern Window appear to have been extensively recrystallised by the 35 to 40 my greenschist to amphibolite facies regional metamorphism (A in Fig. 1.3), which correlates with the Lepontine phase in the Central Alps. For the most part, this appears to be the only metamorphism to have affected the Peripheral Schieferhülle cover sequence, although some cover rocks in the Central Tauern preserve evidence of earlier high pressure metamorphism (B in Fig. 1.3) in the form of relict eclogite and glaucophane-schist-facies assemblages (Miller, 1974, 1977; Holland, 1977 and in press).

It is generally agreed that the tectonic burial of the Pennine Zone beneath the East Alpine thrust sheets was largely responsible for both Alpine deformation and metamorphism in the Eastern Alps. What is not clear, however, is whether burial alone could have been sufficient to induce amphibolite facies conditions in the buried rocks, or whether it is necessary to postulate some additional source of heat. Karl (1959) and Exner (1964) argue that a large body of magma existed beneath the present level of erosion during metamorphism. This is geologically quite reasonable because several Tertiary tonalitic plutons have been identified (the Peri-Adriatic suite of intrusives), one of which outcrops at Wöllaner Kopf only 4 km south of the margin of the Tauern Window.

Recently, several workers (e.g. Oxburgh and Turcotte, 1974; Bickle et al., 1975; England, 1976, 1978) have approached the problem quantitatively by applying one-dimensional heat-conduction models to the Eastern Alps. All the models predict, other factors being equal, higher equilibrium temperatures at higher lithostatic pressures for a given heat flow. Norris et al. (1971c) calculated a pressure of 5 kb and a temperature of 550°C for the base of the Peripheral Schieferhülle; Oxburgh and Turcotte substituted these values into their thermal model and concluded that Alpine metamorphism was accompanied by an enhanced heat-flow from below, possibly through a higher-than-average contribution from the mantle. The validity of the above pressure estimate, which was calculated by summing the thicknesses of the overlying units, was questioned by Waters (1976) who deduced a pressure of about 5.5 kb for the base of the Altkristallin during the 80 to 100 my metamorphism (implying that the East Alpine thrust sheets must, at that stage, have been thicker than was previously thought). From this, Bickle et al. (1975) concluded that burial alone could have been sufficient to induce amphibolite facies metamorphism in the Pennine Zone. Nevertheless, the problem remains essentially unresolved, due to the large uncertainties in the calculated pressure and temperature conditions.

B. THE AREA STUDIED IN THIS THESIS

The principal aim of this thesis is to establish the pressure and temperature conditions attained by the Alpine amphibolite facies metamorphism in the Pennine Zone.

The area chosen for study is located in the south-east corner of the Tauern Window (Fig. 1.1) where some of the deepest structural levels in the Eastern Alps are exposed. The area lies entirely within the Pennine Zone, and Peri-

pheral Schieferhülle, Zentralgneis and Inner Schieferhülle are all represented.

The basement complex outcrops in a series of dome structures (Kernen) generally occupying central positions in the eastern Tauern. Peripheral Schieferhülle rocks form an envelope surrounding these domes, but are also exposed in tight synformal structures (Mulden) between them. Parts of three major gneiss-domes outcrop within the area of study: the Hochalm Kern, Hölltor Kern, and Sonnblick Kern. The geographical relationship between them is shown in Fig. 1.4. The Hochalm Kern is the main massif of Penninic basement in the eastern Tauern, and contains both Inner Schieferhülle and Zentralgneis components. These are not distinguished in Fig. 1.4 because they have complex geometrical inter-relationships. The other Kernen are composed predominantly of Zentralgneis.

The Mallnitzer Mulde, Mölltal Schieferhülle and Ankogel Mulde occupy the structural depressions between the Kernen mentioned above. The first two are laterally continuous and trend north-west, while the third has a northerly or north-easterly trend; they appear to intersect near the town of Mallnitz.

Towards the south-east the Sonnblick Kern tapers into an elongate tongue, the Sonnblick Lamella; this body extends for some 15 km along the Mölltal parallel to the edge of the Tauern Window, and appears to be concordant to the Peripheral Schieferhülle rocks which flank it. Exner (1962) believes that several other thin lamellae of gneiss (not shown in Fig. 1.4) also occur within the cover sequence overlying and flanking the Sonnblick Kern. A tongue-like body of syenitic Zentralgneis, the Romate-decke, flanks the south west side of the Hölltor Kern and is separated from it by a screen of mica-schists of uncertain age known as the Woisgen Zone.

The selected area has already been mapped by Angel and Staber (1952), Exner

(1962, 1964) and Cliff et al. (1971) (Fig. 1.5). Cliff et al., who worked in the central part and southern margin of the Hochalm Kern, also studied the structural geology and geochronology in detail, and made preliminary investigations into the metamorphic petrology. Thus, although parts of the area have received comparatively little attention, a great deal is already known about it as a whole; this is the main reason for choosing it to study metamorphism.

The following geological problems pertinent to the selected area must be solved before any attempt can be made to estimate metamorphic conditions and substitute them into thermal models:

(i) to what extent to the preserved mineral assemblages reflect the 35 to 40 my Alpine metamorphism, and to what extent are early Alpine high pressure parageneses and pre-Alpine assemblages preserved in the Peripheral and Inner Schieferhülle?

(ii) What are the relative time relationships between deformation and mineral growth?

(iii) Does any lateral variation in Alpine metamorphic grade exist, and if so, how is it related to macroscopic structures?

(iv) Is there any evidence of widespread migration of metamorphic fluids, e.g. through veins, that could invalidate the commonly made assumption that heat transfer was purely conductive?

C. METHODS OF STUDY

Three field seasons (1975, 1976, 1977) were spent mapping and collecting structural data and samples for laboratory study. About 300 specimens were examined in thin section. Photographic enlargements of the 1:50000 Österreichische Draukraftwerke map at a scale of 1:10000 were used in mapping.

The results of the lithological mapping are presented in maps 1 and 2. Map 1, which was begun in 1975, covers about 30 square kilometres and lies entirely within the area studied by Cliff et al. (1971) (Fig. 1.5); it is essentially an agglomeration of small areas in which fairly detailed mineralogical mapping was undertaken. However, the structures in this area were found difficult to interpret, and in 1976 the decision was taken to begin mapping in the adjacent area (map 2) where, it was felt, the structures were more clearly displayed. Map 2, which covers about 100 square kilometres, lies mostly to the west of Mallnitz and within the area studied by Exner (1964); it includes parts of the Sonnblick Kern, Mallnitzer Mulde, Hölltor Kern, Ankogel Mulde and Hochalm Kern.

The lithologies encountered in the two map areas are described in Chapter 2, with emphasis on field characteristics and petrography. A list of mineral assemblages is given for each lithology. A limited amount of geochemical work was undertaken to compare the compositions of highly deformed metagranites in the gneiss lamellae with those of relatively undeformed metagranites from the Sonnblick Kern; this is also presented in Chapter 2.

The next part of the thesis attempts to define a coherent tectono-metamorphic history for the rocks in the map areas by drawing together material from a combined structural and petrographic study. Chapter 3 deals with the analysis and correlation of mesoscopic and macroscopic deformation structures throughout the two map areas. The emphasis here is on the establishment of

a relative time sequence of structural events for the south-east Tauern rather than the calculation of finite strains involved. One important conclusion arising from it is time-relationships between the various large-scale structures (gneiss domes, lamellae, Mulden, etc.). In Chapter 4, the structural history deduced in Chapter 3 is used to construct a metamorphic history. Mineral textures are examined in more detail here than in Chapter 2, and sequences of mineral growth are determined by observing the textural relationships between the constituent minerals and microfabric in critical rocks, and relating the microfabrics to mesoscopic structures of known relative age.

The next stage in the development of the thesis is the estimation of metamorphic conditions during late Alpine metamorphism. A qualitative approach is employed in Chapter 6; this examines the spatial variation in Alpine metamorphic grade as revealed by the distribution of metamorphic zones and isograds in metapelites and, to a lesser extent, metabasites. This work is based on mineralogical mapping both within the two map areas and elsewhere in the south-east Tauern; the results are presented in Map 3. In Chapter 7, the pressures and temperatures which prevailed during each major episode of metamorphic mineral growth are calculated, but with emphasis on the thermal peak of the Alpine metamorphism. A thermodynamic approach was used for this. As none of the thermodynamic theory is original, most of it is treated as reference material and summarised in Appendix II. Chapters 6 and 7 both require knowledge of mineral compositions; mineral analyses are listed in Appendix I and discussed in Chapter 5.

The oxygen isotope studies discussed in Chapter 8 provide not only additional estimates of Alpine metamorphic temperature but also information on the variation of metamorphic fluid composition in space, and hence on the scale over which fluid migration has taken place. In Chapter 9, the results of a

combined oxygen isotope and fluid inclusion study on one late-stage vein are used to calculate the pressure and temperature conditions late in the Alpine metamorphic cycle.

The various conclusions of this thesis are drawn together in the final chapter.

CHAPTER 2

Description of lithologies

CHAPTER 2

DESCRIPTION OF LITHOLOGIES

A. INTRODUCTION

The aim of this chapter is to describe the field relations and list the mineral assemblages and important petrographic features of lithologies encountered in the area of study.

To reduce the size of this task, and to avoid repetition of the observations of Exner (1957, 1964) and Cliff et al. (1971), the lithologies described here will, in general, include only those found in the two map areas. Emphasis will be placed upon features which have not been described elsewhere. The contents of this chapter should not, therefore, be regarded as comprehensive, and the reader is referred to the papers mentioned above for details of lithologies not covered here.

The textural relationships of minerals in thin-section will be amplified in Chapter 4.

The petrographic work is based almost entirely on the study of thin-sections. Minerals have been identified optically with the exception of some of the white micas. The presence of paragonite could only be established by X-ray diffractometry or microprobe analysis. Not all micaceous rocks have been analysed using these techniques, but in those that were, paragonite was found rarely and only in aluminous pelites. In the absence of evidence to the contrary, it will be assumed that the white-mica is muscovite. A list of mineral abbrevia-

tions is given in Table 2.1.

Grainsizes are classified as follows: fine-grained: up to ~1 mm; medium-grained: ~1 to ~2 mm; coarse-grained: larger than 2 mm.

B. INNER SCHIEFERHÜLLE LITHOLOGIES

Introduction

This very heterogeneous group represents the country rocks into which the Zentralgneis granites were intruded. They are therefore the oldest rocks in the Tauern Window. Cliff et al. (1971) have accumulated much evidence to show that the Inner Schieferhülle had undergone at least one phase of metamorphism and one, possibly two, phases of deformation prior to intrusion. Furthermore, these phases of deformation and metamorphism must have happened in or before Permian times as even the latest and least deformed members of the Zentralgneis give Rb/Sr whole rock ages of around 240 my (Cliff and Oxburgh, 1971).

The evidence that any particular rock type belongs to the Inner Schieferhülle has been based primarily on the recognition of an older foliation truncated by an intrusion of granitic material which may itself be variously deformed. Discordant angular relationships are best seen at low structural levels within the basement complex (i.e. towards the centre of the Hochalm Massif) where subsequent deformation has not been intense. However, much of the present study relates to rocks at higher structural levels where subsequent deformation has been intense and any such angular relationships will have been destroyed. Exner (1954) refers to these interbanded schists and granite-gneisses as the Randgneis. At high structural levels it is often difficult to

distinguish between basement and cover rocks. In places, particularly in the mica-schists in the Mallnitz area, the author has adopted rather subjective criteria to do this.

In the absence of discordant intrusive granites, these are:

(i) the presence of concordant bands of granite-gneiss (excluding quartz-phengite-schist);

(ii) the occurrence of relict minerals attributable to pre-Alpine metamorphism;

(iii) the presence of intercalated bands of non-calcareous graphitic quartzite (part B(g), below);

(iv) the absence of calcareous lithologies such as marble, calc-schist and calcareous quartzite (part D(c), (d) and (f), below);

(v) the occurrence of quartz + garnet veins (part B(f), below).

Rocks for which criterion (ii) applies must, be definition, belong to the basement, but unfortunately clear-cut examples are rare. Criterion (iv) holds generally for the Tauern Window (Frasl and Frank, 1966; Norris et al., 1971a). Criterion (i) is a logical extrapolation to strongly deformed terrains of the intrusive criterion mentioned above. Criterion (iii) is based on the fact that graphitic quartzites occur within undisputable Inner Schieferhülle migmatites in the Hochalm Kern, and (v) is purely empirical.

Field Relations and Petrography of Inner Schieferhülle Lithologies

(a) Banded Gneiss

This lithology is the main constituent of the Inner Schieferhülle in the area around Reisseck but in the map areas it is only important in Dösener (north-east corner of Map 1). Varieties can be identified with compositional banding of different thicknesses. Finely-banded gneiss is a distinctive rock in the field, with dark and light bands alternating on a scale from several cm to 1 mm or less (Fig. 2.1). At many localities these are truncated by relatively undeformed granite-aplite veins and must therefore have been generated in a pre-Alpine deformational event. Grey banded gneisses are characterised by less distinct banding repeated on a 10 to 100 cm scale. Deformed K-feldspar mega-crysts (augen) commonly occur, elongated in the plane of foliation.

The following assemblages appear to be in equilibrium:

(1) biot + ep $\pm$ chl + plag + qz + sph + ore

(2) biot + ep + plag + ksp $\pm$ musc + qz + sph + ore

The first is typical of the finely-banded varieties. Quartz and oligoclase-andesine form a fine-grained granulose matrix in which evenly sized aligned biotites and highly birefringent epidotes define a schistosity. This is usually but not always parallel to the banding, produced by variation in modal biotite.

(b) Dark Biotite-Schist

This lithology has only been identified in Map 2, and is probably a melano-cratic type of banded gneiss. It outcrops on the north-eastern margin of the

Sonnblick Kern along a single horizon to the south-east of Böseck, but is not represented on the ridge between Rissieck and Türl Kopf. It reaches a maximum thickness (ca. 200 m) 600 m E.S.E. of Rissieck.

The rock is dark, medium grained, poorly-foliated and rich in biotite and chlorite, the former usually more abundant. Finely banded rocks of the same mineralogy occur locally. Concordant layers of fine-grained biotite-granite-gneiss and fine-grained finely-banded biotite-amphibolite are occasionally found, although the latter is the dominant lithology in the exposures to the east of Steiner Alm.

(c) Quartzo-Feldspathic Gneiss with Amphibolite Lenses

A single band of this occurs along strike from the dark biotite-schist, extending north-west from Böseck to beyond Feldsee Scharte (Map 2).

The matrix is a massive, fine-grained, unbanded, off-white rock, superficially resembling leucocratic granite-gneiss, composed mainly of quartz and feldspar with subordinate muscovite providing a weak schistosity. However, the dominant coloured mineral is chlorite, and tiny garnets are occasionally visible. The only specimen studied in thin section contains the assemblage:

$$gt + chl + biot + ep + musc + qz + ab \pm cc + mt + ap$$

The minerals appear to be in equilibrium. Texturally they bear a close resemblance to those in the paragneisses described in part (d), the main difference being the presence of small amounts of calcite and a larger proportion of quartz and feldspar in this case.

Within this matrix, bands and lenses of metabasic rock commonly occur ranging

in thickness from a few centimetres to several metres. Most thin bands are composed of fine-grained black amphibolite (Fig. 2.2) whereas thicker ones tend to be of chlorite-schist with abundant albite porphyroblasts. The latter probably represent amphibolites reconstituted during intense Alpine deformation under fairly low-grade metamorphic conditions (see Chapter 4, part C(b)).

Exner (1964) maps this lithology together with the dark biotite-schist as "Amphibolitische Schollenmigmatit".

(d) Schistose Feldspathic Paragneiss

This lithology outcrops in a band about 300 m thick directly below the Peripheral Schieferhülle in Kaponigtal and on the Gröneck ridge (Map 1). It also occurs as thin concordant bands within basement amphibolites.

These paragneisses tend to be rich in ellipsoidal plagioclase porphyroblasts, often concentrated in layers or rods. Biotite is ubiquitous and forms a schistosity usually accompanied by muscovite. Small garnets and needles of hornblende may be visible in hand specimen.

The following assemblages have been recorded:

(1) biot + hbl + ep + chl + musc + qz + plag + mt

(2) gt + biot + hbl + ep + musc + qz + plag + mt + tour $\pm$ ap

(3) gt + biot + hbl + ep + chl + musc + qz + plag + mt

(4) gt + biot + ep + chl + musc + qz + plag + mt + ap

In thin section the garnets form ~1 mm sub/euhedral crystals, generally free of inclusions but sometimes enclosing quartz and epidote. The plagioclase and quartz together make up roughly 80% of each rock. The plagioclase (oligoclase-

andesine) forms irregular patches associated with small granules of iron-rich epidote or lenticular porphyroblasts enclosing quartz, biotite and epidote. In some rocks the biotite and muscovite schistosity "flows round" lenticular plagioclases and garnets, but in others the garnets appear to be cross-cutting. Euhedral prisms of blue-green hornblende occur in localised bands. Chlorite appears to have grown in these rocks as an alteration product of biotite, but in the gneiss (part (c) above) it appears to be primary.

Coexisting hornblende and muscovite are rare in the south-east Tauern and may be restricted to unusual bulk compositions. These rocks combine characteristics of pelites and amphibolites and may therefore represent meta-tuffs, which would also explain their presence as bands within amphibolite.

(e) Amphibolite

This lithology forms a major part of the Inner Schieferhülle in all three map areas. Norris et al. (1971a) subdivide the amphibolites into the following groups:

(i) Fine-grained banded amphibolites,

(ii) Coarse feldspar-amphibolites,

(iii) Coarse massive biotite-amphibolites,

(iv) Garnet-amphibolites,

(v) Actinolite-amphibolites.

In addition, a sixth group is recognised in the present study:

(vi) Eclogitic amphibolites.

The relative abundance of these types varies enormously. Most of those encountered in the map areas belong to (1), (4) and (6) so only these will be

described here.

Fine-grained banded amphibolites: These are the most abundant in the areas examined. They tend to be massive, black, homogeneous rocks composed mainly of small 1 to 2 mm hornblende prisms forming a good schistosity and, often, a pronounced mineral lineation. Thin bands of leucocratic amphibolite and garnetiferous feldspathic gneiss (qv.) occur parallel to the schistosity.

Amphibolites of this type outcrop in the north-east corner of Map 2. Part of this outcrop extends southwards into Dösener Tal (Map 1) forming a band nearly 2 km thick abruptly terminated by an east-west trending vertical fault. Fine grained banded amphibolites predominate in all but the middle part of this band.

Garnet-amphibolites: These strongly resemble those described above (within which they are often interbanded) but carry garnet and tend to be coarser grained. The garnets measure up to 4 mm across and usually have a spongy appearance. In some rocks they have clearly developed reaction rims of plagioclase.

Rare, thin (5-10 cm) bands of dense, pink leucoamphibolite, extremely rich in garnet, occur within normal garnetiferous and finely banded amphibolites towards the eastern end of the Dösener Tal section.

Eclogitic Amphibolites: These are restricted to an area of about 2 sq. km within the large amphibolite mass in Dösener Tal (Map 1). They occupy roughly the middle portion of the thickest part of this mass. They are best exposed in the footpath and adjacent Dösen Bach gully 500 m east of the Konrad Hütte. Bands of eclogitic amphibolite, usually no more than 5 m thick, are intercalated within more "normal" garnetiferous amphibolites.

In hand specimen, eclogitic amphibolites are dense, fine-grained and generally paler than other types. They tend to be either homogeneous or finely banded on a 1 to 2 mm scale (Fig. 2.3). Very abundant tiny (0.5 mm) garnets are set in a groundmass of black hornblende and pale greyish green material (an alteration product of omphacite) or, less commonly, grass-green omphacite. Poikiloblasts of albite up to 4 mm diameter occur in all these rocks. Varieties with abundant fresh omphacite tend to contain less hornblende and fewer albites than those with symplectite.

At one locality (alt. 1640 m in Dösen Bach gully) the compositional banding is cut at a low angle by an unfoliated biotite-granite-aplite dyke 1 m thick. This suggests that these eclogites were produced before the intrusion of the Zentralgneis.

Other types of amphibolite occur rarely in the map-areas. An outcrop of coarse ophitic meta-gabbro-pegmatite occurs about 300 m west of the Schmidt-Haus (Map 1). This is cut by 30 to 50 cm dykes of fine-grained amphibolite, and both are cut by granite-aplite veins. The meta-gabbro-pegmatite body (300 x 150 m) is enveloped almost completely by streaky feldspar-amphibolite, but it is not clear whether it is the undeformed equivalent of the latter or whether it represents an intrusion into already deformed basement gneisses (i.e. an early basic phase of Zentralgneis magmatism) as is suggested by monometamorphic textures in thin section described below.

Petrography of the amphibolites:

The following assemblages have been recorded:

(1) gt + cpx + hbl + plag + qz + rut + pyr + ore + ap

(2) gt + cpx + hbl + ep + plag + qz + rut + pyr $\pm$ cre $\pm$ ap

(3) gt + cpx + hbl + ep + plag + qz + sph + pyr + ore + ap

(4) gt + cpx + hbl + ep + plag+ qz + rut + sph + pyr + ore

(5) gt + cpx + hbl + ep + qz + rut + sph + pyr

(6) gt + hbl + ep + qz + rut + sph + ore

(7) gt + hbl + ep + qz + sph

(8) gt + hbl + ep + plag + qz + rut $\pm$ sph + ore $\pm$ ap

(9) gt + hbl + ep + biot + plag + qz + rut + ore + ap

(10) gt + hbl + ep + biot + chl + plag + qz $\pm$ sph + ore $\pm$ ap

(11) gt + hbl + ep + plag + qz + sph + pyr + ore + ap

(12) gt + hbl + ep + biot + plag + qz + pyr + ore

(13) gt + hbl + act + ep + sph + mt + pyr

(14) act + biot

(15) act + zo + qz + pyr + ore

(16) act + zo + plag + qz

(17) act + ep + biot + chl + pyr

(18) hbl + plag + pyr + ore

(19) hbl + ep + qz + rut + pyr

(20) hbl + biot + chl + plag + qz + ore

(21) hbl + ep + biot + plag + qz $\pm$ sph + pyr $\pm$ ap

(22) hbl + biot + plag + qz + sph + ore

(23) hbl + ep + biot + chl + plag + rut + sph + pyr + ap

(24) hbl $\pm$ chl + plag + qz + sph + pyr + ore

Assemblages (1) to (8) are characteristic of eclogitic amphibolites. In these, garnet commonly exceeds 20 mode percent. (1), (2) and (5) contain omphacite and preserve the primary eclogite mineralogy; garnet, omphacite, quartz and rutile. This assemblage without additional phases was observed in a small patch in one slide (GD 136a). Here random 1 mm prisms of pale green omphacite, altering to dark material along grain boundaries, enclose numerous tiny 0.1 to 0.5 mm interconnected garnets (ca. 30 mode percent) and sporadic grains of quartz and rutile (Fig. 2.4). Omphacites in other rocks

tend to be smaller (0.2 to 0.5 mm) prisms set between euhedral garnets, apparently also in equilibrium, and have been partly replaced by very fine-grained symplectitic intergrowths of clinopyroxene and albite in which green hornblendes have grown. In GD 136a, the fresh eclogite patch is surrounded by rock rich in hornblende + albite symplectite in which relict omphacites are clearly rimmed by coronas of garnet (Fig. 2.5). These garnets are clearly of a different generation to those in the fresh patch.

Poikiloblasts of albite occur in eclogitic amphibolites both with and without fresh omphacite. They are more abundant in the latter where they are associated with euhedral garnets with spongy cores (Fig. 2.6). Garnets are as abundant as inclusions within albites as outside them, whereas hornblende and pyroxene only occur as small ragged inclusions or not at all and appear to have broken down to produce albite.

Assemblages (3) and (4) occur in eclogitic amphibolites with no fresh omphacite but with abundant clinopyroxene + albite symplectite (Fig. 2.7). These varieties grade into types with assemblages (6) to (9) differing from the garnet-amphibolites described below only in their greater abundance of garnet, lack of hornblende schistosity and the presence of patches of hornblende-albite-symplectite.

Assemblages (9) and (10) are characteristic of most garnet-amphibolites. The hornblende usually has a well developed schistosity and the garnets tend to be ragged, often partially replaced by oligoclase-andesine either as reaction rims (e.g. GD 135, Fig. 2.8) or as cores to atoll garnets (e.g. GD 476).

Fig. 4.2 attempts to summarize the textures and correlate phases of mineral growth in rocks ranging from fresh eclogite through to garnet amphibolite collected from the same area. Its main conclusion is that basement amphi-

bolites contain two distinct generations of garnet, both of which are to some extent breaking down.

Assemblages (11) and (12) were formed in garnet-rich leucoamphibolites. The modal proportion of garnet ranges from about 20% to over 60% and two generations are often present. Rounded 1 mm diameter garnets are accompanied by numerous darker 0.1 to 0.2 mm ones, and are themselves zoned. These two generations have a different textural relationship than those described above in that here the small postdate the large. Euhedral hornblende poikiloblasts enclose both types (Fig. 2.9).

Also restricted to rare horizons within the Düsener Tal outcrop are magnesian amphibolites with assemblages (14) to (17). Actinolite is the characteristic amphibole and appears to be in equilibrium with uniaxial (+ ve) orthozoisite in GD 105 and GD 112. The chlorite in GD 130 (assemblage (17)) is a colourless Mg-rich variety, evidently part of the stable assemblage. In all other occurrences in the Inner Schieferhülle amphibolites the chlorite is a green variety and has grown from the retrogressive breakdown of garnet, biotite or hornblende.

Assemblages (20), (24) and (21) were found in rocks with well-preserved igneous textures. The first two are from the meta-gabbro-pegmatite mentioned above. They are coarse-grained, undeformed and retain ophitic textures (Fig. 2.10a), but the original pyroxene is now pseudomorphed by hornblende. The original euhedral plagioclases, while retaining their shape and twinning, have become filled with tiny zoisite needles (Fig. 2.10b) presumably as a result of readjustment to lower calcium content during metamorphism. Assemblage (21) was found in a fine-grained granulose amphibolite containing random euhedral 1 to 1.5 cm plagioclase megacrysts (Fig. 2.11a). The rock (GD 96) probably originally a feldspar-phyric basalt, was collected from an isolated exposure near

the meta-gabbro and may represent one of the basic dykes. The matrix is an equigranular mosaic of hornblende, labradorite (An 62%), biotite and quartz grains. The plagioclase megacrysts have recrystallised marginally to clear polygonal labradorite but elsewhere contain needles of zoisite (Fig. 2.11b), so the original phenocrysts must have been even more calcic. These two types of basement amphibolite differ from others in that they appear to have only experienced one phase of metamorphism and no deformation. In this respect they resemble parts of the Zentralgneis.

(f) Mica-Schist

Within the Hochalm massif, mica-schist commonly occurs in thin (~50 m) screens between masses of granite-gneiss, tonalite, etc. Large outcrops of schist up to 700 m wide occur near the Zwenberger Seen and in the upper Gössgraben (north east of Map 1). Many of these schists have been heavily injected by the Zentralgneis granites and merit the name migmatite. Norris et al. (1971a) describe a gradation in rocks from the Obere Zwenberger See from schists with virtually no granitic material to migmatites consisting of about 80% granite in which thin wisps of schist are suspended.

Another important outcrop occurs in the area around Mallnitz (Map 2). As a whole, it appears to contain schists from both Peripheral and Inner Schieferhülle. The former are found at high structural levels associated with calc-schists and marbles. The latter occur immediately above augen-gneisses of the Zentralgneis and form the Randgneis, together with thick sheets of syenite-gneiss. The boundary between these two types of schist is not obvious in the field. Generalisations will therefore be restricted to the lowest unit (the Woisgen Zone, Exner, 1956) between the augen gneisses and the main syenite-gneiss outcrop. Here, coarse-grained schists and feldspar-rich paragneisses

contain occasional 5 m thick sheets of biotite-granite-gneiss with contacts generally parallel to the schistosity in the host rock (Fig. 2.12). Small angular discordances have been found (less than 5°); at one locality a bifurcating gneiss band was observed.

Flattened quartz veins and stringers are abundant in the schists of the Randgneis (and the cover), but rare in the migmatites. In the Mallnitz area, many of these quartz veins contain numerous 1 mm or smaller garnets either distributed with a uniform concentration (up to 30%) throughout the veins or concentrated towards the margins. Quartz-garnet-veins are almost entirely restricted to garnet-bearing host-rocks. In rare instances, the host-rock adjacent to thick garnet-free quartz veins or lenses is greatly enriched in garnet (50 to 80% in GD 667).

Petrography of the mica-schists:

The following assemblages have been recorded:

(1) biot + chl + qz + ep + plag + ap

(2) biot + chl + musc + qz + ep + plag $\pm$ ap

(3) gt + biot + chl + musc + qz $\pm$ ep + plag + ilm $\pm$ tour + ap

(4) gt + biot + musc + qz + ep $\pm$ plag + ilm $\pm$ tour + ap

(5) ky + biot + chl + musc + qz + plag + sph

(6) gt + biot + musc + ep + plag

(7) gt + ky + biot + musc + qz + plag + ksp + ilm

(8) gt + st + biot $\pm$ musc + qz + ep + plag + ilm + mt + ap

(9) gt + st + ky + biot + chl + musc + qz + ep + plag + ilm $\pm$ tour

(10) gt + st + ky + biot + chl + musc + parag + plag + ilm + ap

(11) st + ky + parag + qz + plag

(12) gt + st + ky + biot + chl + musc + ilm + ap

(13) gt + st + biot + chl + musc + ilm + ap

(14) gt + musc + parag + qz + plag + ilm

(15) biot + musc + cc + ep + ore + ap

Of these, only (1), (3), (4), (13) and (14) appear to be in equilibrium. Assemblages (1) to (4), representing low-alumina pelites and semipelites, are most common.

Rocks with assemblages (1) and (2) tend to be dark, coarse grained and rather flaggy, with prominent flakes of biotite forming a good schistosity. They are particularly abundant in the Ankogel Mulde. Schists containing assemblage (3) from the Mallnitz area tend to have a very well developed muscovite and biotite schistosity contrasting with the rather random mica fabric of those from the Hochalm massif. Two generations of garnet are discernible in schists from both areas. Small (0.1 to 0.5 mm) euhedral garnets coexist with larger (2 to 5 mm) ones which show evidence of two stages of growth; rounded cores and euhedral rims. In GD 1192 the boundaries between cores and rims are sharp enough to give a Becke line, the cores having higher relief (richer in iron or manganese). Some schists from the Mallnitz area contain only large garnets, often rimmed or partially pseudomorphed by chlorite. Plagioclase generally forms small polygonal grains texturally similar to the coexisting quartz, and appears to be in equilibrium (Fig. 2.13). Chlorite occurs either as flakes aligned in the schistosity or as cross-cutting sheaves.

One schist (GD 1230b, assemblage (3)) from Klein Woisgen Kopf (Map 2) contains an isoclinally folded quartz-garnet vein. In the host rock, small (0.5 mm) pale, elongated garnets, containing granules of opaque material lie either parallel to the schistosity or at an angle to it between quartz pressure shadows. In the vein, random elongated garnets with aligned inclusion trails coexist with granular zoned (orange cores, thin pale rims) garnets with no

inclusion fabric (Fig. 2.14). Both garnet and quartz are more abundant within the vein than without; micas and chlorite are less abundant. Modal analyses are presented in Table 2.2. The conclusions drawn from these textures are: (a) some garnets in the vein originally grew in schist and have been "digested out", (b) other garnets, possibly primary vein minerals, also occur and (c) the vein was formed before D_a^1 isoclinal folding (see Chapter 3).

Assemblage (5) was found in one rock (C 523) from south of Kaponigtörl (Map 1). It represents a bulk composition much richer in magnesium than iron. Pale phlogopitic biotite forms abundant large (up to 3 mm) unstrained flakes mimetically recrystallised round a small fold. Long blades of kyanite, containing phlogopite inclusions, have been broken and bent round the same fold and are altered marginally to chlorite, muscovite and oligoclase. The chlorite, a colourless variety, forms in radiating sheaves associated with muscovite. An initial assemblage of kyanite and phlogopite has therefore been replaced by phlogopite, Mg-chlorite, muscovite and oligoclase. The presence of numerous aligned muscovite inclusions in oligoclase suggests that muscovite breakdown might have contributed to its formation. Another magnesian schist (GD 80b, assemblage (2)) resembles C 523 except that texturally it seems to have attained a greater degree of equilibrium, and kyanite is absent.

Assemblages (8) to (12), representing aluminous pelites, have only been found in a few localities in Map 1: on the eastern shore of Dösener See on the ridge 500 m E.N.E. north of Ochladin Spitze, in the west wall of Kaponigtal. Other localities have been found within the schist outcrop south of the Villacher Hütte, within the main schist band in the Gössgraben and in the migmatites in upper Riekengraben (see Plate 1, Cliff et al., 1971).

In hand specimen, these schists are fine-grained silvery-grey rocks in which patches of plagioclase and 1 cm sheaves of chlorite are the most obvious por-

phyroblasts. Large garnets with diameters up to 1.5 cm are conspicuous in some specimens (fig.2.15). In many, quartz is rare or absent, and in these the mica is decussate. In rocks with more quartz, the micas tend to have a well developed schistosity cross-cut by flakes of mica in random orientation. Two generations of garnet are present: tiny, subhedral "pin-prick" grains (all samples) and large porphyroblasts (most samples). The latter consist of rounded cores containing straight trails of biotite or quartz inclusions at an angle to the external schistosity, and thin knobbly rims dusted with opaque material (Figs. 4.3 and 2.16). The rims are about as wide as the "pin-prick" grains (often also dusted) and probably represent the same growth phase. Staurolite and kyanite occur as randomly orientated euhedral prisms overgrowing "pin-prick" garnets and cutting across any existing mica schistosity (Fig. 2.18). In three slides, composite crystals were found in which staurolite is twinned with kyanite, the composition plane being {100} in kyanite and {010} in staurolite (Fig. 2.17). Deer et al. (1963) mention that this is quite a common phenomenon. In GD 208d two textural types of kyanite coexist, apparently in equilibrium. Individual euhedral prisms are scattered throughout the rock, while lenticular aggregates (2 x 0.5 mm) of tiny aligned needles lie in the plane of schistosity (Fig. 2.18), reminiscent of the wisps of fibrolite found in pelites of high metamorphic grade.

Large patches of calcic oligoclase have grown after garnet, staurolite and kyanite apparently only at the expense of muscovite. Radiating sheaves of chlorite also enclose the high-grade phases (Fig. 2.21).

Ilmenite is the characteristic ore mineral in these schists, although in a few it is accompanied by magnetite. It has a platy habit and its texture mimics that of the surrounding mica.

One sample from the Kaponigtal locality (GD 209a) consists largely of

assemblage (10). Textures are as described above except that quartz is totally absent and paragonite contributes to the decussate mica fabric. Assemblage (11) forms a 1 cm band along one side of the specimen. Most of this is fine-grained, monomineralic schistose paragonite overgrown by euhedral prisms of staurolite (up to 1 cm long) and kyanite (Fig. 2.19). A single stringer of quartz is armoured from the paragonite in places by large ragged oligoclase porphyroblasts (Fig. 2.20). Assemblage (14) was found in a fine-grained white micaceous quartzite (GD 208e), also from this locality. Quartz and oligoclase form polygonal grains elongated parallel to the schistosity. In this rock, plagioclase is clearly in equilibrium with muscovite and paragonite which are present in about equal amounts.

One rock with assemblage (9) was found 1.3 km W.N.W. of Mallnitz in the Woisgen zone, 5 m above the contact with underlying augen-gneisses. The kyanite forms ragged blades, breaking down to muscovite and chlorite (Fig. 2.22), between rounded 1 cm fractured garnets, also altering to chlorite. 2 mm flakes of orange biotite and tiny anhedral staurolites are associated with kyanite and have also been altered to chlorite and muscovite. All sheet silicates have been severely kinked.

(g) Graphitic Quartzite

This lithology outcrops in bands up to 5 m thick within mica schists of the Ankogel and Mallnitzer Mulde. In the latter, it does not occur in units structurally higher than the Feldwand-lamella (part C(d), below).

Two 5 m thick parallel bands outcrop on Ebeneck and probably correlate with the two exposed 1¾ km N.N.E. of the Jamnig Alm. The graphitic quartzite exposed on the path leading northwards from the Hagener Hütte may also

correlate with one or both of these. A 50 cm band of this lithology was found within the migmatitic schists in Riekengraben (Map 1). This is the only a priori evidence for placing these quartzites in the Inner Schieferhülle.

These rocks are very distinctive in the field owing to their dark-grey to coal-black colour. They invariably have a fine-grained granulose, almost hornfelsic texture, but localised slightly micaceous partings give them a well-spaced (1 to 5 cm) non-penetrative fissility which results in their characteristically flaggy appearance where exposed. Flattened white quartz veins, sometimes isoclinally folded, commonly occur parallel to this fissility.

One specimen was studied in thin section. It has the assemblage:

qz + graph + mt + ep + musc + biot

The quartz grains are remarkably uniform in size (0.3 mm) and shape; polygonal and slightly elongated parallel to the mica schistosity. Tiny granules of graphite are concentrated along quartz-quartz grain boundaries. Rounded grains of magnetite are abundant in some layers. Magnetite, epidote and micas occupy no more than 5%.

ZENTRALGNEIS

Introduction

Norris et al. (1971a) take the Zentralgneis to include the granitic and granodioritic component of the basement complex. These authors have shown that, in the Hochalm massif, most of these acid plutonic rocks were intruded into schists and gneisses of the Inner Schieferhülle around 270 my bp. These rocks generally show evidence of only one phase of metamorphism, and

nearly all contain relict or pseudomorphed igneous minerals, particularly feldspars.

In this section description is restricted mainly to the Zentralgneis in the are area covered by Map 2. As any intrusive relationships here have largely been obliterated, lithologies have been assigned to this group purely on the basis of their similarity, in outcrop and in thin section, to rocks from "type localities" in the Hochalm Kern. This task is made more difficult by the occurrence of intermediate and basic meta-igneous lithologies peculiar to this area.

Field Relations and Petrography of Zentralgneis Lithologies

(a) Augen Gneiss

This lithology forms much of both the Sonnblick Kern and the Hölltor Kern (Map 2), and is also an important constituent of the Hochalm Kern (Map 1). On Map 2 no distinction is made between these and the subordinate augen-free biotite granites and granite-gneisses.

The augen gneisses are characterized by 1-2 cm K-feldspar crystals of significantly coarser grain size than the quartzo-feldspathic matrix in which they are embedded. They vary in texture from coarse-grained, poorly foliated rocks, in which biotite is the dominant coloured mineral, to medium-grained, schistose rocks with biotite absent, or subordinate to, phengitic muscovite. In the latter there is a more pronounced difference in grain size between the matrix and the K-feldspar augen. These more schistose gneisses appear to have undergone a greater degree of strain.

Within the Sonnblick Kern, these varieties outcrop with a systematic zonal distribution. The more massive types are restricted to the central parts while the schistose types are found mainly towards the margins. Approaching Rissieck from the south (Map 2), massive biotite-augen-gneisses gradually give way to types richer in phengite with a good, penetrative lineation. North of Rissieck, 200 m from the contact with meta-sediments of the Mallnitzer Mulde, biotite becomes rare and the rocks developed a good schistosity, crenulated in places. About 100 m from this contact both biotite and large K-feldspar augen disappear, and the rocks can best be described as quartz-phengite-schists. Similar mineralogical changes associated with increasing deformation have been described by Norris et al. (1971a) for each of the wide range of acid plutonic parental rock types found in the Hochalm Kern. The range in bulk composition in the Sonnblick Kern is narrow by comparison with the Hochalm Kern. The monotonous augen-gneisses could well have all originated within a single adamellite pluton.

Two-mica augen gneisses with l-tectonite fabric are common where the outcrop of the Sonnblick Kern thins to become the Sonnblick gneiss-lamella.

Coarse, even grained biotite granite-gneiss forms part of the granitic terrain to the west and south of the Mindener Hütte (Map 2). The quartz in this lithology has a purplish-grey colour in hand specimen and contrasts strongly with the white feldspar.

Petrography of the Augen-Gneisses:

The following assemblages have been recorded:

(1) biot + musc + qz + ksp + plag + ep $\pm$ all + chl $\pm$ cc $\pm$ ap $\pm$ sph

(2) biot + musc + qz + ksp + plag $\pm$ gt + chl + ap (leucogranite)

(3) musc + qz + ksp + plag + cc + mt.

The augen-gneisses contain roughly equal amounts of K-feldspar and plagioclase. The former occur as ½ to 2 cm long single crystals usually with Carlsbad twinning. They tend to be rounded with granulated margins and are commonly enveloped by muscovite. Nowhere do they show textures indicative of growth after deformation, common for porphyroblast minerals in other lithologies (see Chapter 4). On the other hand there is very good evidence for their primary igneous origin (see also Frasl, 1954):

(1) In most rocks they are orthoclase perthites, indicating temperatures of formation considerably greater than those deduced for metamorphic conditions (see Chapter 7). Some orthoclase megacrysts contain patches of microcline, often near their margins suggesting that the orthoclase has inverted to microcline in response to strain.

(2) The exsolved plagioclase stringers are sometimes arranged in a pattern suggestive of oscillatory zoning, a feature that is unlikely to have been produced by crystal growth in a solid medium.

(3) Small, 1 mm, euhedral plagioclase crystals are commonly aligned parallel to crystallographic directions within host K-feldspars, and to the oscillatory zones described above.

Four textural types of plagioclase may occur together in the granites and granite-gneisses:

(1) The inclusions described above.

(2) Large, blocky grains of albite or oligoclase commonly full of tiny, random, epidote needles and muscovite

platelets. This "gefüllte plagioklas" texture has been described by Frasl (1953), Karl (1959) and Ackermand and Karl (1972) and is particularly well developed in meta-tonalites. It is ascribed to the breakdown of the anorthite component of the original igneous plagioclases during metamorphism, possibly due to the net reaction:

$$4CaAl_2Si_2O_8 + KAlSi_3O_8 + 2H_2O = 2Ca_2Al_3Si_3O_{12}(OH) + KAl_3Si_3O_{10}(OH)_2 + 2SiO_2$$

plagioclase — K-feldspar — fluid — epidote — muscovite — quartz

(2.A)

The plagioclase inclusions ((1) above) commonly have "gefüllte" cores.

(3) Small (~0.2 mm) inclusion-free grains of recrystallised albite/oligoclase. In the more schistose gneisses this type is more abundant at the expense of type (2).

(4) Rounded blebs of myrmekitically intergrown plagioclase (~90%) and quartz (~10%) replacing K-feldspar (Fig. 2.23).

In most granite-gneisses granular iron-rich epidote is an important constituent of the recrystallised matrix. Large grains with dark brown cores of allanite are abundant in some rocks.

The phengite is faintly pleochroic in thin section and generally defines a good schistosity. Dark brown biotite with both schistose and random textures are common, even in the same rock. Grains of epidote, not part of the igneous assemblage, are commonly included within the random, unstrained, biotites indicating that these biotites are the product of recrystallisation and are not relict igneous crystals (Fig. 2.24).

The textural and mineralogical changes, associated with increasing degree of deformation, are summarized in Fig. 2.23 for three augen-gneisses from the Sonnblick Kern.

Garnetiferous granite-aplites and pegmatites form a very minor part of the Zentralgneis, and in the areas mapped, only occur in upper Dösener Tal. Ackermand and Karl (1972) and Holland (1977) describe garnetiferous meta-tonalites from the Grossvenediger area, Central Tauern. There, the garnet seems to be part of the stable metamorphic assemblage, but here it is more likely that the garnet is magmatic. Pegmatite GD 77 demonstrates this very clearly. It has the assemblage:

gt + musc + qz + plag (+ ore + chl)

Coarse grained plagioclase and muscovite form a matrix in which are set large (5 cm+) patches of intergrown garnet and quartz. The garnet is all interconnected and, in any given cross section, forms a net-like structure enclosing 3 mm diameter areas of quartz. This texture is remarkably uniform, with a quartz:garnet proportion of about 43:57, and may represent magmatic coprecipitation of quartz and garnet, analogous to quartz and K-feldspar in graphic granite. Tiny garnets in leucogranite GD 468 (assemblage 2) are tentatively interpreted as products of metamorphic crystallisation.

(b) Quartz-Phengite-Schist

This lithology occurs in sheets within the Peripheral Schieferhülle succession in the Mallnitzer Mulde (Map 2). These sheets, termed "lamellae" by Exner (1964), are absolutely concordant to the banding in the metasediments and are very persistent laterally. The largest (called the Lonza lamella, or gneiss-lamella 2 by Exner) can be traced for 9 km, from south of the Hagener Hütte to Lassach. It appears to wedge out to the north, but may continue south-

eastwards off the map for a considerable distance. It reaches maximum thickness of over 50 m near the Häusler Alm. The other two important lamellae (both considered by Exner to correlate with the Rote Wand-Modereck Decke, gneiss lamella 4) are, at the most, 5 m thick but nonetheless traceable for several kilometres. Each one outcrops between two calcareous quartzite units and these 'sandwiches' are symmetrically disposed about the axis of the Mallnitzer Mulde. (The structural and tectonic significance of this outcrop pattern is discussed in Chapter 3.)

Exposures of quartz-phengite schist are distinctly grey and rather massive, contrasting with the buff colour and characteristic thinly banded appearance of associated quartzites.

In the thin lamellae these schists are medium grained and dark coloured due to large amounts of green-brown phengite. Small, 0.2 to 0.5 cm, feldspar augen are common. These tend to be elongated parallel to a strong phengite schistosity and an indistinct millimetre to metre scale banding in which phengite-rich bands alternate with more massive augen-rich bands.

Many of the schists of gneiss-lamella 4 and the northwest end of gneiss-lamella 2 are characterised by abundant rounded 2 to 4 mm albite porphyroblasts forming up to 20% of the rocks. Similar albite-porphyroblastic quartz-phengite-schists were found near the Schüster Hütten (3 km south of Böseck) in a 100 m wide shear-zone in otherwise massive augen-gneisses.

The schists of gneiss-lamella 2 are massive, fine-grained, greenish-white rocks. In the middle of the thickest part, however, they have significantly coarser grain size with prominent 1 cm K-feldspar augen, and resemble the schistose augen-gneisses, suggesting that the quartz-phengite schists are merely granite-gneisses which have undergone intense deformation. This is

supported by the similarity between the albite-porphyroblastic quartz-phengite-schists in the gneiss lamellae and those of undisputable granitic origin in the shear-zone described above. A considerable degree of overlap therefore exists between rocks labelled "augen-gneiss" and "quartz-phengite schist".

Exner (1964) discusses the problem of the origin of the gneiss-lamellae in some detail. He suggests three possible alternative modes of origin:

(a) Sheared and mylonitised slices of Zentralgneis granite.

(b) Originally concordant flows of keratophyre or porphyry or tuff bands.

(c) Originally concordant bands of arkose or sandstone.

While the occurrence of augen-gneiss in the middle of gneiss-lamella 2 convinces Exner of its granitic origin, he is less sure that the other lamellae originated in the same way, but nevertheless favours hypothesis (a). Frasl (1958), on the other hand, postulates a sedimentary origin for lamellae in the Glockner depression. There is no doubt that (a) applies in the case of the Sonnblick gneiss-lamella (lamella 1) as this 200 m thick sheet of augen-gneiss and quartz-phengite-schist can be traced north-westwards along strike into the Sonnblick Kern itself (Fig. 1.4).

The main objections against hypothesis (a) are the dissimilarity in the modal composition (particularly the quartz/feldspar ratios) of quartz-phengite-schist and more "normal" granite-gneiss (see below), the shape of the lamellae themselves, and their concordance with enclosing metasediment bands. In the absence of evidence supporting hypotheses (b) or (c), the author also favours (a). An attempt will be made to reconcile these objections in later sections.

Petrography of the Quartz-Phengite-Schists

The following assemblages have been recorded:

(1) pheng + qz + ksp + plag $\pm$ cc + mt + ap $\pm$ ep $\pm$ chl + zirc

(2) pheng + biot + plag + qz + chl + ap + ore

(3) pheng + qz + ksp + ore + ap + ep + zirc

All these rocks contain high modal proportions of quartz and phengite. The phengite is strongly pleochroic and generally has a well developed schistosity. Fine-grained, strained, granulose quartz is accompanied by rare 2 mm long strained quartz augen in some samples.

The K-feldspar in assemblage (1) is invariably microcline and often turbid due to incipient sericitisation. These features contrast with the fresh perthitic orthoclase of the augen-gneisses (sensu stricto), and probably reflect a greater degree of strain. The quartz fabric supports this.

Schists with assemblage (2) are characterized by 2-4 mm albite porphyroblasts, forming up to 30% of the rock. These are clear, unstrained and cut across the phengite schistosity. They must, therefore, have grown during metamorphism and cannot be relict igneous feldspars. Small quantities of very fine-grained plagioclase can be found in most of the rocks, but it is difficult to distinguish from quartz.

In Fig. 2.23 the petrographies of three schist samples taken from different bands within a single lamella (the lower outcrop of gneiss-lamella 4) 1 km S.S.E. of Jamnig Alm are compared with those of the Sonnblick Kern augen-gneisses. The quartz-phengite-schists show a more extreme development of the trends observed in the gneisses, i.e. increase in phengite and quartz at the expense of biotite, igneous feldspars and sometimes recrystallised plagioclase.

The virtual absence of feldspars in extreme cases (assemblage (3)) suggests that the bulk composition has, in some way, been altered during deformation.

To test this hypothesis, chemical compositions have been determined for the six meta-granites described in Table2.3 (three augen-gneisses and three quartz-phengite-schists) and are presented in Fig. 2.25 and Table 2.4. All major element analyses, with the exception of sodium, were determined by standard X-ray fluorescence spectroscopic techniques (Norrish and Chappell, 1967) using glass beads prepared from fused rock powders. Sodium concentrations were determined by atomic absorption spectroscopy.

The following arguments are based on the assumption that the original compositions of the augen-gneisses and quartz-phengite-schists were similar. This cannot be proved in this geological situation as the Sonnblick Kern and the gneiss-lamella from which the samples were collected show no evidence of being physically connected. However, the samples <u>within</u> each group are likely to have had similar original chemistries. The three augen-gneisses were obviously derived from adamellites, probably from a single pluton. The compositional range of adamellites is limited with $Na_2O + K_2O$ ranging from 5 to 10 wt. %, SiO_2 from 60 to 78 wt. % and Na_2O/K_2O from 0.3 to 2 (Le Maitre, 1976). The quartz-phengite-schist samples were collected from within a metre of one another, so it is reasonable to assume that they began with similar bulk compositions.

The object of this exercise is to relate the present bulk compositions of these rocks to some process that might have caused initial compositions to change. The hypothesis is that intense deformation can cause such changes. The six samples were collected to represent a range in the amount of strain experienced However, in general, this quantity is very difficult to assess and can only be gauged subjectively by the degree of anisotropy of the rock fabric. In

Chapter 3 it is established that the penetrative strain associated with D_a^1 structures in the Mallnitzer Mulde was very large, but that in the core of the Sonnblick Kern, large amounts of strain were confined to localised shear zones. The three samples from the gneiss-lamella must therefore have undergone more strain than those from the Sonnblick Kern. With these assumptions in mind, the following generalisations can be made. Sodium, calcium and strontium contents of the three augen-gneisses are similar to one another and compare well with typical calc-alkaline adamellite compositions (Table 2.4). In the quartz-phengite-schists, however, Na, Ca and Sr fall off drastically with increasing intensity of deformation to 5 or 10% of the augen-gneiss values. This ten-fold range is well outside the range of granitic compositions. The minimum Na_2O and CaO contents of granites and granodiorites in the central Sierra Nevada batholith are 1.7 and 0.9 wt. % respectively (Bateman and Dodge, 1970), several times the values from quartz-phengite-schists. The latter also have abnormally low Na/K ratios (0.027 to 0.1). Metagranite Na, Ca and Sr values correlate well with the modal contents of albite + epidote or calcite, the recrystallised products of the original igneous plagioclase. This suggests that plagioclase was removed during the intense shearing associated with D_a^1. Hanson et al. (1970) record a similar depletion of Ca, Na and Sr in a tectonised granite in Switzerland.

Potassium and magnesium appear to show an inverse relationship with one another, but do not correlate with degree of deformation. Variation in these two elements may reflect original compositional differences. Within the quartz-phengite-schists alone, however, Mg concentrations increase with the degree of deformation. This trend is the same as that observed by Hanson et al. and is attributed by them to an increase in modal phengite. This might also be the cause of the gradual increase of both iron and titanium with deformation over all six samples.

The most abundant major elements, silicon and aluminium, appear to remain constant. This in itself could be taken as evidence that the quartz-phengite-schists were originally granites and not arkoses.

It seems probable, therefore, that the chemistry and mineralogy of the quartz-phengite-schists differ from those of "normal" granite-gneisses because they have experienced a larger amount of strain. During shearing, the rocks in the lamellae may have been ground into particles small enough to allow a large proportion of the feldspars to dissolve in the fluid phase. The accompanying decrease in Na, Ca and Sr can only be explained if these elements were not subsequently reprecipitated but remained in solution and were carried elsewhere by the fluid. Similarly, the extra Fe and Ti can only have been obtained from the fluid. The K content may have been roughly preserved because, although modal K-feldspar decreased with increasing deformation, phengite increased. The mineral reactions involved are discussed in Chapter 7.

The main conclusion from this study is that deformation may have been responsible for inducing metasomatism in metagranites.

(c) Syenite-Gneiss

This distinctive lithology outcrops in a single sheet, the Romate-decke (Exner, 1957) within mica-schists and paragneisses along the northern flank of the Mallnitzer Mulde (Map 2).

Its most easterly exposures are found at the bottom of the Mallnitz-Hadsler Alm path. There, the sheet is already over 200 m thick. At the northern edge of Map 2 it is about 500 m thick. Exner maps it as extending as far as

Böckstein, 7 km north of Map 2. It appears to become interfolded with overlying granite-gneiss (the "Siglitzlappen") on a large scale on Radhausberg 3 km north of Map 2. However, along its whole length it remains physically separated from the underlying granite gneisses of the Hölltor Kern by the schists of the Woisgen zone. A small window through the syenite-gneiss into these schists occurs in upper Weissenbachtal.

The syenite-gneiss is characterised by high modal proportions (60-70%) of ½ to 1 cm blocky white feldspar crystals (mostly simple-twinned K-feldspars) set in a finer grained dark matrix rich in biotite and epidote. In some rocks, these two minerals occur together in small clots or streaks. Quartz is not visible in hand specimen.

Throughout its outcrop in Map 2, the syenite-gneiss possesses some degree of foliation. This is formed by a biotite-schistosity which tends to wrap round elongated and aligned feldspars giving the more strongly foliated rocks a "total augen" texture. In the area around Romate Spitze, the syenite-gneiss is very poorly foliated except in localised planar zones parallel to the boundaries of the sheet itself. Lenses of biotite-rich schist flattened in the plane of the foliation were found at a few localities.

In the least well foliated rocks, the clots of biotite and epidote are very striking (Fig. 2.26) and appear to be pseudomorphs after euhedral crystals of a prismatic ferromagnesian mineral such as hornblende or clinopyroxene. They measure up to 1 cm long and show hexagonal cross-sections more suggestive of an amphibole. Exner (1957) interprets them as hornblende pseudomorphs.

Petrography of the Syenite-Gneiss:

The following assemblages have been recorded:

(1) biot + ep + ksp + ab + qz + cc + ap

(2) biot + ep + cpx + hbl + ksp + ap + qz + cc + sph + ap

(3) biot + cpx + hbl $\pm$ ksp + ab + qz + sph + mt + ap

Assemblage (1) is by far the most common. K-feldspar is more abundant than plagioclase in thin section. Their textures are similar to those in the augen-gneisses described above, except that the euhedral gefüllte plagioclase inclusions in K-feldspars are often large and form "cores", and no quartz + plagioclase myrmekitic intergrowths have been observed replacing K-feldspars. A more fundamental difference between these two lithologies is the conspicuous absence of muscovite in the syenite-gneiss, even in schistose varieties. Whereas muscovite commonly accompanies epidote as inclusions in "gefüllte" plagioclases of granite-gneisses, epidote alone occurs in the syenite-gneiss.

The syenite-gneiss contains about 10 mode % of quartz which, with clear recrystallised albites, forms a polygonal mosaic between the feldspar megacrysts. The biotite is dark greenish-brown and associated with $\frac{1}{2}$ to 1 mm wide blocky grains of strongly coloured, highly birefringent epidote.

Both the syenite-gneiss and the various granite-gneisses crystallised two feldspars during their igneous cooling histories (i.e. they are "sub-solvus" rocks) which argues that they may all be part of the same suite of magmas. The syenite-gneiss could be more accurately described as quartz-monzonite-gneiss, but the name will be retained because of the priority of the terms "granosyenite-gneiss" of Angel (1954), Exner (1957) and "syenite-gneiss" of Becke (1912).

Assemblage (2) is restricted to the types carrying recognisable pseudomorphs. The pseudomorphs themselves contain the assemblages:

(a) biot + ep + sph + ap

(b) biot + ep + cpx + sph + ap

(c) biot + ep + cpx + hbl + sph + ap

The constituent minerals have a fine-grained equigranular texture and all appear to be in equilibrium with one another (Fig. 2.27). Hornblende may be an exception as it is rather rare and may have grown at the expense of pyroxene. The approximate modal composition of a typical pseudomorph of type (b) would be: biotite, 70%; epidote, 20%; clinopyroxene, 7%; sphene, 3%; apatite, <1%. The clinopyroxene is obviously not a relict igneous phase. It is unzoned, slightly pleochroic ($\alpha > \gamma$ (pale green) $> \beta$ (pale straw)) with a $\gamma:z$ extinction angle of 53° and a $2V_{\gamma}$ of 70°. Unlike the other phases, it only occurs within the pseudomorphs.

The high modal proportion of biotite in the pseudomorphs implies that the replacement of the original amphibole or pyroxene during metamorphism was accompanied by an influx of potassium.

Assemblage (3) was found in one sample (GD 1234b) from a lens of medium-grained biotite-schist within the syenite gneiss on the path east of Romate Spitze. The lens probably represents a deformed cognate inclusion of igneous origin. Clinopyroxene and biotite, optically identical to those described above, occur in equal amounts and together make up half the rock. Most of the remainder is composed of albite and subordinate quartz in stringers parallel to the biotite schistosity. These four essential minerals all occur as clear, unstrained, unzoned polygonal grains and probably represent an equilibrium assemblage (Fig. 2.28).

The occurrence of clinopyroxene in metamorphic rocks of "medium grade" (see Chapters 6 and 7) other than marbles is very unusual. Even the least foliated

types of syenite- and granite-gneiss show textures indivative of recrystallisation due to hydration reactions analogous to 2.A. This suggests that the growth of pyroxene in the syenite-gneiss was not caused by a restricted supply of water. A more likely cause is an unusual bulk chemistry. The absence of muscovite and presence of pyroxene suggests that these rocks have an excess of Ca, Fe and Mg over K and Al, The phase relations of syenite- and granite-gneisses are discussed in more detail in Chapter 7.

(d) Biotite-Epidote-Feldspar-Schist

This lithology forms a thick concordant sheet, parallel to and structurally above the Romate-decke, termed the Feldwand-lamella by Exner (1964). Along much of its outcrop its thickness is about 100 m, but north of the Jamnig Alm it seems to exceed 200 m. On the ridge crest east of Greil Kopf it is only 50 m thick. It is not clear from the field evidence whether this sheet merges with the Romate-decke or simply pinches out in the vicinity of Mallnitz.

The rock is a medium-grained schist rich in biotite and epidote, suggesting an affinity with the syenite-gneiss. This is supported by the common occurrence of small K-feldspar augen. Some varieties contain a significant amount of muscovite. Although Exner maps this unit as amphibolite, amphibole is completely absent except in the associated lithologies described in parts (e) and (f).

A 1 m thick sheet of biotite-epidote-feldspar-schist cuts the syenite-gneiss about 150 m from its western contact in the road section below the Jamnig Alm. The sheet has distinctly finer-grained margins (Fig. 2.29) and the schistosity in both itself and the host-rock are about 20° discordant to the contacts. Together, these features suggest that it is a pre-tectonic dyke

intrusion.

Only two specimens were studied in thin section. They have the assemblages:

(1) biot + ep + musc + ksp + plag + qz + cc + chl + sph

(2) biot + ep + ksp + qz + chl + ap + pyr

The first comes from the Feldwand-lamella and the second from the dyke. Iron-rich epidote and dark brown biotite altering to chlorite are common to both. In (1) the K-feldspar is perthitic orthoclase and occurs as very ragged augen apparently breaking down to fine-grained aggregates of quartz and albite streaked out parallel to the mica schistosity. Muscovite is concentrated along localised schistosity surfaces. In (2), clear poikiloblasts of microcline overgrow quartz and epidote aligned parallel to the schistosity and appear to be the product of post-deformational recrystallisation.

(e) Amphibolite

A 100 m wide patch of unfoliated amphibolite occurs within the Feldwand-lamella about 1 km north-west of the Jamnig Alm.

Random, discrete 3 to 8 mm long "prisms" of hornblende are set in a yellowish medium-grained matrix rich in epidote and biotite. The rock resembles a metamorphosed intermediate to basic plutonic rock and contains the assemblage:

hbl + biot + ep + ab + cc + pyr + ap

In thin-section, the texture is thoroughly metamorphic and none of the minerals appear to be igneous relicts. The hornblende is a blue-green variety. Most of the hornblende "prisms" are polycrystalline aggregates of acicular interpenetrating grains. The epidote forms abundant $\frac{1}{2}$ to 1 mm

yellow unzoned granules, locally enclosed by random poikiloblasts of biotite. The albite occurs as small aggregates of clear, twinned, polygonal grains.

(f) Hornblendite

Two small masses of hornblendite were found within the Feldwand-lamella. The larger, outcropping 1 km south of Ebeneck, is lens-shaped and measures 200 m by 60 m. The other is only 20 m thick and outcrops on the ridge between Greil Kopf and Ebeneck. At the second locality, the hornblendite is brecciated and veined by the enclosing biotite-epidote-feldspar-schist.

In hand-specimen, the hornblendites are coarse, black, equigranular almost monomineralic rocks with grainsizes ranging from 0.3 to 2 cm. One specimen was studied in thin-section. It contains the assemblage:

hbl + act + biot + chl + ep + cc + sph + ore + ap

About 95% of the rock is composed of amphibole. Most of this is actinolite which forms large, ragged grains interlocking in random orientation. Some have patchy cores of brown hornblende partially or completely enclosing euhedral areas of actinolite in optical continuity (Fig. 2.30). The brown hornblende is interpreted as a primary igneous mineral. The actinolite is evidently part of the stable metamorphic assemblage as it appears to coexist with epidote, biotite, etc., and it seems to have replaced mineral grains (not amphibole) enclosed within brown hornblende. Calcic clinopyroxene is commonly pseudomorphed by actinolite in low and medium grade metabasites such as the "epidiorites" of the Scottish Dalradian (Wiseman, 1934), suggesting that this rock may have originally been a hornblende pyroxenite.

D. PERIPHERAL SCHIEFERHÜLLE

Introduction

Norris et al. (1971b) distinguish rocks of the Peripheral Schieferhülle from those of the basement complex on the grounds that the former generally

(i) show evidence of only one major phase of deformation;

(ii) may be cut by a variety of low temperature hydrothermal veins, but are not cut by veins of granite-aplite or -pegmatite;

(iii) contain abundant calcareous units throughout the sequence.

The applicability of criterion (i) is questioned in Chapter 3, and only (ii) and (iii) have been used here.

In the area covered by Map 1, Peripheral Schieferhülle rocks surround the southern margin of the Hochalm Kern and trend north-west/south-east, parallel to the Mölltal. In the area covered by Map 2, they can be subdivided into two parts: those occupying the Mallnitzer Mulde and those forming the core of the Ankogel Mulde. The former are laterally continuous with the more south-westerly (structurally high) units in the Mölltal, and trend in exactly the same direction. The latter pass into units that occupy structural positions closer to the basement-cover interface, but have a more northerly trend.

Field Relations and Petrography of Peripheral Schieferhülle Lithologies

(a) Epidote-Amphibolite and Greenschist

Metabasic rocks form an important part of the cover sequence, although the proportion varies considerably. In the Kaponig area (Map 1), metabasites form about 20% of the sequence, but in the Mallnitzer Mulde they are volumetrically insignificant and only occur in a few 1 to 5 m thick bands, within calc-schist. A large mass outcrops to the east of Mallnitz, extending from the railway station north-eastwards to Wasenle Kopf. It passes into the Kaponigtal bands towards the south-east, but appears to be enclosed by mica-schist in all other directions. The significance of this outcrop pattern is discussed in Chapter 3.

There are several consistent differences between these amphibolites and those of the Inner Schieferhülle. In direct contrast to the latter, Peripheral Schieferhülle amphibolites tend to be finer grained and to contain abundant chlorite, epidote and carbonate, but very rarely garnet. They also contain numerous 1 mm to 3 cm thick bands rich in epidote, contrasting with the feldspathic bands in the latter.

Holland (1977) has shown that the epidote-banded amphibolites are probably intensely deformed and metamorphosed pillow lavas, the bands representing stretched-out remnants of the altered glassy pillow-margins. This is consistent with the chemical data of Bickle and Pearce (1975) who conclude that the Peripheral Schieferhülle metabasites were probably derived from ocean-floor tholeiitic basalts.

The epidote-amphibolites have a well developed foliation formed by a penetrative chlorite + amphibole schistosity parallel to intermittent epidote banding.

Most are dark green in colour but distinctly paler varieties, "greenschists", also occur, especially in the Mallnitzer Mulde. Coarse-grained unfoliated meta-gabbroic epidote-amphibolite forms a slice 5 to 20 m thick and 1 km long associated with serpentinite and talc-schist in a shear-zone extending south of Sickerkopf (Map 1).

Petrography of the Epidote-Amphibolites and Greenschists:

The following assemblages have been recorded:

(1) gt + hbl + ep + biot + plag + dol + cc + qz + ilm

(2) hbl + ep + chl + plag + cc + qz $\pm$ rut + ilm $\pm$ pyr

(3) hbl + ep + plag + qz + ore

(4) hbl + ep + chl + biot + plag + dol + cc + qz $\pm$ mt + ilm + ap

(5) hbl + ep + biot + plag + ?cc + qz + ore

(6) hbl + ep + chl + biot + plag + qz $\pm$ rut $\pm$ ilm $\pm$ ap

(7) hbl + ep + chl + biot + plag + dol + qz + mt + ilm + pyr

(8) hbl + act + ep + biot + plag + qz + sph + ilm + ap

(9) hbl + ep + chl + plag + qz + rut + mt + ap

(10) hbl + ep + chl + plag + dol + qz + rut + mt + ilm + ap

(11) act + ep + plag + cc + qz + sph

(12) act + hbl + ep + chl + plag + cc + qz + sph + pyr

(13) act + ep + chl + biot + plag + dol + cc + qz + rut + ore

(14) ep + chl + biot + plag + qz + rut + ore + ap

(15) ep + chl + plag + cc + qz + ilm + tour + ap

(16) ep + chl + biot + musc + plag + qz + ilm + ap

(17) ep + chl + biot + plag + cc + qz + ilm

(18) hbl + act + ep + chl + plag + rut

Assemblages (1) to (10) are from epidote-amphibolites, (11) to (17) from

greenschists and (18) from the meta-gabbro. All assemblages except (1), (12), (13) and (18) appear to be in equilibrium.

The hornblende has a pleochroic scheme: α (pale brown) < β (grass green) $\approx$ γ (blue green), and usually forms sub/euhedral acicular prisms aligned in the plane of the foliation. In GD 193 (assemblage (6)) decussate hornblende poikiloblasts enclose epidote grains aligned parallel to a chlorite schistosity (Fig. 2.31). Actinolite, the characteristic amphibole of the greenschists, is defined here as having a pleochroism: $\alpha = \beta$ (colourless) < γ (pale blue-green). It tends to form as fine needles with a good schistosity or lineation.

Generally, the amphibole is not optically zoned. However, some of the actinolites in GD 551 (assemblage (12)) have very thin rims of hornblende. Hornblendes in GD 179c and GD 359 (assemblages (18) and (8) respectively) have indistinct actinolite cores (Fig. 2.32). In these cases, hornblende has grown after actinolite, suggesting that the differences between epidote-amphibolites and greenschists may be partly attributable to metamorphic grade. In GD 230i (assemblage (13)) tiny needles of actinolite are preserved within albite porphyroblasts in otherwise amphibole-free rock.

The chlorites tend to be pale and only slightly pleochroic (pale blue-green to pale straw-yellow), suggesting rather magnesian compositions. The epidote-amphibolites rarely contain strongly-coloured chlorite.

Epidote is an essential mineral in all Peripheral Schieferhülle metabasites. It is generally either an unzoned yellow, highly birefringent variety or zoned from an iron-rich core out to a clinozoisite rim. Reverse zoning was observed in only one specimen (GD 558, assemblage (6)). Unzoned clinozoisite was found in some of the greenschists.

Metabasites from the Mallnitzer Mulde commonly contain conspicuous 0.5 to 5 mm porphyroblasts of albite which give them a knobbly appearance in hand-specimen. These albites tend to be idiomorphic towards chlorite, displaying characteristic rhombic cross-sections with rounded corners, but enclose aligned prisms of epidote and/or amphibole. A few of the amphibolites in the Kaponig area also conform to this pattern, but more usually the plagioclase is not porphyroblastic. In some it is zoned from albite cores to oligoclase-andesine rims (see also Norris et al., 1971b) and in others unzoned oligoclase-andesine, or coexisting albite + oligoclase not obviously in zonal relationship (e.g. GD 193) have been identified. The paucity of albite-twinning makes compositional determination difficult without recourse to the microprobe.

Garnet was only found in one rock (GD 22b), from a 50 cm thick band within calc-schists in mid-Zwenberger tal (Map 1). It is a typical epidote-amphibolite except that it contains garnet and has a coarser than average grainsize. The garnet forms ragged 2 mm grains, some of which are "atoll" forms being replaced by biotite and appears to be a relict from an early stage in the rock's history. This rock shares with a few others the distinction of containing two coexisting carbonates. Here, irregular patches of iron-stained dolomite are surrounded by smaller polygonal grains of calcite.

Amphibole-free greenschists (assemblages (14) to (17)) with high modal proportions of epidote, were found in a 20 m wide zone between calc-schists and amphibolites in lower Kaponigtal. The presence of muscovite in one (GD 152e) suggests that these metabasites contain a small proportion of pelitic sedimentary material i.e. that they are meta-tuffs.

(b) Serpentinite and Talc-Schist

Ultrabasic rocks form a very minor part of the Peripheral Schieferhülle. Serpentinites tend to occur in lens-shaped bodies generally within calc-schists. In the areas studied, the two largest, nearly 100 m thick, outcrop on the ridge between the Hagener Hütte and Geisel Kopf (Map 2) and on the south-east slope of Labiges Köpfl (Map 1). Smaller bodies also occur in the vicinity of Geisel Kopf, in Kaponigtal and Zwenberger graben and on the ridge between Auernig and Törl Kopf. The latter is unusual in that the surrounding rock is epidote-amphibolite.

The serpentinites sensu stricto are massive, fine-grained, pale or dark green unfoliated rocks carrying occasional dolomite or magnetite porphyroblasts. In every serpentinite outcrop examined, massive serpentinite is associated with talc-tremolite- or talc-carbonate-schist to a greater or lesser extent. These ultrabasic schists commonly occur at or near the serpentinite/calc-schist contacts, or along shear planes cutting through massive serpentinite. The proportion of serpentinite to ultrabasic schist varies considerably and correlates with the size of the body. In the largest masses there is very little associated schist, whereas the smallest ones (e.g. the 4 m wide lens on the Böseck-Törl Kopf ridge) tend to be composed entirely of it (Fig. 2.33).

Frasl (1958) and Norris et al. (1971b) point out that the serpentinite pods in the Peripheral Schieferhülle tend to occur along strike from one another. Local remapping has revealed the existence of a continuous shear zone extending from Sicker Kopf to south of Labiges Köpfl (Map 1). It varies in width from 10 to 50 m, and contains lenses and slices of serpentinite, meta-gabbro (see above) and siliceous dolomite in a talc-schist matrix. It occurs within calc-schists close to their contact with epidote-amphibolites and is parallel to the macroscopic lithological banding (Fig. 2.34).

It is generally accepted that these serpentinites represent fragments of peridotite originally part of the oceanic lithosphere. Exner (1957) and Hawkesworth (1974) record the occurrence of serpentinites containing both orthopyroxene and clinopyroxene relics, indicating that in those cases the precursors were lherzolites.

The emplacement of masses of oceanic peridotite into a sequence of metasediments constitutes a geological problem analogous to the origin of the gneiss lamellae described above except that the metagranites represent continental basement while the serpentinites represent oceanic basement. Lockwood (1971; 1972) argues that many Alpine-type serpentinites could be large blocks incorporated into the enclosing sediments during deposition. This might conceivably be the case for a few of the isolated lenses within calc-schists, but cannot apply in the case of the shear-zone mentioned above or the serpentinite pod enclosed within epidote-amphibolite. All available evidence, therefore, points to a tectonic mode of emplacement.

Petrography of the Serpentinites and Talc-Schists:

The following assemblages have been recorded:

(1) serp $\pm$ dol $\pm$ mt

(2) serp + tc $\pm$ mt

(3) serp + tc + dol $\pm$ mt

(4) serp + trem + dol + tc + mt + chrom

(5) tc + dol

(6) tc + dol + mag

(7) trem $\pm$ chl $\pm$ phlog

(8) trem + tc $\pm$ chl

(9) trem + tc + dol

Assemblage (1) is typical of massive unveined serpentinite from the largest outcrops. Assemblages (2) and (3), representing slightly carbonated serpentinite, are more common. In all of these, the dominant serpentine mineral is antigorite which occurs as random interlocking flakes euhedral to both magnetite and dolomite. Assemblage (4) was found in one rock (GD 868) from the large outcrop on Geisel Kopf. Random euhedral tremolites up to 1 cm long and ragged patches of dolomite are set in a very fine- but even-grained matrix of antigorite. Thin veneers of talc rim some of the tremolites (Fig. 2.35). Magnetite forms irregular grains up to 2 mm across, the largest of which contain rounded inclusions of chromite.

Assemblages (5) to (9) are found in ultrabasic schists, (5) and (8) being the most common. The tremolite usually occurs as well lineated 2 to 5 cm long needles, apple-green in hand specimen, although white fibrous varieties have also been found. It commonly exceeds 70 mode % in assemblages (7) to (9). Talc normally forms a fine-grained felted mass but in assemblages (5) and (6) it is occasionally coarse grained and schistose. The dolomite is usually an iron-rich variety and euhedral to talc. Colourless chlorite is a common accessory, although rare chlorite-rich schists with the assemblage:

$$chl + mt \pm tc \pm ap$$

are associated with talc-schists in the Kaponigtal shear-zone. Green pleochroic chlorite forms a monomineralic schistose matrix enclosing euhedral octahedra of magnetite up to 4 mm across. Talc where present occurs in "strain shadows" round the magnetites.

(c) Calc Schist

This is by far the most abundant lithology in the Peripheral Schieferhülle

and forms 50 to 70% of the succession. The average calc-schist contains a high modal proportion of calcite giving fresh specimens a blue-grey colour and exposed rocks a characteristic buff colour. Calc-schists all contain abundant muscovite and graphite, but the proportion of calcite to sheet-silicate varies considerably. They grade into marbles with decreasing graphite and silicates and into graphitic quartz-mica-schists with decreasing calcite. Consequently, the distinction between these lithologies is arbitrary. Calc-schists were mapped as those rocks with both essential calcite and essential silicates, but where calc-schist and graphitic mica-schist were interbanded on a 20 m scale or smaller, they were given the label of the dominant component.

All calc-schists have an excellent penetrative muscovite and calcite schistosity making them extremely fissile. A millimetre-to metre-scale compositional banding is often present which is generally sub-parallel to this schistosity except in the noses of folds.

Black porphyroblasts of plagioclase up to 1 cm across, are conspicuous in some varieties.

Petrography of the Calc-Schists:

The following assemblages have been recorded:

(1) cc + ep + biot + musc + plag + qz (+ graph $\pm$ rut $\pm$ sph $\pm$ ore $\pm$ tour)

(2) cc + ep + chl + biot $\pm$ musc + plag + qz (+ graph + sph + ore)

(3) cc + gt + ep + chl + biot + musc + qz (+ graph + ore + tour + ap)

(4) cc + ep + biot + marg + musc + plag + qz (+ graph + sph + ore)

(5) cc + dol + ep + biot + musc + plag + qz (+ rut + pyr + tour + ap)

Of these, only (1) and (2) appear to be in equilbrium and (1) is by far the

most common. Well-twinned, polygonal grains of calcite usually make up 50 to 75% of the rocks. The characteristic epidote mineral is an unzoned clinozoisite but it is never very abundant. Plagioclase porphyroblasts (andesine-labradorite) generally appear to have grown at the expense of micas along certain bands and commonly contain trails of tiny graphite inclusions which give them their black colour. The biotite is generally an orange variety and the ore mineral is invariably pyrite.

Assemblage (3) was found in one rock (GD 15) from near the basement-cover-interface in Zwenberger Tal. 2-3 mm reticulate poikiloblasts of garnet with quartz pressure shadows enclose quartz and calcite and have been partially replaced by sericite.and chlorite.

Assemblage (4) was found in two rocks (GD 154, 154a) from lower Kaponig Tal. The margarite occurs in fine-grained wisps parallel to a muscovite and biotite strain-slip schistosity. The latter also occur as cross-cutting flakes. The margarite appears to have been partially replaced by small plagioclase grains.

(d) Marble and Dolomite

These lithologies contain essential carbonate (calcite, dolomite or both), but only minor amounts of silicates. Fresh marbles can be white, bluish-grey or, more rarely, banded, and dolomites are generally white or buff. Muscovite is usually the most conspicuous silicate.

Bands of marble are more abundant in the Mallnitzer Mulde than in the Möllital Schieferhülle. They are normally between 5 and 20 m thick, but a 100 m thick band, dipping parallel to the ground surface, outcrops over a large area in

Tauerntal (Map 2). Marble is frequently associated with calcareous quartzite and quartz-phengite-schist. Dolomite bands occur sporadically within normal marbles.

Marbles and dolomites contain the following assemblages:

(1) cc + musc $\pm$ qz $\pm$ ore

(2) cc + chl + ep + ore

(3) cc + chl + ep + musc + qz + sph + ore

(4) cc + chl + musc + qz $\pm$ graph + sph

(5) cc + dol + musc + qz $\pm$ ore

(6) dol + musc $\pm$ biot + qz $\pm$ plag

(7) dol + ep + qz $\pm$ plag + ap

All assemblages appear to be in equilibrium. Mineralogically, these rocks closely resemble the calc-schists except that they contain a much higher modal proportion (80 to 100%) of carbonate. In the majority of cases, this is calcite.

The carbonate usually forms a mosaic of interlocking twinned, polygonal grains, which are sometimes elongated or flattened parallel to the mica schistosity. In a few marbles, the calcite is strained, i.e. displays undulose extinction and curved twin lamellae.

(e) Siliceous Dolomites

These rocks are considered separately from the dolomites (sensu stricto) because they are free of muscovite and rich in CaMg silicates (particularly tremolite-actinolite), and thus have mineralogical affinities with the ultra-basic schists with which they are often associated. They do not outcrop over

large enough areas to be mapped. In Kaponigtal, assemblages (3) and (6) were found in a 5 m wide slice in the thrust zone mentioned in part D(b), above. Assemblages (1), (2), (4) and (5) were collected from several 1 mm to 10 cm thick bands in calc-schists both in the vicinity of this thrust zone and in the Lonza Köpfl - Törl Kopf area (Map 2). (7) and (8) occur together in a 1 mm wide band in calc-schist 750 m E.S.E. of Geisel Kopf (Map 2).

Petrography of the siliceous dolomites:

Assemblages:

(1) trem + cc + qz + chl $\pm$ biot - sph + pyr

(2) trem + cc + qz + tc + chl + biot + sph + pyr + ap

(3) trem + cc + qz + ep + plag - biot - sph - pyr

(4) trem + dol + qz + chl + biot + sph + pyr

(5) trem + dol + qz + tc + chl + biot - sph - pyr

(6) di + trem + cc + qz + ep + plag + biot + sph + pyr + ap

(7) di + trem + cc + plag + sph + ap

(8) di + trem + cc + chl + ep + sph + pyr + ap

All assemblages except (2), (5) and (6) appear to be in equilibrium.

Tremolite is the most abundant silicate. It is generally pale green in hand specimen, colourless in thin section (though sometimes with a faint absorption parallel to 2) and usually has a strong linear fabric. Tremolite appears to be in equilibrium with quartz and carbonate in assemblages (1) to (5) (e.g. Fig. 2.36).

Radiating sheaves of colourless magnesian chlorite, silvery grey in hand-specimen, are common and apparently stable in most assemblages, as are spora-

dic flakes of pale orange biotite. Talc, where present always appears to be replacing tremolite. Epidote and albite (An 3-5%) are abundant in some rocks and seem to be in equilibrium.

In GD 160b (assemblage (6)) 1 mm prisms of diopside, altering patchily to tremolite, enclose euhedral epidotes in a calcite albite quartz matrix. The diopside in GD 1085 a and b (assemblages (8) and (7)) forms 1 to 3 mm clots in which it has a radiating habit. It appears to coexist stably with tremolite in the former but is breaking down to tremolite in the latter.

The diopside-bearing rocks might be expected to outcrop in a region of higher metamorphic grade but there seems to be no logical arrangement of assemblages in space. Instead, apparently incompatible assemblages such as (3) and (6), between which chemical reactions can be written are commonly found close together, in this case within 1 m. An extreme case is illustrated by GD 230d where 2 to 8 mm bands with assemblages (1) and (4) alternate with one another (Fig. 2.37). On the other hand, identical or equivalent assemblages can occur miles apart. The significance of these relationships is discussed in Chapter 7.

(f) Graphitic Mica-Schist

Two 300 m thick and several thinner bands of this lithology outcrop in the Mallnitzer Mulde on the south side of Tauerntal. In the Kaponig area, however, it has not been mapped as a separate lithology, because it only occurs as occasional thin bands within calc schists. West of Mallnitz, calc schists and graphitic mica schists are locally interbanded on a 1 to 10 m scale, and in these situations are not separated.

Mineralogically, the graphitic mica schists are unexciting, consisting mainly of quartz and graphite-stained muscovite. Much of the quartz occurs as veins parallel to the schistosity or as rods parallel to small-fold axes. 1 to 2 mm black porphyroblasts of plagioclase are sometimes visible, and calcite is permitted as an accessory phase. Garnetiferous varieties occur rarely and are confined mainly to the Kaponig area. These are considered together with other garnet-mica-schists.

Only one rock was studied in thin section. It has the assemblage:

ab + musc + qz + graph + pyr

The muscovite has a well-spaced strain-slip schistosity superimposed on an older fabric. Between strain-slip partings, 2 mm albite porphyroblasts enclose strainght trails of graphite at an angle to the older schistosity, and have evidently been rotated by later movements.

(g) Calcareous Quartzite

These quartzites are totally different from those described in part B(g), above. They are pale, buff-weathering rocks with a pronounced compositional banding in which 1 to 30 cm thick massive mica-free layers are separated by 1 mm thick micaceous partings, giving rise to a characteristically flaggy appearance in outcrop. This banding has enabled them to respond to deformation by producing striking fold structures (Figs. 3.5, 3.7 and 3.19).

Units of calcareous quartzite are generally between 10 and 30 m thick but outcrop over considerable distances. They are common in the Mallnitzer Mulde where they are associated with graphitic mica schists, marbles and quartz-

phengite-schists. A 200 m thick unit extends south-eastwards into the Keponig area. Norris et al. (1971b) have described it as dolomite-quartzite-breccia as in places it consists of deformed dolomite clasts embedded in a quartzose matrix. A pair of quartzite bands about 60 m apart extend from east of Böseck north-westwards past the Hagener Hütte into Höllkar. Between them, immediately below the upper, is a 5 m wide quartz-phengite-schist band. Collectively, these units form a useful marker horizon in mapping out large fold structures. Significantly, this pattern is repeated in reverse a few hundred metres towards the south-west.

The flaggy nature of these quartzites and the common occurrence of graphite in the micaceous partings serve to distinguish them as metasediments from the quartz-phengite-schists. Both lithologies contain calcite; the former as an essential mineral, the latter only as an accessory.

The typical mineral assemblage of calcareous quartzites is:

$$qz + cc \pm dol + musc + ksp \pm graph + ore$$

Quartz totals 70-80%. Individual grains tend to have irregular grain boundaries and undulose extinction, indicative of strain without subsequent recrystallisation. Calcite (ca. 10%) forms small unstrained, twinned polygonal granules. Dolomite has only been found in the coarsely clastic types where it forms the clasts themselves. Both microcline and orthoclase are usually present. Flakes of muscovite (ca. 5%) are scattered throughout the rocks and define an indistinct schistosity.

(h) Pelitic Mica-Schist

This lithology forms an important part of the Peripheral Schieferhülle

succession. Two main outcrops extend along the floor of Tauerntal one of which is connected laterally with the schists surrounding the Auernig epidote-amphibolite mass. These in turn continue south-eastwards to form the base of the cover succession in the Kaponig area.

In hand specimen the majority of these schists contain numerous 0.1 to 1.0 cm porphyroblasts of garnet set in a fine-grained schistose matrix of white mica with or without chlorite and/or biotite. Thin veins of quartz commonly occur parallel to the schistosity, and this foliation is often crenulated. Other conspicuous porphyroblast minerals sometimes present include plagioclase, chloritoid, chlorite, biotite and magnetite.

Petrography of the pelitic schists:

About a hundred of these were studied in thin section. They contain the following assemblages:

(1) gt + biot + chl + musc + qz $\pm$ ep $\pm$ plag $\pm$ graph $\pm$ rut $\pm$ ore $\pm$ tour $\pm$ ap

(2) gt + chl + musc + qz $\pm$ ep $\pm$ plag $\pm$ graph $\pm$ sph $\pm$ ore $\pm$ tour $\pm$ ap

(3) biot + chl + musc + qz + ep + ore + ap

(4) chl + musc + qz $\pm$ ep $\pm$ plag + ore $\pm$ tour $\pm$ ap

(5) chl + musc + parag + ep + plag + ore

(6) gt + ctd + chl + musc + qz $\pm$ ep $\pm$ plag $\pm$ ore $\pm$ tour $\pm$ ap

(7) ctd + chl + musc $\pm$ qz $\pm$ ep + ore $\pm$ tour $\pm$ ap

(8) gt + ctd + biot + musc + qz $\pm$ ep $\pm$ plag + ore $\pm$ tour

(9) ctd + biot + chl + musc + qz $\pm$ ep $\pm$ plag + ore $\pm$ ap

(10) ctd + biot $\pm$ parag + musc + qz + ep + ore $\pm$ ap

(11) st + biot + musc + qz + ep + plag + graph + ore + tour + ap

(12) gt + st + ctd + chl + musc + qz + ep $\pm$ plag + ore $\pm$ tour $\pm$ ap

(13) gt + st + biot + chl + musc + qz $\pm$ ep $\pm$ plag $\pm$ graph + ore $\pm$ tour $\pm$ ap

(14) gt + st + ctd + biot + chl + musc + qz ± ep ± plag ± graph + ore ± tour ± ap

(15) gt + ctd + biot + chl + musc + qz ± plag ± ore

(16) gt + st + ky + biot + chl + musc + qz + ep ± graph ± ore + tour + ap

(17) gt + st + ky + ctd + biot + musc + qz + ep + plag + rut + ore + tour+ap

(18) gt + st + ky + ctd + biot + chl + parag + musc + qz + ep + plag + ore+ap

(19) gt + biot + marg + musc + qz + ep + plag + ore + tour + ap

(20) gt + biot + chl + chl + marg + musc + qz + ep + plag + rut + ore + tour + ap

(21) gt + biot + musc + ep + plag + ore + tour

(22) gt + musc + ep + plag + graph + ore

Assemblages (1) and (2) are most common, and (5), (10), (11), (17) to (23) are rare. Nearly all rocks with assemblages (13) to (21) display some textural features of disequilibrium. Some of the assemblages are discussed in more detail in Chapter 6 where they are related to chemical mineralogy and metamorphic zonation. A few salient features are listed below, hopefully without generalising to the point of uselessness.

A variety of textural types of garnet have been found. In quartz-rich schists, they tend to have poorly developed crystal faces and abundant quartz inclusions, while those in fine-grained quartz-poor schists are generally euhedral. In either case, the garnets may have grown before, during or after the development of the schistosity in the micaceous matrix, giving rise to a variety of forms: rounded with quartz pressure -shadows, sigmoidal, cross-cutting, etc. (Fig. 4.1b), and inclusion trails may be straight, folded, sigmoidal, convergent, or concordant or discordant to the matrix schistosity. In certain schists rich in quartz stringers, flat garnets have grown parallel to and cutting across the schistosity. Detailed description of these garnet textures is left until Chapter 4 by which stage the various generations of

schistosity will have been identified.

The most common inclusions in garnet are of ilmenite, chloritoid, graphite and quartz, although staurolite, chlorite epidote and tourmaline also occur.

Flakes of biotite, typically a pale reddish brown variety, generally lie parallel to the white mica schistosity or form 1 to 2 mm cross-cutting porphyroblasts. Both types may be found together in the same rock. Biotite is partially altered to chlorite with anomalous blue interference colours in a few rocks. However, in many parageneses, abundant chlorite with brownish interference colours has grown parallel to the schistosity and appears to be primary, even in some of the staurolite and kyanite bearing assemblages (12) to (16). Although common in the pelites of the Inner Schieferhülle, cross-cutting sprays of chlorite are unknown in those of the cover.

Chloritoid, pleochroic with α : greyish green $\approx \beta$: greyish blue violet $> \gamma$: pale yellow; forms blocky plates up to 2 mm across with various relationships with the mica schistosity. Pre-tectonic chloritoid tends to form large, ragged, sometimes bent porphyroblasts with lamellar twinning. Occasionally these are wedged at an angle to the schistosity and quartz pressure shadows have developed. Syn-tectonic chloritoid is most common and occurs as small plates lying in the schistosity (Fig. 4.20). Post-tectonic plates cutting across the schistosity are rare and are restricted to assemblages (9) and (15). Types (a) and (b) or (b) and (c) have been found together in some slides. In some rocks with assemblages (14), (17), (18) and (20) the only chloritoid present occurs as inclusions within garnet and seems to have been preserved from reaction beyond the point where it is no longer stable in the presence of the minerals outside the garnets (Fig. 6.13).

The staurolites in Peripheral Schieferhülle pelites are pale compared with

those from other terrains. They are mustard-coloured in hand-specimen and pale watery yellow in thin section. They generally form cross-cutting euhedral inclusion-free prisms, often with knee-twins (Fig. 4.20). No staurolite-bearing assemblages have been found by the author west of Mallnitz.

Kyanite was found in four Peripheral Schieferhülle pelites, in all of which it seems to be in equilibrium with garnet, staurolite and biotite. The textural relationships of kyanite to the mica schistosity are unclear but seem to mimic those of staurolite. It has not been found as inclusions within garnet. In GD 18 and GD 438 (assemblages (18) and (16) respectively) some of the kyanite has been replaced by oligoclase.

Rounded porphyroblasts of albite, similar to those in metabasites, occur in some of the less aluminous pelites and include trails of graphite and quartz. Many have been rotated so that these trails lie at high angles to the schistosity. More common, however, are patchy porphyroblasts of oligoclase-andesine or albite up to 7 mm long which appear to have grown at the expense of muscovite. These occur individually or as overgrowths round albite cores (Fig. 4.10) and in some cases enclose biotite in random orientation.

Paragonite has been identified in three schists. In all of them it has a very fine-grained schistose texture (Fig. 4.19). Margarite occurs in two pelites (GD 49, GD 50a) from Zwenberger tal. As in the calc-schists, it is being replaced by plagioclase, and is not found in contact with quartz.

The characteristic ore mineral is magnetite which forms octahedra or skeletal crystals up to 1 mm. Ilmenite is less common. It usually forms plates aligned in the schistosity and sometimes also in trails included within garnet.

(i) Weiss-schiefer

This term has been used by Exner (1957) and Norris et al. (1971b) to describe very fissile mica schists composed essentially of muscovite and quartz. They are associated with pelites outcropping along the Mölltal where they are, if anything, more abundant, but are rare in the Mallnitzer Mulde area. On Maresen Spitze and in Seebach tal 10 and 50+ m respectively of weiss-schiefer lie immediately above basement amphibolites and below pelites. This suggests that these rocks mark a zone of intense shearing above autochthonous basement. Norris et al. consider them to be of metasomatic origin, in which case they may bear a similar relationship to pelites as the quartz-phengite-schists do to augen-gneisses.

(j) Coarse-grained Chloritoid-Schist and other odd rocks

The coarse-grained chloritoid-schist outcrops in a 2 m thick band within white marbles on the east side of the Tauern Bach gully at an altitude of 2000 m. Large chloritoid porphyroblasts up to 1·5 cm across (Fig. 2.38) are abundant in all varieties (approaching 50% in some) and may be accompanied by rounded porphyroblasts of albite, garnet or needles of hornblende (Fig. 2.39).

Some of the assemblages are:

(1) ctd + gt + chl + ab + musc + mt + ap

(2) ctd + gt + biot + chl + ab + musc + mt + ap

(3) ctd + gt + hbl + chl + ab + musc + qz + ilm

(4) ctd + hbl + biot + chl + musc + qz + cc + mt + ap

(5) ctd + chl + ab + musc $\pm$ qz + ilm

The chloritoid has strong pleochroism and dispersion. It appears to be in equilibrium with garnet, rounded albite, biotite, quartz, ore minerals, and hornblende. The garnets are large, somewhat fractured, and enclose albite and aligned plates of chloritoid and ilmenite. Blue-green hornblendes form 2 cm long euhedral prisms which also enclose aligned ilmenite and chloritoid, but are intergrown with the latter in GD 821b (assemblage (4)). In most specimens, albite forms rounded sub/euhedral 1 to 5 mm grains enclosing trails of ore. Some have been rotated. In GD 821a, granulose albite and magnetite occur together in flattened patches making up half the rock. Late stage albite replaces white mica in some rocks and enclosed all other minerals. The biotite is dark chocolate-brown. The chlorite, an optically negative variety with grass-green to straw yellow pleochroism and anomalous greenish-yellow interference colours, has grown at the expense of chloritoid, garnet and biotite. Calcite occurs in a quartzitic facies, apparently in equilibrium with the other phases. In the rocks with little or no quartz, fine-grained schistose muscovite is cross-cut by patches of decussate muscovite and chlorite. These rocks seem to be extremely iron-rich aluminous pelites (possibly meta-clay-ironstones) with variable amounts of admixed calcareous material introduced either during deposition or from the surrounding marbles during metamorphism. Similar chloritoid-rich rocks have been described by Liou and Chen (1978) who concluded that they represent metamorphosed laterites.

Another unusual lithology was found in a 20 cm band between calc-schist and pelite on Wasenle Kopf (Map 2). It is a typical garbenschiefer, with conspicuous sprays of hornblende needles set in a schistose white-mica matrix (Fig. 2.40). It has the assemblage:

gt + hbl + ctd + biot + chl + ep + marg + musc + qz + plag + ore + ap

It is a complex rock, and certainly not in equilibrium. Texturally it has many similarities with those described above, but is unusual in that it

contains a large amount of fine-grained schistose margarite. This, together with muscovite, is being replaced by andesine-labradorite overgrowths.

E. LATE STAGE VEINS AND RELATED ROCKS

(a) Late-Stage Veins

Veins cutting across foliation are a common feature in Zentralgneis, Inner and Peripheral Schieferhülle rocks. They fall into two categories : (a) quartz veins, (b) chlorite-albite veins. The latter are often associated with gentle buckle folding (veins generally parallel to fold axial planes) or boudinage (vein minerals grow in boudin necks). Fine examples of these features can be found, respectively, at the summit of Auernig, where veins cut epidote-amphibolite, and in the pelitic schists exposed on the Jamnig Alm road section.

The following assemblages were recorded:

(1) qz

(2) chl

(3) chl + ab

(4) chl + ab + qz

(5) chl + ab + qz + ilm

(6) chl + ab + qz + sph

(7) chl + ab + qz + cc

(8) chl + ab + qz + cc + rut + sph

(9) chl + ab + sph

(10) chl + ab + rut + sph

(11) chl + ab + cc + ilm + sph

(12) chl + ab + cc + act

(13) chl + ab + qz + cc + ilm

Albite and quartz mostly occur as euhedral crystals about 1 cm long. Fine-grained massive chlorite fills the interstices between these euhedra, or, in vuggy veins, forms a thin coating over and between them. Chlorite occasionally occurs as discrete barrel-shaped rosettes. The sphene is pale yellowish green and rims rutile where both are found together. The minerals appear to have crystallised in the order: rut, ab, ilm, qz/sph, chl/act. The position of calcite in the sequence is variable.

Alteration of the country rock is sometimes observed immediately adjacent to these veins. In granite-gneisses and mica-schists biotites have become altered to chlorite up to 2 cm either side of even the thinnest hair-line fractures. Amphibolites tend not to be so badly affected, and it almost seems as if the vein assemblages are in equilibrium with Peripheral Schieferhülle epidote-amphibolites. Fig. 2.41 illustrates the mineralogical changes observed in a specimen of garnet amphibolite from Maresen Spitze where it has been cut by a thin vein with assemblage (6).

(b) Pseudotachylite:

Two loose blocks of finely banded gneiss containing pseudotachylite were found in a large boulder field in middle Düsener Tal. They were probably derived from somewhere on the ridge between Zaguten Spitze and Wabnig Spitze. In both specimens, the pseudotachylite forms a 2 to 5 mm thick band of brownish black glass roughly parallel to the banding. In one, the rock adjacent to the glass is fractured (Fig. 2.42) and glass fills short 1 cm veins perpendicular to the main band. The coincidence of the glass bands with small scale features

(microfaulting, etc.) indicative of relative movements suggests that they have been produced by melting of the host rocks by frictional heat released during fracture and slip along isolated foliation surfaces.

In thin section, the bands are zoned (Fig. 2.43) with thin margins of dark brown glass and thicker central zones of partially devitrified glass. Both parts contain numerous angular to rounded fragments of quartz and plagioclase which, in the central zone, are rimmed by dark brown glass. The host rock is coarse grained but quartz and feldspar have been finely crushed along low- and high-angle microfractures.

CHAPTER 3

Structural Geology

CHAPTER 3

STRUCTURAL GEOLOGY

A. INTRODUCTION

Aims of this Chapter

This chapter aims to describe the mesoscopic and macroscopic structures encountered in the map areas. This is undertaken qualitatively and no attempt is made to calculate the strain represented by individual structures. The information will be used to establish a time sequence of deformation events and hence a relative time scale in which to place phases of mineral growth. The relationship between small-scale structures and the textures of minerals in thin section are examined in Chapter 4.

The metamorphic and tectonic evolution of an orogenic belt are closely interrelated. Structures within a metamorphic terrain provide information on large-scale relative movement of rock masses which may be responsible for the initiation, development and termination of the metamorphism itself, while mineral parageneses in the rocks (see Chapter 7) provide information about temperature and structural depth at certain stages during these movements.

Implications of the structural geology for the interpretation of the stratigraphic succession in the Peripheral Schieferhülle are discussed. The data are also used to examine spatial variation in style and orientation of isochronous structures of over the south-east Tauern.

Previous Work

Few studies involving structural work have been undertaken in the area examined in this thesis. Prior to the work of Norris et al. (1971b) no detailed investigations into structural sequences were attempted.

Angel and Staber (1952) and Exner (1957; 1964) mapped many principal foliation dips and lineation plunges over an area covering much of the Sonnblick and Ankogel groups and including the area of Map 2. However, no mention was made by them of any but the most prominent minor structures. Although they noted one or two localities in which an older lineation is deformed about a younger one, structural age relationships were not discussed in any detail.

In their study in the Reisseck and southern Ankogel groups, Norris et al. (1971) recognised one, possibly two, phases of isoclinal folding (D_v) in the Inner Schieferhülle, prior to the intrusion of the Zentralgneis, and a subsequent phase of deformation (D_a^1) common to all the Pennine units. This later event produced asymmetric folds in the Inner Schieferhülle, a strong penetrative foliation in much of the Zentralgneis and asymmetric tight to isoclinal folds in the Peripheral Schieferhülle of the Möll tal. The intensity of D_a^1 in the basement complex was shown to decrease structurally downwards from the basement-cover interface. This, and the observation that D_a^1 folds have subhorizontal axes and similar senses of asymmetry in both Inner and Peripheral Schieferhülle, were taken as evidence that D_a^1 structures were produced during the emplacement of the Altkristallin sheet as it was thrust in a northeasterly direction over the Penninic rocks.

Norris et al. also recognised a late minor set of kinks and joints (D_a^2) which they attributed to movements during the regional updoming of the Tauern. The rotation of the regional foliation into a subvertical orienta-

tion parallel to the south-east trending fault running the length of the Mölltal (the Mölltal line) has also been ascribed to this event.

Methods of Study

The interpretation of structural observations in the field entails two distinct steps - the measurement and description of structural elements at individual exposures and the correlation of the elements between exposures on all scales. In the present study, more emphasis was given to the detailed study of critical localities where time-space relationships could be established than in the accumulation of very large numbers of measurements.

The existence of more than one phase of deformation can only be established when structures corresponding to one phase are seen to be deformed by another. For instance, to prove that two fold generations exist, it is necessary to see the axial surface of one bent round another. Generally, exposures showing clear-cut overprinting relationships of this sort are rare in the area studied. Often, only one set of structures is represented. Early structures may be totally masked by later ones. Later structures may not be visible if the associated deformation was inhomogeneous and not very intense, (eg. localised boudinage, kinking, etc.). It is often not possible to recognise the oldest of two or more sets of lineations on the same unfolded foliation surface.

In areas of good exposure individual structures can be traced for considerable distances. Unfortunately, in the areas mapped, exposure only becomes semi-continuous at altitudes greater than about 2000 m. The ridge crests afford completely continuous exposure, but access is often difficult; in

these places 'binocular mapping' of large scale structures from vantage points lower down has proved very useful.

At altitudes less than 2000 m, particularly on steep, forested valley sides, adjacent exposures may be hundreds of metres apart, and minor structures can only be correlated indirectly. Correlations are made on the basis of grouping structures with similar style and orientation, coupled with a knowledge of their age relationships. Great care must be taken in applying such criteria because (a) the orientation of some planar and linear features may change abruptly if large folds are common, (b) in this kind of metamorphic terrain, large scale folds are commonly non cylindrical, (c) the style and orientation of structures are dependent, not only on the regional stress environment, but also on the nature and geometry of the preexisting fabric (Norris, 1970; Bickle, 1973; Holland, 1977), and (d) fold style groupings do not always correspond to fold generations (Park, 1969; Norris, 1970; Hobbs et al., 1976). These are however the only methods available. The main test of any interpretation is internal consistency and compatibility with petrological and other constraints.

Inherent in the above described are the assumptions (a) that deformation in the rocks of the south-east Tauern occurred in discrete episodes separated by relatively long periods of static mineral growth, and (b) that these episodes were not diachronous over the mapped area on the time-scale of the static periods.

These are not unreasonable assumptions. Cheeney (1971) has calculated that it is theoretically possible to form a large flattened fold of amplitude 700 times the initial perturbation in as little as 0.1 my. The overthrusting of the Altkristallin sheet, probably responsible for one deformation phase, may have been completed in 1 to 5 my if geologically reasonable subduction

rates of 3 to 15 cm per year are assumed. On the other hand, geochronological studies suggest that the duration of the Alpine metamorphic cycle was greater than 30 my (Cliff et al., 1971b) or 50 my (Hawkesworth et al., 1975).

It is unlikely that the deformation phases are diachronous because the small scale structures which define them, at least in the case of D_a^2, are clearly related to megascopic folds on the scale of the map area or larger.

The map areas have been subdivided into six separate subareas for the purpose of presenting structural data. Individual subareas were chosen to group together localities at which the orientation of principal fold axes are broadly consistent. These are as follows (Fig. 3.1):

1. Böseck subarea (southern margin of Map 2)
2. Romate subarea (north-west corner of Map 2)
3. Auering subarea (north-east corner of Map 2)
4. Kaponig subarea (central and southern parts of Map 1)
5. Sickerkopf subarea (north-west corner of Map 1)
6. Dösener subarea (north-east corner of Map 1).

Rocks of the Inner Schieferhülle, Zentralgneis and Peripheral Schieferhülle are represented in all except the Dösener subarea which is located entirely within the basement complex. Consequently, no fundamental differences in structural sequence are expected between subareas. In the following sections the macroscopic and mesoscopic structures present in the subareas are described. Most attention is given to examples where structures of different generations have been superimposed.

Following the convention of Norris et al. (1971b), phases of deformation will be given the label D. Similarly, phases of folding and generations of

schistosity will be labelled F and S respectively. The subscripts 'v' and 'a' serve to distinguish phases attributable to Variscan and Alpine events. The superscripts 1, 2, etc., refer to the first, second, etc., recognisable phase in each event. Thus, D_a^1 refers to the first Alpine phase of deformation and F_a^2 to the second Alpine phase of folding. Confusion may be avoided if it is emphasised that a deformation phase labelled, for instance, D_a^1 by one author in one area may not be the same as that given the same label by another in a different area.

The attitude of planar and linear features were measured using a Clar Gefügekompass (Clar, 1954). The errors in the clinometer and compass readings are estimated as $\pm 2^o$ and $\pm 1^o$ respectively. Structural measurements are plotted on equal area, lower hemisphere, stereographic projections with north at the top. Directional data are abbreviated to x/y, where x is the dip or plunge and y is the compass bearing measured clockwise from grid north.

The terminology of Fleuty (1964) is used in describing the shape of folds.

B. BÖSECK AND ROMATE SUBAREAS

The Böseck and Romate subareas are described first because structural time relationships are displayed more clearly within them than in the others. Together, they constitute a 'type area' to which structures in other subareas can be compared. Principal fold axes have shallow north-westerly of south-easterly plunges in the Böseck subarea, and shallow south-south-easterly plunges in the Romate subarea.

The areas of outcrop of the main structural units can be summarised as follows:

Peripheral Schieferhülle	: Mallnitzer Mulde
Zentralgneis	: North-east margin of the Sonnblick Kern. Romate-decke and Feldwand-lamella (only north-west parts). Gneiss lamellae within Mallnitzer Mulde.
Inner Schieferhülle	: Thin zone of Randgneis lithologies flanking the Sonnblick Kern. Bands of Randgneis schist below the Romate-decke and Feldwand-lamella.

(a) D_a^1 Structures

A striking feature of the rocks in these two subareas is their large-scale lithological banding (Fig. 3.3). Rocks of all three structural units show this. Map 2 shows that individual bands tend to maintain a constant thickness (often only a few metres) and are usually traceable for several kilometers. The main exceptions to this are the slices of calc-schist, marble,

dark biotite-schist, quartzo-feldspathic gneiss and others that together form the 100 to 200 m wide unit, hereafter called the Böseck-lenticle zone, that flanks the Sonnblick Kern.

Nearly all the lithologies possess, on hand-specimen scale, a strongly developed foliation parallel to the compositional banding which, in calcareous and graphitic quartzites takes the form of thin micaceous laminae separating 1 to 30 cm thick layers of massive unfoliated quartzite (Figs. 3.2, 3.5) In calc-schists it is generally a millimetre- t- metre-scale colour-banding. In pelitic and graphitic mica-schists, quartz-phengite-schist and biotite-epidote-feldspar-schist, the foliation is a strong mica-schistosity which, in the first two, is accompanied by concordant quartz veining. In metabasites, it is an epidote-banding and a chlorite + amphibole schistosity. Ultrabasic schists develop a talc-schistosity but serpentinites generally have no internal planar fabric.

This foliation has been called S_a^1 because it is the earliest deformational planar feature generally recognisable in Peripheral Schieferhülle lithologies. It cannot be solely a primary depositional bedding in the metasediments because

(i) certain units (such as the quartzites on the west slope of Geisel Kopf and the greenschist band north-east of Rissieck) have been stretched out into discontinuous pods, i.e. boudinaged on a large scale;

(ii) isoclinal folds (F_a^1) occur within the banding, and their axial surfaces have been folded with it in subsequent deformation events.

By virtue of its concordance with S_a^1 in overlying cover rocks, the foliation

in the Zentralgneis lithologies is assumed to be the same feature. In the more massive augen-gneisses and syenite-gneiss it is a poorly developed flattening and attenuation of quartz + feldspar aggregates with a subordinate mica schistosity. Towards the margins of both the Sonnblick Kern and the Romate-decke the foliation becomes much more obvious and the schistosity becomes dominant.

The more common type of F_a^1 folds are isoclinally folded quartz veins in mica-schists. The S_a^1 schistosity is axial planar to these isoclines (Fig. 3.4). In places their limbs have become so attenuated that only isolated fold hinges remain, producing a quartz rodding lineation.

Less common are examples in which an earlier compositional banding, possibly the original bedding, has been folded. Some are illustrated in 3.6, 3.7, and 3.8. In these examples, the F_a^1 isoclines have been overprinted by a subsequent fold generation (F_a^2). In quartzite, the later folds tend to be open, and relationships are clear; in the calc-schists shown in Fig. 3.8 they are tight and give rise to complex interference patterns. D_a^1 mineral lineations have been found in ultrabasic schists, where tremolites are bent round F_a^2 small fold hinges (Fig. 3.9).

No large-scale F_a^1 folds have been identified. It is possible that the two quartzite bands which outcrop 300 to 400 m north of Zedöllnik represent the limbs of an F_a^1 isocline, particularly as they appear to merge south-west of the Hagener Hütte, but adjacent units are not symmetrical about these 'limbs', so this seems unlikely. If, however, the gneiss-lamella which outcrops between the two quartzites represents a slice of Zentralgneis basement, it must have been thrust into the Peripheral Schieferhülle cover sequence before or during the initial stages of D_a^1 as it, too, has been deformed by this event. Many of the units in the Mallnitzer Mulde, therefore, may have been

repeated by thrusting prior to D_a^2. The serpentinite pods were probably also emplaced during D_a^1. The Romate-decke and Feldwand-lamella are interpreted as flattened thrust slices cut from one syenite-complex within the original pre-Alpine basement.

(b) Pre-D_a^1 Structures

Two pieces of evidence suggest that in places, the compositional banding folded by F_a^1 was deformed previously.

Within one of the graphitic quartzite bands which outcrop 1.3 km south-east of Ebeneck, asymmetric tight to isoclinal F_a^1 folds, with axial planes parallel to S_a^1, deform a banding containing flattened white quartz veins (Fig. 3.10). As this graphitic quartzite probably belongs to the Inner Schieferhülle (see Chapter 2, part B(g)), the early foliation may have been formed in a pre-Alpine phase of deformation, D_v. Waters (1976) found graphitic quartzites in the Altkristallin which contain structures corresponding to early deformation events not preserved elsewhere, and attributed this to the extremely competent nature of this lithology.

The second piece of evidence comes from the F_a^1 isocline in Peripheral Schieferhülle calcareous quartzite, shown in Fig. 3.7. The nose of the isocline appears to contain a type II, or 'boomerang', interference pattern (Ramsay, 1962b). If this is the case, then F_a^1 must refold a still earlier set of Alpine folds. The pre-F_a^1 foliation in the graphitic quartzite may also be of this generation.

Taken by themselves, these two occurrences are insufficient evidence for extra phases of deformation.

(c) Orientation of S_a^1 in relation to the Mallnitzer Mulde and Sonnblick Kern

The recognition of the Mallnitzer Mulde as a large synform (Exner, 1957) was based largely on the observation that the macroscopic lithological banding is symmetrically disposed about the main calc-schist unit (Geisel Kopf - Zedöl-nik - Törl Kopf) and that certain units can be traced from one limb to the other around a fold that closes downwards.

Exner recognised the following structural succession in the Mallnitzer Mulde:

(5) Calc-schist, greenschist zone

(4) Quartzite, dolomite, gneiss-lamella zone

(3) Middle graphitic mica-schist ('black phyllite') zone

(2) Quartzite, dolomite, marble, calc-schist, greenschist zone

(1) Basal mica-schist and lower 'black phyllite' zone

(0) Sonnblick Kern (south-west), Romate-decke and Hölltor Kern (north-east).

The present study largely supports the above conclusions. The Peripheral Schieferhülle units occupying the central part of the Mallnitzer Mulde (i.e. between the Böseck zone and the Lonza-Jamnig Alm gneiss-lamella) have an excellent symmetry (see Map 2). These correspond to units (3), (4) and (5) above. In particular, the core of calc-schist (5) is separated from graphitic mica-schist (3) on both limbs by a sandwich of two calcareous quartzite units with a 'filling' of quartz-phengite-schist, marble, etc. (4). The quartz-phengite-schist (Exner's gneiss-lamella 4) is immediately below the upper quartzite in the lower limb, but above the lower quartzite in the upper limb.

A single large recumbent isoclinal fold, closing downwards and towards the south, can be traced in the main calc-schist unit in the cliff 1.5 km south-south-east of the Jamnig Alm. As it occupies the central two-thirds of the unit, and has a half-wavelength of over 100 m, it must represent the core of the Mulde itself. However, contrary to previous ideas, the banding at high structural levels within the Mulde, on Geisel Kopf, is not folded into a simple synform; it is more accurately described as a synclinorium (Map 2A). This is discussed further in section (e).

There is very little correlation between units in the Böseck lenticle zone and those north of the Lonza-Jamnig Alm gneiss lamella (covered largely by the Romate subarea). The latter amount to over 1 km thickness whereas the former total about one tenth of this. Although calc-schist, marble and mica-schist are common to both areas, there is no indication of syenite-gneiss or biotite-epidote-feldspar-gneiss in the Böseck lenticle zone. It is possible that the bands of dark biotite-schist and quartzo-feldspathic gneiss (neither of which outcrop in the Romate subarea) could be correlated with the Romate-decke. The author's interpretation of the structural succession in the Mallnitzer Mulde is summarised in Table 3.1 and can be compared with the cross sections in Map 2A.

The attitude of D_a^1 structures in the Böseck and Romate subareas are plotted stereographically in Figs. 3.11 and 3.12. The Romate subarea includes most of the units that form the lower limb of the Mallnitzer Mulde. Within it, S_a^1 has a uniform dip towards the south-west, resulting in the tight clustering of poles to S_a^1 in Fig. 3.12a. The curious outcrop pattern of the main marble band in Map 2 is caused by S_a^1 dipping almost parallel to the ground surface on the northern slope of Tauerntal (Fig. 3.3).

Within the Böseck subarea, S_a^1 in the lower limb (i.e. north-east of the

main calc-schist unit) dips parallel to S_a^1 in the Romate subarea. However, in the upper limb, the dip is demonstrably steeper (Fig. 3.13) and, locally, much more variable, due to subsequent folding. At three places, all topographically high (eastern slopes of Rissieck, Böseck and Feldsee Kopf), units in the Böseck lenticle zone are folded round the nose of the Sonnblick granite-gneiss Kern. At low altitudes the bands dip towards the south-west as before, but become steeper, vertical, and eventually north-east dipping as they climb. On the summit of Feldsee Kopf, the foliation is horizontal, and a flat cap of graphitic mica-schist rests on top of augen gneisses (Fig. 3.15).

The S_a^1 foliation within the granite-gneisses of the Sonnblick Kern follows a similar pattern (Figs. 3.16 and 3.17), and the boundary between the gneisses and surrounding units appears to be exactly concordant. Cross-sections through the Sonnblick Kern and Mallnitzer Mulde (Map 2A) illustrate these relationships.

The rotation of S_a^1 in the south-west of the Böseck subarea results in the spreading out of poles to S_a^1 along a great circle in Fig. 3.11 a. The centre of this great circle represents the axial direction of both the Mallnitzer Mulde and the Sonnblick Kern, and plunges roughly 0/135, which agrees well with the plunge of their hinges as deduced from Maps. 2 and 2A.

To conclude, the Mallnitzer Mulde and Sonnblick Kern form an asymmetric synform-antiform-pair with subhorizontal axes trending north-west/south-east. Their formation must have post-dated D_a^1 because the principal foliation, S_a^1, has been folded round them.

(d) Orientation of F_a^1 Fold Axes

Of those studied, the Romate subarea has been least affected by post F_a^1 folding. The constant dip of the banding in the lower limb of the Mallnitzer Mulde provides an ideal opportunity to study the orientation of F_a^1 fold axes and quartz-rodding. These are plotted in Fig. 3.12 b. In general they have a shallow southerly or south-westerly plunge, but a few depart significantly from this direction.

The few definite D_a^1 generation quartz-rods found in the Böseck subarea plunge gently towards the north west or south-east, parallel to F_a^2 small-fold axes (Fig. 3.11 b). Many may have been mistaken for F_a^2 quartz rods. Also, the orientation of F_a^1 rods may have, to some extent, induced F_a^2 small folds and rods to develop parallel to them. In Figs. 3.4 to 3.7, the interference patterns generated by the superposition of F_a^2 folds on F_a^1 isoclines are of type III (Ramsay, 1962b; 1967) which are only produced when the axes of both fold generations are sub-parallel.

It is conceivable that the initial angular relationships between F_a^1 and F_a^2 axes are preserved in the Romate subarea but unlikely because the Mallnitzer Mulde is a flattened similar fold (Ramsay, 1962 a). Although the lower limb is comparatively free of later folds, it may have undergone considerable stretching in a north-easterly direction, i.e. perpendicular to the axis.

(e) D_a^2 Structures

The overprinting of D_a^1 structures by D_a^2 is clearly displayed in the Romate and Böseck subareas where the main effect of D_a^2 is to fold S_a^1 without obliterating it. The majority of recognisable folds belong to F_a^2.

Calcareous quartzites respond to D_a^2 by developing tight to gentle folds with wavelengths of a metre or more (Figs. 3.2, 3.7 and 3.18), and millimetre-scale axial crenulations in the micaceous partings. Fold hinges tend to be sharp in finely-banded varieties.

In rocks with a strong penetrative S_a^1 schistosity, the style of F_a^2 folds is very variable. Slightly flattened, metre-scale, open asymmetric buckles are common throughout both subareas, but towards the core of the Mallnitzer Mulde, tighter chevron-folds are developed on a smaller scale (Fig. 3.4). The former are usually accompanied by a parasitic, millimetre-scale, axial crenulation. Where D_a^1 lineations are present, they are folded round F_a^2 hinges (Fig. 3.21). A strain-slip schistosity, S_a^2, has begun to develop in the tightly crenulated cores of a few F_a^2 folds. On flat, unfolded S_a^1 surfaces, the presence of a faint microripple lineation is often the only manifestation of D_a^2.

Mica-schists with abundant quartz veins in S_a^1 develop an axial quartz-rodding where tightly crenulated by F_a^2. The distinction between this and the D_a^1 rodding is not always clear.

Mesoscopic F_a^2 folds in calc-schists and marbles are only developed in the colour banded varieties. Most are tight to isoclinal and strongly flattened, with orthogonal band thickness at the hinges many times greater than in the limbs. The dominant muscovite-schistosity in carbonate rocks is axial planar to these folds. The trace of the colour banding on schistosity surfaces defines an axial intersection lineation. Quartz and calcite rods are also aligned in this direction. It is rare to find small folds in augen gneisses with poorly developed S_a^1 schistosity, but where they do occur they form gentle or open buckles (Fig. 3.17). Normally, the only evidence of D_a^2 is a mineral lineation on foliation surfaces.

The orientations of F_a^2 quartz rods, small fold axes and intersection lineations are plotted in Figs. 3.11 d and 3.12 d. In the Romate subarea, most plunge gently towards the south-south-east. In the Böseck subarea, they are tightly clustered about a mean direction ca. 0/135, which coincides roughly with the pole to the π-pole girdle of Fig. 3.11 a. As this is also the inferred trend of the Mallnitzer Mulde and Sonnblick Kern, it seems likely that these structures are merely very large F_a^2 folds. This is confirmed by (a) the presence on Böseck, Geisel Kopf, Törl Kopf and elsewhere of intermediate scale (100+ m wavelength) asymmetric folds on the same axes, and (b) the fact that within each limb of the Mallnitzer Mulde, S_a^1, S_a^2 and small fold asymmetry have a consistent and complementary relationship to one another, i.e. nearly all F_a^2 folds in the Romate subarea are clockwise (looking north-west) with S_a^2 steeper than S_a^1, consistent with their being situated on the lower limb of a large F_a^2 synform. In the Böseck subarea, south-east of Zedöllnik, the same relationship applies north-east of the axial trace of the Mallnitzer Mulde, but south-west of this line, S_a^2 is shallower than S_a^1 (Fig. 3.13) and most folds are either anticlockwise looking north-west (Fig. 3.14) or symmetrical. The orientation of S_a^2 schistosity and F_a^2 axial surfaces are plotted in Figs. 3.11 c and 3.12 c.

The situation in the northern part of the Böseck subarea along the ridge between Feldsee Kopf and the Hagener Hütte is less straightforward. Except in one or two short stretches, F_a^2 folds are clockwise (looking north-west) along the whole length of this section. There is no switch in the sense of asymmetry within the main calc-schist band to mark the axial trace of the Mallnitzer Mulde, and the upper limb of the Mulde appears to be completely absent.

Fig. 3.20 shows the south side of Geisel Kopf. The top of this mountain is composed of calc-schists, greenschists and quartzites at a very high struc-

tural level. The place of the single synform that forms the Mallnitzer Mulde at lower structural levels is taken by a cascade of asymmetric 100+ m wavelength folds. The isolated stretches of symmetrical or anticlockwise small folds correspond to the short limbs of these larger folds.

The Mallnitzer Mulde and Sonnblick Kern are very large F_a^2 folds containing parasitic folds with wavelengths ranging from hundreds of metres down to millimetres. The former is a tight to isoclinal synform at low structural and topographic levels, but opens up and eventually dies out at higher levels.

(f) D_a^3 Structures

A variety of locally developed minor structures occur in the Böseck and Romate subareas; these have been grouped together as D_a^3 irrespective of the fact that they are almost certainly polyphase.

Three main types of D_a^3 structure have been found:

(i) Microripple lineation - wavelength 2 mm to 1 cm - mostly crenulating S_a^1.

(ii) Boudinage - boudin necks take the form of small veins filled with chlorite, albite, quartz, etc. - deforming S_a^1.

(iii) Kinks.

The relationships between some of these and earlier structures is best seen in the mica-schists exposed in the Jamnig Alm road section. At one locality (hairpin bend, altitude 1520 m) three sets of lineations are visible on one

S_a^1 surface: a D_a^1 rodding, a D_a^2 crenulation and chlorite lineation, and a gentle microripple. At nearby exposures, the F_a^2 lineation (parallel to axes of isolated F_a^2 folds) is bent by the boudinage, while the microripple is parallel to the boudin axes. The boudinage and microripple seem therefore to be related, and these lineations are plotted together in Fig. 3.12 e. In the Romate subarea, most D_a^3 lineations plunge gently towards the east or west, but in the Böseck subarea, they are less well defined.

In both subareas, the orientations of kink axes are poorly clustered (Figs. 3.11 f and 3.12 f) and partly overlap with the field of other D_a^3 lineations. The relationship of the kinks to other D_a^3 structures is not clear.

C. AUERNIG SUBAREA

The sequence of structural events recognised in this subarea is the same as that described in the previous section. To avoid repetition, only the important differences will be discussed, the main one being the orientation of F_a^2 folds, which plunge in a south-westerly direction (Fig. 3.1). Fold axes appear to swing gradually into this orientation approaching the Auernig subarea from the adjacent Romate and Sickerkopf subareas, but the boundary between the Auernig and Böseck subareas is marked by a sharp 90° discordance in axial direction. The major structural units outcrop as follows:

Peripheral Schieferhülle: core of Ankogel Mulde (mostly epidote-amphibolite). Part of the north-east margin of the Mallnitzer Mulde.

Zentralgneis: Höltor Kern (augen-gneisses). Eastern extremities of Romate-decke and Feldwand-lamella.

Inner Schieferhülle: Randgneis schists in the Ankogel Mulde and overlying the Höltor Kern. Western margin of the Hochalm Kern (amphibolites, etc.)

(a) D_a^1 Structures

The dominant foliation in the Auernig subarea is S_a^1. This is shown by the observation that units such as the Romate-decke and underlying schists can be traced directly from the adjacent Romate subarea. Although many of the rocks here belong to the Inner Schieferhülle, the foliation cannot be merely a slightly modified S_v because the same features pass continuously into Peripheral Schieferhülle rocks.

F_a^1 folds are less abundant than in the Böseck and Romate subareas. The mica-schists of the Seebach Mulde contain abundant D_a^1 quartz-rods and rare isoclinally folded quartz veins (Fig. 3.19). The only example of an F_a^1 isocline folding a compositional banding occurs within a thin marble band refolded by F_a^2 (Fig. 3.22).

(b) Orientation of S_a^1 in relation to the Höltor Kern and Ankogel Mulde

On the south-facing slope of Lisgele Spitze, S_a^1 dips parallel to the ground surface, and the triangular patch of Inner Schieferhülle schists (Map 2) is

merely a thin covering on top of augen-gneisses of the Höllton Kern (Fig. 3.23). This is demonstrated by the existence of two erosional windows through the schists into the underlying gneisses. However, on the east side of Hindenburg Höhe, S_a^1 swings round to dip eastwards (Fig. 3.26), steepens and becomes overturned at low altitudes in the west flank of Seebachtal. Thus the Höllton Kern is an asymmetric antiform analogous to the Sonnblick Kern in that S_a^1 is deformed round it, but differs in that its axis plunges towards the south-west.

The Ankogel Mulde occupies the structural depression between the Höllton Kern, and the Hochalm Kern, taking the form of a recumbent, asymmetric synform plunging towards the south-west and closing towards the north-west. The limbs are of Inner Schieferhülle schists and dip towards the east, but the core is formed of Peripheral Schieferhülle, mostly epidote-amphibolites (the Auernig mass) and pelitic schists. The following unusual geometrical features of the Auernig mass can be explained if, and only if, it occupies the core of the Mulde:

(i) The epidote-amphibolite outcrop can be traced south-eastwards into the main metabasite unit in the Kaponig subarea, but in all other directions it abuts directly against pelitic schist.

(ii) The epidote amphibolite structurally overlies pelitic schist on Wasenle Kopf but underlies it west of Mallnitz railway station. At both localities S_a^1 dips westwards. The contact can be traced from one to the other round the west side of Auernig where the dip of S_a^1 passes through vertical.

The Peripheral Schieferhülle core is not exposed on the north side of

Seebachtal because the synform plunges more steeply than the topography.

Poles to S_a^1 surfaces in the Auernig subarea are spread out along a well-defined girdle in Fig. 3.25 a. The approximate pole to this girdle plunges towards the south-west, confirming that this is the axial direction of the Hölltor Kern and Ankogel Mulde.

(c) D_a^2 Structures

D_a^2 structures are more strongly developed here than in the Lösseck and Komate subareas. In the mica-schists immediately adjacent to the augen-gneisses of the Hölltor Kern, F_a^2 merely bends S_a^1 (Fig. 3.24) and D_a^1 quartz-rodding can often be traced round fold noses. In such cases, D_a^1 rods sometimes fall close to small circles round F_a^2 axes when plotted on a stereogram (e.g. Fig. 3.28). Elsewhere, it is more usual to find a penetrative strain-slip schistosity, S_a^2, developed with an equal intensity to S_a^1 but at an angle to it, producing an intersection lineation parallel to F_a^2 axes. S_a^2 is axial planar to small F_a^2 folds, where they occur. Rarely, S_a^2 is more strongly developed than S_a^1 and the D_a^1 quartz rodding is obliterated altogether. These relationships are illustrated in Fig. 3.29.

The dominant hornblende lineation in the epidote-amphibolites is parallel to F_a^2 axes, although a few examples have been found in which an older lineation has been folded. An axial epidote-rodding occurs in epidote-rich varieties. Calc-schists and marbles possess a strong S_a^2 muscovite schistosity. In Fig. 3.22, this cuts both limbs of the F_a^1 isocline.

D_a^2 lineations are plotted stereographically in Fig. 3.25 d. They fall almost exclusively in the south-west quadrant and cluster round the approxi-

mate centre of the π-pole girdle of Fig. 3.25 a, indicating that F_a^2 small folds are parasitic on the Hölltor Kern and Ankogel Mulde.

The 90° angular discordance between the F_a^2 axial directions of the Auernig and Böseck subareas could have one of two possible explanations:

(i) Deformation phases identified as D_a^2 in the two subareas might not be the same.

(ii) D_a^2 is the same in both subareas, but F_a^2 folding is grossly non-cylindrical on a large scale.

The second possibility is favoured for the following reasons:

(a) The axial surfaces of F_a^2 folds are sub-parallel in all the subareas studied so far (Figs. 3.11 c, 3.12 c and 3.25 c).

(b) D_a^2 lineations fan out in the Romate subarea, and dip towards the south along its eastern margin.

(c) There is no interference between folds on different axes around the Hallsler Alm (where the Böseck and Auernig areas meet) as would be expected if different fold generations were involved. Instead, there is a sharp discontinuity in axial direction. A smilar discontinuity has been mapped by Exner (1957) in the area around Badgastein, to the north of Map 2.

(d) D_a^3 Structures

D_a^3 structures in this subarea are most clearly developed in the epidote-amphibolites on the summit of Auernig itself. Similar features have been observed in the metabasites of the Sickerkopf subarea, and these also will be considered here.

The main effect of D_a^3 here is a gentle buckling of the S_a^1 banding. Generally, the D_a^2 hornblende lineation is oblique to F_a^3 axes and folded round them. A millimetre-scale microripple is locally developed parallel to F_a^3 axes. Occasionally, kinks have formed on axes oblique to both the F_a^3 microripple and the D_a^2 lineation, suggesting that these were not produced at the same time.

Most of the abundant chlorite + albite veins on Auernig are closely associated with D_a^3 structures. Many have developed sub-parallel to F_a^3 axial surfaces (Fig. 3.30) but others have been folded after emplacement. Both these situations require rapidly varying local stress conditions. In the first example, compression was followed by tension, but in the second the sequence is reversed. The spatial and temporal association of veins with D_a^3 structures is demonstrated by the amphibolite block shown in Fig. 3.32, where coarse-grained chlorite and albite have grown in the cavity produced by small-scale décollement in a disharmonic F_a^3 fold. The cavity-filling must have precipitated from a locally derived fluid. The same conclusion can be drawn from the D_a^3 boudinage described in Section B (f).

In all these veins and segregations, the lithostatic pressure, P_e, cannot have exceeded the pore fluid pressure, P_f, by more than the tensile strength of the rock, T_o (~200 bars), otherwise the cavities could not have remained open to allow euhedral crystals to grow. Local fluid pressures high enough

to satisfy the relationship $P_f > P_l + T_o$ may have caused veins to form through hydraulic fracturing (Hubbert and Willis, 1957; Secor, 1965; Norris and Henley, 1976) although this is not a likely explanation for the boudinage and décollement structure mentioned above.

D_a^3 microripples have also been recognised in other parts of the Auernig subarea. Amphibolites and mica-schists often possess a faint crenulation in S_a^1 oblique to the rodding (e.g. Fig. 3.31). Usually it is not possible to prove that this is F_a^3, except where S_a^2 is crenulated. There is very little consistency in the orientation of D_a^3 lineations over the subarea (Figs. 3.25 e and 3.25 f), presumably because this is largely dependent on the attitude of S_a^1, which is extremely variable.

D. KAPONIG AND SICKERKOPF SUBAREAS

In these two subareas, D_a^1 structures are scarcely recognisable because they are almost completely obliterated by intense D_a^2 deformation. In fact, Norris et al. (1971 b) concluded that the dominant foliation in this subarea was S_a^2 (author's notation, and believed it to be the earliest deformational fabric.

The main structural units outcrop as follows:

Peripheral Schieferhülle : Flanking the found-west margin of the Hochalm Kern.

Zentralgneis : Sonnblick gneiss-lamella (off Map 2 to the south-west. Other thin gneiss-lamella. Parts of the Hochalm Kern.

Inner Schieferhülle : South-west margin of Hochalm Kern.

(a) D_v Structures

In the Sickerkopf subarea, Inner Schieferhülle rocks possess a steep westwards plunging lineation not observed in the cover rocks. This is probably a pre-Alpine structure. In amphibolites, it is a hornblende lineation parallel to a mullion flattened by D_a^2, and, in schistose gneisses, a quartz and feldspar rodding. The large-scale lithological banding is probably an inherited Variscan structure as it appears to be truncated by the thrust-plane at the base of the Peripheral Schieferhülle.

(b) D_a^1 Structures

The epidote-amphibolites, calc-schists and calcareous quartzites of the Peripheral Schieferhülle all possess a distinct compositional banding folded by F_a^2. This is taken to be S_a^1 by analogy with the subareas already described. Only two independent pieces of evidence have been found to suggest that this foliation is a deformation structure:

(i) In certain pelites and talc-schists, S_a^1 is a flaky mineral schistosity folded by F_a^2. This is particularly obvious in thin section (e.g. Fig. 4.19).

(ii) The deformed pebbles in the Kaponig dolomite quartzite breccia have been flattened in the plane of the foliation prior to F_a^2.

No F_a^1 folds have been recognised in these two subareas. A D_a^1 lineation

was confirmed at only one locality, 233, where a tremolite elongation lineation is bent round a small F_a^2 fold (Fig. 3.33 b).

Norris et al. (1971 b) draw attention to the abundance of shear zones within the Peripheral Schieferhülle of this area, and refer to the whole sequence as a mélange. A certain amount of translational inhomogeneous strain (thrusting) must, therefore, have occurred at some time during its deformational history. Many of the shear zones separate areas of homogeneous D_a^2 deformation and appear to have undergone shearing during this event, but others are probably D_a^1 structures. A good example of the latter is described in Chapter 2, Section D(f); it contains slices of serpentinite and metagabbro in a talc-schist matrix. The talc schistosity has been crenulated by F_a^2, indicating that the tectonic introduction of ophiolite lenses into the succession occurred before or during D_a^1. The thrust zone separating the Peripheral Schieferhülle from the underlying basement is probably also of this age, as the schistosity in some of the weiss-schiefer associated with it is crenulated by F_a^2.

(c) D_a^2 Structures

D_a^2 structures are much more strongly developed along the south-western flank of the Hochalm Kern than in other subareas. A manifestation of this is the change in shape of the Sonnblick Kern as it passes south-eastwards from a broad asymmetric antiform into a thin parallel-sided sheet, the Sonnblick gneiss-lamella. This progression is matched by changes in the fabric of the constituent augen-gneisses. In the western part of Böseck subarea, S_a^1 foliation is dominant and D_a^2 only produces gentle crenulations. However, in the Gugganig area, where the Sonnblick Kern outcrop has thinned to about 400 m, an S_a^2 schistosity has developed at a high angle and with

equal intensity to S_a^1 in the middle of the outcrop, giving the gneisses a strong linear fabric. In the Mölltal, where the Sonnblick lamella is about 200 m thick, S_a^1 is unrecognisable and the steep S_a^2 schistosity is dominant. These relationships are illustrated in Fig. 3.35.

Tight to isoclinal F_a^2 folds are developed in all Peripheral Schieferhülle lithologies except quartzite where they tend to be more open. In mica-schists, epidote-amphibolites and calc-schists a penetrative S_a^2 schistosity is developed axial planar to these folds (Fig. 3.36). Generally, within fold limbs, the S_a^1 foliation has been rotated parallel to S_a^2 and the two are indistinguishable. For the purposes of data presentation, all foliations have been plotted as S_a^1 except where it is clear that they are axial planar to F_a^2. Even making this distinction, poles to S_a^1 plot in the same place as poles to S_a^2 in Figs. 3.33 a and 3.33 c.

In calcareous and pelitic schists, earlier calcite and quartz veins have been folded and intensely flattened parallel to S_a^2, giving rise to a strong rodding lineation. Hornblendes in metabasites are also aligned in this direction.

D_a^2 lineations have a variable orientation in the Kaponig subarea, but their mean direction of plunge, ca. 0/135, is the same as in the Böseck subarea (Fig. 3.33 d). The regional trend of D_a^2 lineations swings gradually as the outcrop of the Peripheral Schieferhülle is traced from the Möll tal north into the Ankogel Mulde, so that in the Sickerkopf subarea, most plunge gently towards the south or south-south-east (Fig. 3.34 d).

E. DÖSENER SUBAREA

This subarea lies entirely within the Hochalm Kern so only Inner Schieferhülle and Zentralgneis lithologies are represented.

Much less time was spent in this subarea than in the others, and the structures have not been examined in such detail. A more detailed account is given in Cliff (1968). Compared with other subareas, the effects of D_a^1 are almost negligible, so that abundant pre-Alpine structures are preserved within the Inner Schieferhülle, some of which have been truncated at contacts with intrusive Zentralgneis aplites. This contrasts strongly with the situation in, for example, the Ankogel Mulde where any pre-Alpine structures that may have existed have been obliterated by intense D_a^1 shearing and flattening.

All the structural elements from the eastern part of the subarea have been plotted together in Fig. 3.37.

(a) D_v Structures

The banding in many Inner Schieferhülle lithologies is obviously pre-Alpine as it is cut by Zentralgneis intrusive rocks. Tight to isoclinal folds are common in the finely banded amphibolites and banded gneisses, and at a few localities these, too, are truncated. In the amphibolites, some of the hornblende prisms are aligned parallel to F_v axes. In the Reisseck area Norris (1970) recognised at least two generations of pre-Alpine folds, both deformed by the later F_a^2 folds.

(b) D_a^1 Structures

The only structures possibly attributable to this event are certain localised shear zones within the Zentralgneis. A good example occurs in augen-gneiss on the eastern side of the Mallnitzer Scharte. The shear zone is about 2 m wide and consists of quartz-muscovite schist, presumably phyllonised augen-gneiss. The schistosity has been tightly crenulated by F_a^2 folds, indicating that shearing took place before D_a^2. Elsewhere in the Hochalm Kern, several generations of shear zone can be identified, some of which may be associated with D_a^2 (P. Butler, pers. comm.).

(c) D_a^2 Structures

In the large areas of granite-gneiss, a foliation defined by a preferred orientation of mica flakes and of flattened quartz and feldspar aggregates is usually present. This is often accompanied by a biotite-streak lineation, generally subhorizontal and parallel to axes of tight to open asymmetric folds in nearby banded-gneisses. Norris et al. (1971b) have shown that the foliation in the Zentralgneis is generally parallel to the axial surfaces of these folds and that the deformation phase generating them can be correlated with D_a^2 (author's notation) in the Peripheral Schieferhülle. This is supported by the observation that the lineations in the Zentralgneis of the Dösener subarea are parallel to the D_a^2 lineations in the adjacent Sickerkopf subarea.

F. CONCLUSIONS

(a) Conclusions from the thesis area

The main conclusions drawn from the structures described in this thesis are listed below. The structural history of the area is summarised in Table 3.2.

(i) Deformation has taken place in discrete episodes. Two major phases of Alpine deformation, D_a^1 and D_a^2, can be identified and correlated over the whole area examined.

(ii) The first of these, D_a^1, involved both translational inhomogeneous strain (thrusting) and homogeneous strain (penetrative shearing and flattening). Thrusting occurred early in this event. Slices of continental basement (Zentralgneis granite) and oceanic basement (ophiolite fragments: peridotite and gabbro) were tectonically emplaced into the Peripheral Schieferhülle supracrustal sequence and then deformed to produce the gneiss-lamellae and pods of serpentinite and talc-schist.

The homogeneous strain associated with D_a^1 was very large in Peripheral Schieferhülle rocks, but decreased markedly with structural depth in the basement. Structurally high basement units, such as the allochthonous gneiss-lamellae, were intensely flattened together with surrounding metasediments and metavolcanics.

The concentration of D_a^1 effects in the relatively incompetent cover sequence is consistent with production during large-scale overthrusting of the Altkristallin sheet. Oxburgh (1966), Brewer (1970), Wright (1973), Hawkesworth (1974), Waters (1976) and Bickle and Hawkesworth (1978) have shown that the effects of Alpine penetrative deformation in the Altkristallin are negligible. The D_a^1 deformation can therefore be envisaged as the shearing and flattening of a sandwich of cover rocks caught between two semi-rigid blocks of basement moving relative to one another.

Some F_a^1 fold axial directions are consistent with relative movements in a south-west/north-east direction, others with a more east/west direction.

(iii) As D_a^1 involved both thrusting and intense penetrative deformation, probably accompanied by large-scale isoclinal folding, it is virtually impossible to reconstruct the original stratigraphic relationships within the Peripheral Schieferhülle. This is largely because one cannot tell whether units are repeated by isoclinal folding, thrusting or both, or whether interbanding is a depositional feature. Norris et al. (1971b) go as far as to refer to the Möltal Schieferhülle as a mélange. The author hesitates to apply this term to the Mallnitzer Mulde succession because of the lateral continuity of rock bands, but it is possible that it represents a mélange in the conventional sense which has subsequently been flattened so that stratigraphic and stratal discontinuities are no longer obvious.

Frasl and Frank (1966) put forward a stratigraphic scheme for the Mesozoic Schieferhülle based on work in the Grossglockner area, but this has not been used in the present study because of the objections mentioned above. However, certain lithological associations can be recognised and matched to some extent with Frasl and Frank's classification. The most obvious are:

1. Calc-schist-serpentinite-metabasite-association (e.g. unit 5 in Section B(c)). Correspond to their Glockner facies. Probably pelagic. Incorporates ophiolite.

2. Calcareous quartzite-marble-quartz-phengite-schist-association (e.g. unit 4). Corresponds to their Wüstkogel or Seidlwinkl series. Probably epicontinental. Incorporates granitic basement.

The close proximity of these two associations in the Mallnitzer Mulde emphasises the problems of palinspastic reconstruction. On the south side of Geisel Kopf, only 200 metres separate slices derived from continental and oceanic basement.

(iv) The D_a^2 event involved basement and cover rocks, both to a large extent, and its effects are seen at low structural levels in the basement complex. The Sonnblick Kern, Mallnitzer Mulde, Hölltor Kern and Ankogel Mulde are large-scale asymmetric folds of this generation. The strain associated with D_a^2 was less than that of D_a^1 but more evenly distributed among the Pennine units in the Tauern Window.

(v) The orientation of large and small F_a^2 fold axes varies from north-west/south-east in the Böseck and Kaponig subareas, through north/south in parts of the Romate, Sickerkopf and Dösener subareas, to north-east/south-west in the Auernig subarea. This variation must be due to the non-cylindrical nature of F_a^2 and cannot be attributed merely to the rotation of axes due to later uplift of gneiss domes as suggested by Norris (1970) and Bickle and Hawkesworth (1978). The reason for this non-cylindricality is not clear, but it is conceivable that large-scale heterogeneities within the underlying basement were responsible. A screen of foliated Inner Schieferhülle rocks between two large plutons of Zentralgneis granite could have constituted a zone of weakness, and the trend of such a screen, or the orientation of the pre-Alpine fabric within it, might have governed the axial direction of Alpine folds. The mis-match of basement lithologies exposed on either side of the Ankogel Mulde (granite-gneisses to the west; amphibolites and banded gneisses to the east) is consistent with this hypothesis. Nevertheless, the sharp 90° axial discontinuity between the Böseck and Romate subareas is difficult to reconcile geometrically. It cannot be explained by postulating a large post-D_a^2 thrust fault separating homogeneously deformed masses, because the Auernig amphibolite (containing south-west plunging F_a^2 axes) passes laterally, without evidence of faulting, into amphibolite bands in the Möll tal Schieferhülle (where F_a^2 axes plunge north-west or south east).

The asymmetry of F_a^2 folds suggests that they, too, were produced in a phase of overthrusting, but this time deep-seated asymmetric tongue-like proto-nappes

(e.g. the Sonnblick Kern) were generated in the Penninic domain. Only the folds in the Bösseck and Kaponig subareas are consistent with overthrusting towards the north-east.

(vi) The intensity of D_a^2 deformation in the Peripheral Schieferhülle and the difficulty of recognising D_a^1 structures increase in a south-easterly direction in the part of the Tauern Window examined.

This could be partly a function of metamorphic grade. Peak temperatures at the basement-cover interface in the Mölltal were sufficient for staurolite + biotite growth in pelites (see Chapter 6) but further west, temperatures at this boundary were lower. Holland and Lambert (1968) suggested that the competence of rocks undergoing prograde metamorphism may be reduced as they heat up to amphibolite facies by the large amount of water evolved from the numerous dehydration reactions which occur around 550°C. This mechanism may have been important here.

(vii) A third, probably polyphase deformation event, D_a^3, produced localised boudinage, buckling and kinking. The low degree of strain and semi-brittle nature of these structures contrast strongly with the intense plastic deformation associated with D_a^1 and D_a^2, suggesting that they were generated at shallower levels, perhaps during the uplift of the axial zone. Fractures were opened during the formation of some of these structures and fluids, some derived locally, were admitted and allowed to precipitate vein minerals.

The pseudotachylite described in Chapter 2, part E(b) was probably formed at this stage. Sibson (1975) has calculated that frictional melting can only take place along slip-planes if the rock is devoid of intergranular fluid, the faulting is of seismic magnitude and the overburden is less than 10 km. Similar pseudotachylites have been recorded from the Engadine Window by Masch (1973) who also attributes them to late-stage shallow-level movements.

(b) Correlation of Deformation Phases in other parts of the Tauern Window

Structural sequences have been established in other parts of the Tauern Window by the following authors: Norris et al (1971b) in the Reisseck Group, south-east Tauern; Bickle (1973), in the Grossglockner area, central Tauern; Hawkesworth (1974) along the eastern margin of the Tauern; Bickle and Hawkesworth (1978); Holland (1977) in the Grossvenediger area, central Tauern. In Table 3.3, these are correlated with one another and with the sequence deduced in this study. There is a high degree of correlation between events recognised by different workers and the tectonic interpretations put on them. This suggests that deformation events were distinct even on a regional scale and that it is valid to use them as time markers with which to monitor mineral growth phases.

CHAPTER 4

Metamorphic History

CHAPTER 4

METAMORPHIC HISTORY

A. INTRODUCTION

The previous chapter set out pre-Alpine and Alpine deformation events in sequence to form the basis of a structural history. In this chapter, this structural history is used as a relative time scale with which to correlate phases of mineral growth. A unified picture of the tectonic and metamorphic evolution of this part of the south-east Tauern Window should emerge if structural and mineralogical histories are examined simultaneously. This approach has been employed by Ayrton and Ramsay (1974) in the Central Alps, Zwart (1963) in the Pyrenees, Johnson (1963) and Harte and Johnson (1969) in the Scottish Highlands and Holland (1977) in the Central Tauern.

Pressure and temperature conditions experienced at any stage in a rock's metamorphic history can only be calculated if the equilibrium mineral assemblage corresponding to that stage can be identified. In practice this is difficult, if not impossible, for all but the most recent assemblage because recrystallisation obscures early relationships. If rocks were not subject to kinetic barriers against nucleation during prograde as well as retrograde metamorphism, no equilibrium assemblages formed before the latest temperature peak would be preserved. Many of the rocks in the area covered by Map 1 have undergone much late recrystallisation, making the interpretation of earlier stages of metamorphism, particularly the early part of the Alpine cycle difficult. Overprinting has been less efficient further west where D_a^2 structures are not as strongly developed.

The time of growth of a particular mineral is established from its textural relationships with microscopic structures known to have been formed in a particular deformation. Generally, such relationships can only be seen clearly in thin section. Some of the more useful metamorphic textures are listed below and illustrated in Fig. 4.1 (partly adapted from Spry, 1963; 1969; Misch, 1969; and Zwart, 1962).

In metamorphic rocks, where minerals crystallise from preexisting solids, idiomorphic crystals are likely to post-date their neighbours, whereas the opposite is generally true in igneous rocks.

Matrix platy minerals (e.g. micas, chlorite, ilmenite and chloritoid) and prismatic minerals (e.g. amphiboles) can be identified as:

1. **Pre-tectonic:** (i) bent by microfolds,
   (ii) kinked,
   (iii) microboudinaged.
2. **Syn-tectonic:** (i) in parallel alignment, forming a schistosity,
   (ii) growing parallel to axial surfaces of microfolds.
3. **Post-tectonic:** (i) cutting across a schistosity,
   (ii) mimetically recrystallised round microfolds,
   (iii) decussate.

Porphyroblast minerals (e.g. garnet, staurolite, chloritoid, kyanite, amphiboles and feldspars) can be identified as:

4. **Pre-tectonic:** (i) enveloped by schistosity, with pressure shadows developed,
(ii) containing an inclusion fabric discordant to the matrix fabric.

5. **Syn-tectonic:** (i) containing a convergent concordant inclusion fabric (growth during development of a schistosity)*;
(ii) containing a concordant inclusion fabric straight in the core but folded in the rims (growth during development of a crenulation),
(iii) containing a sigmoidal concordant inclusion fabric (growth during rotation, e.g. due to shearing in limb regions of a larger fold).

6. **Post-tectonic:** (i) euhedral, cutting across a schistosity,
(ii) enclosing a straight concordant inclusion fabric (growth after development of a schistosity),
(iii) enclosing a folded concordant inclusion fabric (growth after development of a crenulation).

Individual optically zoned crystals of garnet, feldspar or amphibole can combine features of 4, 5 and 6. Replacement textures, such as pseudomorphs and reaction rims are useful for determining the direction of metamorphic

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*Misch (1971) interprets this texture as post-tectonic crystallisation, where the growing mineral pushes aside the enclosing matrix minerals by 'force of crystallisation' and then overgrows them. Taken on its own, it could be ambiguous.

reactions. Early-formed minerals may be preserved as inclusions within garnet porphyroblasts.

The reader is referred to Table 4.3 for a summary of the metamorphic history discussed in this chapter.

B. PRE-ALPINE METAMORPHIC HISTORY

Although the emphasis in this thesis is placed upon Alpine structure and metamorphism, pre-Alpine mineral assemblages are identified here so that the effects of Alpine metamorphism can be recognised in the Inner Schieferhülle. Only then will it be possible to study fully the regional variation in Alpine metamorphic conditions across the terrain.

(a) Inner Schieferhülle Metabasites

There is strong evidence in support of a pre-Alpine, probably Variscan, amphibolite facies metamorphic event in basic rocks in the Dösener subarea. Many rocks contain two generations of amphibole; the first as discrete porphyroblasts, the second as smaller prisms lying in the foliation and bent round the first. In GD 117, a 1 x 2 cm hornblende megacryst contains straight aligned trails of garnet and epidote inclusions at right angles to the matrix ?S_v hornblende schistosity, implying one, possibly two, generations of pre-Alpine hornblende growth. The occurrence of D_v foliation in rocks cut by aplites and in xenoliths enclosed by undeformed Zentralgneis meta-granite (Norris et al., 1971a, p. 139) is in itself evidence for at least one pre-Alpine metamorphism.

Primary eclogite assemblages in the eclogitic amphibolites probably formed before or early in the Variscan cycle because

(i) Both the eclogites and nearby garnet-amphibolites contain a pre-Alpine foliation cut by aplite veins. In the former this is a compositional banding, in the latter a banding and hornblende schistosity.

(ii) Primary eclogite minerals, fine-grained garnet and omphacite, have been partially or completely replaced by hornblende, coarse-grained garnets and symplectite, probably equivalent to the dominant assemblage in the garnet amphibolites. There is no proof of this, however, and it is possible that the replacement products are Alpine. The ambiguity arises because hornblende in the eclogites has crystallised at all angles to the banding. This banding, which is parallel to S_v in surrounding rocks may have been produced before S_v, and the random texture due to the competent behaviour of this lithology during D_v and D_a^2.

(iii) Definitely pre-Alpine garnets occur in certain other amphibolites, e.g. garnets with ?Alpine reaction, rims of plagioclase in GD 135, and garnets containing inclusion trails discordant to S_a^2 in GD 183c.

Details of these textures have been described in Chapter 2. The preferred scheme of correlation between mineral growth phases in the Dösener Tal eclogites and garnet-amphibolites is summarised in Table 4.1.

In the Eastern Alps, eclogites have been described from the Altkristallin (Waters, 1976) and from the Peripheral Schieferhülle of the central Tauern (Miller, 1974, 1977; Holland, 1977) but until now, none have been discovered in the Pennine basement complex. As shown above, the basement eclogites were

formed in a pre-Alpine, probably pre-Variscan metamorphism and are therefore not related to the early Alpine eclogites in the Peripheral Schieferhülle.

(b) Inner Schieferhülle Metapelites

Basement metapelites preserve relics of a Variscan amphibolite facies event, but contain no eclogite facies parageneses.

Most Inner Schieferhülle metapelites possess some sort of schistosity. In the Randgneis schists of the Mallnitz area this is a strongly developed S_a^1 (see Chapter 3, Section C) but within the Hochalm Kern a weak S_a^2 is usually present, sometimes cutting across an indistinct and largely recrystallised S_v muscovite and biotite fabric.

The large garnets in the pelites of the Hochalm Kern commonly enclose straight trails of quartz or biotite inclusions discordant to S_a^2. In some the trails traverse the entire grains (Fig. 4.2) indicating that the whole of each garnet is pre-Alpine, but in most, discordant trails are confined to the large, rounded core regions (Fig. 4.3). Norris et al. (1971a) described large garnets incorporated as xenocrysts into granite injected intc migmatitic schists, confirming that they grew before the intrusion of the Zentralgneis.

In Randgneis schists, large garnets generally have euhedral rims and rounded cores which rarely contain an inclusion fabric so that in most cases their pre-Alpine origin is not obvious. In GD 1192, the rims clearly cut across S_a^1 but S_a^1 appears to have been deflected round the cores before the rims grew. In GD 652, large 1 cm diameter garnets are unzoned, but transected by a dense network of interconnected fractures, a texture common in the sheared pre-Alpine rocks at the base of the Altkristallin (Waters, 1976). This rock

also contains both Variscan staurolite and kyanite. Both minerals have been extensively altered to chlorite and micas (see Chapter 2).

Pre-Alpine kyanite has only been confirmed in one other rock, C 523. The kyanite has been fractured and bent round a small F_a^2 fold and partially altered to oligoclase, chlorite and muscovite. Most of the phlogopitic biotite has been mimetically recrystallised round the fold, but some is enclosed within the kyanite and altering with it, suggesting that the original assemblage included both minerals.

Curious 1 x 0.2 x 0.2 cm aggregate of chlorite, biotite and quartz occur in the metapelites immediately overlying amphibolites 2 km west of Maresen Spitze (Map 2). The rocks are characterised by assemblage 2 (Chapter 2, Section B (g)) and contain Variscan garnets and a strong S_a^1 muscovite, biotite and chlorite schistosity. The aggregates lie at low angles to S_a^1, and quartz pressure shadows have developed round some. The chlorite and biotite forming them are spongy and have grown in random orientation, but the quartz within them occurs as flattened grains aligned in straight trails at high angles to S_a^1. These aggregates are interpreted as pseudomorphs after a prismatic mineral, such as staurolite, which crystallised at the same time as the garnets and overgrew an S_v fabric.

Textural evidence for the overprinting of the Variscan regional metamorphism by a higher temperature contact metamorphism associated with the intrusion of the Zentralgneis, is provided by three migmatitic schists, RN 245, RN 246 and RN 247, which show progressive degrees of granite injection. They were collected by R.J. Norris from the Obere Zwenberger See area north-east of Map 1, and are described in Norris et al. (1971a) who conclude that

(i) the mineralogical and textural differences between the three

samples are directly related to the introduction of granitic material.

(ii) The sample of 'normal' mica-schist (RN 245), with no injected granitic material, contains a predominantly Variscan assemblage; garnet, muscovite, biotite and plagioclase.

(iii) With increasing introduction of granitic material, quartz and plagioclase increase, K-feldspar appears, large muscovite flakes break down and new garnets are produced at the expense of biotite.

The author's interpretation differs from that of Norris et al. only on the following points:

(i) The fine-grained mesh identified as muscovite + chlorite replacing large muscovite flakes in RN 246 and RN 247 is muscovite + kyanite. Also, muscovite mesh occurs without kyanite, and lenticular aggregates of fine-grained sheaves of kyanite with subordinate biotite are found.

(ii) Elongated needles identified as zoisite in RN 247 (Fig. 4.4) are pseudomorphs of kyanite (identified by microprobe analysis) after acicular crystals with square cross-sections. The original crystals appear to have nucleated within the large Variscan muscovite grains, particularly where surrounded by granitic material.

(iii) Some of the muscovite and biotite grains in RN 247 have vermiform embayments (Fig.s 4.5 and 4.6) and appear to have been partially digested by the granitic melt.

Probably the acicular pseudomorphs were originally sillimanite needles which grew as the large muscovites were thermally decomposed by the surrounding

melt according to the reaction:

$$KAl_3Si_3O_{10}(OH)_2 + SiO_2 = Al_2SiO_5 + KAlSi_3O_8 + H_2O \qquad (4.A)$$

muscovite — quartz or melt — sillimanite — K-feldspar or melt — fluid or melt

The kyanite + biotite mats are likely to have been wisps of fibrolite + biotite. While biotite was dissolving in the melt it was, at the same time, breaking down to form garnet in the remaining solid rock. One possible reaction is:

$$KFe_3AlSi_3O_{10}(OH)_2 + KAl_3Si_3O_{10}(OH)_2 + 3SiO_2 = Fe_3Al_2Si_3O_{12} + 2KAlSi_3O_8 + 2H_2O \qquad (4.B)$$

biotite — muscovite — quartz — garnet — K-feldspar — fluid

This is supported by the occurrence of the following features in a specimen of migmatitic schist from the Rieken Graben (GD 404): (i) tiny garnets rim large muscovites as well as biotites, and (ii) where large numbers of garnets have grown round biotite, the latter has clearly recrystallised to a mosaic of paler more magnesian crystals (Fig. 4.7). Pseudomorphs of biotite + muscovite after pre-Alpine ?cordierite have been recorded by Frasl (1967) from granitic rocks of the central Tauern. The original cordierite could have crystallised at this time.

C. EARLY STAGES OF ALPINE METAMORPHISM

The effects of the early stages of Alpine metamorphism are recognisable only in the Peripheral Schieferhülle and Randgneis, particularly in the area covered by Map 2 (see Chapter 3, Sections B and C) where early Alpine structures persist.

(a) Syn-D_a^1 Mineral Growth

The main reason for assuming that the pre-S_a^1 minerals mentioned above crystallised during a pre-Alpine metamorphic event is that none have been found in unquestionable Peripheral Schieferhülle. All the Zentralgneis lithologies in the area covered by Map 2 contain some minerals as deformation relics within S_a^1 (e.g. feldspar augen, hornblende-pseudomorphs) but all cases appear to represent shearing of an original igneous texture (i.e. they predate the deposition of the cover rocks). The earliest Alpine metamorphic minerals recognisable in this part of the south-east Tauern are therefore those which constitute S_a^1 itself.

In most quartzites, graphitic mica-schists, granite gneisses and quartz-phengite schists the only syn-D_a^1 mineral to escape subsequent recrystallisation is potassic white mica, but this is only apparent in specimens containing micas bent by F_a^2 crenulations (see Fig. 4.1a, top left).

Pelitic mica-schists provide a more useful selection of syn-D_a^1 minerals for assessing metamorphic grade. These include muscovite, paragonite, margarite, biotite, chlorite, chloritoid, epidote and ilmenite, but few of them are ever found together. Judging from the abundance of quartz veins and stringers folded and flattened parallel to S_a^1, quartz was present in most assemblages during D_a^1. In iron-rich aluminous pelites of the Mallintzer Mulde, chloritoid, chlorite, muscovite, quartz and ilmenite are sometimes aligned parallel to the principal foliation. Some of the chloritoid is usually cross-cutting, and often these same minerals are also aligned parallel to an S_a^2 strain slip schistosity in the same rock. In such cases it is difficult to be sure that the minerals in S_a^1 were not mimetically recrystallised during D_a^2, but the presence of chloritoid inclusions in some pre-D_a^2 garnets argues in favour of earlier growth.

All species of mica except biotite are bent round microscopic F_a^2 folds (e.g. muscovite in Fig. 4.12 and paragonite in Fig. 4.19). Syn-S_a^1 margarite and paragonite generally occur as fine-grained, sometimes monomineralic, layers in which individual crystals are almost indistinguishable owing to their almost perfect parallelism.

Chlorite, epidote and actinolite define S_a^1 in the greenschists of the Mallnitzer Mulde. In most cases, rotation of later pre-D_a^2 albite porphyroblasts which overgrow actinolite and epidote (Fig. 4.8) betrays a slight modification of matrix S_a^1 schistosity by D_a^2 shearing. In some of the epidote- amphibolites of the Mölltal, where hornblende is the dominant amphibole and S_a^2 the main schistosity, actinolite is preserved as discordant inclusion trails of aligned needles within rotated albite porphyroblasts (Fig. 4.9) or as cores to hornblende crystals. Hornblende is very rarely found as inclusions, which suggests that it was not part of the syn-D_a^1 metamorphic assemblage stable in basic rocks. It is therefore surprising to find that hornblende has grown in S_a^1 in the epidote- amphibolites of the Auernig mass and further west. However, hornblende C-axes are nearly always aligned parallel to F_a^2 fold axes, suggesting that crystallisation occurred after D_a^1 and during D_a^2 and that the prisms were constrained to lie within the foliation presumably because the fold axes were contained within that plane. D_a^1 hornblende lineations have been found at a few localities, but these are rare.

(b) Post-D_a^1, pre-D_a^2 Mineral Growth

A variety of porphyroblast minerals grew in the interval between the two main Alpine deformation phases, the most widespread ones being albite and garnet.

An unusual feature of this generation of albite porphyroblasts is the wide spectrum of rock-types in which they are found: metabasites, semi-pelitic mica-schists, graphitic mica-schists, quartz-phengite-schists and coarse-grained chloritoid-schists. In all these lithologies, the albites form rounded subhedral grains sometimes with imperfectly developed crystal faces. Poikiloblastic grains tend to be less idiomorphic than inclusion free ones. They range in size from 0.2 mm in some semi-pelites and metabasites to 5 mm in the coarse-grained chloritoid-schist, and are normally twinned in the Carlsbad law or not twinned at all.

The albite porphyroblasts in the quartz-phengite-schists tend to be devoid of inclusions and merely cut across S_a^1. In other micaceous rocks they may contain aligned blebs of quartz, or straight trails of fine-grained opaque material, probably graphite or ore. In many, these trails are discordant to the matrix S_a^2 (Figs. 4.10 and 4.11), evidence of syn-D_a^2 rotation. The porphyroblasts must, therefore, have grown before D_a^2. The inclusion trails represent an earlier schistosity, presumably S_a^1, which was fully developed before albite growth. In the coarse-grained chloritoid-schist, the trails are discordant to S_a^1 but, as explained above, this is probably due to D_a^2 movements. The albites in metabasites display features identical to those in micaceous lithologies except that their syn-S_a^1 inclusion trails are of aligned actinolite and epidote prisms.

The simultaneous crystallisation of albite in lithologies of widely different bulk composition is probably not the result of a single discontinuous reaction because it is unlikely that they all contained the same reacting assemblage at one time. This mode of albite growth may be a form of recrystallisation in which the finely-divided remains of pre-existing albites, intensely sheared during D_a^1, have been re-assembled in new, unstrained porphyroblasts, implying that the deformation actively promoted a phase of mineral growth.

occurrence of certain albite-porphyro-
fied as the sheared equivalents of
ifferent rocks which have not grown large
albites. Equivalent pairs include:

Quartz-phengite-schist and Augen-gneiss (meta-granites)
Albite-porphyroblastic chlorite-schist and Amphibolite
(metabasite bands and lenses associated with
quartzofeldspathic gneiss).

In Chapter 2 a positive correlation between the degree of deformation and the amount of sodium depletion in quartz-phengite-schists which had not crystallised albite was shown suggesting that the sodium was dissolved in the fluid phase and removed. It is conceivable that the albite porphyroblasts represent the reprecipitation of this dissolved sodium. The ionic reactions which govern feldspar-mica-equilibria are discussed in Chapter 7.

Pre-S_a^2 Alpine garnet porphyblasts have only been confirmed in Peripheral Schieferhülle pelites. The garnet shown in Fig. 4.12 is one of the very few which have been found containing S_a^1 inclusion trails. The muscovite in this rock (GD 561) defines an S_a^1 schistosity which has been bent by F_a^2 into round-hinged microfolds. The garnet is euhedral and contains slightly sigmoidal but unfolded inclusion trails of ore mineral at right angles to S_a^1, indicating that it overgrew S_a^1 while being slightly rotated, but that growth was largely complete when the crenulations were developed.

In graphitic mica-schists, garnets are occasionally found with rims which clearly post-date D_a^2 (see Fig. 4.1 b, right (ii) and (iii) but whose included S_a^2 fabric or F_a^2 axial surfaces bend round inclusion-free early Alpine cores (Fig . 4.13).

Most early Alpine garnets are free of aligned inclusions and can only be recognised by the fact that the S_a^2 strain-slip schistosity is deflected round them, allowing quartz pressure-shadows to develop. In aluminous pelites, these garnets commonly contain random equant chloritoid inclusions but in semi-pelites clinozoisite and quartz inclusions are more abundant.

Variscan or early Alpine garnets rimmed or completely pseudomorphed by chlorite are common in the area covered by Map 2. Some of the garnets in GD 561 have been altered in this way. One pelite (GD 496) from the Peripheral Schieferhülle 1 km south-west of Mallnitz contains some unusual 2 mm diameter spherical aggregates consisting of radiating plates of chloritoid and chlorite (Fig. 4.14). The dominant schistosity, S_a^2 bends round them, and quartz pressure-shadows have developed. It is possible that these, too, are pseudomorphs after early Alpine garnet. Garnet and chloritoid porphyroblasts cut across the schistosity in the same rock, implying that if the aggregates are garnet-pseudomorphs then two discrete phases of garnet growth must have occurred. Also, for the late garnets to nucleate at new sites rather than rim the early ones, the early garnets must have had compositions distinctly out of equilibrium with the later mineral assemblage.

The atoll garnets in the garnetiferous-epidote-amphibolite GD 22b may have grown before D_a^2 as they appear to be unstable with respect to biotite and hornblende of the later syn-D_a^2 assemblage.

(c) Eo-Alpine High Pressure Assemblages

The earliest recognisable stages of Alpine metamorphism in the south-east Tauern are characterised by the following greenschist facies assemblages

act + ep + chl + ?ab (metabasites)

ctd + chl + musc + qz (pelites)

In the Peripheral Schieferhülle of the central Tauern, eclogite and glaucophase-schist facies assemblages were developed early in the Alpine metamorphic cycle (Miller, 1974, 1977; Holland, 1977). Both authors agree that the eclogites crystallised before the growth of glaucophane, and that these high-pressure parageneses were then overprinted by later assemblages of greenschist to epidote-amphibolite facies. Abraham et al. (1974) concluded that the Tauern eclogites and greenschists formed simultaneously, but now believe otherwise (Raith et al., 1977). Eo-Alpine high pressure assemblages have also been recorded from the Western Alps (Bearth, 1973, 1974; Desmons, 1977), and the sequence of overprinting (1. eclogite 2. blueschist 3. greenschist) appears to be identical. Holland established that only part of the Tauern Schieferhülle had experienced eclogite facies metamorphism, but that the whole cover sequence had been subjected to blueschist facies conditions because pseudomorphs after porphyroblastic lawsonite occur in greenschists at high structural levels near the boundary with the overlying Matrei zone (see also Fry, 1973). As lawsonite and glaucophane are thought to have grown syn- to post-D_a^1 (notation of Holland and the author: Fig. 3.40), it appears that rocks in the central Tauern were being subjected to glaucophane-schist facies while those in the south-east Tauern were crystallising greenschist facies assemblages. Unaltered lawsonites have recently been discovered in the Engadine Window (Oxburgh pers. comm.) but their relationship to minor structures is not known.

There is a piece of evidence, however, which suggests that the structurally lowest parts of the Peripheral Schieferhülle, experienced blueschist facies conditions early in their history. The pre-D_a^2 garnets in a pelitic mica-schist (GD 49, assemblage (19) in Chapter 2, Section D (g)) collected from

400 m south of the basement-cover interface in Zwenberger Tal (Map 1) contain polycrystalline inclusions with the following assemblages:

(1) czo + wm + qz

(2) marg + wm + qz

(3) czo + marg + qz + plag

Some have rhomboid cross-sections, but the largest, with assemblage (3), is distinctly rectangular (Fig. 4.15). Margarite and clinozoisite make up a large proportion of this inclusion but quartz and plagioclase are present only in minor amounts. Its regular outline contrasts with the random orientation of the minerals within it, suggesting that it is a pseudomorph after a euhedral crystal. The precursor must have been a Ca, Al rich silicate with a form capable of giving both rhombic and rectangular cross-sections; lawsonite is the most obvious candidate.

To test this hypothesis, grain boundaries were traced from an enlarged version of Fig. 4.15 and the areas on the tracing corresponding to each mineral species were cut out and weighed to give a crude estimate of the modal composition of the pseudomorph. A chemical composition was then calculated assuming the minerals to be pure CaAl end-members. The results are listed in Table 4.2 and the composition is plotted on a CaO-Al_2O_3-SiO_2 phase diagram in Fig. 4.16. The agreement between the compositions of the pseudomorph and lawsonite is good, suggesting that lawsonite was the precursor and that it has been replaced almost isochemically. The metamorphic history of the pseudomorph can be summarised as:

(i) Growth of euhedral lawsonite.

(ii) Garnet grows, enclosing lawsonite, thus isolating it and preventing it from reacting with other phases.

(iii) Isochemical breakdown of lawsonite by the reaction:

$$5CaAl_2Si_2O_7(OH)_2.H_2O = 2Ca_2Al_3Si_3O_{12}(OH) + CaAl_4Si_2O_{10}(OH)_2 + 2SiO_2 + 8H_2O \quad (4.C)$$

lawsonite — clinozoisite — margarite — quartz — vapour

(iv) Formation of plagioclase by the reaction:

$$2Ca_2Al_3Si_3O_{12}(OH) + CaAl_4Si_2O_{10}(OH)_2 + 2SiO_2 = 5CaAl_2Si_2O_8 + 2H_2O \quad (4.D)$$

clinozoisite — margarite — quartz — plagioclase — vapour

In the matrix of the rock, stage (iv) happened after D_a^2 because fine-grained margarite was folded by F_a^2 before being partially replaced by calcic andesine.

D. LATE STAGES OF ALPINE METAMORPHISM

(a) Syn-D_a^2 Mineral Growth

This phase of crystallisation mainly involved the platy minerals such as micas, chlorite, chloritoid and ilmenite which now form S_a^2, and the prismatic minerals, such as hornblende and epidote which form mineral lineations parallel to F_a^2 axes.

In micaceous rocks of the Peripheral Schieferhülle and Randgneis, D_a^2 affected the older schistosity in two ways, (i) by crenulating it, and (ii) by flattening it further without crenulating it, possibly rotating it into the position of the new schistosity. In general, (i) will have applied in the hinge zones of large folds and (ii) in the limbs. Growth of new platy minerals parallel

to S_a^2 is easily detectable in the first case, provided that flattening has not been too intense, but not in the second. In gently crenulated mica-schists, it is common to find either that localised syn-D_a^2 muscovite flakes have grown across S_a^1 parallel to F_a^2 axial surfaces (Fig. 4.12) or that S_a^1 has been folded without new micas growing (Fig. 4.21). However, an intense strain-slip schistosity is normally developed in tightly crenulated schists (Fig. 4.19) so that earlier platy minerals are only preserved in tight or isoclinal microfolds in the bands between closely spaced foliation surfaces. In mica-schists of the Mölltal, even these have largely been obliterated, with the result that most of the biotite, muscovite, chlorite and chloritoid are aligned parallel to a penetrative S_a^2 schistosity. Micas in virtually all calcareous rocks are aligned in this direction, with the exception of folded syn-D_a^1 margarite.

Muscovite, paragonite and, to a lesser extent, biotite crystallised at this stage to form S_a^2 in the Inner Schieferhülle pelites of the Hochalm Kern.

Although most of the syn-D_a^2 crystallisation involved sheet-silicates, some synkinematic garnets can be recognised in Peripheral Schieferhülle schists. Many of these have grown as anhedral 'S'-shaped grains closely associated with coarsely crystalline quartz, sometimes forming pressure shadows. In places the outer parts of garnets have grown along quartz grain boundaries forming sieve-textures. In GD 572 all the garnets have grown in the hinge zones and short limbs of small asymmetric F_a^2 folds. They are distinctly 'S'-shaped, and follow the folds round, sharing their sense of asymmetry. Part of the porphyroblast shown in Fig. 4.17 appears to have replaced some of the schistose matrix of the rock during rotation, while the rest has grown along the junction between two quartz stringers. Garnets with this vein-like habit have also been recorded by de Wit (1976) who attributes the texture to precipitation along thin tension gashes. Whether or not fractures

were actually developed, the garnets in this example appear to have nucleated at sites of low stress.

Synkinematic garnets have also been found in the fine-grained micaceous parts of certain schists. They are normally euhedral and contain either convergent or sigmoidal trails of opaque granules concordant to the external S_a^2 schistosity.

Some of the euhedral hornblendes in the garbenschiefer GD 480 appear to have been rolled during growth (Fig. 4.18).

(b) Post-D_a^2 Mineral Growth in Peripheral Schieferhülle Lithologies

Up to this stage, mineral growth in all the Mesozoic rocks has been consistent with an increase in metamorphic grade through time. Greenschist assemblages have been succeeded by epidote amphibolite in metabasites, lawsonite has broken down and garnet has grown in pelites. However, the effects of prograde post-D_a^2 neomineralisation are confined mainly to the area east of Mallnitz.

All the staurolite in the Peripheral Schieferhülle pelites grew after D_a^2. It forms euhedral porphyroblasts which cut across S_a^2 and any preserved F_a^2 crenulations (Figs. 4.19 and 4.20). Kyanite textures are similar but not so clear. Kyanite has never been found included within garnet, and the only staurolite inclusion found occurs very near the rim of its host. The growth of staurolite and kyanite in pelites marks the Alpine metamorphic temperature peak since no minerals indicative of higher grades occur.

Post-D_a^2 garnets are common and assume a variety of shapes. In some rocks

euhedral crystals merely cut across D_a^2 microstructures (Fig. 4.19) but more usually they overgrow straight trails of aligned quartz blebs, chloritoid plates or fine-grained opaque material parallel to S_a^2 in the matrix (Fig. 4.20 4.20). In GD 492, fine-grained muscovite and ore aligned in S_a^1 have been folded into millimetre-scale F_a^2 chevron folds. Euhedral garnets grew after F_a^2 in such a way as to replace mica but not ore so that the trace of S_a^1 is preserved within them as ore inclusion trails. The fold outlines pass continuously from the matrix into the garnets without changing shape or orientation (Fig. 4.21).

Flat garnets are common in schists containing abundant quartz veinlets parallel to S_a^2. The garnets appear to have grown along the micaceous partings between the veinlets, contacting the quartz on either side, suggesting that the veinlets have acted as barriers against growth perpendicular to S_a^2.

In some rocks, post-D_a^2 garnets are the first generation to have crystallised, but in others they rim earlier grains.

Other post-D_a^2 minerals in pelites include biotite, chloritoid and magnetite. Areas of recrystallised muscovite, biotite or chlorite, forming random interlocking flakes, occur in most mica-schists (Fig. 4.22), even those west of Mallnitz.

Large, ragged porphyroblasts of plagioclase (An_{20-30}) crystallised after D_a^2 at the expense of muscovite in a large proportion of calcareous and pelitic schists. In the hornblende-garbenschiefer and the more calcic pelites, both muscovite and margarite have been partially replaced by plagioclase with up to 60 mole-percent anorthite. In pelites, plagioclase occasionally encloses post-D_a^2 biotite or staurolite. Oligoclase commonly forms ragged rims round rotated albites (Fig. 4.10), but not all early albites are rimmed.

Albites in metabasites are also commonly rimmed by oligoclase, but unzoned polygonal albite and oligoclase have also grown between amphiboles aligned in S_a^2. The random hornblende poikiloblasts in GD 193 postdate this schistosity but hornblendes in other epidote-amphibolites show very little evidence of post-D_a^2 recrystallisation.

Most siliceous dolomites have granulose textures, implying that the diopside, tremolite, quartz, carbonate and albite crystallised or were extensively recrystallised close to the peak of metamorphism. Epidotes probably grew at an earlier stage as they are enclosed by both albite and diopside. Tremolites bent during F_a^2 have reformed into trains of contiguous but optically distinct crystals.

(c) Post-D_a^2 Mineral Growth in Basement Lithologies

In the granite-gneisses of the Hochalm Kern and part of the Sonnblick Kern, a certain amount of late recrystallisation has occurred, giving rise to random growth of unstrained micas and aggregates of polygonal quartz and feldspar. Replacement of K-feldspar by myrmekitic intergrowths of plagioclase + quartz probably occurred at this stage. Much of the phengite in the gneiss-lamellae has recrystallised, but quartz, particularly in the larger grains, still retains a strained appearance.

The pyroxene, biotite, epidote, quartz and albite of the syenite-gneiss crystallised after D_a^1, and possibly after D_a^2, because compositional banding and pseudomorphs flattened in S_a^1 are composed of unstrained polygonal crystals.

The large albite porphyroblasts in the eclogitic amphibolites probably grew

during the Alpine metamorphism and some of the hornblende may also have recrystallised then. The feldspars replacing Variscan garnets in GD 135 (Table 4.1) and the recrystallised parts of the plagioclase megacrysts in the meta-gabbro and meta-basalt amphibolites are among the few that are definitely Alpine. By analogy, the clear, unzoned, unstrained plagioclases in other amphibolites are taken to be Alpine. The only Alpine garnets recognised in Inner Schieferhülle meta-basites occur in the garnetiferous leucoamphibolites where the 'pin-prick' garnets (garnet 3 in Table 4.1) form rims round larger ?Variscan grains. The hornblende prisms which enclose both generations (Fig. 2.9) must also be Alpine.

The muscovite, biotite and, locally paragonite in basement pelites have totally recrystallised to a fine-grained decussate network. In samples where this recrystallisation is incompletely developed, random mica flakes cut across those forming S_a^2. The abundant 'pin-prick' garnets must be Alpine as they cut across S_a^2 where it is preserved, and enclose aligned ilmenites (Fig. 4.23). Euhedral staurolite and kyanite prisms overgrow the garnets where they come in contact, and also cut across S_a^2 (Figs. 2.18, 2.19 and 4.23) The occurrence of twinned staurolite/kyanite crystals (Fig. 2.17) indicates that these minerals crystallised simultaneously. Kyanite presumably pseudomorphed the sillimanite in the migmatitic schists at this stage. Thus garnet, staurolite and kyanite grew in these rocks at the same time as they crystallised in Peripheral Schieferhülle pelites. A further similarity exists in that here also large oligoclase porphyroblasts have grown at the expense of white mica and enclose the peak-temperature AFM minerals.

(d) Retrogressive Alteration

Most rocks studied have suffered very little retrogressive alteration. The basement pelites are an exception. Large sprays of late stage chlorite have grown from biotite, muscovite and, rarely, staurolite and normally enclose garnet, staurolite and kyanite. Their relationship to the plagioclase porphyroblasts is not clear; the two minerals may have grown simultaneously.

Biotite in some granite-gneisses and mica-schists has altered locally to chlorite, but wholesale chloritisation of biotite is confined to rocks immediately adjacent to late veins and fractures. Apart from the veins themselves, no other syn-D_a^3 minerals have been found.

One kinked mica-schist was examined in thin section. Micas and chlorite are sharply bent or cracked along D_a^3 kink planes, but no new flakes have grown.

E. CONCLUSIONS

A summary of the metamorphic history of the area, as exemplified by the phases of mineral growth in pelitic and basic rocks, is presented in Table 4.3.

The main conclusions of this chapter are as follows:

(1) Three distinct phases of pre-Alpine metamorphism can be recognised in the Inner Schieferhülle. Arranged in sequence, these are:

(i) an eclogite facies metamorphism (M_e) only recognisable in metabasites in Dösenertal;

(ii) a widespread Variscan amphibolite facies metamorphism (M_v);

(iii) a contact metamorphism (M_z) associated with the intrusion of Zentralgneis granites, only recognisable in localised migmatitic mica-schists.

A single phase of Alpine metamorphism (M_a) of greenschist to amphibolite facies, overprints much of the earlier assemblages in the Inner Schieferhülle and produces metamorphic assemblages in the Zentralgneis intrusives and the Peripheral Schieferhülle cover sequence.

(2) No equivalents of the central Tauern eclogites occur in this area.

(3) Much of the cover sequence in the south-east Tauern crystallised greenschist facies assemblages early in the Alpine cycle while cover rocks in the central Tauern were being subjected to blueschist facies. Lawsonite may have grown locally near the structural base of the Peripheral Schieferhülle along the Mölltal.

(4) East of Mallnitz, the temperature peak of the Alpine metamorphism, as represented by the growth of staurolite and kyanite in pelites, occurred after D_a^2 but before D_a^3. Most of the staurolite and kyanite in the Inner Schieferhülle is Alpine. West of Mallnitz, post-D_a^2 neomineralisation is less widespread. This is probably because the M_a peak metamorphic grade was lower further west (see Chapter 6) and fewer discontinuous mineral-producing reactions occurred. The occurrence of post-D_a^2 oligoclase rims round albite cores argues against the likelihood of an earlier M_a temperature peak further west.

(5) The growth of post-D_a^1 porphyrobalsts of albite in a wide range of lithologies is more likely to be the result of recrystallisation than of neomineralisation. This phase of recrystallisation may have been induced by the preceding deformation itself, possibly because the deformation reduced the grain-size and produced numerous crystal defects in preexisting albites.

CHAPTER 5

Chemical Mineralogy

CHAPTER 5

CHEMICAL MINERALOGY

In order to use various equilibrium univariant reactions effectively as geothermometers and geobarometers (see Chapter 7) it is necessary to know the chemical compositions of the participating phases. Similarly, a knowledge of the mineral chemistry, particularly Mg/Mg+Fe ratio, is essential for understanding the chemographic relationships and isograd reactions between AFM phases in pelitic rocks (see Chapter 6). With these objectives in mind, 28 rocks containing useful mineral assemblages were selected for mineral analysis from the 1,200+ samples collected.

Major-element mineral compositions were studied using a Cambridge Instruments Microscan-9 electron-microprobe. About 1000 analyses were performed altogether, of which roughly two-thirds are listed in Appendix I , together with short notes on analysed rocks. The aim of this chapter is to present these analyses graphically, and to compare, for each mineral, the compositions of different textural types in similar lithologies and of the same generations in different lithologies.

Garnets

Garnet compositions are plotted in terms of mole percent pyrope, almandine + spessartine and grossular + andradite + Ti-andradite ($Ca_3Fe^{III}_2Ti_{1.5}Si_{1.5}O_{12}$) in Figs. 5.1, 5.2 and 5.3.

Pre-Alpine garnets of the eclogitic amphibolites have a restricted range of

compositions (Fig. 5.1), and there is little chemical difference between the tiny M_e garnets, the dark orange inner rims of M_v garnet coronas and large M_v garnet rims. The pyrope content of the M_e garnets (maximum 12.6%) is low in comparison with values from eclogites from other metamorphic terrains, e.g. 28% in the amphibolite zone of the Sanbagawa belt, Japan (Banno, 1964) and 15 to 40% in the Peripheral Schieferhülle (Miller, 1974, 1977; Holland, 1977). Possible reasons for this are discussed below.

The M_a garnets in the Inner Schieferhülle garnetiferous feldspathic paragneiss (Fig. 5.2) have the highest Ca contents of any analysed, with up to 36% grossular + andradite. Although the garnets in this lithology are small (diameter < 1 mm) they are distinctly zoned (Fig. 5.4). The analyses plotted in this profile are not listed in Appendix II as they all have unacceptably high SiO_2 values and totals, but it is assumed that the cation partial analyses are correct, at least relative to one another. Ca decreases and Mg increases sharply near the rim. Mn decreases gradually from core to rim to less than half the core value in a smooth bell-shaped profile similar to those predicted theoretically by Hollister (1966) and Atherton (1968) assuming that Mn is partitioned into garnet and that there is no intracrystalline diffusion.

The large M_v garnets in the Inner Schieferhülle aluminous pelites are also zoned (Fig. 5.5b), the main variations being an increase in Fe^{2+}/Mg ratio from core to rim. The cores are among the most magnesian garnets analysed, with up to 20% pyrope. Norris et al. (1971a) noted an increase in Mn towards the rim in four garnets from Inner Schieferhülle migmatites, but Mn remains virtually constant in the non-migmatitic pelite analysed here.

'Pin-prick' M_a garnets have variable compositions (Fig. 5.2) which overlap with those of M_v garnets in the same rock, with the result that there is no

convincing chemical evidence to show whether or not the thin rims of M_v garnets are equivalent to the M_a 'pin-prick' grains. The small size of the latter and their great abundance in these pelites suggests that material has only diffused over short distances to nucleation sites, which in turn implies that the composition of these grains may be sensitive to small-scale variations in bulk composition (e.g. the presence or absence of an adjacent flake of biotite). The core and rim compositions of one such grain are presented alongside the M_v garnet profile in Fig. 5.5 for comparison.

The large M_a garnets from Peripheral Schieferhülle pelitic schists show a fairly narrow range in composition as plotted in Fig. 5.3. The clustering in this diagram is due to the fact that few rocks were analysed, and garnets within each rock have similar compositions. The most striking differences between garnets in different rocks are seen in the $Mn/Mn+Fe^{2+}$ ratios, which presumably reflect variations in bulk rock Mn content. For instance, the ratio ranges from 0.292 to 0.377 in the post-S_a^2 flat garnet in GD 308c but from 0.040 to 0.150 in the post-S_a^2 garnet in GD 168b.

Peripheral Schieferhülle garnets show a variety of zoning profiles. Complete rim-to-rim traverses were made across the two garnets mentioned above, as well as a core-to-rim traverse across a garnet in GD 18 which has a post-S_a^2 rim and a core which may be pre-S_a^2 as it contains rotated trails of ore and chloritoid inclusions. In the latter, the profile for each element is kinked (Fig. 5.5a) with a low concentration gradient in the core and a high gradient, with the reverse sense in the case of Mg and Ca, in the rim. Mn decreases from core to rim as predicted by the fractionation model. The large post-S_a^2 garnet in GD 168b is zoned smoothly perpendicular to S_a^2 (Fig. 5.6) showing the same trends as the rim in GD 18. Traverses were made both parallel and perpendicular to S_a^2 across the flat garnet in GD 308c (Fig. 5.7). The element profiles of the latter are irregular, asymmetric and difficult to

interpret. The profiles parallel to S_a^2 are smoother and show some unusual trends. The Mn concentration is high and more or less uniform except approaching the rims where it increases sharply. Not only is this the opposite of the normal situation, but the effect is not observed in the profile perpendicular to S_a^2, implying that growth in that direction had finished by the time the Mn concentration began to increase. This supports the textural interpretation of flat garnet growth proposed in Chapter 4, part D(b). A possible mechanism for the increase in Mn is suggested in Chapter 6. A concomitant decrease in Fe^{2+} is also observed.

Post-S_a^1 garnets in the coarse-grained chloritoid-schist are very rich in Fe (87% almandine) and have low Mn and Mg contents (Fig. 5.3). The post-S_a^2 garnets in the hornblende-garbenschiefer are richer in Ca than those of normal pelites.

Clinopyroxenes

Clinopyroxene analyses are recast using a scheme similar to that of Cawthorn and Collerson (1974). Titanium pyroxenes are only calculated if there is some Al^{IV}. Titanium is allotted to $NaTiAlSiO_6$ unless there is insufficient sodium, in which case the remainder is cast as $CaTiAl_2O_6$. Calcium Tschermak's pyroxene is calculated next using the remaining Al^{IV} and an equal amount of Al^{VI}. If there is insufficient Al^{VI}, the excess Al^{IV} is allotted to ferrian Tschermak's pyroxene $CaFeAlSiO_6$. The remaining Al^{VI} is then cast as jadeite, and the excess sodium is combined with Fe^{3+} to form acmite. Next, the manganese is allotted to johannsenite, $CaMnSi_2O_6$, and the remaining magnesium, Fe^{2+} and calcium are used to form enstatite, ferrosilite and wollastonite.

The compositions of all analysed clinopyroxenes are plotted in Fig. 5.8.

All M_e clinopyroxenes from eclogitic amphibolites plot within the omphacite field (classification after Essene and Fyfe, 1967) although some have nearly 20% acmite component and fall almost in the chloromelanite field. The acmitic pyroxenes have the lowest jadeite contents (~27%) while those with low acmite contents (~4%) contain most jadeite (up to 45%). The M_e pyroxenes are unusually rich in iron, and Fe^{2+}/Mg ratios range from 0.25 to 0.66 compared with the range 0.035 to 0.45 plotted by Miyashiro (1973) as representative of pyroxenes from eclogites in amphibolite and granulite facies terrains.

The fact that both M_e garnets and pyroxenes have high Fe^{2+}/Mg ratios indicates that the eclogitic amphibolites themselves are richer in iron than most metabasites. In this respect they resemble the superferric eclogites of the Voltri group, Liguria (Mottana and Bocchio, 1975; Ernst, 1976). The abundance of garnet in other Inner Schieferhülle metabasites could be explained if they too are iron-rich compared with those of the cover which are virtually free of garnet and have more normal bulk compositions (Abraham et al. 1974; Miller, 1974).

The tiny grains of $?M_v$ symplectitic pyroxene in GD 122 plot in the augite field in Fig. 5.8. They have similar acmite contents but much less jadeite (11% and 16%) than the M_e pyroxenes.

M_a clinopyroxenes from the syenite-gneiss and the biotite-schist inclusion also plot in the augite field. They contain between 0 and 9% jadeite, the lowest values being found in the inclusion. The average pyroxene composition from the syenite-gneiss falls close to $di_{54}hed_{33}jd_6acm_{5.5}jo_{1.5}$ (Fe^{2+}/Mg ratio = 0.61) while compositions from the inclusion cluster round $di_{54}hed_{28}jd_{0.5}acm_{14}CaTs_2jo_{1.5}$ (Fe^{2+}/Mg ratio = 0.52). Pyroxenes in neither rock are zoned.

The M_a diopside in the siliceous dolomite GD 160 has the composition di_{63} $hed_{35}acm$ etc.$_2$.

Amphiboles

All analysed amphiboles belong to the tremolite-actinolite-hornblende series. The number of Ca atoms per 23 oxygens ranges from 1.34 to 1.92, but falls below 1.50 in only three analyses, so that virtually all can be classed as calcic amphiboles (Ernst, 1968). Atomic sites are labelled in accordance with the idealised formula $A_{0-1}X_2Y_5Z_8(OH)_2$.

The lowest calcium contents are found in the hornblendes in eclogitic amphibolites. The ?M_v subcalcic pargasites in GD 136a all contain over 1.0 Na atoms per formula unit (Fig. 5.9), and A-site occupancies are very high (0.79 to 0.90). The maximum sodium content of the X-sites, calculated assuming that all potassium is in the A-site, range from 0.25 to 0.52. These high values may be indicative of high metamorphic pressures (Brown, 1977), consistent with their occurrence in eclogitic rocks. Fe^{2+}/Mg ratios are close to unity.

The M_a hornblendes in the garnetiferous feldspathic paragneiss have high Fe^{2+}/Mg ratios (1.53 to 2.11) and A-site occupancies of about 0.6, but virtually no alkalis in X-sites. They are classified as hastingsites.

The compositions of M_a amphiboles in Peripheral Schieferhülle metabasites have a distinctly bimodal distribution (Figs. 5.10 and 5.11). In GD 551 there is a distinct compositional gap between the actinolites and the hornblende rims. The former generally contain little sodium, no potassium and empty A-sites. Total aluminium ranges from 0.15 to 0.68, and silicon from 7.66 to 7.98. In the hornblendes, however, the A-sites are about one third

full, and silicon and aluminium values cluster about 6.7 and 2.2 respectively. The significance of this gap is discussed in Chapter 6. The post-S_a^2 hornblendes in GD 193 are not zoned.

The tremolites replacing the diopside mentioned above have compositions very similar to the metabasite actinolites.

The post-S_a^2 hornblendes in the garbenschiefer GD 480 are extremely aluminous (total Al ranged from 3.6 to 3.8), and are similar to the gotthardites of the Lepontine Alps (Wenk et al., 1974). They plot close to tschermakite in Fig. 5.10. Raase (1974) concluded that the aluminium content of hornblende increases with metamorphic pressure, but it is unlikely that abnormally high pressures are involved here because the host rock was found less than 6 km from both GD 193 and GD 551 mentioned above, neither of which contain particularly aluminous hornblendes. It is more likely that the high Al-contents in this case reflect an aluminous bulk rock composition. These hornblendes contain almost as much Al^{VI} (1.62 to 1.71) as the most aluminous tschermakites listed in Deer et al. (1963) well in excess of the empirical maximum of 1.40 for hornblendes containing $Si_6Al_2^{IV}$ per half unit cell (Leake, 1971; Kemp and Leake, 1975). Fe^{2+}/Mg ratios are unusually high (3.1 to 3.5), and the A-sites are about one third full.

Epidotes

All analysed epidotes are Alpine with the possible exception of those in GD 122 which may be Variscan. They show a wide range in Fe^{3+} content, and the mole fraction of Al_2Fe-epidote ranges from 0.24 to 0.85 (Fig. 5.12). There seems to be little systematic variation of iron content with lithology. Epidotes from metabasites span nearly the whole range, as do those from meta-

sediments. However, the most iron-rich analysed epidotes occur in the syenite-gneiss.

There is no indication of a miscibility gap in the monoclinic epidote series between 55 and 63% Al_2Fe-epidote as postulated by Raith (1976) for epidote-amphibolites in the central Tauern.

Feldspars

Plagioclase is an important constituent of nearly all rocks within the area studied, but K-feldspar has only been identified in Zentralgneiss lithologies and orthogneisses of the Inner Schieferhülle.

A histogram of all plagioclase compositions determined by microprobe is given in Fig. 5.13. The distribution is distinctly bimodal, with 87% of all analyses falling in the ranges an_{0-6} and an_{18-32}. Optically determined compositions have not been included, but they conform to the same pattern. The near-empty compositional gap between an_6 and an_{18} coincides with similar gaps found in other terrains, commonly ascribed to the peristerite solvus (e.g. Cooper, 1972).

Most plagioclase compositions appear to have been set during M_a, even when they retain the morphology of igneous primocrysts in Zentralgneis rocks (see Chapter 2). Possible exceptions to this are the large inclusion-free plagioclases (An 11) in the leucogranite GD 468, which plot in the middle of the peristerite gap, but it is more likely that they too have recrystallised because M_a myrmekitic plagioclases replacing microcline in the same rock have the same composition.

All analysed pre-S_a^2 M_a porphyroblasts are albites. Albite is also the characteristic plagioclase of the syenite-gneiss and siliceous dolomite, both of which contain clinopyroxene and calcite. The most calcic plagioclases analysed are the post-S_a^2 porphyroblasts replacing margarite in the hornblende-garbenschiefer. Other lithologies contain a variety of post-S_a^2 plagioclase compositions but these are dependent to some extent on metamorphic grade and bulk composition, and are discussed further in Chapter 6.

Most plagioclases contain less than 0.7% orthoclase, but those in GD 468 contain as much as 1.6%.

K-feldspar compositions in the leucogranite and pyroxene-biotite-schist range from $or_{88}ab_{12}$ to $or_{91}ab_9$. In the former, there is no chemical difference between large relict igneous microclines and small clear recrystallised M_a grains.

White micas

Three white micas, muscovite, paragonite and margarite, are found in the rocks of the south-east Tauern. All analyses are presented together in an Na-K-Ca plot (Fig. 5.14).

Margarites were probed in only one rock, the hornblende-garbenschiefer GD 480, in which they form a fine-grained matrix with an S_a^2 schistosity. They show a small degree (0 to 7%) of muscovite substitution, but contain more sodium than coexisting muscovites. This agrees with the results of Höck (1974).

Paragonite has only been identified in aluminous pelites in the area studied. Individual paragonites contain up to 17% muscovite and 6% margarite components.

Paragonites coexisting with margarite in amphibolites of the Greiner Schiefer Series, western Tauern (Ackermand and Morteani, 1973) have similar potassium contents but over twice as much calcium.

Single crystals of anomalously potassic paragonite and sodic muscovite occur in many paragonite-bearing pelites. These are interpreted as mixed layer micas and are not listed in Appendix I .

Muscovites show a wide range in degree of paragonite substitution (Fig. 5.14). The highest values (ca. 30%) are found in Inner Schieferhülle paragonite-bearing pelites, and the lowest (2 to 5%) in the leucogranite GD 468. Most muscovites contain less than 0.5% margarite component, but some in GD 468 contain over 2%.

Muscovite compositions have been plotted on an S.A.F. diagram (Lambert, 1959) to show the degree of $Si(Mg, Fe) = Al^{IV}Al^{VI}$ substitution (Fig. 5.15). The highest celadonite contents (ca. 40%) are found in GD 468 and an epidote-rich pelite, GD 43a, but the strongly pleochroic phengites in the quartz-phengite-schists can be expected to have even higher values. Muscovites from aluminous pelites contain as little as 10% celadonite. Fe/Mg ratios range from 0.4 to 4.3.

The scatter of points off the muscovite-celadonite join in Fig. 5.15 may be partly due to Fe^{3+} = Al substitution but the extent of this is not known as no wet chemical Fe_2O_3 determinations were made.

Distinct compositional differences between different generations of muscovite were only observed in the coarse-grained chloritoid-schist, GD 821d, where the syn-S_a^1 flakes are less phengitic and contain about 1½ times as much sodium as those cutting across S_a^1.

Many analysed white micas are deficient in alkalies, with Na+K+Ca totalling between 1.7 and 1.9 atoms per 22 oxygens. The significance of low alkali content of white micas has been discussed by Butler (1967), Cipriani et al. (1971) and Waters (1976). This phenomenon is generally attributed to H_3O^+ ions substituting for K^+ and Na^+ in the interlayer sites, but the polish on the probe-slide may also be an important factor. A large proportion of the present analyses for which Na+Ca+K total 1.95 to 2.0 were measured on slides in which the mica surfaces have an unusually smooth polish (e.g. GD 465b).

Biotites

All analysed biotites contain significantly larger amounts of aluminium than phlogopite and annite. The substitution reaction

$$Al^{VI}Al^{IV} = (Mg,Fe)^{VI}Si^{IV},$$

resulting in the displacement of micas on the phlogopite-annite join towards end-members of the eastonite-siderophyllite series, is probably responsible (Rutherford, 1973). However, the analyses all fall well to the A'S' side of the phlogopite+annite - eastonite+siderophyllite join when plotted on an S.A.F. diagram (Fig. 5.16), an expression of the fact that the Si/Mg+Fe+Mn ratio is always greater than unity. Substitution of Fe^{3+} for Al cannot be the cause as this would tend to lower the ratio.

All analysed biotites contain significant titanium concentrations, generally between 0.1 and 0.3 atoms per 22 oxygens. However, if Ti^{4+} only follows the substitution

$$Mg^{VI} + 2Si^{IV} = Ti^{VI} + 2Al^{IV}$$

as suggested by Robert (1976b), the Si/Mg+Fe+Mn ratio will be further reduced. The only substitution reaction which explains the spread in Fig. 5.16 is

$$2Al^{IV} = 3(Mg,Fe)^{IV}$$

i.e. towards dioctahedral micas of the muscovite-celadonite series. This effect has also been noted by Robert (1976a) in natural and synthetic phlogopites.

Biotites from the syenite-gneiss plot closest to the ideal phlogopite-annite composition. The highest degrees of dioctahedral mica substitution occur in biotites of the leucogranite and low-alumina Inner Schieferhülle pelites, while the highest eastonite+siderophyllite contents are found in more aluminous pelites.

Fe/Mg ratios vary with host rock composition. The highest values (4.25) are found in the leucogranite, and the lowest in the kyanite-schist, C 523. Biotites in most pelites have ratios close to 1.0.

Chlorites

Most analysed chlorites are magnesian ripidolites according to the classification of Hey (1954) but the chlorites in C 523 are sheridanites (Fig. 5.17). Syn-S_a^1 and -S_a^2 chlorites from Peripheral Schieferhülle metabasites have higher Si/Al ratios than syn-S_a^2 and post-S_a^2 chlorites from pelites. The latter resemble the majority of pelitic chlorites in that they plot close to the composition $(MgFe)_9Al_6Si_5O_{20}(OH)_{16}$ (Velde and Rumble, 1977).

Chloritoids

Mg/Mg+Fe ratios of chloritoids in pelitic schists (sensu stricto) range from 0.13 to 0.25, but much lower values (ca. 0.06) are found in the coarse grained chloritoid-schist GD 821d. Appreciable manganese concentrations occur in some chloritoids (e.g. over 5% Mn end-member in GD 308c), but manganese is barely detectable in those of GD 821d.

No significant differences in composition were detected between the cores and rims of large pre-S_a^2 porphyroblasts in GD 168.

Staurolites

Staurolite analyses have been recalculated to both 46 and 47 oxygens in order to compare them with the idealised compositions $(Fe,Mg)_4Al_{18}Si_8O_{44}(OH)_2$ and $(Fe,Mg)_4Al_{18}Si_{7.5}O_{44}(OH)_4$ proposed by Naráy-Szabó and Sasvári (1958) and Richardson (1968) respectively. Cation totals agree well with predicted values (30.0 and 29.5) in each case, but silicon and aluminium stoichiometries generally agree better with the 4:18:8 formula. The staurolites in some rocks are deficient in Fe+Mg+Mn, but in others total ca. 4.0 atoms even when recalculated to 46 oxygens. There seems to be an inverse relationship between Si and total Fe+Mg+Mn content.

Mg/Fe+Mg ratios range from 0.13 to 0.21.

Carbonates

Analyses of coexisting calcites and dolomites from a calc-schist and an

epidote-amphibolite are presented in Appendix I , complete with values for CO_2. The divalent cation concentrations were measured directly but the carbon concentrations were calculated by difference during the Z.A.F. correction procedure assuming a carbon to oxygen ratio of 1:3.

The carbonates are plotted together in ternary $CaCO_3$-$MgCO_3$-$FeCO_3$ (Fig. 5.18). The dolomites contain appreciable quantities of iron (8-20 mole % siderite) and fall slightly to the calcium-rich side of the $CaMg(CO_3)_2$ - $CaFe(CO_3)_2$ join. The calcites contain small amounts of both magnesium and iron.

CHAPTER 6

Metamorphic Zonation

CHAPTER 6

METAMORPHIC ZONATION

A. INTRODUCTION AND DEFINITIONS

Textrual criteria for distinguishing generations of mineral growth in various rock types were established in Chapter 4. This chapter examines some mineral assemblages in more detail and uses them to provide information on spatial variation in grade attained during Alpine metamorphism. The problem will be tackled qualitatively by studying the metamorphic zonation in rock types whose parageneses are sensitive to differences in physical conditions such as pressure and temperature, i.e. predominantly metapelites but also metabasites. No attempt will be made here to compute absolute values of pressure, temperature and thermal gradient; this aspect is dealt with in Chapter 7.

(a) Metamorphic facies and grade

The equilibrium mineral assemblage in a rock is a function of both its bulk composition and the external conditions such as temperature and pressure imposed upon it during metamorphism. For a given bulk composition the assemblage may remain essentially unaltered over a range in imposed conditions, but if the conditions are varied by more than a certain amount new phases appear at the expense of pre-existing ones. The presence of a specific assemblage in a rock can therefore be used to estimate, within limits, the metamorphic conditions under which it equilibrated. This principle forms the basis of both this and the following chapter.

The range in imposed conditions over which a group of assemblages remain unaltered in a large range in bulk compositions (e.g. pelitic rocks, basic rocks, all rocks) is defined as a metamorphic facies.

In this chapter, the term 'grade' is broadly synonymous with temperature. Thus, high grade rocks are interpreted as having equilibrated at relatively higher temperatures than low grade rocks. The term implies nothing about differences in pressure.

(b) Some implications of the phase rule

A general relationship between chemical composition and the number of minerals in equilibrium in a rock is given by the Gibbs phase rule (derived in Appendix II, part (c)):

$$p + F = c + 2$$

where c is the number of components necessary to define the system, p the number of physically separable crystalline or fluid phases, and F the number of degrees of freedom.

If the rock system is closed, i.e. there is no exchange of material between it and its surroundings, pressure and temperature are the only variables controlled externally. For a given phase assemblage to characterise a metamorphic facies, as defined above, it must be stable over a range in pressure and temperature, i.e. stable within a finite volume of rock or exposed as an area on the ground surface. The variance of such an assemblage must be greater than or equal to two so that pressure and temperature can be varied independently of one another. Consequently, the maximum number of mineral phases that can stably coexist over a range in external conditions is equal to (c-1)

(Goldschmidt, 1912).

It is sometimes reasonable to assume that a rock system is open to mobile components such as H_2O or CO_2. These are called "perfectly mobile components" after Korzhinskii (1965) and Thompson (1970). Those for which the system is closed are called "fixed components". The numbers of each are labelled C_m and C_x, respectively. Thus, the externally controlled variables include not only temperature and pressure but also the chemical potential of each mobile component (see also Equation A.14, Appendix II), so that the variance of the system must be equal to or greater than $2+C_m$ for the assemblage to be stable over a range in imposed conditions. Thus, the maximum number of minerals that can coexist stably in an open system over a finite volume of rock is equal to C_m

In a closed system, the components of the fluid phase are not perfectly mobile and the composition of the fluid is buffered by the coexisting minerals and not controlled externally. Many metamorphic rock systems may be more realistically represented by this model than by the open system model (Greenwood, 1975).

Metamorphic zones and isograds

In order to examine spatial variations in externally imposed conditions, it is necessary to compare only those rocks whose compositions can be represented by the same major components, and whose parageneses can be plotted on the same phase diagram. Metamorphic zones are areas on the ground surface characterised by different mineral assemblages in rocks with similar compositions. Each such zone corresponds to a distinct metamorphic facies.

The line separating two metamorphic zones has been termed an "isograd" by Tilley (1925) who defines it as a line joining points marking the first appearance, proceeding upgrade, of a mineral characterising a higher grade zone.

Two distinct types of isograd can be recognised. The first marks the trace of a discontinuous reaction which is univariant, or nearly so, in pressure-temperature-space, e.g. the staurolite isograd in the Scottish Dalradian. In this case, mineral assemblages in adjacent zones are incompatible in the sense that they result in different basic topologies (crossed tie-lines, etc.) when plotted on equivalent phase diagrams. This is the definition of "isograd" adopted by Carmichael (1970) and Frey (1974). The position of such an isograd is the same in rocks with different proportions of major components (e.g. in pelites with different Mg/Fe ratios) and therefore represents a facies boundary.

The second type of isograd marks the first appearance of a mineral as the result of a continuous or multivariant reaction, in which the phase diagrams maintain the same topologies but the compositions of the constituent minerals change gradually so that paragenetic fields migrate across the phase diagrams. If, for instance, the composition of a rock plots in a 2-phase field at low grade but the adjacent 3-phase field moves across the composition at higher grades, the third phase will appear suddenly. The outer limit of the phase will be mappable as a univariant line on the ground surface even though the reaction generating it was multivariant e.g. the garnet isograd in the Scottish Dalradian (Atherton, 1964). The grade at which the mineral appears will differ for rocks containing different proportions of major components so that this line is only a facies boundary for rocks with identical bulk compositions. As the composition on the scale of a hand specimen can be expected to vary considerably even in a seemingly monotonous lithological unit, this type of isograd is not as useful as the first.

The above definitions imply that the assemblages on either side of an isograd equilibrated under the locally imposed conditions. A line separating assemblages formed at different times (i.e. a metamorphic front) is not an isograd because it is controlled by kinetic factors and not equilibrium phase relations.

B. PELITES

(a) Critical Assemblages

Most of the pelitic schists in the area studied can be represented in the system FeO-MgO-Al_2O_3-SiO_2-K_2O-H_2O provided that the following assumptions are made:

1. The abundant late overgrowths of oligoclase grew metasomatically after the peak of metamorphism at the expense of white mica, and need not be considered as part of the peak metamorphic assemblage. CaO and Na_2O can be treated as minor components introduced at a later stage.

2. The minor components CaO, Fe_2O_3, TiO_2 and Na_2O can be ignored because each occurs only in significant amounts in one equilibrium mineral, epidote, magnetite, ilmenite and albite or paragonite, respectively. The phase rule predicts that the equilibrium relations among other phases will not be affected if each of these components and the phase containing it are both excluded

from the equation. In addition, it is assumed that MnO is sufficiently dilute in all phases as to have no significant effect on stability.

The mineral parageneses in muscovite+quartz-bearing pelites can therefore be treated using the AFM projection of Thompson (1957). Ignoring retrogressive chlorite and inclusions isolated from other minerals by the enclosing porphyroblasts, the following AFM assemblages (excluding excess components) can be distinguished:

(1) gt + chl

(2) gt + biot + chl

(3) gt + ctd + chl

(4) ctd + chl

(5) ct + st + ctd + chl

(6) ctd + biot

(7) gt + ctd + biot

(8) gt + ctd + biot + chl

(9) ctd + biot + chl

(10) gt + st + ctd + biot

(11) gt + st + ctd + biot + chl

(12) st + biot

(13) gt + st + biot

(14) gt + st + biot + chl

(15) gt + st + ky + biot

(16) gt + st + ky + biot + chl

In most of these assemblages, the chemographic relationships are the same, with Mg/Mg+Fe ratios increasing in the order garnet<staurolite<chloritoid<biotite<chlorite. This is identical to the sequence found by Albee (1972) and Rumble (1974) in pelites from many different localities. In some rocks with

assemblages (15) and (16), garnet rims have virtually identical Mg/Mg+Fe ratios as coexisting staurolite.

Two other assumptions are inherent in the use of this projection: (i) muscovite is present as a pure phase and (ii) H_2O is a perfectly mobile component. The first of these is clearly not true, as all muscovites analysed so far have some degree of celadonite and paragonite substitution, but this will be ignored because pelitic muscovites have a fairly narrow compositional range. The topology of an AFM diagram in which mineral compositions are projected from the pure muscovite component will be the same as one in which the real composition is used as the projection point. Topological differences between phase diagrams of different rocks are not, therefore, likely to be due to this systematic discrepancy. Only a few examples have been found of pelites in which the second assumption may be at fault; these are discussed in later sections. Unless otherwise stated, it will be assumed that the composition of the fluid in equilibrium with pelites is uniform on a regional scale.

A divariant assemblage in the AKFMSH-system consists of three minerals in addition to the excess components muscovite and quartz. In reality, the compositions of AFM minerals, particularly Fe/Mg ratios, can vary within certain limits so that compositionally univariant two-phase-fields and compositionally divariant single-phase-fields occur. Those of the parageneses listed above having a variance of two or more (i.e. capable of being pressure-temperature-facies defining assemblages) are plotted on schematic AFM diagrams in Fig. 6.1, where, for the sake of clarity, mineral compositions are restricted to points and two-phase-fields reduced to lines. Assemblages (1), (2), (3) and (4) are isofacial, as are (4), (6), (7) and (11), and (12) and (13).

Several explanations could account for the low variance assemblages (5), (8), (9), (10), (14), (15) and (16). They could be:

(i) truly univariant or invariant equilibrium assemblages collected from discontinuous reaction isograds;

(ii) divariant equilibrium assemblages in which an extra component ignored in this system is important in stabilising a mineral (e.g. Ca in garnet);

(iii) equilibrium assemblages which are univariant in P-T-μH_2O-space but appear to be divariant when projected into P-T-space (i.e. the assumption that μH_2O is externally controlled is wrong, and the fluid composition is buffered by the reacting minerals);

(iv) disequilibrium assemblages in which the unreactive nature of a mineral (e.g. garnet) has prevented it from being exhausted in the course of a reaction when it otherwise would have been, so that the effective reacting bulk composition of the rock plots in an adjacent three-phase-field.

The first possibility is very unlikely on account of the large number of such assemblages and the very small probability of sampling a univariant line or invariant point. Garnets in three-phase assemblages have very similar CaMn, etc., contents to those in four- and five-phase assemblages, so there is no particular evidence here in favour of the second explanation, although it has been invoked in other terrains (Chinner, 1965 and Hudson, 1975). It is unlikely that five-phase assemblages can be stabilised in this way. The third possibility has been used by Waters (1976) to explain the common occurrence of pelites with assemblage (15) in the Altkristallin, and may apply in certain cases here. The fourth is, in the author's opinion, the most likely overall, in view of the proximity of isograds to one another (see part (b)), and has also been invoked by Hollister (1969), Atherton and Brotherton (1972) and Fox

(1975) to account for overdefined assemblages in pelites. The abundance of pre-Alpine garnets in Inner Schieferhülle rocks lends credibility to this.

(b) Metamorphic zones and isograds in the south-east Tauern

Three distinct metamorphic zones can be mapped in quartz-bearing pelitic schists (see Map 3):

(i) Garnet+chlorite zone (lowest grade - assemblages (1) to (4) covering the area west of Mallnitz and the south-western part of the Mölltal Schiferhülle.

(ii) Staurolite+biotite zone (highest grade - assemblages (10) to (16) covering the structurally lowest parts of the Peripheral Schieferhülle and basement of the Hochlam Kern.

(iii) Chloritoid+biotite zone (intermediate grade - assemblages (6) to (9)) occupying a thin 100-200 m wide area between the other two but apparently wedging out towards the south-east.

In addition, a rather ill-defined area characterised by assemblage (5), the garnet+staurolite+chlorite subzone, occurs in the highest grade part of the garnet+chlorite zone, but only where the chloritoid+biotite zone is absent.

These metamorphic zones are bounded by isograds which mark the first appearance of the defining assemblage proceeding upgrade. Consequently, overdefined assemblages are included on the high-grade side of the appropriate isograd (e.g. the gt + ctd + biot + chl assemblage is placed in the chloritoid+biotite zone and not the garnet+chlorite zone).

The isograds can be modelled on the following discontinuous reactions:

$$gt + chl + musc = ctd + biot + qz + H_2O \qquad (6.A)$$

(chloritoid+biotite isograd)

$$gt + chl + musc = st + biot + qz + H_2O \qquad (6.B)$$

(staurolite+biotite isograd north of Zwenberger Tal)

$$ctd + musc = gt + st + biot + qz + H_2O \qquad (6.C)$$

$$ctd + qz = gt + st + chl + H_2O \qquad (6.D)$$

(garnet+staurolite+chlorite isograd)

$$gt + chl + musc = st + biot + qz + H_2O \qquad (6.E)$$

(staurolite+biotite isograd south of Zwenberger Tal)

The relationship between these univariant reactions and the associated divariant assemblages can be represented by a Schreinemaker's bundle (Zen, 1966) in which they are arranged in P-T space around an invariant point gt + st + ctd + biot + chl (+musc + qz) (Fig. 6.2). The temperature axis can be orientated knowing that H_2O is evolved on the high entropy and therefore high temperature side of each reaction, and that staurolite is stable at higher temperatures than chloritoid, as shown experimentally by Halferdahl (1961), Hoschek (1967,1969) Ganguly (1968), Ganguly and Newton (1968) and Richardson (1968). The orientation of the pressure axis is less obvious and is discussed later. The arrangement of isograds on the ground surface is entirely consistent with this topology, in particular the mutual exclusion of the chloritoid+biotite zone and the garnet+staurolite+chlorite subzone.

All along the Mölltal, therefore, low grade garnet+chlorite assemblages pass north-eastwards into high grade staurolite+biotite assemblages. In the south-

eastern section, chloritoid breaks down at lower temperatures than garnet+chlorite, but towards the north-west the opposite is true (Map 3). The invariant point is located somewhere on the hillside south of Gröneck.

The scarcity of pelitic units in the Mölltal Schieferhülle makes it impossible to locate the isograds exactly, except in rare cases. In Zwenberger Tal, for instance, the staurolite+biotite isograd occurs between two pelitic schist horizons about 300 m apart. However, the isograds are constrained sufficiently well to show that they dip roughly parallel to the S_a^2 schistosity, but not well enough to show whether more or less steeply. D_a^3 updoming seems to have rotated the isograds as well as the schistosity into a subvertical position. No discontinuity in metamorphic grade can be detected at the basement-cover interface.

(c) Progressive mineral reactions

There are two ways in which the sequences of reactions can be studied in rocks undergoing prograde metamorphism. The first is by examining the facies changes along a traverse from lowest to highest grade. The assumption that the high-grade rocks in one place were subjected, at some earlier stage, to the conditions under which peak metamorphic assemblages crystallised in lower grade rocks in a different place, is likely to be false in overthrust terrains (England and Richardson, 1977). Despite this limitation, this method provides valuable information about the line in P-T-space linking peak metamorphic assemblages, irrespective of whether or not individual rocks have followed it and whether they have attained those conditions simultaneously. The second approach is to identify the actual reactions experienced by any given rock by examining the relics of earlier stages, particularly zoning profiles and inclusion assemblages in garnets, in the same rock. This method has been

employed by Tracy et al. (1976) and Thompson et al. (1977b). As pointed out in Chapter 4, its main disadvantage is that early assemblages are always incomplete. Both methods are employed in the present study.

Garnet-forming reactions:

Garnet is present even in the lowest grade pelitic rocks, so the main garnet-producing reaction cannot be studied with reference to an isograd. It is generally thought (e.g. Atherton 1964; Miyashiro, 1973) that the first appearance of garnet in pelites is due to the continuous reaction:

$$chl + qz = gt + H_2O \qquad (6.F)$$

Mineral textures in the Tauern pelites are consistent with this reaction in that chlorite appears to be the only reactant ferromagnesian phase. The growth of garnet in schists containing neither biotite nor chloritoid indicates that these minerals do not participate in the reaction. In many schists containing the assemblage gt + ctd + chl, the garnets have clearly grown at the expense of chlorite but have overgrown chloritoid in such a way that the latter appear not to have been resorbed at all (Fig. 6.3), so that chloritoid inclusions are abundant, whereas chlorite ones are rare. The result is the gradual displacement of the three-phase triangles gt + ctd + chl and gt + biot + chl to more Mg-rich compositions (Fig. 6.15).

The gradual increase in the $Mg/Mg+Fe^{2+}$ ratio from core to rim in post-D_a^2 garnets of the chloritoid+biotite zone schist GD 168b and the garnet+staurolite +chlorite subzone schist GD 308c (Fig. 6.5) are consistent with reaction (6.F), but the reverse trend in the ?pre-D_a^2 garnet cores in GD 18 is not.

In nearly all pelites, garnet has replaced muscovite to some extent, indicating that the actual path by which reaction (6.F) took place was more complex

than just in situ replacement of chlorite by garnet. An extreme case of this, GD 492, has already been mentioned in Chapter 4, Section D(b) (Fig. 4.21). Garnets have grown at the expense of muscovite only in chlorite- and quartz-free parts of the rock, while chlorite and muscovite have locally recrystallised in thin quartzose bands. These textures suggest that the net garnet-producing reaction is the sum of several localised ionic reactions and that there has been centimetre-scale migration of soluble ions between widely separated nucleation sites, as proposed by Carmichael (1969). In particular Fe^{2+} and Mg^{2+} must have been transported to garnet nucleation sites and K^{+} and H_2O away from them. The textures therefore set a lower limit (0.5 to 1 cm) on the distances over which such ions can diffuse during a continuous prograde reaction. A simplified cyclic reaction scheme based on idealised mineral compositions is presented in Fig. 6.6. Some diffusion of Al is implied if the reactions are balanced as in Fig. 6.6, but the necessity for this can be removed if a sufficiently subaluminous phengite composition is used instead of muscovite.

Intermediate grade reactions - north-west Möltal:

In this area, the first discontinuous reaction encountered is (6.A). Chloritoid+biotite zone schists provide good textural evidence for new growth of these minerals. In aluminous pelites with an original assemblage gt + ctd + chl, in which chloritoid and chlorite are usually alighed in S_a^2, biotite forms cross-cutting flakes. Similarly, cross-cutting chloritoid plates are prominent in less aluminous schists (Fig. 6.4). In rocks of this zone, chloritoid and biotite are very commonly found in contact with one another with no signs of alteration, and there is little doubt that they are in equilibrium. On the other hand, primary chlorite never seems to form cross-cutting flakes, and is always subordinate to the other minerals in the assemblage gt + ctd + biot + chl suggesting that much of it has been used up.

However, there is no textural evidence to suggest that garnets have been resorbed during reaction (6.A).

The occurrence of the two-phase assemblage ctd + biot is unusual because either garnet has never been formed or it has been completely exhausted by reaction (6.A). If the former is true, a different reaction must be called upon to produce this assemblage. It is possible that the continuous reaction

$$\text{chl} + \text{musc} = \text{ctd} + \text{biot} + \text{qz} + H_2O \qquad (6.G)$$

could result in low grade ctd + chl schists giving rise directly to ctd + biot schists. This reaction would have to take place above the chloritoid+ biotite isograd for garnet not to form, implying that at the isograd chlorite coexisting with chloritoid would be very iron rich (Mg/Mg+Fe ratio of 0.3 or less). There is no evidence in support of this. The alternative explanation seems at variance with the persistence of garnet in the four-phase assemblages, so this problem remains unresolved.

A combined AFM plot of the mineral compositions in three chloritoid+biotite zone rocks is shown in Fig. 6.7.

Since no schists have been found with compositions more aluminous than the ctd + chl tie line, the first appearance of staurolite in this area is heralded by reaction (6.B). In the few st + biot schists collected immediately upgrade of the chloritoid+biotite zone, both garnet and chlorite persist, giving rise to assemblages with five AFM phases. For this reason, reaction (6.C) is totally obscured.

Two rocks containing assemblage (14) have been collected from upgrade of the chloritoid+biotite zone. The garnets in one, GD 202b, contain chloritoid inclusions not present in the matrix, indicating that chloritoid is no longer

stable in the presence of muscovite, quartz, etc., but that conditions did not exceed the stability limit of the pure mineral. The other, GD 189, shows no evidence of ever having grown chloritoid and yet contains abundant garnet and chlorite, suggesting that staurolite formed by reaction (6.E), not (6.B). This discrepancy can only be resolved if it is postulated that the P-T trajectories of these higher grade rocks did not pass through the stability field of ctd + biot.

Intermediate grade reactions - south-east Möltal:

In this area, the first discontinuous reaction encountered involves only schists which contain chloritoid (i.e. low grade assemblages (3) and (4)). There is good textural evidence for the breakdown of chloritoid according to reaction (6.D). Tiny euhedral prisms of staurolite and sprays of chlorite nucleate around and in chloritoid laths (Fig. 6.8). The formation of pro-grade chlorite at this grade is noteworthy. Taken on its own, this texture suggests the continuous reaction:

$$\text{ctd} + \text{qz} = \text{st} + \text{chl} + H_2O \qquad (6.H)$$

The effect of this reaction, proceeding from left to right, would be to move the three-phase triangle st + ctd + chl towards more iron-rich compositions so that aluminous and magnesian ctd + chl schists would gain staurolite as an extra phase (Fig. 6.15). The first appearance of staurolite would be an isograd of the less useful composition-dependent type.

However, all the st + ctd + chl schists found also contain garnet. In GD 43b, the original AFM assemblage was gt + ctd + chl. Fig. 6.9 shows part of this rock where chloritoid, partially enclosed by garnet, is being replaced by staurolite and chlorite. As garnet is already present, reaction (6.D) and not reaction (6.H) must be responsible. It is not clear whether the rare flat

garnets in GD 308c grew before or during chloritoid breakdown, as they have not nucleated beside the staurolite and chlorite. However, breakdown of the comparatively Mn-rich chloritoid after growth of flat garnet cores could a account for the unusual Mn-enrichment of garnet rims (Fig. 5.7). The compositions of AFM minerals coexisting in GD 308c are plotted in Fig. 6.10. Pelites whose compositions plot on the low-alumina side of the garnet-chlorite join are unaffected by the above reactions. One of these, GD 43a, is also represented in Fig. 6.10. Unfortunately none have been found containing the critical subzone assemblage gt + st + chl in which reaction (6.D) has gone to completion.

Discontinuous reaction (6.E) affects schists with compositions plotting both above and below the gt + chl tie-line. GD 50c is a good example of the rock but some inclusions are preserved within large post-S_a^2 garnets. Biotite is abundant and some has formed reaction rims round garnet. Small euhedral staurolites contact biotite but do not occur as inclusions within garnet. Chlorite is rare and seems to be secondary after biotite. In other rocks, staurolite prisms are clustered round the perimeters of garnet grains, confirming that garnet breakdown has occurred.

High grade reactions:

Staurolite+biotite assemblages persist to the structurally lowest parts of the basement exposed in the floor of the Gössgraben. Kyanite is present as an additional phase in many schists of this zone (assemblages (15) and (16)), but its occurrence seems to be sporadic and there is no evidence of a distinct kyanite zone as in the Scottish Highlands (Barrow, 1912; Tilley, 1925; Atherton, 1977). All recorded Alpine kyanite-schists from the thesis area also contain staurolite, garnet and biotite.

There are several possible kyanite-forming reactions. If not all the chlorite has been exhausted by reactions (6.A), (6.B) and (6.E) owing to the non-reactive nature of garnet, some could remain to react with staurolite in the univariant reaction:

$$st + chl + musc + qz = ky + biot + H_2O \qquad (6.J)$$

so that the stable assemblage in the matrix of the rock is st + ky + biot. This reaction would have to follow closely after (6.E) because kyanite has been found in rocks immediately above the staurolite+biotite isograd. Alternatively, kyanite could form by the reaction:

$$st + qz = gt + ky + H_2O \qquad (6.K)$$

As staurolite, garnet and kyanite are virtually colinear (Figs. 6.11 and 6.12) this reaction is likely to be univariant.

There is little textural evidence in support of either (6.J) or (6.K). In GD 18 (assemblage (15)), kyanite blades have grown round, sometimes completely enclosing, staurolite prisms clustered round garnets, suggesting that the kyanites post-dated the staurolites, but the latter do not seem to be corroded (Fig. 6.13). On the other hand, staurolite and kyanite appear to have grown simultaneously in some Inner Schieferhülle pelites (Fig. 2.17) suggesting that reactions (6.J) and (6.K) do not apply. It is difficult to see how two such aluminous phases can form on the same side of a reaction. Some of these rocks contain paragonite, so it is possible that reaction (6.E) occurred at the same time as:

$$NaAl_3Si_3O_{10}(OH)_2 + SiO_2 = NaAlSi_3O_8 + Al_2SiO_5 + H_2O \qquad (6.L)$$

mica quartz plagioclase kyanite fluid

However, the importance of this reaction is rather doubtful because all the

plagioclase occurring in kyanite-schists grew after staurolite and kyanite crystallisation was complete (Figs. 2.20 and 4.23).

The very scattered occurrence of kyanite throughout the staurolite+biotite zone suggests that it has not been produced by reactions discontinuous in P-T-space. Reactions (6.J) and (6.K) could become continuous in P-T-space if the composition of the fluid phase were buffered by the reacting solids. In both, the reaction products are stabilised by low water concentrations (see Chapter 7). If the fluid coexisting with a given rock were sufficiently dilute in water, the stability field of the products would be extended to low enough grades to be stable under the existing P-T conditions. Consequently, low variations in initial fluid composition could determine whether or not kyanite is present as an extra phase. Waters (1976) points out that the position of reaction (6.K) in P-T-space is very sensitive to changes in fluid composition. Perhaps the most convincing piece of field evidence in support of this comes from the Peripheral Schieferhülle schists in Zwenberger Tal. GD 50c, with assemblage (13), was collected from a schist band 50 m wide while GD 18 with kyanite in addition, was sampled from a 10 cm band in nearby calc-schists which probably equilibrated with a fluid rich in CO_2.

AFM diagrams of two kyanite-schists are shown in Figs. 6.11 and 6.12. In one (GD 18) the garnets are compositionally zoned from cores ($Mg/Mg+Fe^{2+}$ = 0.10) through Mg-poor near-rims ($Mg/Mg+Fe^{2+}$ = 0.06) to Mg-rich rims ($Mg/Mg+Fe^{2+}$ - 0.14) and contain a variety of inclusions which, together with the zoning profiles and matrix mineral compositions, can be used to deduce the reaction history of the rock:

1. Bulk of garnets grow by a continuous reaction, resulting in slight zonation outwards to more Fe-rich compositions. Garnets enclose chloritoid grains. Equilibrium AFM

assemblage: gt + ctd + chl.

2. Chloritoid breaks down by discontinuous reaction (6.D). Garnet continues to grow enclosing staurolite, chlorite and relict chloritoid inclusions.

3. Abundant staurolite and biotite produced in matrix of rock by discontinuous reaction (6.E). No evidence exists of garnet resorption.

4. Mg-rich garnet rims (free of inclusions) grow and staurolites become poorer in Fe and Al by continuous reaction:

$$St + musc + qz = gt + biot + H_2O \qquad (6.M)$$

Kyanite grows near staurolite by continuous (Fe-Mg, H_2O) reaction (6.K), which may also contribute to garnet rims.

A schematic pseudobinary $T\text{-}X_{Fe,Mg}$ section has been derived for the south-east Mölltal area using many of these reactions (Fig. 6.15b), and a similar section has been drawn for the north-west Mölltal area (Fig. 6.15a) (cf. Albee, 1972; Thompson et al., 1977b; Atherton, 1977).

(d) Metamorphic facies series in the south-east Tauern

The two anomalous occurrences of ctd + biot schist west of Mallnitz (Map 3) were not included in the chloritoid+biotite zone because both specimens were collected from near (~5 m) to bands of marble and the assemblage has probably been stabilised in both cases by an abnormally CO_2 rich fluid. This is supported by the coexistence of ctd + biot in the hornblende-garbenschiefer and coarse-grained chloritoid-schist, both of which were found in close

association with carbonate rocks. Calcite, chloritoid and biotite actually coexist in one rock (see Chapter 2, part D(j)). However, it is unlikely that the bulk of the chloritoid+biotite zone has been generated in this way because ctd + biot - schists and gt + st + ctd + chl schists have been sampled from units along strike from one another in apparently identical geological settings.

In the absence of evidence to the contrary, it is assumed that the three main metamorphic zones and the garnet+staurolite+chlorite subzone constitute distinct facies in P-T-space at fixed μH_2O. The sequences of prograde reactions in the north-west and south-east Möltal constitute different facies series (Miyashiro, 1961). Fig. 6.2 shows that the difference between them is that rocks in one were being subjected to higher lithostatic pressures than those at the same temperatures in the other, but as yet no evidence for which of the two series is the higher pressure type has been found in the present study.

Both types of facies series can be identified in other metamorphic terrains. Many textbooks (e.g. Winkler, 1974) imply that the south-east Möltal type is more common and that gt + chl are broken down by reaction (6.5). This is certainly true for parts of New England (Thompson et al., 1977b) and the classic Barrovian sequence in the Scottish Dalradian (Harte, 1975). Nevertheless, ctd + biot assemblages have been recorded from several terrains, including Naxos, Greece (Jansen and Schuiling, 1976) various parts of southern Switzerland (Krige, 1917; Bossard, 1929; Frey, 1969) and Stonehaven, Scotland (Harte, 1975). In addition, Albee (1972) analysed specimens from seven different localities in the U.S.A. and Canada. In his study of the aluminous pelites of the Lukmanier area, Switzerland, Fox (1975) documented a series of reactions in which (6.B) and (6.C) were represented. Ctd + biot assemblages would have developed had the rocks been less aluminous. Thus, despite Winkler's (1974) assertion that "all previous recordings of coexisting

chloritoid and biotite have been shown to be incorrect", it appears that the north-west Mölltal (Stonehavian) facies series is more common than previously thought.

The Schreinemaker's bundle in Fig. 6.2 forms a small part of the petrogenetic grid linking all stable invariant points and univariant reactions involving the common pelitic AFM phases: gt, st, cord, ctd, biot, chl, as (as = one of ky, and or sill). It should therefore be possible to orientate Fig. 6.2 uniquely in P-T-space knowing the stable topology of the whole grid. Unfortunately, opinion is divided as to what this is. All published grids agree that the invariant point [as, cord]* is stable. The stability field of ctd + biot extends from [as, cord] to lower pressures in the grids of Albee (1965) for conditions of $P_{H_2O} \approx P_{TOT}$, Hoschek (1969) and Harte (1975), but towards higher pressures in those of Albee (1965) for $P_{H_2O} \approx 0.5\ P_{TOT}$, and Kepezhinskas and Klestov (1977). I favour the former configuration for the following reasons:

(i) As kyanite is the only aluminosilicate found in the south-east Tauern [as, cord] must lie within the kyanite stability field. Schists containing andalusite, chloritoid and biotite, apparently in equilibrium, have been recorded (Albee, 1972) implying that the ctd + biot field extends to lower pressures.

(ii) In eastern Scotland ctd + biot assemblages of the north-west Mölltal facies series are developed in the Stonehaven area, geographically situated between the type areas of Barrovian medium-pressure kyanite-sillimanite (≡ south-east Mölltal) and Buchan low-pressure andalusite-cordierite facies series (Harte, 1975).

(iii) The assemblage gt + chl + ky, metastable in the grids of the latter

*The invariant point of Fig. 6.2 is labelled [as, cord] here in accordance with the normal practice of indentifying invariant and univariant assemblages by the phases absent from them.

configuration, is stable and characteristic of high pressure terrains, e.g. in the Grampian Highlands of Scotland, where P $\approx$ 12 kb (P. Wells, pers. comm.).

The north-west Mölltal and south-east Mölltal facies series are shown in relation to the preferred petrogenetic grid in Fig. 6.16.

Knowing the distribution and pressure-temperature relationships of the different pelite mineral facies, it should be possible to construct a set of isotherms and isobars, representing peak Alpine metamorphic conditions, on the erosion surface in the manner of Chinner (1978).

The relationship between isotherms, isobars and isograds in the Mölltal region is shown schematically in Fig. 6.18. Isotherms have been drawn assuming isograds to be approximately isothermal, and have been assigned arbitrary temperature values in order to show that temperature increases with structural depth towards the north-east. Pressure must also increase with structural depth, but isobars must be orientated at an angle to isotherms in such a way as to produce a pressure gradient along any given isotherm (isograd) from lower values in the north to higher ones in the south. The absolute values of pressure, temperature and their gradients will be studied in Chapter 7.

Few records of pelite assemblages exist for parts of the eastern Tauern window outside the area examined here, making it difficult to extend the P-T -map far. Exner (1957) records the assemblages ctd + ky in schists on Klein Silberpfennig, 10 km north-west of Romate Spitze (Map 2). The author has collected samples with assemblages (3) and (4) from immediately above the basement-cover interface in Grossarl Tal, northern Tauern. Staurolite and kyanite coexist in parts of the Silbereck Mulde (Angel, 1939; Angel and Staber, 1952) but apparently not with biotite. With these few constraints

in mind, metamorphic zones have been tentatively extrapolated over the whole eastern Tauern (Fig. 6.17). To a first approximation, the staurolite+biotite isograd forms a closed loop enveloping the Hochalm Kern. The highest grade rocks therefore occupy the core of a thermal dome which coincides with the area of highest basement elevation, forming a culmination. A similar culmination occurs in the western Tauern in the Zillertal and Tuxer Alps, where st + biot assemblages are also developed (Hoernes, 1973), and the Glockner depression, characterised by lower grade assemblages (Cornelius and Clar, 1939; Frasl and Frank, 1966) lies between them. The coincidence of thermal domes with structural domes indicates that the main gneiss massifs in the Tauern Window were generated after the peak of Alpine metamorphism. The bending of S_a^2 over the Hochalm Kern (Norris et al., 1971b) supports this. However, the trends of the isograds are indifferent to irregularities in the basement-cover interface, such as the Sonnblick Kern, which were generated during D_a^2 i.e. the isograds are not folded round large D_a^2 structures. This is consistent with a post-D_a^2 peak of metamorphism. This may also be why the staurolite+biotite isograd intersects progressively lower structural levels towards the north, as seen in a north-south cross-section through the Hochalm Kern (Fig. 6.19).

C. METABASITES

(a) Critical assemblages

The basic rocks of the Peripheral Schieferhülle can be represented by the system Na_2O-CaO-Al_2O_3-FeO-MgO-SiO_2-H_2O-CO_2 if it is assumed that MnO can be ignored because it occurs in very small amounts and substitutes ideally for FeO, and that K_2O, Fe_2O_3 and TiO_2 can be ignored because they only occur in

significant quantities in the accessory minerals biotite, magnetite and either ilmenite, sphene or rutile, respectively:

Metabasite assemblages projecting in this system include:

(1) amph + ep + chl + plag + qz

(2) amph + ep + chl + plag

(3) amph + ep + plag + qz

(4) amph + ep + plag + cc + qz

(5) amph + ep + chl + plag + cc + qz

(6) amph + ep + plag + cc

(7) amph + ep + chl + plag + cc + dol + qz

(8) amph + ep + chl + plag + dol + qz

(9) ep + chl + plag + qz

(10) ep + chl + plag + cc + qz

All of these except (2) and (6) contain the component SiO_2 as a pure phase, so, ignoring those two assemblages, one can project from SiO_2 into the subsystem Na_2O-CaO-Al_2O_3-FeO-MgO-H_2O-CO_2. Unfortunately, it is not possible to assume that the fluid phase in equilibrium with the minerals is either H_2O-rich or compositionally constant over the terrain because most assemblages contain carbonate minerals.

Harte and Graham (1975) have demonstrated how carbonate-bearing assemblages are dependent upon X_{CO_2} of the fluid by considering an isobaric T-X_{CO_2} plot of albite-bearing assemblages in ternary AC (FM) (Fig. 6.20). Dolomite-bearing assemblages are restricted to high X_{CO_2}, and amphibole-free calcite+chlorite assemblages to intermediate X_{CO_2}. By ignoring (7), (8) and (10) and assuming that the presence or absence of calcite in the remainder is due merely to slight variation in X_{CO_2} and does not affect the Fe/Mg ratios of coexisting silicates, it should therefore be possible to examine how

silicate phase relations are affected by metamorphic grade in rocks which equilibrated with a comparatively H_2O-rich fluid. With these restrictions in mind, the following assemblages remain:

(11) amph + ep + chl + plag $\pm$ cc

(12) amph + ep + plag $\pm$ cc

(13) ep + chl + plag

These are plotted schematically in the CaO-Al_2O_3-$NaAlO_2$-(FeMg)O (C-A-NA-FM) tetrahedron of Ernst (1976) in Fig. 6.21. All three assemblages are compatible.

Since there are no discontinuous reactions relating the above assemblages, the effects of grade will only be apparent in continuous reactions, involving changes in mineral compositions. Two minerals which are clearly affected by such reactions in the area studied are calcic-amphibole (actinolite to hornblende) and plagioclase (albite to labradorite).

(b) The albite-oligoclase isograd

Norris et al. (1971a) mapped an isograd delineating the transition from albite to oligoclase in basic and pelitic rocks. This isograd is roughly parallel to and locally coincident with the staurolite+biotite isograd in pelites. Similar breaks in plagioclase composition in other metamorphic terrains have been observed by De Waard (1959), Wenk (1962), Wenk and Keller (1969), Cooper (1972) and Streckeisen and Wenk (1974), and are generally attributed to the effects of the peristerite solvus.

In the present study, plagioclases have been subdivided according to host-rock mineralogy and textural type in order to eliminate the effects of bulk

composition and time in the assessment of metamorphic grade. In general, plagioclase compositions were determined by the Michel-Levy method and/or by comparing refractive indices with those of quartz. A few were analysed with the electron-microprobe. Map 3 shows the distribution of plagioclase compositions coexisting with calcic amphibole, epidote and quartz in metabasites and both in equilibrium with and replacing white micas in pelites.

The present results agree with those of Norris et al. in that oligoclase and andesine predominate in the Hochalm Kern while albite is more common further south or west. However, they are not well enough grouped for one to be able to draw an isograd marking the first appearance of oligoclase in any of the three groups because oligoclase does occur sporadically, usually as rims round pre-existing albites, in the area west of Mallnitz, and has been recorded from close to the western margin of Map 2.

(c) Progressive mineral reactions

The mineral chemistries of two Peripheral Schieferhülle metabasites were studied using the microprobe: GD 551, a greenschist from 2½ km south-west of Mallnitz and GD 193, an epidote-amphibolite from close to the staurolite+ biotite isograd in Kaponigtal. Both rocks contain assemblage (11) (this section). The former contains abundant actinolite needles, locally with very thin rims of hornblende, while the latter contains only coarse, unzoned, prismatic hornblende. In GD 551, oligoclase (an_{22-28}) forms thin rims round large albites (an_1), while in GD 193, oligoclase (an_{22-30}) and albite (an_{0-3}) appear to form separate polygonal grains. In both cases, the gap between an_3 and an_{22} probably represents the peristerite solvus. The zoning relationship in GD 551 suggests that the albite and oligoclase are not in equilibrium. The composition ranges in GD 193 are almost identical to those in GD 551,

suggesting that disequilibrium applies there also, despite lack of textural evidence.

Both rocks contain albite, epidote and quartz, so their parageneses can be examined using the epidote projection of Harte and Graham (1975). They are plotted together in Fig. 6.22. It is clear that the mineralogical differences between them, particularly the predominance of actinolite in GD 551 as opposed to hornblende in GD 193, cannot be explained by differences in bulk composition because the composition of GD 551 (estimated from modal proportions of essential AFM minerals) can be represented by both sets of amphibole-chlorite tie lines. The two sets of tie-lines have distinctly different slopes and intersect at a high angle (i.e. the distribution coefficients for Mg/Fe between chlorite and amphibole are different) implying that a change in grade is responsible. However, the tie-lines between chlorite and hornblende rims in GD 551 are parallel to those in GD 193.

The actinolites in GD 551 are much less aluminous than hornblendes in either rock. The hornblende rims in GD 551 have similar alumina contents to those in GD 193, so that there appears to be a sharp jump in composition in the zoning profile. Similar compositional gaps are known from many other low- and medium-pressure metamorphic terrains, and are generally attributed to the existence of a miscibility gap (e.g. Hallimond, 1943; Klein, 1969; Cooper and Lovering, 1970; Brady, 1974). However, Graham (1974) and Grapes and Graham (1978) argue that the abrupt change from actinolite to hornblende is the result of kinetic factors and abrupt changes in the compositions of, or the appearance or disappearance of, other phases involved in the following continuous reactions occurring in the actinolite-hornblende transition (see also Holland and Richardson, in press):

$$\underset{\text{albite}}{NaAlSi_3O_8} + \underset{\text{tremolite-actinolite}}{Ca_2(MgFe)_5Si_8O_{22}(OH)_2} = \underset{\text{edenite}}{NaCa_2(MgFe)_5AlSi_7O_{22}(OH)_2} + \underset{\text{quartz}}{4SiO_2} \qquad (6.N)$$

$$Ca_2(MgFe)_5Si_8O_{22}(OH)_2 + 14(MgFe)_5Al_2Si_3O_{10}(OH)_8 + 24Ca_2Al_3Si_3O_{12}(OH) + 28SiO_2$$

tremolite-actinolite chlorite epidote quartz

$$= 25Ca_2(MgFe)_3Al_4Si_6O_{22}(OH)_2 + 44H_2O \qquad (6.P)$$

tschermakite fluid

$$6(MgFe)_5Al_2Si_3O_{10}(OH)_8 + 12Ca_2Al_3Si_3O_{12}(OH) + 14SiO_2 = 10Ca_2(MgFe)_5Al_4Si_6O_{22}(OH)_2$$

chlorite epidote quartz tschermakite

$$+ 4CaAl_2Si_2O_8 + 20H_2O \qquad (6.Q)$$

anorthite fluid

Reaction (6.Q) is probably responsible for the general increase in the An-content of plagioclase with grade. In some rocks (e.g. GD 482, Fig. 6.23) there is textural evidence of thereaction:

$$2Ca_2Al_3Si_3O_{12}(OH) + CO_2 = 3CaAl_2Si_2O_8 + CaCO_3 + H_2O \qquad (6.R)$$

epidote fluid anorthite calcite fluid

but the common occurrence of oligoclase-bearing calcite-free epidote-amphi-bolites indicates that this cannot be universally applicable.

The co-production of tschermakite and anorthite by reaction (6.Q) is supported by the positive correlation between modal hornblende and oligoclase in the two analysed rocks. In GD 551 both occur only as thin rims while in GD 193 both are much more abundant. There also exist greenschists without hornblende or oligoclase, and epidote-amphibolites without albite. The generalisation does not apply universally because some epidote-amphibolites appear to contain only albite, but Grapes and Graham (1978) point out that in many other terrains the actinolite-hornblende transition is accompanied by an albite-oligoclase transition, and argue that the sudden appearance of oligoclase as

the result of the peristerite solvus rapidly increases the Al-content co-existing amphibole, thus causing the apparent actinolite-hornblende miscibility gap.

In general, the occurrence of actinolite is likely to be dependent on the bulk rock Mg/Fe ratio. At any one grade, actinolite is restricted to rocks with high Mg/Fe ratios and hornblende to rocks with low ratios, while at higher grades the paragenetic field of hornblende is extended to progressively more magnasian compositions (Harte and Graham, 1975), i.e. that the reactions (6.N) and (6.P), involving reaction of actinolite to form hornblende, are continuous in FeMg space and could be more realistically written in terms of Fe-end-members. Thus the absence of a well-defined actinolite-hornblende isograd may be due to local variations in bulk-rock Mg/Fe ratios. If reactions (6.N) and (6.P) are dominant over (6.Q) in iron-rich bulk compositions, low-Mg albite-epidote-amphibolites and high-Mg greenschists could be isofacial.

D. CONCLUSIONS

The main conclusions of this chapter can be summarised as follows:

(1) In the area studied, pelitic rocks are the most useful indicators of metamorphic grade, despite their restricted area of outcrop. The main pelite metamorphic zones constitute different facies in P-T-space at approximately fixed μ_{H_2O}, but localised chloritoid+biotite- and kyanite-bearing assemblages may have been stabilised by fluids rich in CO_2.

Basic rocks are of limited use because they respond to changing grade by undergoing only multivariant continuous reactions, so that the nature of peak meta-

morphic assemblages is dependent upon bulk-rock chemistry. Also many metabasite assemblages have equilibrated with fluids of different X_{CO_2}. The mineral assemblages in siliceous dolomites are virtually useless as monitors of grade (see Chapter 2, part (D(e)) for this reason.

(2) The late Alpine metamorphism conforms broadly to the medium-pressure kyanite-sillimanite-type facies series found in the Scottish Dalradian, but within the south-east Tauern the lowest and highest zones are missing. In fact, by analogy with the Dalradian zones listed by Atherton (1977), only the garnet, chloritoid and staurolite zones are represented here. Nevertheless, two distinct baric types of facies-series can be recognised: a high-medium-pressure type (Barrovian) characterised by the gt + st + chl subzone, in the south-eastern part of the area, and a low-medium-pressure type (Stonehavian) characterised by the ctd + biot zone, in the north-western part of the area. One corollary of this is that the traces of peak metamorphic isobaric and isothermal surfaces intersect on the erosion surface.

The 40 my Lepontine metamorphism in certain parts of the central Alps also conforms to the Stonehavian facies series, suggesting that the late Alpine metamorphisms in the eastern and central Alps are related both temporally and genetically.

(3) The Hochalm Kern marks the site of a thermal and structural culmination. The updoming producing it must have occurred after the peak of metamorphism, i.e. during D_a^3.

CHAPTER 7

Metamorphic Conditions

CHAPTER 7

Metamorphic Conditions

A. INTRODUCTION

The principal aim of this chapter is to estimate the temperatures and pressures to which the units in the south-east Tauern were subjected at various stages in their metamorphic history. The emphasis is placed on the Alpine metamorphism (M_a) because (a) most assemblages crystallised or recrystallised then (see Chapter 4), and (b) the implications of M_a conditions are likely to be more numerous than for earlier events, as much is already known about Alpine structure and geochronology. Nevertheless, inferences about metamorphic conditions during the M_e, M_v and M_z events from the few remaining pre-Alpine assemblages are included despite the fact that their significance is obscure.

In Chapter 4, different generations of M_a mineral growth were distinguished. If P-T-conditions can be calculated for each generation, a P-T-time curve can be constructed from which relative rates of uplift and heating can be deduced.

The spatial variation in Alpine peak metamorphic grade has already been discussed qualitatively in terms of metamorphic zonation (see Chapter 6). In this chapter, an attempt is made to quantify the temperature gradient perpendicular to isotherms preserved in the Hochalm Kern, assuming the configuration of Fig. 6.17.

Pressures and temperatures are calculated using a thermodynamic approach. The advantages of this over merely equating equilibrium assemblages in rocks with

the reactants or products of experimentally reversed reactions and using the equilibrium curves as facies boundaries are, (a) it is more rigorous and takes into account variations in natural mineral compositions, (b) it increases the selection of available geothermometers and barometers to include many which have not been experimentally investigated, and (c) it enables one to estimate the composition of the fluid phase. The amount by which the fluid composition varies, and the distances over which it does so, place constraints on the extent to which the fluid composition is externally controlled.

An outline of the thermodynamic theory employed in subsequent sections is presented in Appendix II.

B. APPLICATION OF THERMODYNAMIC METHODS

(a) Applicability of equilibrium thermodynamics to mineral assemblages

The thermodynamic theory outlined in Appendix II can only be applied to metamorphic mineral assemblages in which the preserved compositions were in equilibrium at the pressure, temperature and μ_{H_2O} of interest. It is not possible to prove that equilibrium has occurred, but the assumption is strengthened when two or more P-T-μ solutions agree. In general, equilibrium will be assumed unless the mineral textures suggest reaction relationships or very different times of growth (see Chapter 4). In coarse-grained "equilibrium" rocks, near-rim compositions are used for calculations.

In order to solve thermochemical equations for the equilibrium condition $\Delta G_{P,T} = 0$, it must be possible to write a balanced chemical reaction between components occurring in phases of the equilibrium assemblage:

$$jJ + kK = lL + mM \text{ (components)}$$
$$A \quad B \quad C \quad D \text{ (phases)}$$

Thus, if phases A, B, C and D coexist in a rock, apparently in equilibrium, the calculated equilibrium curve should, within error, pass through the point marking the conditions under which the assemblage equilibrated. However, if the rock only contains phases A, B, and C, the above reaction can still be used as a P-T-μ indicator. In this case the equilibrium curve calculated for real A, B and C compositions and pure D (i.e. $a_D^M = 1$) delimits the maximum possible stability field of A+B+C, i.e. the assemblage equilibrated somewhere to the J+K side of the curve.

An assemblage in which phases A and B have been partially replaced by C and D indicates that the above reaction has gone to the right, and thus provides information on the direction of the P - T trajectory. The calculated equilibrium curve will only pass through the P - T conditions under which the reaction occurred if all participating phases were in equilibrium. In many examples this is clearly not so.

(b) Solving for P,T and X_{H_2O}

In general, the method involves solving equation (A.43) for pressure, using known or assumed values for temperature and X_{H_2O}, X_{CO_2}, etc., and known mineral compositions. If the reaction involves fluid, it can only be solved graphically, but a direct solution is possible for a solid-solid reaction.

It is theoretically possible to calculate an equilibrium curve for any desired reaction, but in practice this is limited by the availability of standard-state thermochemical data for pure phases. There already exist

several data sets which include most common metamorphic minerals (Robie and Waldbaum, 1968; Zen, 1972; Powell, 1973; Helgeson, 1976; Holland, 1977; Robie et al., 1978). Although some of these have been made partially self-consistent, there is still disagreement between many experimental and calculated equilibria.

The compositions of coexisting minerals related by a solvus can be used as geothermometers if some estimate can be made of pressure, or if W^v is small.

(c) Application of equation (A.43) - some simplifications

Equation (A.43) is cumbersome to use in its expanded form, but a number of simplifications are possible:

(i) Assume $\Delta C_p = 0$ (i.e. $\Delta G^\circ_{1,T} = \Delta G^\circ_{1,298} - T\Delta S^\circ_{298}$) (7.1)

This is not accurate where vapour is involved but reasonable for solids where C_p tends to 3R at high temperatures. ΔC_p values of similar combinations of solids are likely to deviate from 3R by similar amounts, so that this assumption can be used for solid-solid reactions.

(ii) Assume ΔC_p of solids ($\Delta C_{p(s)}$) = 0

$$\text{i.e. } \Delta G^\circ_{1,T} = \Delta G^\circ_{(s)1,298} + \Delta G^\circ_{(v)1,298} + \int_{298}^{T} d\Delta G_{(V)} \qquad (7.2)$$

For fluid-bearing reactions, this is a vast improvement on (i) but is still inaccurate, and only holds over small temperature ranges.

(iii) Assume ΔC_p of formation of solids ($\Delta_f C_{p(s)}$) = 0 (approximation of Fisher and Zen, 1971). $\Delta_f C_{p(s)}$ relates to the entropy of formation from pure

elements, $\Delta_f S$, and not the third-law entropy, S. The latter are related by

$$\Delta_f S(\text{mineral}) = S(\text{mineral}) - S(\text{elements}) \qquad (7.3)$$

In this case:

$$\Delta_f G_{P,T} = 0 = \Delta_f G^{o}_{(s)1,298} - \Delta_f S_{(s)}(T-298) + \Delta V_{(s)}(P-1) + G^{*}_{(v)P,T} \qquad (7.4)$$

where $G^{*}_{(v)P,T} = \Delta_f G^{o}_{(v)1,T} + G_{(v)P,T} - G_{(v)1,T}$ (7.5)

This approximation holds over large temperature ranges. The computer program FZ, written by Dr. S.W. Richardson, incorporates this method, and has been used here in cases where heat capacity data are lacking.

(d) <u>Calculating equilibrium curves directly from experimentally determined equilibria</u>

If the equilibrium of interest has been experimentally investigated for pure phases, the difference between the pressure of the experimental curve, P_1, and that of the equilibrium curve for impure phases, P_2, at a given temperature, T, can be calculated by combining (A.23) and (A.40):

At equilibrium, $\Delta G^{o}_{P_2T} = -RT\ln K$ (A.40)

But, $\Delta G^{o}_{P_2T} = \Delta G^{o}_{P_1T} + \int_{P_1}^{P_2} V\,dP$ (A.23)

$$\Delta G^{o}_{P_2T} = \Delta G^{o}_{P_1T} + \Delta V_S(P_2-P_1) + RT\ln\left(\frac{f_{P_2}}{f_{P_1}}\right)^2 \qquad (7.6)$$

Substituting in (A.40) and rearranging:

$$P_2 = P_1 - \frac{RT\ln K.}{\Delta V_S}\left(\frac{f_{P_2}}{f_{P_1}}\right)^2 \qquad (7.7)$$

If the equilibrium of interest (c) has not been experimentally investigated, but can be expressed as the sum of other reactions (a and b) that have, its equilibrium position can be calculated for pure phases, by first evaluating $\Delta G^{o}_{1,T}$ of the source reactions from (A.43) for a variety of temperatures and known values of P, f, ΔV_S. $\Delta G^{o}_{1,T}(C)$ can then be obtained from:

$$z\,\Delta G^{o}_{1,T}(c) = \Delta G^{o}_{1,T}(a) + y\,\Delta G^{o}_{1,T}(b) \qquad (7.8)$$

One can then substitute $\Delta G^{o}_{1,T}(c)$ into (A.43) and solve for P(equil) graphically at each temperature.

The advantage of direct calculation methods is that, apart from molar volumes, no standard state thermochemical data for minerals are required. The main disadvantages are (i) they are only applicable to moderately pressure-sensitive reactions, and (ii) there is no immediate way of telling whether the source equilibria are correct.

(e) Extraction of thermochemical data from experimentally reversed equilibria

For a few calculations it has been necessary to extract data from experiments. The same equations are used for this as for calculating equilibrium curves. For instance, making the Fisher and Zen approximation and applying (7.4), $\Delta_f S_{(s)}$ and $-\Delta_f G^{o}_{(s)1,298}$ of a reaction can be readily obtained as the slope and intercept of a plot of $\Delta V_s(P-1) + G^{*}_{(v)}$ against (T-298).

In order to minimise error, the same assumptions are made in calculating equilibrium curves as were made in extracting the source data.

(f) Fugacity and $G^{\bullet}$ data for gases

Many of the reactions investigated in part F involve either H_2O, CO_2 or both, so that fugacity data are required in order to apply (A.43). The H_2O fugacity data used in hand-calculations are taken from Burnham et al. (1969) and the CO_2 fugacity data from Mel'nik (1972). In the program FZ, $G^{\bullet}_{H_2O}$ and $G^{\bullet}_{CO_2}$ are derived for a range in pressure and temperature from the modified Redlich-Kwong equation (Holloway, 1976) and the 1 atm. $\Delta_f G_{(v)T}$ data of Robie and Waldbaum (1968). Holland (1977) has compared the fugacity data sets of Burnham et al. and Mel'nik with those calculated from the modified Redlich-Kwong equation and found them to be in close agreement.

C. ACTIVITY-COMPOSITION RELATIONSHIPS IN MINERALS

(a) Garnet

Activity-composition relationships in almandine(1)-pyrope(2)-grossular(3)-spessartine(4) solid solutions have been studied by Ganguly and Kennedy (1974) who derive the following regular solution parameters for $630 \pm 40^{\circ}C$:

$$W_{12} = 2979 \pm 369 \text{ cal/mol.} \qquad (7.9)$$

$$W_{23} = 3830 \pm 220 \qquad (7.10)$$

$$W_{13} < 1355 \qquad (7.11)$$

$$W_{24} = 3200 \pm 600 \qquad (7.12)$$

$$W_{14} \approx W_{34} \approx 0 \qquad (7.13)$$

These values, taken to one significant figure, have been used in subsequent calculations. Temperature dependencies of W are ignored because temperatures of interest (lower amphibolite facies) fall close to 630 ± 40°C.

The W_{23} value (7460 - 4.3 T cal/mol.) of Hensen et al. (1975) agrees very well with the value given above.

Substituting the above values into the quaternary version of (A.67),

$$10^{-3}RT\ln {}^{VIII}_{Fe} = 3(X_{Mg})^2 + (X_{Ca})^2 + (X_{Ca}.X_{Mn}) \qquad (7.14)$$

$$10^{-3}RT\ln {}^{VIII}_{Mg} = 3(X_{Fe})^2 + 4(X_{Ca})^2 + 3(X_{Mn})^2 + 6(X_{Fe}.X_{Ca}) + 7(X_{Ca}.X_{Mn}) + 6(X_{Fe}.X_{Mn}) \qquad (7.15)$$

$$10^{-3}RT\ln {}^{VIII}_{Ca} = (X_{Fe})^2 + 4(X_{Mg})^2 = 2(X_{Fe}.X_{Mg}) + (X_{Fe}.X_{Mn}) + (X_{Mg}.X_{Mn}) \qquad (7.16)$$

(b) Clinopyroxene

The components diopside, hedenbergite, jadeite and acmite total between 96 and 100% in all analysed clinopyroxenes (see Chapter 5, part (b)). Activity-composition relationships are required for both jadeite and diopside components.

To a first approximation, diopside and hedenbergite mix ideally (Mueller, 1961) so that in binary $di_x hed_{1-x}$ solid solutions, $a^{cpx}_{di} = X^{M2}_{Mg}$. (7.17)

In natural clinopyroxenes, local charge balance is likely to be maintained in the coupled substitutions $Ca(Mg,Fe^{2+}) = NaAl$ and $Ca(Mg,Fe^{2+}) = NaFe^{3+}$, so that

random mixing can occur in the combined M site, but not in individual M_1 and M_2 sites. Applying (A.56):

$$a_{jd} = X_{jd}, \tag{7.18}$$

$$a_{jd} \neq X_{Na}^{M1} . X_{Al}^{M2} \tag{7.19}$$

Ganguly (1973) has applied the regular solution model to ternary (di+hed)-jd-acm (1-2-3) solid solutions and has estimated the following interchange free-energies:

$$W_{12} = -1400(\pm 1400) + 1.16T(^{o}C) \text{ cal/mol.} \tag{7.20}$$

$$W_{13} \approx W_{23} \approx 0 \tag{7.21}$$

Substituting these values, (A.67) reduces to:

$$RT\ln\gamma_{jd} = W_{12}(X_{di} + X_{hed})(1-X_{jd}) \tag{7.22}$$

$$\text{and } RT\ln\gamma_{di,hed} = W_{12}(X_{jd})(1 - X_{di} - X_{hed}). \tag{7.23}$$

Unless otherwise specified, the mean value of W_{12}, i.e. -1400 + 1.16T, will be used in calculations.

(c) Epidote

The activity of zoisite in epidote is estimated assuming that Fe^{3+} and Al mix ideally but that Fe^{3+} substitutes in only one of the three Al sites (Kerrick, 1977), i.e.:

$$a_{zo}^{ep} = (1 - X), \text{ where } X = \text{mole\% } Al_2Fe \text{ epidote.} \tag{7.24}$$

The data of Dollase (1973) indicate that the above assumption holds for epidotes with less than 80% Al_2Fe end-member, but that in more iron-rich compositions some Fe^{3+} (usually about 10% total Fe^{3+}) enters the M_1 site. Thus, for

Fe-rich epidotes a two-site ionic mixing model of the form

$$a_{zo}^{ep} = (1 - 0.9X).(1 - 0.1X) \qquad (7.25)$$

may be more appropriate. However, as the two models give very similar activities in the region X = 0.8 and few epidotes analysed in this study contain more iron than this (Fig. 5.12), the single-site model will be used throughout.

(d) Plagioclase

No activity-composition data exist yet for low-temperature albites and intermediate plagioclase, but Orville (1972) studied high-temperature plagioclases at 700°C and 2 kb and obtained γ_{ab} and γ_{an} values of 1.0 and 1.28 respectively for compositions in the range An_{0-55}. Saxena and Ribbe (1972) have analysed the 650°C, 1 kb high-plagioclase data of Seck (1971) and derive higher values (ca. 1.5) for γ_{an} in the same composition range. Neither are strictly applicable to the low-plagioclase series. Unless otherwise stated, Orville's values will be used in calculations.

(e) Alkali feldspar

Mixing properties of alkali feldspars have been determined by Thompson and Waldbaum (1969) for the sanidine-high albite series, and by Bachinski and Müller (1971) for the microcline-low albite series. The latter are more applicable to the equilibrium alkali feldspars in the rocks studied. Whitney and Stormer (1977) calculated the following asymmetric solution parameters from the data of Bachinski and Müller assuming a symmetrical W^v of 0.11 cal/mol.:

$$W_{NaK}^{ab} = 7973.1 + 0.11P - 6.48T \qquad (7.26)$$

$$W_{NaK}^{micr} = 7490.9 + 0.11P - 2.17T \qquad (7.27)$$

(f) White mica

The position of the solvus in the binary system muscovite-paragonite has been determined at 2.07 kb by Eugster et al. (1972) who derived the following expressions for interchange free-energies at 2.07 kb using the subregular solution model:

$$W_{NaK}^{parag} = 3252.0 + 0.1698T \qquad (7.28)$$

$$W_{NaK}^{musc} = 4424.1 + 0.3954T \qquad (7.29)$$

Eugster et al. also used available molar volume data to derive expressions for W's involving pressure-dependent W^V terms. Chatterjee and Froese (1975) applied more recent volume data to the solvus of Eugster el al. and derived very different expressions for the pressure-dependence of W. The latter generate almost symmetrical solvi in the pressure range 6 to 8 kb. As the M_a metamorphic pressures probably exceeded 5 kb (see part F), and as $X_{Na}^{musc} \approx 2X_K^{parag}$ in coexisting white micas (Fig. 5.14), Chatterjee and Froese's equations must be incorrect. The analyses of Karamata et al. (1970), Henley (1971) and Thompson and Thompson (1976) support this conclusion.

Recently, however, Blencoe (1977) has shown that the excess volumes of mixing, V^{XS}, of $2M_1$ and 1M polymorphs differ considerably and are themselves pressure dependent, so that the pressure dependence of W is complex and subject to large errors. For this reason, the W^V terms are ignored in this study and (7.28) and (7.29) are used throughout.

Ideal ionic mixing is assumed on all other sites, i.e.:

$$a^{wm}_{parag} = (X_{Na} \cdot \gamma_{Na}) \cdot (X_{Al})^3 \cdot (X_{Si})^3 \qquad (7.30)$$

For margarite-paragonite solid solutions, Holland (1977) has applied the sub-regular solution model to the data of Franz et al. (1977) and extracted temperature dependent asymmetric solution parameters:

$$W^{parag}_{(NaSi)(CaAl)} = -2608 + 6.515T \qquad (7.31)$$

$$W^{marg}_{(NaSi)(CaAl)} = 1602 + 2.470T \qquad (7.32)$$

Assuming ideal ionic mixing on all other sites, as above:

$$a^{wm}_{marg} = (X_{CaAl} \cdot \gamma_{CaAl}) \cdot (X^1_{Al})^3 \cdot (X_{Si})^2 \qquad (7.33)$$

where X^1_{Al} = (Al-Ca)/6 in a formula calculated to 22 oxygens.

(g) Carbonates

Bickle and Powell (1977) estimate the following regular solution parameters for carbonates in the ternary system $CaCO_3$-$MgCO_3$-$FeCO_3$:

$$W_{CaMg} = 5360 - 0.032P \text{ cal mol.}^{-1} \text{ (in the region } 400^\circ C \text{ to } 600^\circ C) \qquad (7.34)$$

$$W_{CaFe} = 3800 \text{ cal mol.}^{-1} \qquad (7.35)$$

$$W_{MgFe} = 0 \qquad (7.36)$$

(h) Other minerals

As quartz and kyanite occur as virtually pure phases, their activities are

taken to be unity. Ideal ionic mixing is assumed for biotite, staurolite, chloritoid, chlorite and amphiboles. There is some evidence (Mueller, 1972) that this assumption is valid for biotite.

D. PRE-ALPINE METAMORPHIC CONDITIONS

In this section, the thermodynamic principles outlined in Appendix II and parts B and C above, are applied to the few recognisable pre-Alpine assemblages. The approximate pressure and temperature conditions during the M_e, M_v and M_z events are estimated and compared.

(a) The M_e event

The primary M_e assemblage in the eclogitic amphibolites is gt+cpx+rut+qz, which, without rutile, is one of the critical assemblages used by Banno (1966) to define the eclogite facies.

The geothermometer of Råheim and Green (1974) based on the exchange reaction

$$\underset{\text{garnet}}{Fe_3Al_2Si_3O_{12}} + 3\underset{\text{clinopyroxene}}{MgSiO_3} = \underset{\text{garnet}}{Mg_3Al_2Si_3O_{12}} + 3\underset{\text{clinopyroxene}}{FeSiO_3} \qquad (7.A)$$

can be used to estimate the temperature of equilibration. For this reaction,

$$K_D = \frac{(Fe^{2+}/Mg^{2+})gt}{(Fe^{2+}/Mg^{2+})cpx} .$$

K_D values for near-rim compositions of adjacent garnet and omphacite grains in the fresh patch in GD 136a, range from 10 to 19 but cluster between 11 and

and 14 (Fig. 7.1). Similar partly amphibolitised eclogites associated with schists and gneisses in the Bohemian Massif, to which the Inner Schieferhülle eclogites may be genetically related, have K_D values ranging from 2 to 10 (Dudek and Fediuková, 1974).

A minimum pressure for the M_e event can be calculated using the reaction:

$$NaAlSi_3O_8 = NaAlSi_2O_6 + SiO_2 \qquad (7.B)$$

$$\text{plagioclase}_{\text{clinopyroxene}}\text{quartz}$$

This equilibrium has been experimentally investigated by Birch and Le Comte (1960), using high-albite, Newton and Smith (1967) and Johannes et al. (1971), both using low-albite. The 16.3 kb 600°C equilibrium point of Johannes et al. falls on the curve of Newton and Smith, so the latter is used here.

As albite is absent from the assemblage, a_{ab}^{plag} is set at 1.0. Taking a clinopyroxene composition of $(di+hed)_{50}jd_{40}acm_{10}$ as representative (Fig. 5.8), and substituting in equation (7.89) and (7.59),

$$K = \frac{a_{jd}^{cpx}.a_{qz}}{a_{ab}^{plag}} = 0.36 \quad \text{(at 773 K)}.$$

Using the data of Robie et al. (1978) to calculate ΔV_S (= -0.4059 cal/bar), equation (7.81) can then be used to calculate the equilibrium curve. The intersection of this with the K_D curves (Fig. 7.1) gives a minimum temperature and pressure for the M_e event of 540°C and 11 kb.

(b) The M_v event

Pressure-temperature conditions of the M_v regional metamorphism can be esti-

mated crudely from the following M_v assemblages:

(i) gt + st + ky + qz (GD 652)

(ii) ky + Mg-biot + qz (C 523)

The lower limit of pressure is marked by the equilibrium:

$$\underset{\text{kyanite}}{Al_2SiO_5} = \underset{\text{sillimanite}}{Al_2SiO_5} \qquad (7.C)$$

This has been investigated by Richardson et al. (1968, 1969) and Anderson and Kleppa (1969) whose data are compared in Fig. 7.2.

Assuming that garnet and staurolite have similar Mg/Fe+Mg ratios in GD 652, the assemblage probably equilibrated to the low-temperature side of the reaction:

$$\underset{\text{staurolite}}{6Fe_4Al_{18}Si_{7.5}O_{44}(OH)_4} + \underset{\text{quartz}}{25SiO_2} = \underset{\text{garnet}}{8Fe_3Al_2Si_3O_{12}} + \underset{\text{kyanite}}{46Al_2SiO_5} + \underset{\text{fluid}}{12H_2O} \qquad (7.D)$$

calculated for K = 1, as X_{H_2O} may not have been unity.

The stability field of kyanite+biotite is bounded at low temperatures by:

$$\underset{\text{chlorite}}{3Mg_5Al_2Si_3O_{10}(OH)_8} + \underset{\text{muscovite}}{5KAl_3Si_3O_{10}(OH)_2} = \underset{\text{biotite}}{5KMg_3AlSi_3O_{10}(OH)_2} + \underset{\text{kyanite}}{8Al_2SiO_5}$$

$$+ \underset{\text{quartz}}{SiO_2} + \underset{\text{fluid}}{12H_2O} \qquad (7.E)$$

but expands with decreasing X_{H_2O}, so that some assumption about the fluid composition is necessary in order to apply this reaction.

The experimental reversals of Bird and Fawcett (1973) have been used to

extract the following data for (7.E) using the method of Fisher and Zen (1971)

$$\Delta_f H^o_s = 1081.3 \text{ kcal}. \Delta_f S^o_s = 736.4 \text{ cal } K^{-1}. \quad \Delta V_s = -4.7996 \text{ cal bar}^{-1}.$$

The equilibrium curve for this reaction has been calculated for K = 1.26 and $X_{H_2O} = 1.0$ and is plotted in Fig. 7.1. Thus, M_v temperatures during the M_v event were probably between 625°C and 710°C at pressures greater than 6 kb.

Incidentally, as chlorite, muscovite and some of the biotite in C 523 are Alpine, (7.E) can also be used as an M_a temperature maximum for the basement in the Kaponigtörl area (Fig. 7.21).

(c) The M_z event

During the intrusion of the Zentralgneis, M_v muscovites in the Inner Schieferhülle migmatite RN 247 were partially melted to produce sillimanite. M_z contact metamorphic conditions are therefore constrained to lie within the sillimanite stability field at temperatures above the water-saturated granite solidus and above the equilibrium:

$$\underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{quartz}}{SiO_2} + \underset{\text{vapour}}{H_2O} = \underset{\text{sillimanite}}{Al_2SiO_5} + \text{MELT} \qquad (7.G)$$

The experimental results of Storre and Karotke (1972) for reaction (7.G) indicate a minimum temperature of 730°C for the M_z event, and suggest that the pressure was probably between 6 and 8 kb (Fig. 7.1).

E. M_a METAMORPHIC CONDITIONS

In this section, thermodynamic principles are applied to post-D_a^2 equilibrium mineral assemblages to estimate temperature and pressure conditions at the thermal peak of the M_a metamorphic event (a) at very low structural levels in the Hochalm Kern, (b) near the basement-cover interface in the area of Map 2, and (c) in parts of the low-grade area west of Mallnitz.

The evidence from different lithologies will be discussed separately partly because it is convenient to treat equilibria in each chemical system together, and partly to emphasise differences between deduced X_{H_2O}/X_{CO_2} ratios of fluids in equilibrium with the assemblages in different lithologies.

(a) Pelites

Aluminosilicate equilibria

Kyanite is the only aluminosilicate found within the Tauern Window. The experimental results of Richardson et al. (1968; 1969) are similar to but not entirely consistent with those of Holdaway (1971) for the kyanite-andalusite inversion (Fig. 7.2), but the kyanite-sillimanite inversion curve of Anderson and Kleppa (1969) occurs at 0.5 to 1 kb lower pressures than that of Richardson et al. between 600°C and 800°C. On the basis of these experiments, the absolute lower limits of Alpine metamorphic pressure at 500°C and 600°C are 3.6 kb and 4.9 kb respectively.

Kyanite has been found at very high structural levels in the Peripheral Schieferhülle of the Glockner Depression (Frasl and Frank, 1966) suggesting that it was stable throughout the entire Tauern Window during the M_a peak.

If this is true, the pyrophyllite dehydration curve:

$$Al_2Si_4O_{10}(OH)_2 = Al_2SiO_5 + 2SiO_2 + H_2O \qquad (7.H)$$

pyrophyllite kyanite quartz fluid

forms a temperature minimum for the base of the Altkristallin sheet, assuming $P_{H_2O} = P_{tot}$. This equilibrium has been studied experimentally by Althaus (1966) and Kerrick (1968) but using andalusite instead of kyanite. Their results differ by about 100°C. Those of Althaus suggest that the reaction products are not stable below 550°C, which is rather high in view of the similar temperatures estimated for the basement-cover interface (see below). Accepting Kerrick's data for the moment, the minimum temperature at the peak of the M_a event is 430°C (Fig. 7.2).

ANaK phase relations

The following K_2O-Na_2O-Al_2O_3-SiO_2-H_2O phase assemblages appear to have been stable during Alpine metamorphism:

(1) musc + plag + qz

(2) musc + parag + plag + qz

(3) musc + parag $\pm$ qz

(4) musc + parag + ky + qz

(5) musc + ky + qz

(6) musc + parag + ky

The limiting assemblages (2) and (4) are compatible only if they belong to ANaK facies types 2b or 2c in the petrogenetic grid of Thompson and Thompson (1976). The lower limit of pressure for these facies types is defined by the stability limit of kyanite (see above), the upper limit of pressure by reaction (7.B), and the upper limit of temperature by the reaction:

$$NaAl_3Si_3O_{10}(OH)_2 + SiO_2 = NaAlSi_3O_8 + Al_2SiO_5 + H_2O \qquad (7.I)$$

mica quartz plagioclase kyanite fluid

Chatterjee (1972) has reversed this equilibrium in the pure system using high albite, and his results have been used to extract the following data for this reaction using the method of Fisher and Zen (1971):

$\Delta_f H^o_s = 83.15$ kcal. $\Delta_f S^o_s = 55.01$ cal K^{-1}. $\Delta V_s = -0.2437$ cal bar^{-1}.

These have been made consistent with Chatterjee's (1970) results for the breakdown of paragonite to albite and corundum (Fig. 7.3). Molar volumes of corundum, kyanite and quartz were taken from Robie et al. (1978), and that of paragonite from Chatterjee (1974b).

The equilibrium curve for reaction (7.I) has been calculated using mineral compositions in GD 208e (assemblage (2)), which was sampled from a structural height of $2\frac{1}{4}$ km below the basement-cover interface in the Hochalm Kern. As the equilibrium temperature at a given pressure decreases with decreasing X_{H_2O}, the maximum temperature for this assemblage is given by the condition $X_{H_2O} = 1.0$. The calculated curve is shown in Fig. 7.20.

Muscovite-paragonite thermometry

Pelites containing the assemblage muscovite+paragonite have been found at structural levels ranging from the deepest exposed levels of the Hochalm Kern to the lower limb of the Mallnitzer Mulde. The sodium and potassium contents of coexisting white micas have been used to estimate peak metamorphic temperatures at the various localities using the 2.07 kb solvus of Eugster el. al. (1972) (Fig. 7.4). The advantages of using a solvus as a geothermometer are two-fold: (a) each mineral pair gives two temperatures (one for each

limb of the solvus) which, if they agree, are good evidence that the minerals were in equilibrium; (b) equilibration temperatures are independent of fluid composition. The main disadvantage of this particular geothermometer is that its pressure sensitivity is not known (see part C(f))and must be assumed to be negligible, so that absolute temperature values may be considerably in error. On the other hand, calculated temperature differences are probably of the correct order of magnitude.

Mean Na/Na+K values and calculated temperatures for all analysed muscovites and paragonites in each of GD 18, GD 208e, GD 209a, GD 465b and GD 526 are listed in Table 7.1. Results are also tabulated for the high-Na muscovites in the paragonite-free schist C 523. In most schists, both micas show a large range in Na/Na+K values, giving rise to large ranges in calculated temperatures. tures. Very low temperatures could arise from retrogressive re-equilibration or failure to equilibrate with adjacent grains. Very high temperatures (up to 720°C) are common, particularly for paragonites. These could be caused by (a) high metamorphic temperatures (very unlikely as paragonite is no longer stable at such extremes), (b) the presence of finely intergrown muscovite-paragonite mixtures, or (c) the assumption that the thermometer is pressure-independent being wrong. If the muscovite-paragonite intergrowths are exsolution features, then very high temperatures are still implied, so (c) seems to be the most likely explanation.

The highest mean temperatures for paragonite (650°C, GD 465b) and muscovite (633°C, C 523 and 607°C, GD 465b) are from the deepest structural levels in the Hochalm Kern while the lowest mean temperature (460°C, GD 526 muscovite) is from the Mallnitzer Mulde. These trends are consistent with metamorphic zonation in pelites (Chapter 6).

The muscovite temperatures are considered the most reliable because (a) the muscovite limb of the solvus has a higher dX_{Na}/dT than the paragonite limb; (b) analyses of ? mixed layer muscovite-paragonite intergrowths could be easily distinguished by their high Ca-contents from characteristically Ca-poor single-phase muscovites, but not from single-phase paragonites,. as the latter contain significant Ca (Fig. 5.14); (c) the projection of paragonite cationic compositions from Ca on to the Na+K join is probably unjustified because the presence of muscovite-margarite and paragonite-margarite solvi in the ternary NaKCa system indicates that at any given temperature, Na/Na+K of paragonite in equilibrium with muscovite increases with increasing Ca/Ca+Na+K.

Eleven pairs of analyses were obtained for coexisting muscovite and paragonite grains in contact (Fig. 7.4). Of these, only three give temperatures that agree to within 30°C:

rock	paragonite		muscovite	
GD 208e	PAK 3	570°C	MUK 4	600°C
GD 208e	PAK 12	605°C	MUK 10	625°C
GD 465b	PAM 4	605°C	MUM 9	615°C

These agree well with the mean muscovite-limb temperatures for each rock.

Reactions involving FASH phases

In Chapter 6 it was established that the trace of the [as, cord] invariant assemblage in 3-D space intersects the erosion surface close to the structural base of the Möltal Schieferhülle. Harte (1975) estimates the position of this invariant point as ca. 600°C at 6 kb, but Atherton (1977) argues that it may be up to 40°C lower than this.

Experimental data indicate that natural staurolite+quartz (Hoschek, 1967) and pure Fe-staurolite+quartz (Richardson, 1968) are only stable with respect to aluminosilicate+chloritoid above ca. 540°C. As this reaction (7.L, 7.M) must occur at lower temperatures than any of the staurolite-producing isograd reactions in the Peripheral Schieferhülle (Fig. 6.16), 540°C can be considered a minimum temperature for the basement-cover interface in the area covered by Map 1.

The actual temperature of the staurolite in isograds can be estimated from the assemblage gt+st+ctd+qz in the garnet+staurolite+chlorite subzone pelite GD 308c. The reaction relating these phases, (7.O), is one of the many listed in Table 7.2 involving minerals in the system $FeO-Al_2O_3-SiO_2-H_2O$. These reactions were balanced using the 4:18:7.5 staurolite formula of Richardson (1968) and half a molecule of water per unit cell in cordierite. The latter falls within the range of water contents estimated by Newton (1972) for synthetic Mg-cordierite and by Weisbrod (1973) for synthetic Fe cordierite.

An attempt was made to extract a self-consistent data set for FASH reactions from experimental reversals using the method of Fisher and Zen (1971). However, as it was not possible to make all the curves pass through the relevant brackets, the preferred data set (Table 7.3) was chosen so that the equilibrium curves for reactions (7.M) and (7.D) were consistent with their experiments. The calculated curves are compared with the experimental data in Fig. 7.6.

The position of reaction (7.O) was recalculated for real mineral compositions in GD 308c (K = 0.025) assuming $X_{H_2O} = 1.0$. The curve has a large dP/dT and indicates an equilibrium temperature of between 564°C and 577°C in the range 5 to 10 kb (Fig. 7.19). If X_{H_2O} is reduced from 1.0 to 0.8, the equilibrium temperature only falls 8°C, so that unless H_2O is very dilute in the fluid,

the conclusions remain essentially the same.

Although chloritoid has broken down in the presence of quartz near the base of the Peripheral Schieferhülle, the presence of chloritoid inclusions in the garnets of staurolite+biotite-zone pelites indicates that reaction (7.J) was never overstepped. This can therefore be used as a temperature maximum for the basement-cover interface. Equilibrium temperatures calculated for (7.J) using K = 0.45 and X_{H_2O} = 1.0 are about 20°C higher than for the chloritoid+ quartz stability limit calculated above.

Geobarometry

Many equilibria in the system $CaO-Al_2O_3-SiO_2-H_2O$ have low dP/dT and can be used as geobarometers.

Coexisting garnet, epidote and plagioclase were analysed in pelitic schist GD 43a, garnetiferous paragneiss GD 145a and garbenschiefer GD 480a. In the first two, plagioclase appears to have been in equilibrium with other phases, but in GD 480a it has replaced mica. As all three rocks contain quartz, the reaction:

$$4Ca_2Al_3Si_3O_{12}(OH) + SiO_2 = Ca_3Al_2Si_3O_{12} + 5CaAl_2Si_2O_8 + 2H_2O \qquad (7.T)$$

epidote quartz garnet plagioclase fluid

can be used to estimate M_a peak metamorphic pressures. This reaction has been experimentally investigated for pure phases by Newton (1966) and Boettcher (1970), whose results are compatible, and Ackermand and Karl (1972) whose curve lies at lower temperatures. The curve used in subsequent calculations passes through the brackets of Newton and Boettcher (Fig. 7.7). Equation 7.7 was used to calculate the pressure correction on the equilibrium

position of (7.T) for impure mineral compositions substituting a ΔV_s value of 1.449 cal bar^{-1}.

Fig. 7.8 shows the position of this equilibrium calculated for $X_{H_2O} = 1.0$ and average mineral compositions in each rock. The curves for GD 145a and GD 43a agree to within 0.8 kb at any temperature below 650°C. Equilibrium pressures at 570°C are 7.0 kb and 6.3 kb, respectively. Strictly speaking, these should be considered pressure minima because P(equil) could have been greater if X_{H_2O} had been smaller. However, the coexistence of epidote (50-80% Al_2Fe end-member) and plagioclase (20-30% anorthite) without calcite suggest that the left-hand-side of reaction (7.AD) was stable (see part (d), below), which could only have been possible if X_{H_2O} were greater than 0.9. At $X_{H_2O} = 0.9$, the equilibrium pressures for reaction (7.T) are only 0.2 kb higher than at $X_{H_2O} = 1.0$. Changing the value of γ_{an}^{plag} also has very little effect. If Saxena and Ribbe's (1972) value of 1.5 is used instead of Orville's (1972) value of 1.28, equilibrium pressures are lowered by about 0.5 kb.

Also shown in Fig. 7.8 are equilibrium curves calculated for different groups of mutually contacting mineral grains in GD 145a. These indicate a within-sample error of $\pm$1 kb on the pressure calculated from average mineral compositions. Originally, the author's intention was to use reaction (7.T) to measure the pressure gradient along isograds in the Mölltal Schieferhülle (see Chapter 6). However, as any such pressure differences are likely to be much smaller than the error, not even the sense of the gradient can be determined with confidence.

In GD 490a, margarite appears to have been replaced by plagioclase, particularly near quartz grains. One reaction that could account for this texture is:

$$2Ca_2Al_3Si_3O_{12}(OH) + CaAl_4Si_2O_{10}(OH)_2 + 2SiO_2 = 5CaAl_2Si_2O_8 + 2H_2O \quad (7.U)$$

epidote mica quartz plagioclase fluid

The position of this equilibrium has been calculated using the thermodynamic data of Holland (1977). The calculated curves (pure phases, $X_{H_2O} = 1.0$) for some reactions involving margarite are compared with experimental data in Fig. 7.7. The reaction:

$$4CaAl_4Si_2O_{10}(OH)_2 + 3SiO_2 = 2Ca_2Al_3Si_3O_{12}(OH) + 5Al_2SiO_5 + 3H_2O \quad (7.V)$$

mica quartz epidote kyanite fluid

cannot be made consistent with the data of Storre and Nitsch (1974) without making other reactions inconsistent. The calculated curve has a low negative dP/dT, while the experimental reversals suggest that it is extremely high. The calculations of Chatterjee (1976) also predict a back-sloping equilibrium curve.

The equilibrium position of (7.U) for $X_{H_2O} = 1$ and K = 0.365 (mineral compositions in GD 480a) suggests a pressure maximum of 7.8 kb at 570°C for the products. Little confidence can be placed on this value because (a) P(equil) increases with decreasing X_{H_2O}, and (b) the assemblage may never have been in equilibrium.

Several pelitic schists contain M_a garnet, plagioclase and sodic white-mica, apparently in equilibrium. The following reaction can be written between these phases:

$$Ca_3Al_2Si_3O_{12} + 2NaAl_3Si_3O_{10}(OH)_2 + 3SiO_2 = 3CaAl_2Si_2O_8 + 2NaAlSi_3O_8 + 2H_2O \quad (7.W)$$

garnet mica quartz plagioclase fluid

The equilibrium curve for pure phases was calculated in two ways:

(i) Linear combination of reaction (7.I) with:

$$Ca_3Al_2Si_3O_{12} + 2Al_2SiO_5 + SiO_2 = 3CaAl_2Si_2O_8 \qquad (7.X)$$

grossular kyanite quartz anorthite

The equilibrium curve of Schmidt et al. (1978) for (7.X) passes through the experimental reversals of both Hays (1967) and Hariya and Kennedy (1968), and was used to extract the following data using the method of Fisher and Zen (1971):

$$\Delta_f H^o_s = 9.39 \text{ k cal.} \quad \Delta_f S^o_s = 35.6 \text{ cal } K^{-1}. \quad \Delta V_s = 1.5823 \text{ cal bar}^{-1}$$

which, when combined with data from (7.I) yield:

$$\Delta_f H^o_s = 175.69 \text{ k cal.} \quad \Delta_f S^o_s = 145.62 \text{ cal } K^{-1}. \quad \Delta V_s = 1.1121 \text{ cal bar}^{-1}$$

for reaction (7.W).

(ii) Linear combination of the following reactions:

$$Ca_3Al_2Si_3O_{12} + 2KAl_3Si_3O_{10}(OH)_2 + 3SiO_2 = 3CaAl_2Si_2O_8 + 2KAlSi_3O_8 + 2H_2O \qquad (7.Y)$$

grossular muscovite quartz anorthite sanidine fluid

$$NaAl_3Si_3O_{10}(OH)_2 = NaAlSi_3O_8 + Al_2O_3 + H_2O \qquad (7.Z)$$

paragonite albite corundum fluid

and

$$KAl_3Si_3O_{10}(OH)_2 = KAlSi_3O_8 + Al_2O_3 + H_2O \qquad (7.AA)$$

muscovite sanidine corundum fluid

The following data were extracted from the experiments of Ackermand and Karl (1972) for (7.Y), Chatterjee (1970) for (7.Z) and Chatterjee and Johannes (1974) for (7.AA):

(7.Y) : $\Delta_f H^o_s$ = 139.54 k cal. $\Delta_f S^o_s$ = 91.86 cal K^{-1}. ΔV_s = 1.0919 cal bar^{-1}.

(7.Z) : $\Delta_f H^o_s$ = 85.02 k cal. $\Delta_f S^o_s$ = 57.11 cal K^{-1}. ΔV_s = -0.1442 cal bar^{-1}.

(7.AA): $\Delta_f H^o_s$ = 83.95 k cal. $\Delta_f S^o_s$ = 53.68 cal K^{-1}. ΔV_s = -0.1543 cal bar^{-1}.

The relationship (7.W) = (7.Y) + 2(7.Z) - 2(7.AA) was used to derive the following data for (7.W):

$\Delta_f H^o_s$ = 141.68 k cal. $\Delta_f S^o_s$ = 98.72 cal K^{-1}. ΔV_s = 1.1121 cal bar^{-1}.

The source equilibria are compared with experimental reversals in Fig. 7.3. Also shown here are the two calculated pure-phase equilibrium curves for (7.W); the agreement between them is poor as regards both equilibrium pressure, in the temperature range of interest, and slope.

The equilibrium position of (7.W) has been calculated for the assemblage musc+gt+plag+qz in GD 145a, and for parag+gt+plag+qz in GD 208e using both methods. Curves calculated using method (ii) correspond closely to those for reaction (7.T) and yield equilibrium pressures between 6 and 7 kb in the range 570°C to 600°C (Figs. 7.19 and 7.20), but method (i) gives pressures of between 9 and 11 kb in this range, which are unreasonably high compared with the other estimates.

Estimates of X_{H_2O}

In Chapter 6 part B(c), it was suggested that the presence of kyanite in staurolite+biotite zone pelites might be caused by low X_{H_2O} in the fluid. Now that some estimate has been made of M_a peak metamorphic conditions (ca. 570°C, 7 kb at the basement-cover interface in the Möltal) it should be possible to calculate the X_{H_2O} for which reaction (7.D) is in equilibrium

under these conditions. A value of K = 0.01 (7.D) was derived from analyses of coexisting garnet and staurolite in GD 18. Fig. 7.9 shows a projection, into P-T-space, of the calculated equilibrium surface contoured for values of X_{H_2O}. At 570°C and 7 kb, the assemblage in GD 18 appears to have equilibrated with a fluid of X_{H_2O} = 0.4. This low value might be explained by the close proximity (ca. 5 to 10 cm) of the sample locality to calc-schists.

Higher values of X_{H_2O} are indicated for the garnet-staurolite-chlorite subzone pelite GD 308c. The 7 kb temperature, 570°C, calculated from reaction (7.0) corresponds closely to the X_{H_2O} - independent mean muscovite temperature, 570°C, from GD 18 and the calcite-dolomite-solvus temperatures, 560°C and 580°C, from GD 22a (see part (c) below). However, had X_{H_2O} in GD 308c been lower than about 0.7, these temperatures would not agree so well.

The assemblages parag+plag+qz and parag+ky+qz in Inner Schieferhülle pelites (for which temperatures of 580°C to 600°C have been estimated) also indicate high X_{H_2O} values. At 590°C and 7 kb, X_{H_2O} of the fluid in GD 208e must have been greater than 0.7 otherwise the reactants of reaction (7.I), paragonite and quartz, would not have been stable (Fig. 7.20).

(b) Zentralgneis lithologies

Syenite-gneiss

The pseudomorph-bearing syenite-gneisses and pyroxene-biotite-schist contain the M_a assemblage cpx+ab+qz apparently in equilibrium (Figs. 2.27, 2.28). Reaction (7.B) can therefore be used to estimate M_a pressure conditions in the Romate subarea.

The equilibrium curve for (7.B) has been calculated for a typical clinopyroxene composition of jd_7acm_5 $(di+hed)_{88}$ using a variety of different activity-composition relationships:

(i) If equation (7.20) (which implies a negative deviation from ideality at low temperatures) is used, the curve gives pressures of 2-3 kb at temperatures between 450°C and 500°C (curve A, Fig. 7.18), well within the stability field of andalusite.

(ii) If mixing in clinopyroxene is assumed to be ideal, i.e. $a_{jd} = x_{jd}$, the calculated curve (B in Fig. 7.18) plots close to the andalusite-kyanite boundary.

(iii) If mixing in clinopyroxene is assumed to show the maximum positive deviation from ideality permitted by equation (7.20), i.e. $W_{12} = 0 + 1.16T$ (°C), the equilibrium curve (C. in Fig. 7.18) falls within the stability field of kyanite.

(iv) If the equation of Currie and Curtis (1976) is used, negative pressures are obtained for temperatures around 500°C.

As kyanite is the only M_a aluminosilicate found in the Tauern, (iii) is considered the best estimate: 5 to 5.4 kb at 450 to 500°C. The jadeite contents of the clinopyroxenes in the syenite-gneiss mostly fall between 5 and 9%, giving a within-sample error of about $\pm$1 kb.

Albite (or_1) and microcline (or_{90}) coexist in the pyroxene-biotite-schist GD 1234b. These compositions therefore lie on the limbs of the low albite-microcline solvus, and the sodium content of the microcline can be used as

a geothermometer.

An equilibrium curve (Fig. 7.18) has been calculated for the composition $or_{90}ab_{10}$ by substituting equations (7.26) and (7.27) into (A.75) and solving for temperature graphically at various pressures. These temperatures agree well with the estimate of 460°C based on the sodium content of muscovite in GD 526 (Table 7.1).

The M_a peak metamorphic conditions in the Romate-decke are estimated as 5 ± 1 kb and 450-500°C.

The assemblage sph+cc+qz (without rutile) provides an estimate of the upper limit of CO_2 content of the fluid phase, as these minerals are related by the reaction:

$$\underset{\text{rutile}}{TiO_2} + \underset{\text{calcite}}{CaCO_3} + \underset{\text{quartz}}{SiO_2} = \underset{\text{sphene}}{CaTiSiO_5} + \underset{\text{fluid}}{CO_2} \qquad (7.AB)$$

The experimental data of Hunt and Kerrick (1977) have been used to extract values of $\Delta_f H^\circ_s$ (112.25 k cal) and $\Delta_f S^\circ_s$ (36.54 cal K^{-1}) for this reaction making the Fisher and Zen approximation and using a ΔV_s value of -0.5447 cal bar^{-1} (Robie et al., 1978). These have been used to construct an isobaric (5 kb) T - X_{CO_2} diagram (Fig. 7.10) which indicates that at temperatures between 450°C and 500°C, sphene can only form if $X_{CO_2} < 0.1$. The data of Schuiling and Vink (1967) predict an even smaller sphene stability field.

Meta-granites

Minerals were only analysed in one granitic rock, leucogranite GD 468 from the Gössgraben, which contains the assemblage gt+musc+ab+olig+ksp.

Ackermand and Karl (1972) have reversed some equilibria involving grossular, zoisite, muscovite, K-feldspar, anorthite and quartz. The zoisite-absent reaction (7.Y) can be used to give an M_a pressure minimum for the lowest structural level exposed in the south-east Tauern. The equilibrium curve for pure phases is tightly bracketed in the temperature range of interest and has an error of only $\pm$0.2 kb (Fig. 7.3).

The equilibrium curve for impure phases has been calculated for a range in X_{H_2O} values. At 610°C (the temperature estimated from muscovite-paragonite geothermometry) the curve for pure water gives a pressure of 5.8 kb. This can be considered a minimum value as lowering the water activity has the effect of increasing the equilibrium pressure. If the leucogranite assemblage crystallised at 610°C and 7 kb, as suggested by other equilibria (see above), X_{H_2O} would have been roughly 0.5. The common occurrence of clinozoisite in calcite-free oligoclase-bearing meta-granites and meta-tonalites suggests that the fluid is generally much richer in H_2O than this, as the assemblage epidote+plagioclase is stable only on the high X_{H_2O} side of reaction curve (7.AD) in Fig. 7.14. This conclusion is supported by the fact that sphene is the characteristic Ti-bearing phase in the Zentralgneis (Norris et al., 1971a) while rutile has not been recorded.

The K-feldspar in GD 468 has a composition almost identical to that in the syenite-gneiss, for which a low albite-microcline solvus temperature of 480°C was calculated for 6 kb. This is unacceptably low in view of other estimates.

Phase relations

Phase relations in albite- and quartz-bearing meta-granites and meta-syenites are plotted in a K - (Al + Fe^{3+}) - Ca - (Mg + Fe^{3+}) tetrahedron in Fig. 7.11.

Stable assemblages include:

biot + musc + ksp biot + musc + ep + ksp	} augen-gneisses
biot + ep + ksp cpx + biot + ep + ksp	} syenite-gneisses
cpx + biot + ksp	pyroxene-biotite-schist

As mentioned in Chapter 2 part C(c), the association cpx+musc does not occur in the area of study, and appears to be unstable with respect to biot+ep+ksp as shown by the reaction:

$$12CaMgSi_2O_6 + 9KAl_3Si_3O_{10}(OH)_2 \qquad (7.AC)$$

clinopyroxene white mica

$$= 4KMg_3AlSi_3O_{10}(OH)_2 + 5KAlSi_3O_8 + 6Ca_2Al_3Si_3O_{12}(OH) + 6SiO_2 + 2H_2O$$

biotite K-feldspar epidote quartz fluid

Thus, clinopyroxene is restricted to the Al-poor syenite-gneisses and related rocks, but is absent from the more aluminous meta-granites.

The equilibrium curve for reaction (7.AC) has been calculated using known mineral compositions and assumed values of 0.4 for the activity of muscovite and 1.0 for X_{H_2O}. Diopsidic clinopyroxene and potassic white mica appear to be compatible only at high pressures and low temperatures so that this curve can be used as (i) a pressure maximum (9 kb at 500°C) for the syenite-gneiss assemblage (Fig. 7.18), and (ii) a pressure minimum or temperature maximum for phengite-bearing eclogites. The syenite-gneiss assemblage is therefore not isofacial with the cpx+pheng+ksp+qz meta-granites of the Sesia-Zone (Compagnoni and Maffeo, 1974).

The leucogranite assemblage garnet+K-feldspar would be incompatible with the

more common assemblage biotite+muscovite+epidote if the garnet could be projected into the K -(Al + Fe^{3+}) - Ca - (Mg + Fe^{2+}) system, but as the garnet contains 26% spessartine component, projection is not justified.

(c) Calc-schists and metabasites

Calcite-dolomite geothermometry

The calcite-dolomite geothermometer of Bickle and Powell (1977) has been used to estimate M_a peak metamorphic temperatures for the two rocks in which co-existing carbonates were analysed: calc-schist GD 992 from the Mallnitzer Mulde (Map 2) and epidote-amphibolite GD 22a from near the basement-cover interface in Zwenberger Tal (Map 1). The results are shown in Fig. 7.12 for the calcite limb of the solvus, and in Fig. 7.13 for the dolomite limb, for a pressure of 5 kb. The thermometer is very insensitive to pressure, and the temperatures are only 6°C lower at 7 kb.

Pairs of calcite and dolomite grains, mutually in contact, give a large spread of temperatures for both rocks. In Fig. 7.12, however, both populations show distinctly linear trends towards the origin, suggesting that the spread is more likely to be due to retrogressive re-equilibration than to errors in measurement. For this reason, the maximum temperature in each population is considered the best estimate for the M_a thermal peak. Temperatures from the two limbs of the solvus agree to within 10°C for each rock. At 7 kb M_a peak metamorphic temperatures in the Mallnitzer Mulde and mid-Zwenberger Tal are estimated as 520 to 540°C and 560 to 580°C respectively. The error of the thermometer is quoted as $\pm$30°C.

Other temperature and X_{H_2O} estimates

All calc-schists contain calcite, epidote, muscovite and quartz, and many also contain andesine. No silicate minerals in calc-schists were analysed in this study, so that the conclusions of this section are only semi-quantitative. Some experimentally determined equilibria relevant to calc-schists are shown in an isobaric T-X_{CO_2} section in Fig. 7.14.

Plagioclase-bearing calc-schists must have crystallised on the equilibrium curve of the reaction:

$$2Ca_2Al_3Si_3O_{12}(OH) + CO_2 = 3CaAl_2Si_2O_8 + CaCO_3 + H_2O \qquad (7.AD)$$

epidote fluid plagioclase calcite fluid

At 7 kb and at temperatures above 550°C, pure zoisite is only stable with respect to anorthite and calcite in the presence of an aqueous fluid with more than 85 mole percent H_2O (Johannes and Orville, 1972). At lower pressures, zoisite is restricted to yet more H_2O-rich fluids (Storre and Nitsch, 1972). Although the feldspars of the calc-schists may contain up to 50% anorthite component, the epidotes appear (an optical grounds) to contain very little iron, so that the equilibrium constant for reaction (7.AD), as it is written above, is likely to be less than unity, which will have the effect of moving the curve to more H_2O-rich fluids. Plagioclase-free calc-schists are likely to be stable only on the high X_{H_2O} side of this curve.

The common occurrence of sphene in calc-schist is consistent with the presence of an H_2O-rich fluid. However, two samples were found containing rutile, calcite, quartz and epidote. The stability field of this assemblage is bounded by equilibria (7.AB) and (7.AD) at temperatures lower than 540°C (Fig. 7.14). One of the samples yields 530°C to 550°C by calcite-dolomite thermometry (see above). At 7 kb, the fluid in equilibrium with this rock probably contained approximately 17 mole percent H_2O.

Most calc-schists contain graphite. French (1966) concluded that the maximum possible P_{H_2O} in a gas in equilibrium with graphite above 500°C and 1 kb is 0.5 P_{tot}. This is inconsistent with the above conclusions.

Many of the Peripheral Schieferhülle metabasites contain epidote, calcite and plagioclase. In some (e.g. GD 482, Fig. 6.23) reaction (7.AD) was obviously proceesing from left to right, i.e. towards more CO_2-rich fluid compositions, but in others these minerals appear to be in equilibrium. In GD 551, the equilibrium constant for reaction (7.AD) is 0.07, supporting the assumption made in Chapter 6 that dolomite-free calcite-bearing metabasites can equilibrate with very H_2O-rich fluids.

The reaction:

$$\underset{\text{white mica}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{calcite}}{4CaCO_3} + \underset{\text{quartz}}{6SiO_2} = \underset{\text{epidote}}{2Ca_2Al_3Si_3O_{12}(OH)} + \underset{\text{K-feldspar}}{3KAlSi_3O_8}$$

$$+ \underset{\text{fluid}}{4CO_2 + 2H_2O} \qquad (7.AE)$$

defines the maximum X_{H_2O} in equilibrium with muscovite and calcite. The intersection of this curve with that of reaction (7.AD) provides a temperature maximum for calc-schists. The experiments of Johannes and Orville (1972) suggest that, at 7 kb, this intersection (the invariant point zo+musc+an+ksp+cc) occurs at 570°C and at X_{CO_2} = 0.17 for pure phases (Fig. 7.14).

The reaction:

$$\underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{calcite}}{CaCO_3} + \underset{\text{quartz}}{2SiO_2} = \underset{\text{orthoclase}}{KAlSi_3O_8} + \underset{\text{anorthite}}{CaAl_2Si_2O_8} + \underset{\text{fluid}}{CO_2 + H_2O} \qquad (7.AF)$$

has been investigated by Hewitt (1973) whose data have been used to extract

values of $\Delta_f H^o_s$ (196.16 k cal) and $\Delta_f S^o_s$ (95.87 cal K^{-1}) making the Fisher and Zen approximation. At 7 kb, the calculated equilibrium temperatures are 15°C higher than those predicted by the data of Johannes and Orville.

(d) Marbles and siliceous dolomites

The stable assemblages in the siliceous dolomites of the Peripheral Schieferhülle are:

(1) di + cc + qz + ep

(2) di + trem + cc + ep

(3) trem + cc + dol (Cliff et al., 1971)

(4) trem + cc + qz + ep

(5) trem + cc + qz + chl

(6) trem + dol + qz + chl

In addition, some micaceous marbles contain the assemblages:

(7) cc + chl + qz $\pm$ ep

(8) dol $\pm$ cc + qz + ep

With the exception of (2), all of these have been found close to the structural base of the lowest calc-schist unit in the Mölltal Schieferhülle. Assemblages (1) and (5) occur along strike from one another in the Sickerkopf shear zone, and (5) and (6) occur in alternate bands in GD 230a (Fig. 2.37) from nearby.

Considering only the CMS (CaO-MgO-SiO_2-fluid) system, i.e. neglecting the aluminous phases epidote and chlorite, three mutually incompatible groups of assemblages can be recognised: A, (1) and (2); B, (3), (4) and (5); C,

(6) and (8). The reaction separating groups A and B is:

$$Ca_2Mg_5Si_8O_{22}(OH)_2 + 3CaCO_3 + 2SiO_2 = 5CaMgSi_2O_6 + 3CO_2 + H_2O \qquad (7.AG)$$

tremolite calcite quartz diopside fluid

and the reaction separating groups B and C is:

$$5CaMg(CO_3)_2 + 8SiO_2 + H_2O = Ca_2Mg_5Si_8O_{22}(OH)_2 + 3CaCo_3 + 7CO_2 \qquad (7.AH)$$

dolomite quartz fluid tremolite calcite fluid

The thermodynamic data of Powell (1973) were used to construct a 6 kb, isobaric T-X_{CO_2} section for CMS-sluid univariant reactions around the [tc, fo] and [di, fo] invariant points (Fig. 7.15). The result is essentially the same as the 5 kb experimental data of Slaughter et al. (1975). The main difference is that the calculated position of [di, fo] is at X_{CO_2} = 0.06 while the experimental position is at X_{CO_2} = 0.2. The experiments of Skippen (1971) and Metz and Trommsdorff (1968) indicate that [di, fo] occurs at very high X_{CO_2} (0.95) at 1 kb, suggesting that the difference in pressure may be partly responsible for this discrepancy.

Since the above assemblages have been found closely associated in space, it is unreasonable to explain the incomaptibility between groups A, B and C in terms of difference in temperature and pressure. However, at any temperature between 485°C and 620°C (the calculated temperatures of [di, fo] and [tc, fo], respectively*), the incompatibilities can be explained by differences in the X_{H_2O}/X_{CO_2} ratio of the coexisting fluid phase. At a given temperature, this ratio must have been larger for group A than for group B, and larger for B than for C. This relationship is not altered by the presence of impurities in the minerals concerned, provided that the composition of each mineral is approximately the same in each assemblage.

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*The data of Slaughter et al. suggest temperatures of 460°C and 625°C for these invariant points.

The equilibrium position of reaction (7.AG) has been calculated for real diopside and tremolite compositions in GD 160a, and differs only slightly from the pure-phases curve (Fig. 7.15). The tremolite appears to be retrogressive so that the peak M_a assemblage, (1), crystallised on the high temperature side of this curve. If T_{max} was 570°C, X_{CO_2} was probably less than o.12, and may have been buffered to even lower values along the curve during retrogression (W-X in Fig. 7.15). Similarly, the occurrence of talc as an alteration product of tremolite in rocks with equilibrium assemblage (6) suggests that the fluid composition was buffered along the path Y-Z (Fig. 7.15) during retrogression.

The position of reaction (7.AG) was also calculated for the assemblage clinopyroxene, calcite, quartz and retrogressive hornblende in the syenite-gneiss assuming a tremolite activity of 0.002 and a calcite activity of 1.0. The primary assemblage is restricted to X_{CO_2} values less than 0.02 below 500°C (Fig. 7.10).

Also shown in Fig. 7.15 are the calculated curves for reactions involving chlorite (clinochlore), zoisite, tremolite, calcite, dolomite and quartz. These reactions are essentially the same as those shown in Fig. 6.20 except that they have been balanced for pure phases. The compatibility fields in $T\text{-}X_{CO_2}$ space can be further subdivided so that X_{H_2O}/X_{CO_2} ratios of fluids in equilibrium with the above assemblages increase in the order (8) < (8), (6) < (6), (7) < (3), (5), (7) < (3), (4) < (1), (2), (3) at any given temperature. However, for <u>all</u> the assemblages to have formed under isobaric (6 kb) isothermal conditions, temperatures must have been between 540°C and 580°C. The position of the trem+ep+chl+cc + dol + qz invariant point, which forms the upper temperature limit, only increases 10-15°C for every 1 kb pressure increment (Fig. 7.15). These limits are in remarkably good agreement with other estimates, despite the fact that no account has been taken of the

the effect of impurities on the positions of the equilibria.

(e) Serpentinites

Serpentinites occur throughout the Peripheral Schieferhülle and at high structural levels in the basement complex (Angel and Staber, 1952; Norris et al., 1971a). The serpentine dehydration reaction:

$$5Mg_3Si_2O_5(OH)_4 = 6Mg_2SiO_4 + Mg_3Si_4O_{10}(OH)_2 + 9H_2O \qquad (7.AI)$$

serpentine olivine talc fluid

can therefore be used as a temperature maximum for the basement-cover interface.

Reaction 7.AI has been studied experimentally in the pure system by Bowen and Tuttle (1949) (using chrysotile) and Kitahara et al. (1966), whose data suggest that serpentine is not stable above 500°C (1 to 3 kb) and 580°C (10 kb) respectively. More recently, Evans et al. (1976) reversed the same reaction using antigorite. Their data indicate slightly higher dehydration temperatures which are more consistent with other temperature estimates from this study. As antigorite appears to be the characteristic mineral of the Tauern serpentinites, this curve is used as a temperature maximum (ca. 600°C at 7 kb) for the base of the cover-sequence (Fig. 7.19).

Equilibria in the system $MgO-SiO_2-H_2O-CO_2$ involving serpentine, forsterite, talc and magnesite have been experimentally investigated at pressures up to 7 kb by Johannes (1967; 1969) who concludes that serpentine is stable only in the presence of extremely H_2O-rich fluid (Fig. 7.16). For values of X_{CO_2} greater than 0.02 (2 kb), 0.08 (7 kb) serpentine is replaced by talc and magnesite:

$$2Mg_3Si_2O_5(OH)_4 + 3CO_2 = Mg_3Si_4O_{10}(OH)_2 + 3MgCO_3 + 3H_2O \qquad (7.AJ)$$

serpentine fluid talc magnesite fluid

However, most of the serpentinites studied contain CaO in addition to the components mentioned above, and the carbonate, where present, is invariably dolomite. The following critical assemblages were found:

(1) serp + tc + dol

(2) serp + trem + dol

(3) trem + tc + dol

Assemblage (2) is incompatible with (1) and (3) and is related to them by the reaction:

$$2Mg_3Si_2O_5(OH)_4 + Ca_2Mg_5Si_8O_{22}(OH)_2 + 4CO_2 = 2CaMg(CO_3)_2 + 3Mg_3Si_4O_{10}(OH)_2 \qquad (7.AK)$$
$$+ 2H_2O$$

serpentine tremolite fluid dolomite talc fluid

The occurrence of the above assemblages indicates that one of the invariant points [di, fo, qz] and [di, fo , cc] is stable. If [di, fo, cc] were stable, it would lie on the stable extension of the reaction (di, fo, cc, serp):

$$Mg_3Si_4O_{10}(OH)_2 + 2CaMg(CO_3)_2 + 4SiO_2 = Ca_2Mg_5Si_8O_{22}(OH)_2 + 4CO_2 \qquad (7.AL)$$

talc dolomite quartz tremolite fluid

However, this is very unlikely becuase reaction (7.AL) extends from [di, fo, serp] towards higher X_{CO_2} (Figs. 7.15 and 7.17) so that [di, fo, cc] would then lie outside the stability field of serpentine as deduced by Johannes (1969). If, however, [di, fo, qz] were stable, it would lie on the stable extension of the reaction (di, fo, qz, serp):

$$2Mg_3Si_4O_{10}(OH)_2 + 3CaCO_3 = Ca_2Mg_5Si_8O_{22}(OH)_2 + CaMg(CO_3)_2 + H_2O + CO_2 \quad (7.AM)$$
talc calcite tremolite dolomite fluid

and hence on the high X_{H_2O} side of [di, fo, serp] and at lower temperatures, i.e. well within the stability field of serpentine. This is considered much more likely and is supported by calculations based on the data set of Helgeson (1976). The preferred topology of the T-X_{CO_2} grid is shown in Fig. 7.17. Thus, at any given temperature, the fluid in equilibrium with (2) has a higher X_{H_2O}/X_{CO_2} ratio than one in equilibrium with (1) or (3). The reaction rims of talc round tremolite in GD 868 (assemblage (2), Fig. 2.39) could have been formed either by falling temperature or by an isothermal increase in X_{CO_2}.

F. EARLY ALPINE METAMORPHIC CONDITIONS

The reaction history of the lawsonite-pseudomorph described in Chapter 4, part C(c) can be used to speculate about the PT trajectory during the early part of the Alpine metamorphic cycle. The pseudomorph, now composed mainly of clinozoisite and margarite with minor quartz and plagioclase, probably formed directly from lawsonite by the reaction:

$$5CaAl_2Si_2O_7(OH)_2.H_2O = 2Ca_2Al_3Si_3O_{12}(OH) + CaAl_4Si_2O_{10}(OH)_2 + 2SiO_2 + 8H_2O \quad (7.AN)$$
lawsonite epidote margarite quartz fluid

However, it is also possible that the lawsonite initially decomposed by the reaction:

$$4CaAl_2Si_2O_7(OH)_2.H_2O = 2Ca_2Al_3Si_3O_{12}(OH) + Al_2SiO_5 + SiO_2 + 8H_2O \quad (7.AO)$$
lawsonite epidote kyanite quartz fluid

Were this the case, it would be reasonable to expect some relict kyanite

to remain (even though kyanite and zoisite could have reacted subsequently to produce margarite and quartz by reaction (7.V)), because kyanite is a fairly unreactive mineral. Also (7.V) would have involved re-hydration (i.e. re-introduction of water from the rock matrix through the enclosing garnet into the inclusion). Thus, assuming that the absence of kyanite indicates that kyanite never formed, the inclusion must have followed a PT trajectory originating on the low-temperature side of (7.AN) crossing the stability field of epidote+margarite+quartz to the M_a peak conditions on, or to the high-temperature side of, (7.U), i.e. the reaction (7.V) forms the upper pressure limit of the trajectory. Some possible trajectories are shown in Fig. 7.22, together with equilibrium curves for (7.U) and (7.V) (calculated for minerals in GD 480a) and the experimental reversals of Nitsch (1973) for (7.AN). A temperature increase of at least 120°C (probably closer to 150°C) separates the stability field of lawsonite from the calculated M_a peak metamorphic conditions, and a maximum pressure drop of 4 kb between lawsonite-breakdown and the M_a peak is indicated. These data suggest that the minimum temperature increase and maximum pressure drop between lawsonite breakdown and the calculated M_a metamorphic peak were 120°C and 4 kb, respectively.

Evidence from zonation in sodic amphiboles from the central Tauern (Holland, 1977; Holland and Richardson, in press) suggests that pressures during the eo-Alpine blueschist event were in the region 10 to 12 kb. The present data are compatible with these estimates, but would also be compatible with much lower early Alpine pressures (Fig. 7.22).

G. METASOMATIC ROCKS

In the preceding sections it has been convenient to assume that H_2O and CO_2

were the only components to which the rock systems could be open. Indeed the low variance of many assemblages suggests that even these were not perfectly mobile. The rocks described below provide evidence of open system behaviour, at least locally, of other major components during regional metamorphism.

(a) Serpentinites

The tremolite+dolomite+talc schists marginal to, and occupying shear-zones within, serpentinite bodies (Chapter 2, part D(b)) have evidently been derived from originally calcium-free ultrabasic rocks through the metasomatic introduction of calcium and silicon. This is illustrated by Fig. 7.23 which shows the stability limits of serpentine, talc, dolomite and tremolite in $\mu_{CaO} - \mu_{SiO_2}$ - space. The two possible configurations, (a) and (b) apply over different ranges of X_{CO_2} and are related by reaction (7.AK). The values of μ_{CaO} and μ_{SiO_2} which define the common schist assemblages trem+dol+tc (point X in Fig. 7.23(a)), and trem+tc (line X-Y) are greater than those under which serpentine can form.

The obvious source of the introduced CaO and SiO_2 is the calc-schists which normally enclose the serpentinite bodies. Values of μ_{CaO} and μ_{SiO_2} are likely to have been very high in calc-schists, which contain free calcite and quartz, in comparison with unaltered serpentinite which contains neither calcic phases nor quartz.

(b) Pelites and metagranites

The replacement of muscovite by ragged porphyroblasts of albite or oligoclase

is very common in pelitic and semi-pelitic schists of the eastern Tauern (see Chapter 2 parts B(f), D(h) and Chapter 4 parts D(b) and D(c)). This texture can be explained by the generalised reaction:

$$\underbrace{a\ Na^{+} + b\ Ca^{2+} + c\ H_5SiO_4^{+}}_{\text{fluid}} + d\ KAl_3Si_3O_{10}(OH)_2 \longrightarrow \underbrace{a\ NaAlSi_3O_8 + b\ CaAl_2Si_2O_8}_{\text{plagioclase}} + \underbrace{d\ K^{+} + e\ H^{+} + f\ H_2O}_{\text{fluid}} \qquad (7.AP)$$

More specifically, assuming that the mica composition is $musc_{75}parag_{25}$ (V_s = 3.31 cal bar^{-1}), that the plagioclase composition is $an_{30}ab_{70}$ (V_s = 2.39 cal bar^{-1}) and that replacement occurred on a constant volume basis, the reaction can be balanced thus:

$$\underbrace{0.717\ Na^{+} + 0.415\ Ca^{2+} + 0.731\ H_5SiO_4^{+}}_{\text{fluid}} + K_{0.75}Na_{0.25}Al_3Si_3O_{10}(OH)_2 \longrightarrow \underbrace{1.382\ Na_{0.7}Ca_{0.3}Al_{1.3}Si_{2.7}O_8}_{\text{plagioclase}} + \underbrace{1.203\ Al(OH)_2^{+} + 0.75\ K^{+} + 0.325\ H^{+} + 1.462\ H_2O}_{\text{fluid}} \qquad (7.AQ)$$

Although most schists contain a potential source of $H_5SiO_4^{+}$ ions in the form of free quartz, very few contain sufficient quantities of sodic and calcic minerals to provide the necessary Na^{+} and Ca^{2+} ions. Some schists contain as much as 30 mode percent of post-D_a^2 plagioclase, but only accessory epidote and pre-D_a^2 albite showing no evidence of resorption. It must be concluded, therefore, that the Na^{+} and Ca^{2+} ions were introduced from elsewhere, presumably via the fluid phase. Norris et al. (1971c, p. 187) also came to this conclusion in connection with the schist C 523 described in Chapter 2, part B(f).

The replacement of K-feldspar by myrmekitic intergrowths of plagioclase and

quartz in poorly deformed meta-granites is chemically analogous to the reactions discussed above. However, these rocks present a much less convincing case for metasomatism because a potential source for the necessary Na^+ and Ca^{2+} ions exists in the form of abundant plagioclase.

From the work of Wintsch (1975) it can be seen that replacement of muscovite or K-feldspar by albitic plagioclase can proceed only under conditions of increasing $(Na^+/H)/(K^+/H^+)$, as shown in Fig. 7.24.

Highly deformed meta-granites, such as the quartz-phengite-schists described in Chapter 2 part C(b), seem to indicate that metasomatism occurred under very different conditions. In these rocks, Na and Ca appear to have been removed, and the modal proportion of phengite increases at the expense of both plagioclase and K-feldspar. Thus, so far as plagioclase-muscovite relationships are concerned, the ionic reaction involved was the exact reverse of (7.AF) and (7.AQ). The complete reaction can be balanced assuming (a) an initial molar ratio of 3 orthoclase:3 albite:1 anorthite, and (b) that as volume need not have been conserved in a high strain environment, replacement occurred on a constant aluminium basis,:

$$16\ H^+ + \underbrace{9\ NaAlSi_3O_8 + 3\ CaAl_2Si_2O_8}_{\text{plagioclase}} + \underbrace{9\ KAlSi_3O_8}_{\text{K-feldspar}} \longrightarrow$$

$$8\ KAl_3Si_3O_{10}(OH)_2 + 36\ SiO_2 + \underbrace{9\ Na^+ + 3\ Ca^{2+}\ K^+}_{\text{fluid}} \qquad (7.AR)$$

$$\text{phengite} \qquad \text{quartz}$$

This reaction also accounts for the following features of quartz-phengite-schist: (a) modal quartz increases with modal phengite (Table 2.3), and (b) the potassium content of the rock is roughly conserved (Fig. 2.25).

A similar reaction was invoked by Norris et al. (1971c) to account for the production of schists rich in quartz and white mica in shear-zones in the

Zentralgneis. Two extra pieces of evidence support the conclusion that essentially bimineralic quartz-phengite-schists were produced metasomatically:

(i) In the system K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O necessary to define the compositions of such rocks the assemblage quartz+phengite+fluid has a high variance (F = 5) suggesting that the chemical potentials of several of the above components were externally controlled.

(ii) Experiments have shown that feldspars exchange ions very readily with both supercritical and subcritical aqueous chloride solutions (Orville, 1963, 1972; Fournier, 1976) and dissolve readily in hot water if the water: feldspar volume ratio is large enough (Morey and Chen, 1955; Currie, 1968).

The occurrence of complementary ionic reactions in different places and at different times (i.e. reaction (7.AR) in highly deformed granitic rocks at all levels in the basement and in allochthonous sheets within the cover - during D_a^1, possibly also D_a^2; reaction (7.AQ) in mica-schists at all levels in the basement and cover - after D_a^2) may be totally unconnected, but it is conceivable that the Na^+ and Ca^{2+} released into the fluid by reaction (7.AR) were transported, or allowed to diffuse, to the sites of reaction (7.AQ) and consumed there. The distances over which the ions would have had to travel range from only a few metres at some localities in the basement complex, to hundreds of metres for structurally high units in the cover.

H. CONCLUSIONS

1. Temperatures and pressures during pre-Alpine metamorphic events in the Inner Schieferhülle are estimated as follows (see Fig. 7.1):

   (i) The eclogite facies event (M_e): 540-600°C; P > 11 kb.

   (ii) The Variscan amphibolite facies event (M_v): 525-710°C; P > 5.5 kb.

   (iii) The contact metamorphism associated with the intrusion of the Zentralgneis (M_z): T > 730°C; P probably between 6 and 8 kb.

2. Temperatures and pressures at the climax of Alpine metamorphism (M_a) are estimated as follows:

   (i) In the lower limb of the Mallnitzer Mulde in the Romate sub-area (north side of Tauerntal): 470 ± 20°C; 5 ± 1 kb (Fig. 7.18).

   (ii) At the basement-cover interface at the south-west margin of the Hochalm Kern (Kaponigtal, Zwenberger Tal): 570 ± 20°C; 7 ± 1 kb (Fig. 7.19).

   (iii) At a structural depth of $2\frac{1}{4}$ km below the basement-cover interface in the Hochalm Kern (Kaponigtal): 590 ± 20°C; 7 ± 1 kb (Fig. 7.20).

   (iv) At a structural depth of 3 to 4 km below the basement-cover interface in the Hochalm Kern (Kaponigtörl, Gössgraben): 610 ± 20°C; P > 6 kb (Fig. 7.21).

The apparent Alpine geothermal gradient preserved in the Hochalm Kern is 10 ± 10°C km^{-1}. However, apart from being very badly constrained, this quantity is unlikely to represent a geothermal gradient which existed at any single point in time because rocks at different structural depths

in a tectonically thickened metamorphic pile are likely to have reached their maximum temperatures or entropies at different times (England and Richardson, 1977).

3. During Alpine metamorphism, most pelites and calc-schists equilibrated with fluids containing more than 70 mole percent H_2O although some kyanite-bearing pelitic assemblages appear to have buffered the fluid composition to lower values. The variety of incompatible siliceous dolomite parageneses can be explained in terms of isobaric, isothermal variation in the H_2O:CO_2 ratio of the coexisting fluid; X_{H_2O} ranges from 0.9 to less than 0.55. Fluids in equilibrium with serpentinite and syenite-gneiss were extremely water-rich, with X_{H_2O} values greater than 0.95.

4. At some time early in the Alpine metamorphic cycle, rocks near the base of the Möltal Schieferhülle passed through the stability field of lawsonite. Temperatures and pressures at this stage were probably lower than 450°C and 11 kb, respectively (Fig. 7.22).

Construction of a P-T-time trajectory for the Alpine metamorphism is left until Chapter 10, where evidence from late-stage veins is incorporated.

5. Localised metasomatic alteration of serpentinite to tremolite+talc$^{+}_{-}$ dolomite schist has occurred at the margins of serpentinite bodies and along shear-zones within them, due to an influx of CaO and SiO_2 from the enclosing calc-schists.

Metasomatism may also have taken place on a much larger (? regional) scale; the Na and Ca apparently released by feldspar-fluid reactions during intense deformation of granitic rocks might have been the same

Na and Ca that subsequently altered some of the muscovite in mica-schists to plagioclase. In this case, the deformation history and geometrical relationships of the lithologies involved would have had a controlling influence on the pattern of metasomatism.

CHAPTER 8

Oxygen Isotope Studies

CHAPTER 8

OXYGEN ISOTOPE STUDIES

A. INTRODUCTION AND DEFINITIONS

Oxygen is the most abundant element in the Earth's crust. Three stable isotopes, O^{16}, O^{17} and O^{18} account for, respectively, about 99.763%, 0.037% and 0.200% of the total oxygen (Garlick, 1969). However, the oxygen isotopic compositions of minerals often differ from one another by small but significant amounts. This applies both between different mineral species, and also between different samples of the same mineral.

Isotopic fractionation has occurred when two coexisting phases have different isotopic compositions. The fractionation factor, α, measures the degree of isotopic fractionation, and is defined by:

$$\alpha_{AB} = \frac{R_A}{R_B} \qquad (8.1)$$

where R_A and R_B are the isotopic ratios in minerals A and B respectively. For oxygen, considering only O^{16} and O^{18},

$$\alpha_{AB} = \frac{(O^{18}/O^{16})_A}{(O^{18}/O^{16})_B} \qquad (8.2)$$

Isotopic compositions of samples are measured relative to a standard, usually standard mean ocean water (S.M.O.W.), and are represented by a δO^{18} value, given by:

$$\delta O^{18}_A = \left[\frac{R_A}{R_{STANDARD}} - 1\right] \times 1000 \qquad (8.3)$$

Deines (1977) defines ΔO^{18}_{AB} as 1000 ln α_{AB}. For small fractionations,

$$\Delta O^{18}_{AB} = \Delta O^{18}_{A} - \Delta O^{18}_{B} \qquad (8.4)$$

(a) Factors affecting isotopic properties of minerals

If two phases in equilibrium contain oxygen, the heavier isotopes tend to be concentrated in the species in which the oxygen atoms are bonded to ions with smaller atomic mass and higher ionic potential (O'Neil, 1977). This effect is greater for larger isotopic mass differences; for this reason only O^{16} and O^{18} are measured for oxygen. In addition, O^{17} occurs in extremely low concentrations.

O^{18} tends to be concentrated preferentially in species where oxygen atoms are highly coordinated. For instance, aragonite, with oxygen atoms each having three nearest neighbour calcium ions, is enriched in O^{18} with respect to coexisting calcite with two-coordinated oxygens (Tarutani et al., 1969). Consequently, different groups of minerals containing different crystal structures tend to fractionate oxygen among themselves in a regular and fixed sequence. Taylor and Epstein (1962, 1963) have drawn attention to the fact that, in igneous rocks, coexisting minerals tend to concentrate O^{18} in the order:

mt (least O^{18}), ilm, ol, biot, hbl, cpx, plag, ksp, qz (most O^{18})

Taylor et al. (1963) and Garlick and Epstein (1967) have deduced a similar sequence for metamorphic rocks:

mt, ilm, chl, biot, gt, ky, wm , plag, qz

These relationships are supported by experimental data (Fig. 8.1). Occa-

sionally, the order of two minerals in the sequence is reversed; this is usually taken to imply disequilibrium (Garlick, 1969; Frey et al., 1976).

Chemical variation within an isostructural mineral group is not likely to affect the O^{18} affinity significantly if all that is involved is substitution of Fe for Mg or Na for K. O'Neil and Taylor (1969; 1967) demonstrated that no oxygen isotope fractionation occurs between coexisting muscovite and paragonite or between K-feldspar and albite at temperatures above 350°C. However, they also concluded that the replacement of Si-O bonds by Al-O bonds in the plagioclase series caused significant decrease in the relative O^{18} affinity. In the present study, plagioclase, garnet, biotite and phengite are the only phases capable of such substitutions. The compositions of all of these except plagioclase vary within narrow limits in the rock types analysed for oxygen isotopes (Chapter 5) and do not significantly differ from those of the minerals run in calibration experiments.

Temperature is a prime factor in determining the isotopic composition of a mineral because it affects the bond vibrational energy. The entropy of a mineral pair in equilibrium is larger at higher temperature, so the oxygen isotopes can be expected to distribute themselves more evenly between them, i.e. the fractionation factor approaches unity. Detailed accounts of the quantum statistical theory of isotopic fractionation between ideal gases are given by Urey (1947) and Bigeleisen and Mayer (1947) who extrapolate the theory to polyatomic molecules. For anhydrous species, the isotopic ratio, R, is predicted to vary with temperature according to:

$$\ln R \propto \frac{1}{T^2} \qquad (8.5)$$

at temperatures of geological interest. However, a more complicated temperature dependence is predicted for hydrous minerals owing to the high stretch-

ing frequencies of the O-H bonds compared with other bonds to oxygen, with the $1/T^2$ relationship only applying at very high temperatures. Although fractionation factors can be deduced theoretically for simple condensed phases such as the polymorphs of carbon (Bottinga, 1969) and quartz (Shiro and Sakai, 1972), detailed knowledge of the lattice dynamics required for such calculations is lacking for more complex minerals, and for this reason, data from mineral-water fractionation experiments are used instead. Some of these data are listed in Table 8.1 and plotted in Fig. 8.1. For most, $\ln \alpha$ shows a simple dependence on $1/T^2$ as predicted by the theory.

As the degree of isotopic fractionation between two minerals is primarily an entropy effect, pressure is not likely to affect it greatly. Hoering (1961) and Clayton et al. (1975) have confirmed that no measurable differences in the large oxygen isotope fractionation between calcite and water exist at 43.5°C from 0 to 4 kb and at 500°C from 1 to 20 kb. Oxygen isotope fractionations can therefore be used as pressure-independent geothermometers.

The absolute isotopic compositions of the minerals in a given rock depend largely on the whole-rock composition, although the fractionations between them are governed by the factors described above. Most igneous rocks have δO^{18} values in the range 5 to 7°/oo (Taylor, 1968), but values from sedimentary and metamorphic rocks range from +2 to +38°/oo (Hoefs, 1973). Limestones, in particular, tend to have very high δO^{18} values.

(b) Present objectives

In this chapter, the principles of oxygen isotope geothermometry are applied to rocks from the south-east Tauern to estimate the following:

(i) Temperatures of metamorphism : An isotopic fractionation can be measured for any pair of minerals separated from the same rock and used to calculate a temperature. A rock from which n phases have been separated can provide $\frac{1}{2}$ (n^2-n) temperature values (provided sufficient experimental data exist) of which n-1 are independent. These values can then be compared with those estimated from phase equilibria (Chapter 7). As oxygen isotope geothermometry can be applied to rock types for which phase-equilibrium geothermometry is difficult (e.g. meta-granites) temperatures may be obtained along semi-continuous profiles, facilitating the calculation of thermal gradients.

(ii) Degree of isotopic equilibrium : If more than one mineral pair from the same rock yield concordant temperatures, then those minerals are likely to have been in isotopic equilibrium when the isotopes were frozen into the lattices. The resulting temperature has a real physical significance in that it represents the temperature at which this happened. On the other hand, if widely discordant values are obtained, the minerals were obviously not in isotopic equilibrium, and the resulting temperatures have no real significance. Isotopic disequilibrium between minerals could arise in three distinct ways in regionally metamorphosed rocks:

1. when isotopic compositions of all mineral species are frozen in at the time of crystallisation: If minerals A and B crystallised together, but before C and D, A-B and C-D fractionations will give true temperatures of crystallisation, but A-C, A-D, B-C and B-D fractionations will give disequilibrium temperatures.

2. when the isotopic composition of mineral F is frozen in at the time of crystallisation, but the compositions of minerals G and H, which may have grown at the same time as F, continue to equilibrate with one another with rising temperature, though not beyond the thermal i.e. entropy maximum: In this case the G-H fractionation will give the peak metamorphic temperature, but F-G and F-H fractionation temperatures will not be realistic;

3. when the isotopic composition of minerals M and N equilibrate with one another under conditions of falling temperature, but those of minerals P and Q do not readjust after the thermal maximum: In this case the M-N fractionation will give a blocking temperature.

Which of these situations applies depends entirely on the relative ease with which oxygen can diffuse through the lattices of the interacting minerals. Close-packed structures such as garnet and spinel might be expected to behave like F, while open framework silicates such as feldspars like M and N. If oxygen diffusion, and hence isotopic equilibration, cannot occur without destruction of crystal lattices, (2) and (3) are inapplicable.

Comparison with textural relationships in thin section can sometimes reveal which of situations (1), (2) and (3) apply.

(iii) Compositions of coexisting aqueous metamorphic fluids : Once an equilibrium temperature has been ascertained for a pair of minerals in a rock from the mineral-mineral fractionation, the isotopic composition of the water coexisting with those minerals at that temperature can be back-calculated using mineral-water fractionations.

The isotopic composition of the water and minerals can be envisaged as equilibrating on different scales:

1. Large scale : If the volume of water interacting with a given volume of rock is large, i.e. in an open system, the isotopic bulk composition of the effective reacting volume of rock will be governed by the composition of the water and will be roughly uniform, even when comparing rocks of initially different isotopic composition.

2. Small scale : If the volume of water interacting with a given volume of rock is small, i.e. if the pore water is essentially static, its isotopic composition will be buffered by that of the effective reacting volume of rock.

In a closed system,

$$\sum_i m_i \delta O^{18}(\text{minerals}) + w\delta O^{18}(\text{water}) = \text{constant} \qquad (8.6)$$

where m_i is the volume proportion of each mineral species i and w the volume proportion of the water. For most metamorphic rocks, $\sum_i m_i > 0.99$ and $f < 0.01$.

If all the minerals in a rock interact continually with a buffered pore water over a period of time, the isotopic composition of the water will vary smoothly with temperature (and hence time) in such a way that all mineral-mineral and mineral-water fractionations record the same temperature at each point in time, and equation (8.6) is satisfied. If, however, some minerals do not exchange oxygen isotopes with the pore water after crystallisation (i.e. as

in (ii)(2), above), then

$$\sum_r m_r \delta O^{18}(\text{reacting minerals}) + w\delta O^{18}(\text{water}) \neq \text{constant} \qquad (8.7)$$

and its isotopic composition is likely to vary erratically in space and time.

Other things (e.g. mineralogy, reaction history, equilibrium temperatures) being approximately equal, the variation in the pore water composition calculated from isotopic fractionations between equilibrium mineral pair U and V in different rocks places some constraint on the extent of fluid migration.

(iv) The origin of the quartz-phengite-schists : The differences between the whole-rock isotopic compositions of many meta-igneous and meta-sedimentary rocks might be useful in ascertaining whether or not these schists could have been originally granites, as proposed in Chapter 2. Any such correlation would be based on the assumption that no significant changes in whole-rock isotopic composition occurred.

B. METHODS OF STUDY

(a) Sampling and preparation

25 samples were collected for oxygen isotope analysis of constituent minerals. Sampling localities are shown in Fig. 8.7. Choice of samples was restricted to rocks which had (a) no obvious polymetamorphic history, (b) grain-size sufficiently coarse for efficient separation, (c) as many different minerals as

possible for which calibrations exist, and (d) no obvious retrogressive alteration to chlorite, etc. They fall into two main categories: granite-gneisses and mica-schists.

Preparation, extraction and analysis techniques are described in Appendix III. B, and the results are listed in Appendix III.C.

17 samples were crushed and powdered for whole-rock oxygen isotope analysis.

(b) Oxygen isotope thermometry

The mineral-water calibrations used here to calculate temperatures are listed in Table 8.1 and plotted in Fig. 8.1. All were derived experimentally except the garnet-water curve which was derived by Friedrichsen et al. (1973) using natural garnet data collected by these authors and by Taylor and Coleman (1968). These curves can be used to construct a set of mineral-mineral fractionation curves using the relationship:

$$\Delta_{A-B} = \Delta_{A-H_2O} - \Delta_{B-H_2O} \qquad (8.8)$$

Some of these are plotted in Fig. 8.2, and can be used to read off temperature directly knowing the fractionation between each pair of minerals. Garlick and Epsein (1967) have shown that, over a wide range in metamorphic grade, ΔO^{18}(qz-ilm) is nearly always 0.8‰ smaller than ΔO^{18}(qz-mt) in adjacent rocks. Magnetite curves can therefore be used for ilmenite if 0.8‰ is subtracted from δO^{18}(ilm). The most temperature-sensitive curves are those with the steepest slopes, e.g. quartz-magnetite. The biotite-magnetite and garnet-muscovite curves are virtually horizontal and have no use as thermometers. For an uncertainty in the ΔO^{18} values of $\pm$0.2‰, the following errors are

obtained:

$\pm$10°C : qz-mt, qz-biot, musc-biot, gt-biot, musc-mt, fsp-biot, fsp-rut, qz-rut

$\pm$20-30°C : gt-mt, qz-gt

$\pm$40°C : gt-fsp

$\pm$60°C : qz-musc

$\pm$100°C : qz-fsp

The temperature data for each rock can be summarised in a graph of fractionation versus temperature. A mineral assemblage which preserves isotopic equilibrium plots as a straight vertical line within the uncertainties listed above. Some of these graphs are shown in Figs. 8.3, 8.4 and 8.5.

Temperatures calculated using the graphical method of Javoy et al. (1970) are also listed in Appendix IIIC for comparison. This method is based on experimentally derived empirical relationships of the form:

$$\Delta_{x\text{-water}} = 1000 \ln\alpha_{x\text{-water}} = A_x + \frac{10^6 B_x}{T^2} \qquad (8.9)$$

$$= \Delta_{x-y} = (A_x - A_y) + \frac{10^6(B_x - B_y)}{T^2} = A'_{x-y} + \frac{10^6 B'_{x-y}}{T^2} \qquad (8.10)$$

These authors calculate a set of A' and B' coefficients for mineral-quartz fractionations but use different mineral-water curves from those used above. A graph of $\Delta_{qz-x} - B'_{qz-x}$ against A'_{qz-x} for all the quartz-mineral fractionations in one rock should give a straight line of slope $10^6/T^2$ if the minerals were in isotopic equilibrium. If the points do not lie on a straight line, a regressed line can be used to give a crude temperature estimate. Fractionation data from some of the Tauern rocks are presented in this form in Figs. 8.3 and 8.5. Hoernes and Friedrichsen (1978) argue that this method gives temperatures that are too low by comparison with experimentally determined reactions.

C. DISCUSSION OF RESULTS - SEPARATED MINERAL DATA

(a) Peripheral Schieferhülle lithologies

All temperature data extracted from Peripheral Schieferhülle rocks must, by definition, relate to the M_a metamorphic event. Of the seven mica-schists in this category from which more than two minerals were separated, only two give concordant temperatures (GD 28b, ~ 540°C (Fig. 8.3(b)) and GD 496, ~ 560°C). The rest, GD 234 (Fig. 8.5(c)), GD 573, GD 992, GD 890 (Fig. 8.5(a)), and GD 903 (Fig. 8.5(b)) give discordant temperatures and did not achieve oxygen isotopic equilibrium on the scale of a hand specimen. In addition, Inner Schieferhülle schist GD 302 gives an excellent concordant temperature of 555°C (Fig. 8.3(a)).

The temperature data from GD 28b, GD 496 and GD 302 were based on fractionations between quartz, muscovite and magnetite. If three minerals in a rock give concordant values, it does not necessarily imply that all the minerals in the rock were in isotopic equilibrium during metamorphism (Hoernes and Hoffer, in prep.). However, it does mean that those minerals were in equilibrium at or near the calculated temperature at some stage in the metamorphic history of the rock. Such temperatures can therefore be used as minimum values for conditions at the peak of metamorphism.

Temperatures from GD 302 and GD 28b agree well with those from other estimates. GD 28b was collected from a locality near the isograd marking the first appearance of staurolite in chloritoid-bearing rocks, for which a temperature of 570°C is estimated (see Chapter 7). GD 496 was found in the low-grade area west of Mallnitz. The isotopic temperature of 560°C probably falls close to the upper-temperature limit of staurolite-free aluminous

pelites. Although GD 302 contains a garnet+chlorite assemblage, it was found at a structurally lower level than chloritoid+biotite zone rocks outcropping on Sicker Kopf and along strike from staurolite-biotite-schists in Kaponigtal and on the Grüneck ridge. A temperature of 570°C is probably reasonable for the staurolite+biotite isograd. Oxygen isotopic fractionations, therefore, appear to have been set close to the peak of metamorphism in these rocks. This is surprising because muscovite has a well developed schistosity in all three (crenulated S_a^1 in GD 496, ?S_a^2 in GD 28b and GD 302) and obviously grew at an earlier stage. This implies that either (a) the deformation events occurred shortly before the peak of metamorphism, or (b) the muscovites exchanged oxygen isotopes after crystallisation, or (c) the muscovites recrystallised mimetically after deformation. Hypothesis (c) is not considered likely owing to the common occurrence of bent muscovites in crenulated schists (see Chapter 4). In GD 496 and GD 28b the magnetites form small euhedral crystals that may well have grown close to the peak of metamorphism. In GD 302 they form 3 mm long pre-tectonic augen, yet still record high temperatures. This suggests that magnetites are capable of exchanging oxygen isotopes after growth.

In the other Peripheral Schieferhülle schists, magnetite and muscovite textures are similar to those in GD 496 and GD 28b. However, in all cases, the fractionations between quartz, muscovite and magnetite yield grossly discordant temperatures. Within any one rock, the muscovite-magnetite fractionation yields the lowest of the three temperatures and the quartz-muscovite fractionation, the highest. While the muscovite-magnetite and quartz-magnetite temperatures are geologically reasonable, the quartz-muscovite temperatures are often in excess of 800°C (1200°C in the case of GD 992), well outside the stability field of muscovite. As there is nothing unusual about the muscovite textures in these rocks, the quartz is probably the disequilibrium phase. In many of these schists, most or all of the quartz occurs as flattened veins up to 1 cm thick within the schistosity. Only a small proportion of the

quartz in such a vein comes into close ($<\frac{1}{2}$ mm) contact with the minerals of the host rock, with the result that the bulk of the quartz may have failed to equilibrate with them. Oxygen isotopic disequilibrium between quartz segregations and host-rock minerals has been recorded by Hoernes and Hoffer (1978, in prep.) from the Damara Orogen, and by Magaritz and Taylor (1976) from the Franciscan Formation. In the former, vein quartz was found to be isotopically lighter than rock quartz, whereas in the latter the opposite was found. In the schists giving concordant temperatures, the quartz occurs as stringers no more than 0.5 mm thick. In the disequilibrium schists, the quartz-muscovite and quartz-magnetite fractionation temperatures for each rock converge upon the muscovite-magnetite temperature when the quartz δO^{18} values are increased by +0.3‰ (GD 234) to +1.6‰ (GD 992), implying that of the three minerals, only the quartz is not in equilibrium and that the muscovite-magnetite temperatures are true equilibrium values. Furthermore, these rocks resemble those described by Hoernes and Hoffer, in that the vein quartzes are isotopically lighter than if they had been in equilibrium. The lowest muscovite-magnetite temperatures (470°C, GD 992 and 480°C GD 575) are from the Mallnitzer Mulde while the highest (570°C, GD 234) is from the southern margin of the Hochalm massif. These agree with the temperature trends inferred from metamorphic zonation in pelites (Chapter 6) and from other geothermometers (Chapter 7).

The garnet-oxide fractionation temperatures in GD 890 and GD 234 are unrealistically high, and the quartz-garnet ones correspondingly low, suggesting that garnets are not in equilibrium with the other phases in these rocks. In both, the garnets grew before D_a^2.

Temperatures calculated for Peripheral Schieferhülle schists using the method of Javoy et al. (1970) agree quite well with the other estimates, provided that quartz-muscovite fractionations are ignored in the disequilibrium rocks. Those for GD 28b, GD 496 and GD 302 are low compared with other estimates.

(b) Zentralgneis lithologies

Temperature data extracted from Zentralgneis rocks could refer either to the original crystallisation and cooling associated with magmatic emplacement or to the Alpine metamorphic event. Hoernes and Friedrichsen (1974) assume that the Zentralgneis rocks in the western Tauern re-equilibrated isotopically during the Alpine metamorphism, at least with respect to quartz and micas. In many of those studied here, the micas have a linear or schistose fabric and have obviously recrystallised during Alpine deformation. Micas in other samples are randomly orientated but are sometimes intergrown with one another and enclose non-allanitic (i.e. metamorphic) epidotes (Fig. 2.24) so that this assumption is probably valid.

Quartz, muscovite and biotite were separated from four metagranites and quartz and biotite from a further seven. Fig. 8.6 is a plot of ΔO^{18} quartz-biotite against ΔO^{18} quartz-muscovite. If the minerals in every rock were in isotopic equilibrium, the points should lie on a straight or slightly curved line passing through the origin (Clayton and Epstein, 1958; Taylor, 1968; Deines, 1977), which should coincide with the line AA' if the calibration experiments shown in Fig. 8.1 are correct. Three of the points lie close to AA' and may represent equilibrium assemblages. In addition, these rocks give good straight line plots on Javoy-diagrams (e.g. Fig. 8.3(c)). The fourth point, GD 334, lies far off the line and obviously represents a disequilibrium assemblage.

The temperatures given by the three equilibrium rocks do not correlate well with other estimates. All three samples give temperatures in the range 460 to 510°C. GD 595 and GD 596 were collected from the Mölltal gneiss-lamella for which those temperatures are reasonable in view of the low metamorphic grade (gt+chl zone), but GD 468 was found in the Gössgraben at a very low structural level, well within the staurolite+biotite zone where 600+°C would

be more appropriate. The quartz-feldspar and quartz-muscovite fractionations in GD 468 give 570°C and 610°C respectively, but these are not very significant owing to the large errors of these thermometers.

Taken together, the quartz-biotite fractionation temperatures of the eleven Zentralgneis rocks show very little spread either side of 500°C and no correlation with inferred peak metamorphic temperatures (Fig. 8.7). The granite-pegmatite, GD 532, gives a value of 610°C, but it is very coarse-grained and its minerals may not have re-equilibrated fully during Alpine metamorphism. This is supported by an unreasonably high quartz-plagioclase temperature of 1220°C. However, if the rest of the quartz-biotite fractionations preserve real metamorphic temperatures, they could have been frozen in either when the biotites first crystallised during prograde metamorphism or at some time during retrograde cooling, in which case they would be analogous to blocking temperatures in unstable isotope systems. Both mechanisms would result in a tight cluster of temperature points as observed here. The common failure of mineral-biotite fractionations to preserve peak metamorphic temperatures has been noted by Bottinga and Javoy (1975) who conclude that biotite is of dubious value to oxygen isotope thermometry. Deines (1977) ascribes this to retrograde exchange. Recently, however, Hoernes and Hoffer (in prep.) have shown conclusively that the oxygen isotopic compositions of biotites from Damaraland pelitic schists and gneisses are fixed during crystallisation.

(c) Inner Schieferhülle lithologies

Other than GD 302, already described, only three Inner Schieferhülle rocks were studied, all from the mica-schists of the Randgneis west of Mallnitz. Equilibrium temperature data extracted from Inner Schieferhülle rocks could refer to either pre-Alpine or Alpine metamorphic events. However, disequili-

brium is to be expected because even if complete equilibrium was attained during the pre-Alpine event(s), it is unlikely that minerals such as garnet will have totally re-equilibrated at a later stage, especially since many Alpine garnets in the Peripheral Schieferhülle schists have not fully equilibrated.

Biotite, muscovite and quartz were separated from GD 642 and GD 1230a, and garnet was separated from the latter. Both rocks give widely discordant temperatures suggesting large departures from isotopic equilibrium. The quartz-muscovite temperatures from the two rocks are identical (680°C) and the quartz-biotite and muscovite-biotite values nearly so (~485°C and ~445°C respectively), indicating that the minerals have behaved the same way in both rocks. The quartz-biotite temperatures agree with values from Zentralgneis rocks and with the quartz-magnetite temperatures in nearby Peripheral Schieferhülle schists.

The garnets in GD 1230a are small and appear to predate S_a^1, and may be pre-Alpine. They do not seem to be in isotopic equilibrium with other minerals.

The third sample, GD 1230b, is the quartz+garnet vein mentioned in Chapter 2, part B(f). The δO^{18} value of the quartz is about 1‰ lower than that of the rock-quartz in GD 1230a, collected from a few metres away. This agrees with the observations made on vein-quartz from Peripheral Schieferhülle schists. However, the vein-garnets have a δO^{18} value 0.2‰ higher than the coexisting vein-quartz. This is a reversal of the order listed in part A(a), and indicates that these minerals and so far out of isotopic equilibrium that no temperature solution is possible for this fractionation. This supports the conclusion arrived at earlier that these garnets are not primary vein minerals. Their δO^{18} value is 3.4‰ higher than that of the rock-garnets in GD 1230a, and is abnormally high compared with other published garnet compositions.

(d) Late-stage veins

Two samples of vein rock from the summit of Auernig were separated into constituent minerals for analysis. Both were loose blocks and were found close to one another in the debris below a 20-30 cm vein conduit.

One sample, AU1, yields a quartz-rutile fractionation temperature of 530±10°C, only slightly lower than the peak metamorphic temperature estimated for that area, ~560°C. Quartz, albite and ilmenite were separated from AU2 and give slightly discordant temperatures. The quartz-ilmenite value of 520±10°C agrees well with the estimate from AU1 but the albite-ilmenite temperature, 560±10°C, is unrealistically high, and the quartz-albite value of 340±100°C correspondingly low. If the data from AU1 and AU2 are combined using only the albite δO^{18} value of AU1, then an excellent set of six concordant temperatures are obtained in the range 520 to 530°C, all of which have errors of only ±10°C (Fig. 8.4). The new quartz-albite temperature of 460±100°C agrees with the others.

As both samples almost certainly came from the same vein, the author considers the combined fractionation temperature of 525°C to be a reasonable equilibrium value, particularly as this would have been close to the best estimate from each sample taken individually. This probably represents the temperature at which the vein minerals precipitated from the fluid rather than some point during subsequent cooling because in the latter case one would expect different minerals to block at different temperatures, which would give grossly discordant results.

The peak metamorphic temperature (560°C) provides a maximum vein equilibrium value only if it is assumed that the temperatures of the vein fluids were buffered by those of the rocks through which they passed. If those fluids

rapidly rose from great depths the vein minerals could have grown at a higher temperature than that achieved during metamorphism of the host-rocks. However, field evidence (Chapter 3, part C(d)) suggests that the vein fluids were not far-travelled.

(e) Isotopic compositions of metamorphic waters

Direct sampling of the fluid phase is impossible unless fluid inclusions are preserved within mineral grains. However, the isotopic composition of the H_2O in the fluid in equilibrium with the minerals can be calculated using the mineral-water curves shown in Fig. 8.1 provided that the mineral-mineral fractionations preserve equilibrium temperatures. This method has been used by Clayton and Epstein (1961), Rumble et al. (1975) and Magaritz and Taylor (1976). Table 8.2 lists the isotopic compositions of waters in equilibrium with muscovite and magnetite/ilmenite in Peripheral Schieferhülle schists, and with quartz and biotite in Zentralgneis rocks. These compositions range from +5 to +8‰ and from +10 to +17‰, respectively. The size of these ranges suggest that little homogenisation has occurred within each group, although this is more convincing for the Peripheral Schieferhülle rocks than the others as they have a larger spread of fluid compositions. Hoernes and Friedrichsen (1974) arrive at similar conclusions for rocks in the western Tauern on the basis of whole-rock isotopic compositions (see below). The only equilibrium Inner Schieferfülle rock analysed here, GD 302, yields a water δO^{18} value of +8.1‰, just at the limit of the Zentralgneis range, but quartz-biotite data from the disequilibrium schists, GD 642 and GD 1230a, yield compositions near +9‰.

The flattened quartz veins in the pelitic schists were originally formed before D_a^1. The temperature at this stage of metamorphism was probably lower

than 400°C (see Chapter 7). It may have been much lower than this, considering the abundance of quartz veins in very low-grade phyllites in many metamorphic terrains (Norris and Henley, 1976). The δO^{18} values of waters in equilibrium with segregation quartz in GD 992, GD 903 and GD 573 are listed in Table 8.3 for various temperatures. The main conclusion is that these early metamorphic waters were isotopically much lighter than those in equilibrium with muscovite and magnetite at the peak of metamorphism. They were also lighter than those that would have been in equilibrium with muscovite and magnetite at lower temperatures (assuming that the bulk isotopic composition of the rocks remained the same through time) because the segregation quartz is isotopically lighter than it would have been if in equilibrium with muscovite and magnetite and not heavier as predicted by the increase in quartz-mineral fractionations with decreasing temperature. At 300°C, the δO^{18} value of water in equilibrium with quartz, muscovite and magnetite in GD 903, bulk composition +13‰, would have been approximately +7.5‰ whereas the water in equilibrium with the segregation quartz would have had a value of +5.75‰. This discrepancy could be explained if part of the water released during this early stage of vein formation was not in equilibrium with the minerals in the rocks, and was not produced by dehydration reactions. This O^{18} depleted disequilibrium fraction might have been ocean water ($\delta O^{18} \approx 0$‰) trapped during sedimentation. Hoernes and Hoffer (in prep.) conclude that the O^{18} depleted concordant quartz segregation in the Damaraland pelitic schists were precipitated from syn-metamorphic fluids that were formed in equilibrium with hotter rocks at greater depths, but the structural time relations rule out this explanation for the Tauern schists.

The δO^{18} value of water in equilibrium with the Auernig chlorite-albite-vein samples, +9.7‰, falls at the lower end of the composition range of metamorphic waters in equilibrium with pelite minerals. However, the host rocks to these veins are epidote-amphibolites, too fine-grained for mineral

separation. Hoernes and Friedrichsen (1974) separated minerals from Peripheral Schieferhülle amphibolites. Waters in equilibrium with these minerals have δO^{18} values ranging from +11‰ to +14‰, about 1‰ higher than corresponding whole-rock values. The whole-rock isotopic composition of the Auernig amphibolite sample, GD 483, is +8.9‰, 0.8‰ less than δO^{18} of the water, suggesting that the vein fluids could have been in equilibrium with amphibolite minerals.

D. DISCUSSION OF RESULTS - WHOLE ROCK DATA

Whole-rock isotopic compositions have been calculated for the 25 separated samples using the mineral compositions and modal analyses, and are included in Appendix II Wherever minerals present have not been analysed, approximate compositions have been estimated assuming them to have been in equilibrium with other minerals. This method has been used by Hoernes and Friedrichsen (1974) who quote the error on calculated whole-rock compositions as $\pm$0.5 to 1.0‰.

The δO^{18} values of the 17 analysed whole-rock samples are listed in Table 8.4.

Calculated whole-rock compositions are listed together with separated mineral compositions in Appendix III.

Zentralgneis whole-rock compositions are well-clustered around +8‰ to +9‰. This is close to the value Taylor (1968) estimated for granites of magmatic origin, indicating that no large-scale re-adjustment of isotopic composition has taken place.

Rocks of the Peripheral Schieferhülle have a much larger range in δO^{18} values than the Zentralgneis, but also appear to preserve their original isotopic compositions: ca. 14-20‰ for pelites (Savin and Epstein, 1970a), 15-25‰ for carbonate rocks (Schwarcz et al., 1970) and 6-9‰ for basic and ultrabasic rocks (Taylor, 1968).

Pelitic mica-schist GD 18 was sampled from a 10 cm band within calc-schists of which GD 50 is representative. GD 18 has a low δO^{18} value (+11.5‰) compared with GD 50 (+15.4‰) and compared with other pelites, and evidently did not exchange oxygen isotopes with the surrounding calc-schists. This places an upper limit of 10 cm on the scale of isotopic homogenisation.

The serpentinite GD 48a and actinolite-rock GD 48b were collected from within 1 m of one another from an ultrabasic pod within the same calc-schist outcrop. GD 48b is 2.0‰ richer in O^{18} than GD 48a, suggesting that water was introduced from the calc-schists together with calcium when the metasomatic alteration of serpentine to actinolite took place.

δO^{18} values of the three quartz-phengite-schists, MG4-6, are slightly higher than those of most Zentralgneis rocks but fall well within the range of Peripheral Schieferhülle metasediment compositions and within the range of primary sandstone compositions: +11-+16‰ (Savin and Epstein, 1970b). Taken by themselves, therefore, they do not preclude a sedimentary origin for the gneiss lamellae. If, however, the lamellae were originally granite, as argued in Chapter 2 part C(b), then whole-rock isotopic enrichment of O^{18} must have occurred during deformation. This hypothesis is supported by the increase in δO^{18} from +11.2‰ in the least deformed schist, MG4, to +13.6‰ in the most deformed schist, MG6.

E. CONCLUSIONS

The main conclusions of this study are as follows:

(i) Most rocks show evidence of disequilibrium on hand-specimen scale.

Garnet and coarse-grained vein quartz, particularly where they appear on textural grounds to be relicts, are commonly out of equilibrium with other phases. It seems likely that the isotopic compositions of garnet and vein-quartz were frozen in early in their metamorphic history and failed to equilibrate with other minerals through the M_a temperature peak. Retrogressive re-equilibration is unlikely because fine-grained vein quartz, which should equilibrate more readily than coarse-grained quartz, commonly give realistic M_a peak temperatures.

Failure of biotite to re-equilibrate with other minerals through the peak of metamorphism may be the reason that quartz-biotite fractionation temperatures are clustered round 500°C and do not vary systematically with metamorphic grade.

(ii) Muscovite and magnetite appear to have equilibrated very readily with one another, even when their textures indicate that they crystallised early in the metamorphic cycle. Feldspars probably also belong in this category.

On the basis of the present results, ΔO^{18} (musc- mt) is considered the most reliable oxygen isotope geothermometer. ΔO^{18} (qz- mt) is only reliable if the quartz occurs as individual grains or thin (<1 mm) veins.

Muscovite-magnetite fractionation temperatures agree well with estimates from

phase equilibria. They range from 470°C in the Mallnitzer Mulde to 570°C in the Mölltal Schieferhülle. However, temperatures are not well enough constrained to be able to draw isotherms or compute geothermal gradients.

(iii) Quartz, albite, ilmenite and rutile crystallised at ca. 525°C in the Auernig chlorite+albite veins. They were precipitated from metamorphic waters that could have been in isotopic equilibrium with minerals in the surrounding epidote-amphibolites.

(iv) The oxygen isotopic composition of Peripheral Schieferhülle and Zentralgneis lithologies are similar to those characteristic of unmetamorphosed rocks of the same bulk chemistries. In general, whole-rock isotopic compositions were not homogenised on a regional scale, and probably not over distances greater than 10 cm. Localised exceptions to this include rocks modified by major-element metasomatism, such as ultrabasic schists derived from serpentinites and possibly quartz-phengite-schists.

The calculated isotopic composition of metamorphic waters in equilibrium with quartz and biotite in meta-granites and with muscovite and magnetite in pelites varies sympathetically with the isotopic composition of the host rocks and appears to have been buffered by them. Except for the metasomatic rocks mentioned above, this precludes the possibility of comparatively large volumes of water having migrated and interacted with individual volumes of rock during Alpine metamorphism.

CHAPTER 9

Fluid Inclusion Studies

CHAPTER 9

FLUID INCLUSION STUDIES

A. INTRODUCTION

Two important conclusions have already been drawn concerning chlorite + albite veins in the south-east Tauern:

(i) they were injected after the Alpine metamorphic peak and during D_a^3 (Chapter 3).

(ii) Quartz, albite, rutile and ilmenite crystallised together at 525 $\pm$ 10°C in the Auernig veins (Chapter 8).

The discovery of primary fluid inclusions in quartz from one of the Auernig veins provides an opportunity to estimate the lithostatic pressure during D_a^3.

The reasoning behind the use of fluid inclusions as barometers is as follows. If inclusions of a single fluid phase become trapped by solid at some temperature T_1 and pressure P_1, the internal pressure of the inclusion is buffered along a curve of equal fluid density (isochore) with falling temperature because the volume of the inclusion remains essentially constant due to the negligible effect of pressure and temperature on the molar volume of the enclosing solid (see Appendix II (d)). Eventually, at some lower temperature T_2, the internal pressure is sufficiently low for a vapour bubble to nucleate, and thereafter it is buffered along the liquid-vapour equilibrium curve with

falling temperature (Fig. 9.1). Thus, the isochore containing P_1 and T_1 can be identified if T_2, the homogenisation temperature of the inclusion (i.e. the temperature at which the bubble disappears during reheating), is measured. P_1 can then be determined if an independent estimate of T_1 exists. In this case, T_1 is assumed to be the oxygen isotope equilibrium temperature, 525 ± 10°C.

B. DESCRIPTION OF MATERIAL

The sample of vein-rock selected for study consists mainly of coarse-grained milky quartz with locally developed crystal faces. A small proportion of albite and chlorite are also present.

The quartz contains abundant fluid inclusions up to 30 μ long which tend to be concentrated towards the rims of individual crystals. Two distinct types of inclusion are recognised:

(i) Primary inclusions: These tend to be isolated spheroidal to ellipsoidal negative crystal forms (Fig. 9.2) containing both liquid and gas phases. One gas bubble occurs in each inclusion and occupies 20-30% of the total volume. Primary inclusions rarely contain daughter crystals, and no more than one crystal occurs per inclusion. The daughters are rhomboid and birefringent and may be albite or, more likely, carbonate. The primary inclusions are interpreted as having been formed during crystal growth at the time of vein precipitation.

(ii) Secondary inclusions: These are much less abundant than the primaries. They tend to have very irregular shapes and to be

aligned in short trails and are interpreted as having formed after quartz crystallisation had ceased. Rhomboid, birefringent daughter crystals, apparently identical to those mentioned above, are very common.

C. ANALYTICAL METHODS

(a) Sample preparation

A 4 cm^2 slice of vein quartz was ground down to a thickness of 0.06 mm and highly polished on both sides, as described by Holland et al. (1978). The thin section was then detached from the glass slide by dissolving the "Lakeside" mounting medium in alcohol, dried and broken into pieces small enough (10-15 mm^2) to fit into the heating and freezing stage.

(b) Microthermometry

Fluid inclusion homogenisation and freezing temperatures were measured on a Chaix-Meca heating and freezing microscope stage. In this apparatus, the sample chip is enclosed in a small thick-walled metal capsule incorporating a heating element and a coiled tube. The temperature of the entire capsule and its contents can be controlled by varying the current through the heating element or the rate of flow of liquid nitrogen through the coiled tube (for temperatures above and below 25°C, respectively). The capsule is equipped with heat-resistant silica-glass windows enabling one to view the inclusions through the microscope as their temperature is varied. The sample is viewed in green monochromatic light in order to increase the definition of phase boundaries. Phase-change temperatures were measured using low heating rates

(ca. 0.5°C per minute) in the critical temperature range.

Although recorded temperatures are precise the 0.1°C, they are not strictly accurate. The apparatus was calibrated using compounds of known melting point in order to ascertain the correction, ΔT, that must be added to the recorded temperature to convert it to the actual temperature. The results are shown in Table 9.1 and Fig. 9.3. Standard compounds were placed between two small cover-slips, but otherwise treated in the same way as the sample chips.

D. RESULTS

The 63 recorded primary fluid inclusion homogenisation temperatures range from 315.8°C to 347.3°C (Table 9.1), with a mean of 324.2°C and standard deviation of 6.0°C. Addition of a ΔT correction factor of -5.7°C (Table 9.1) gives an estimated actual mean temperature of 318.5°C.

The 23 recorded freezing points, mostly from inclusions for which homogenisation temperatures were also obtained, range from -3.3°C to -4.7°C, with a mean of -3.7°C and standard deviation of 0.35°C, giving a corrected mean temperature of -3.9°C. This suggests that the fluid is largely H_2O, but contains sufficient solute to depress the freezing point from 0°C to -3.9°C. In the system NaCl - H_2O, -3.9°C is the freezing point of an aqueous solution of about 8 mole percent NaCl (Roedder, 1971), suggesting that even if NaCl was not the only species present, the trapped vein fluids contain moderately high solute concentrations.

During freezing point studies of some of these inclusions, no evidence of any phase change was observed near -56.6°C, the melting point of CO_2, suggesting

that the solvent is essentially pure H_2O. This has been confirmed by B. Smith (pers. comm.) who dried and heated a piece of the same Auernig quartz sample beyond the decrepitation temperature (345°C) of the inclusions, and analysed the mass spectrum of the evolved fluids. The only gases present in appreciable quantities were H_2O and CO_2, and CO_2 formed not more than 2 or 3 % of the total.

The positions of the isochores and the liquid-vapour equilibrium curve for pure H_2O have been determined experimentally by Kennedy (1950), whose data are shown in Fig. 9.1, and Lemmlein and Klevstov (1961). Both sets of data agree closely to the curves calculated by Holland (1977) from the results of Burnham et al. (1968) using the modified Redlich-Kwong equation (Chapter 7 part B (f)).

Use of the pure water PVT diagram may not be justified if the presence of dissolved solute equivalent to 8 wt. % NaCl has a significant effect on either the slope of the isochores or the position of the liquid-vapour equilibrium curve. However, Haas (1971) concludes that, for a brine containing 10 wt. % NaCI, the liquid-vapour equilibrium pressure is only about 20 bars lower than that for pure water at 320°C. The data of Lemmlein and Klevstov (1961) indicate that the dP/dT of the isochore containing a given pair of P_1T_1 and P_2T_2 points remains essentially unaltered over the range of 0 to 15 + wt. % NaCl. Thus the error introduced by the presence of solute equivalent to 8 wt. % NaCl is negligible compared with the error associated with the spread of homogenisation temperatures.

The lithostatic pressure during D_a^3 is given by the point at which the isochore intersecting the liquid vapour equilibrium curve at 318.5°C reaches 525°C, the oxygen isotope equilibrium temperature. This value is 2.1 $\pm$ 0.2 kb (Fig. 9.4). Although much lower than the M_a peak metamorphic pressure of

7 kb, this is similar to the pressures of 2.5 to 3 kb deduced by Poty et al. (1974) for late Alpine fissures in the Western and Central Alps.

The present results have been combined with evidence from early- and peak-M_a phase assemblages to form a P-T-time trajectory for the Alpine metamorphic cycle (Fig. 10.1). In doing so, it has been assumed that the pressure and temperature conditions of the rocks surrounding the vein were identical to those within it at the time of its formation (i.e. $P_{fluid} = P_{load}$, $T_{vein} = T_{rock}$). However, if the vein had been open to the atmosphere, the pressure within it would have been hydrostatic, and P_{fluid} at any depth would have e equalled ca. 0.35 P_{load}. The head of water required to generate a hydrostatic pressure of 2.1 kb is about 21 km; at this depth, the difference between rock- and fluid pressure (ca. 4 kb) would be many times greater than the tensile strength of the rock, and the vein would collapse. The field relationships described in Chapter 3 part C(d) and the isotopic studies in Chapter 8 part C(d) suggest that the vein fluids were locally derived so that T_{rock} probably equalled T_{vein}.

CHAPTER 10

Conclusions

CHAPTER 10

Conclusions

A. A TECTONO-METAMORPHIC HISTORY FOR THE SOUTH-EAST TAUERN

The conclusions from Chapters 3, 4, 6, 7 and 9 can be combined with evidence from geochronological and other studies to form a brief geological history of the south-east Tauern Window.

The earliest recognisable event in the area studied is an eclogite facies metamorphism (M_e), the products of which are only locally preserved in the middle of a thick fault-bounded block of Inner Schieferhülle amphibolites in Dösener Tal. Garnet, omphacite, quartz and rutile crystallised at pressures greater than 11 kb and at temperatures greater than 540°C (though probably not exceeding 600°C), and a fine compositional banding was generated. The age of this metamorphism is not known, but is almost certainly greater than 245 mybp because the eclogites are cut by members of the Zentralgneis (see below). These are the only eclogites so far recorded from the Inner Schieferhülle; they may be temporally related to those of the Bohemian Massif (e.g. Dudek and Fediuková, 1974) but not to those in the Peripheral Schieferhülle of the Central Tauern (e.g. Miller, 1974, 1977; Holland, 1977, 1979, in press), the Bündnerschiefer of the Central and Western Alps (e.g. Bearth, 1973, 1974) or the Austro-Alpine basement in the Sesia-Lanzo zone (e.g. Ernst, 1973; Dal Piaz et al., 1972). They do not, therefore, indicate any involvement of the Pennine basement complex in the eo-Alpine high pressure metamorphism.

Eclogitic parageneses were subsequently hydrated and overprinted by an amphibolite mineralogy, particularly garnet and hornblende. This overprinting is correlated with the growth of pre-Alpine garnet, muscovite, biotite, kyanite and staurolite in Inner Schieferhülle mica-schists, and ascribed to a Variscan regional metamorphism (M_v). Temperatures during this event were probably between 625°C and 710°C, and pressures over 6 kb. Metamorphism was accompanied by an intense penetrative deformation (D_v) incorporating at least one phase of isoclinal folding.

The next major episode to affect the basement complex was the widespread intrusion of large volumes of tonalite, granodiorite, adamellite, leucogranite and quartz-syenite which now comprise much of the Zentralgneis. Minor amounts of more basic magma crystallised as hornblende-pyroxenite, and (possibly) feldspar-phyric basalt and ophitic gabbro, though the ages of the latter are unsubstantiated. Numerous veins of granite-aplite and pegmatite were injected late in this episode. Rb/Sr whole rock studies in various parts of the Tauern have yielded isochron ages clustering around 255 mybp for the leucogranites (Lambert, 1964; Besang et al., 1968; Cliff, 1968, 1971; Jäger et al., 1969) and an age of $279^{+}_{-}9$ mybp for granitic and granodioritic gneisses (Hawkesworth, 1976) suggesting that the bulk of the Zentralgneis was emplaced in early to middle Permian times.

Inner Schieferhülle rocks now exposed in the Hochalm Kern represent screens and roof-pendants of the country rock into which the Zentralgneis intrusions penetrated. Evidence of a phase of contact metamorphism (M_z) is preserved in pelitic schists that were heavily injected with granitic material. In particular, large Variscan muscovites show textures suggestive of incongruent melting to produce sillimanite, suggesting that metamorphic conditions were close to 750°C and 7 kb.

The events described above were confined entirely to the basement complex. After the intrusion of the Zentralgneis, there was a period of uplift and erosion during which the rocks now exposed were brought close to the earth's surface. Part of this eroded basement (not necessarily the part within the map areas) subsequently formed a substrate on which Mesozoic shallow-water sediments such as limestones, dolomites, ironstones and calcareous sandstones were deposited; these subsequently became the epicontinental part of the Peripheral Schieferhülle succession. Further south, in the Pennine Trough, deep water sediments such as carbonaceous shales and silty (possibly turbiditic) limestones were deposited on an oceanic-type lithosphere containing basaltic, gabbroic and peridotitic components. This group later became the calc-schist-graphitic mica-schist-serpentinite-metabasite association of the Peripheral Schieferhülle. The present work provides no evidence of original stratigraphic relationships; instead, it suggests that attempts to reconstruct the stratigraphy be means of physical correlation based on the principal of superposition are likely to be unsuccessful owing to the inhomogeneous nature of subsequent deformation and the high degrees of associated strain.

In the eastern Tauern, the onset of Alpine orogenesis is marked by the segregation of abundant quartz veins in mica-schists. This phase of segregation probably represents the initial dewatering of the sediments. Part of the evolved water may have been connate, the rest derived from dehydration reactions accompanying very low-grade metamorphism.

The first major structural event to affect the cover sequence was one of thrusting and imbrication; slices of ultrabasic and basic rock were emplaced into the pelagic metasediments, and slices of Zentralgneis granite into the epicontinental series. Rocks from these two essentially different depositional facies may have been juxtaposed at this stage. The thrusting phase appears to have developed into a phase of intense penetrative deformation, confined

mainly to the Peripheral Schieferhülle and structurally high levels in the basement. On a small scale, this phase generated isoclinal folds and an axial planar schistosity; on a large scale it produced a very even lithological banding or pseudo-stratigraphy in which stratal discontinuities generated by thrusting were largely obliterated. The cover rocks experienced very high degrees of strain, probably involving several hundreds, if not thousands, of percent shortening. At low levels in the basement, strain was confined largely to discrete shear-zones.

This major episode of thrusting and penetrative deformation (D_a^1) is tentatively correlated with the onset of continental collision and overthrusting of the Pennine units by the East Alpine sheets. The stratigraphic evidence of Oberhauser (1968) suggests that overthrusting began as early as 80 mybp, while the geochronological evidence of Brewer (1970) suggests that it was finished by 65 mybp, at least in the area surrounding the Eastern Tauern. However, it is possible that the Altkristallin sheet was not emplaced as a single rigid block. Overthrusting may have taken place by the northward shuffling of several semi-independent slices (e.g. the Ötztal and Silvretta masses). If this were the case, then Brewer's date for the end of overthrusting may be of local significance only.

The area studied provides very little evidence of metamorphic conditions during D_a^1. The few assemblages definitely of this age (e.g. the actinolite and epidote inclusions within pre-D_a^2 albites) suggest that greenschist facies conditions prevailed. Most greenschist parageneses were subsequently overprinted by an epidote-amphibolite mineralogy and show no evidence of the syn- to post-D_a^1 glaucophane growth which characterised the eo-Alpine metamorphism in the Central Tauern. Pseudomorphs after lawsonite are, however, preserved within pre-D_a^2 garnets in a specimen of mica-schist from close to the basement-cover interface, suggesting that a few rocks did experience

blueschist facies metamorphism (at temperatures not exceeding 450°C and pressures probably less than 11 kb).

D_a^1 was followed by a relatively static period of increasing temperature during which porphyroblasts were allowed to grow: albite in a wide variety of lithologies, garnet and chloritoid in aluminous pelites. This was followed by a second phase of penetrative deformation (D_a^2) which involved both the basement and the cover to a large extent, although the strain incurred in the latter was less than during D_a^1. Small, intermediate and large scale D_a^2 folds were produced. The largest recognisable folds of this generation are distinctly asymmetric and form the Sonnblick Kern, Mallnitzer Mulde, Hölltor Kern and Ankogel Mulde. Fold axes were developed in directions highly variable on the scale of the map areas, but broadly consistent within each major structure. The intensity of D_a^2 was greatest in the Mölltal Schieferhülle where most evidence of D_a^1 has been destroyed. There, the basement was folded into an isoclinal antiform, the Sonnblick Lamella, while further north-west, the same fold developed into a more open antiform, the Sonnblick Kern.

An important conclusion worth emphasising here is that the various gneiss lamellae exposed in the south-east Tauern were not all formed at the same time, and different generations may have been produced by different mechanisms. As stated above, the Sonnblick Lamella was formed essentially by extreme flattening in a folding deformation (D_a^2). However, the highly non-symmetrical arrangement of units flanking the quartz-phengite-schist bands in the Mallnitzer Mulde suggests that early thrusting (in D_a^1) played an important part in their formation.

The tectonic significance of D_a^2 is poorly understood. The asymmetry of the associated structures suggests that it, too, was produced in a phase of over-

thrusting (Norris et al., 1971b), but no thrusts or allochthonous units of this generation were identified within the map areas. The greatly increased degree of basement involvement in D_a^2 as compared with D_a^1 requires either that the overall stress-field changed or that the competence contrast between basement and cover became smaller with time. Such a contrast would certainly have been reduced by increasing temperature.

D_a^2 was followed by a static period during which a large proportion of rocks in both cover and basement were extensively recrystallised. This was the last major phase of crystallisation to affect the Tauern Window, and is commonly referred to in the literature as the "Tauernkristallisation". This was characterised by the growth of the highest grade assemblages found in the Peripheral Schieferhülle (e.g. garnet+staurolite+biotite$\pm$kyanite in aluminous pelites of the Hochalm Kern) and is therefore assumed to reflect the attainment of maximum Alpine metamorphic temperatures. Peak M_a assemblages show little evidence of subsequent overprinting, suggesting that maximum entropies were also attained at this stage.

In the eastern Tauern, biotite and muscovite K/Ar ages range from 11 to 26 mybp (Oxburgh et al., 1966; Cliff and Oxburgh, 1971) while to the west of the Sonnblick Kern, ages of between 27 and 34 mybp have been obtained (Lambert, 1970). As these have been interpreted as cooling ages, 34 mybp can be taken as the minimum age of the Alpine thermal peak.

Metamorphic conditions at the structural base of the Mölltal Schieferhülle were close to 570°C and 7 kb, but temperatures may have been as high as 610°C in the deepest exposed parts of the Hochalm Kern, and as low as 470°C in the Mallnitzer Mulde. The distribution of metamorphic zones and isograds based on M_a assemblages in pelites support the conclusion that highest temperatures were attained at lowest structural levels. Variations in

pressure are less well constrained; the lowest calculated pressure (5 kb) applies to the area which yielded the lowest temperature (470°C), indicating that the locus of points in PT space connecting the metamorphic conditions preserved in the exposed rocks (the metamorphic geotherm, England and Richardson, 1977) has a positive dP/dT. The shape of the chloritoid+biotite zone indicates that there also existed an isothermal pressure gradient in the Mölltal Schieferhülle from higher pressures to the south-east to lower pressures in the north-west.

The "Tauernkristallisation" was followed by a period of falling temperature, during which minerals no longer re-equilibrated with one another. This was also a time of increased sedimentation in the Molasse (Oberhauser, 1968) and is generally interpreted as one of rapid uplift and erosion of the Axial Zone (e.g. England, 1978). Uplift was accompanied by regional updoming, centred on the Hochalm Kern, which resulted in the rotation of isograds and earlier foliation surfaces away from their assumed original subhorizontal attitudes at the margins of the Tauern Window. A major steeply inclined fault, the Mölltal line, developed along the southern margin of the eastern Tauern, bringing Peripheral Schieferhülle rocks directly into contact with relatively high-levels in the Altkristallin. On the north (i.e. upthrow) side of this fault, isograds and S_a^2 foliation were rotated into a subvertical position. Localised, possibly polyphase, postmetamorphic buckling, kinking, crenulation, jointing, boudinage and chlorite+albite-veining (grouped together as D_a^3) formed in response to stresses generated during updoming. At one locality, chlorite+albite veins formed under estimated temperature and pressure conditions of 525°C and 2.1 kb, respectively. As K/Ar blocking temperatures in micas (ca. 200-350°C for biotite) are likely to be considerably lower than vein formation temperatures, the K/Ar cooling ages for the eastern Tauern (16 to 26 my) can be considered minimum vein ages. Also on the basis of K/Ar data, Cliff and Oxburgh (1971) concluded that temperatures had fallen

to about 250°C by 15 mybp.

Continued uplift, and falling temperature resulted in the exposure of the present-day erosion surface at approximately 0°C.

The tectono-metamorphic history outlined above is summarised in a pressure-temperature-time trajectory for the structural base of the Mölltal Schiefer-hülle (Fig. 10.1). Four consecutive time intervals characterised by different uplift rates can be distinguished:

(i) ?80 to 65 mybp: Rapid burial. Temperature increase not known, but probably no more than 400°C.

(ii) ?65 to 35 mybp: Temperature increase accompanied by either (a) slow uplift, or (b) neither uplift nor burial, or (c) slow burial. The evidence of Holland (1977) from the central Tauern favours (a).

(iii) ?35 to 26+ mybp: Rapid, almost isothermal uplift. A drop of only 50°C for 18 km eroded. (It must be emphasised that this conclusion rests on the assumption that the Auernig veins were in thermal equilibrium with the surrounding rocks, as explained in Chapter 9.)

(iv) ?26+ to 0 mybp: Uplift to present-day erosion surface. Rapid cooling. A drop of 525°C for 7 km eroded.

The general shape of the trajectory is similar to that calculated by England and Richardson (1977) for a tectonically thickened metamorphic pile undergoing concomitant erosion and conductive thermal relaxation. In their model, which is based on an assumed erosion history, geothermal gradients with high dP/dT are produced immediately after tectonic thickening because the rate of

conductive heat transfer through rock is low compared with convective transfer by tectonic movements. Once these movements have ceased, the immediate response of the buried rocks is to heat up either isobarically, or, if erosion has already begun, under conditions of falling pressure.

One significant difference between the theoretical PT trajectory and the observed one is that the former approaches a steady-state geothermal gradient asymptotically, while the latter overshoots it and enters the field of much higher T/P (stage (iii), above). This anomaly can be explained by differences in erosion history. In the theoretical model, erosion begins 20 my after overthrusting and the erosion rate decreases exponentially. In the eastern Tauern, the bulk of the erosion appears to have occurred between 30 and 39 my after overthrusting (i.e. between 35 and 26 mybp). Part of the evidence for this is provided by the geochronological studies mentioned above, and part by Oberhauser (1968) who has shown that the main Molasse deposition occurred in upper Oligocene and lower Miocene times (i.e. ca 26 mybp). Owing to the low thermal conductivity of rock, the effect of very rapid uplift is merely to displace the thermal gradient upwards and "behead" it, giving rise to relatively high temperatures at shallow levels (Oxburgh et al., 1971), which is what is observed during stage (iii).

The erosion of 18 km thickness of crust in 9 my implies an average erosion rate of ca. 2 mm per year. This is an order of magnitude lower than estimates of plate movement, and is not, therefore, geologically unreasonable.

England and Richardson (1977) predict that in a terrain where erosion begins relatively soon after tectonic thickening, and where erosion rate decays smoothly, rocks at different levels in the metamorphic pile reach their maximum temperatures at different times (i.e. the metamorphic geotherm is polychronic). Specifically, the rocks deepest within the pile are predicted

to take longest to reach their thermal maximum. If, however, tectonic thickening is followed by a relatively long period of time during which no erosion occurs and this is terminated abruptly by a phase of very rapid erosion, rocks at different levels in the metamorphic pile might be expected to reach their maximum temperatures at approximately the same time (i.e. at, or shortly after the onset of rapid erosion) and the preserved metamorphic geotherm would be isochronous. Thus, further constraints on the erosional history of the area would be provided if the type of metamorphic geotherm could be established.

Unfortunately, no definite conclusions about this can be drawn from the time relations of mineral growth (Chapter 4). In the low-grade area west of Mallnitz, very little post-D_a^2 recrystallisation and neomineralisation of AFM minerals occurred in pelites, in direct contrast to the Mölltal Schieferhülle and Hochalm Kern. This suggests that the Alpine thermal peak may have been earlier in the areas where lower peak temperatures were attained. Against this argument is the observation that post-D_a^2 oligoclase overgrowths are common to high-grade and low-grade areas, alike.

B. SOME CONSTRAINTS ON FLUID CIRCULATION

The thermal arguments in parts A and C depend upon the assumption that heat transfer during Alpine metamorphism was purely conductive, and that no widespread circulation of metamorphic waters occurred. In general, the conclusions of Chapters 7 and 8 bear this assumption out.

The calculated X_{H_2O}/X_{CO_2} ratios of fluids in equilibrium with M_a mineral assemblages, particularly in siliceous dolomites, vary in space, in some

cases over distances as small as 1 cm. This indicates that the rock systems were largely closed to fluid components during the Alpine metamorphic peak, and that the relative volumes of fluid interacting with given volumes of rock were small. The same conclusions were drawn from the variation in calculated δO^{18} composition of metamorphic waters. Neither set of results, however, precludes fluid movement, but such movement must have been on a small enough scale for the fluid composition to have been essentially buffered by the solid assemblage.

The spread in whole-rock oxygen isotope compositions among chemically and mineralogically similar rocks of the Peripheral Schieferhülle (e.g. mica-schists) suggests that the bulk of the metamorphic pile was not subjected to widespread fluid circulation and isotopic homogenisation at any stage in its Alpine history. However, the data of Hoernes and Friedrichsen (1974) imply that the basement complex was affected by fluid circulation and homogenisation in pre-Alpine times, presumably during the intrusion of the Zentralgneis.

Localised examples of Alpine open-system behaviour have been found in quartz-phengite-schists, serpentinites and mica-schists (Chapter 7, part G). In the first two, cation metasomatism occurred in sites of intense D_a^1 deformation. The shearing associated with D_a^1 probably induced open-system behaviour by (a) severely reducing the grainsize, thereby increasing the rate of ionic reactions, and (b) allowing a large volume of metamorphic fluid to interact with the sheared rocks. In the quartz-phengite-schists, a steady flow of this fluid may have prevented ionic reactions from ever reaching equilibrium by simultaneously supplying reactants and removing products. The effectiveness of shearing as a means of enhancing fluid movement has been demonstrated in other terrains (e.g. in the epidiorites of the Scottish Dalradian, Rast, 1958). However, the replacement of muscovite by post-D_a^2 plagioclase in

pelites was not associated with anydeformation, and, in the absence of evidence of fluid migration near the M_a thermal peak, diffusion appears to be the only possible mechanism of supplying the necessary ionic reactants. If the sources of these ions were deformed granites (as suggested in Chapter 7, part G) diffusion over tens, possibly hundreds of metres is implied, which seems unlikely. This apparent paradox would be resolved if nearer sources could be identified.

Late-stage veins are very abundant. Field and isotopic evidence suggests that the syn-D_a^3 chlorite+albite veins on Auernig were precipitated from fluids probably derived from the amphibolites which they transect. These fluids had very high $X_{H_2O}:X_{CO_2}$ ratios, and contained appreciable quantities of solute, equivalent to about 8 mole % NaCl. Many other veins however, are more likely to have been precipitated from relatively far-travelled fluids, and may have transported significant heat by convection in the post-metamorphic stages of the Alpine cycle. The gold+quartz veins of the Radhausberg (Exner, 1957) and veins which are bordered by wide zones of metasomatic alteration (e.g. Fig. 2.41) are examples of these.

C. SOME THERMAL CONSIDERATIONS

Oxburgh and Turcotte (1974), Bickle et al. (1975) and England (1976, 1978) have applied one-dimensional thermal models to the Eastern Alps in order to assess whether overthrusting of the Austro-Alpine nappes alone could have caused M_a metamorphism of the observed grade in the buried Pennine Zone, or whether it is necessary to invoke a transient high heat-flow regime during the Alpine orogenic cycle. Such a regime might have been generated either by the emplacement of large volumes of magma beneath the present erosion

surface in the Pennine Zone, or by a normal continental heat flow augmented by a higher-than-average contribution from the mantle. The above authors arrived at different conclusions mainly because they considered different ranges in the values of the modelling parameters.

The object of this section is not to provide arguments for or against the use of particular models, but to demonstrate how the results of this thesis help to narrow the range in acceptable values of the principal unknown quantity (the heat flow contribution from the underlying Pennine basement and mantle during Alpine metamorphism) when substituted into any given model.

The model of England (1978) has been chosen as an example mainly because the solutions are presented in terms of variations in all the principal variables (except erosion history) in an easily interpretable summary diagram (England, 1978, Fig. 3 - similar to Fig. 10.2 of this work). This diagram illustrates two important general conclusions common to <u>all</u> models:

(i) The combination of a low mantle contribution to heat flow and high basement contribution can have the same effect as a high mantle heat flow coupled with a low contribution from the basement. No model can distinguish between these two situations.

(ii) The heat flow from below required to raise the temperature of a rock to a given value, by a given length of time after tectonic burial, is lower for rocks deeper in the metamorphic pile.

The model of England (1978) considered a post-overthrusting structural succession composed of three units:

Unit 1 (top): Altkristallin sheet. Variable thickness. Initial post-overthrusting thermal gradient of 20°C/km with zero surface temperature. Uniform heat production of 5 hgu.

Unit 2: Peripheral Schieferhülle and Lower East Alpine sheet. 4 km thick. Negative linear initial thermal gradient with 0°C at base. Uniform heat production of 5 hgu. (All the strain associated with overthrusting is assumed to have been accommodated by this unit).

Unit 3: Pennine basement complex. Initial thermal gradient in equilibrium with mantle heat flow and an internal radiogenic heat production which decreased exponentially with depth. 0°C at top.

Temperature-dependent functions of thermal conductivity and specific heat were used in his calculations, and the following additional assumptions were made:

(i) A period of 30 my elapsed between overthrusting (65 mybp) and the peak of metamorphism, during which time no erosion occurred.

(ii) During erosion, the erosion rate decreased exponentially.

(iii) A temperature of 550°C was attained at the base of Unit 2.

(iv) Heat transfer within Units 1, 2 and 3 was by conduction.

The results of this thesis do not contradict any of the above assumptions except to suggest that the temperature at the base of Unit 2 was ca. 20°C higher than supposed.

The solutions to England's thermal model were presented in terms of variations in the following principal variables: heat flow from below Unit 3 ("mantle" heat flow), total heat flow contribution from Unit 3, and total thickness of Units 1 and 2. On the basis of existing estimates of peak

metamorphic pressure, England took values of 14 km and 31 km as the lower and upper limits, respectively, of the depth to the base of Unit 2. Similarly, the basement heat-production measurements of Hawkesworth (1974) were used to constrain the Unit 3 heat flow contribution to values of between 0.48 and 1.2 hfu.

The present results provide no extra information on heat production but they narrow the range of possible values of metamorphic pressure at the base of Unit 2 to between 6 and 8 kb (i.e. depths of between ca. 21 and 28 km, assuming a conversion factor of 3.5 km/kb). Assuming that the possible 20°C discrepancy has a negligible effect on the results, an Alpine "mantle" heat flow value of between 0.4 and 1.4 hfu is indicated (Fig. 10.2). Commonly quoted "normal" mantle heat flow values are of the order of 0.7 hfu (e.g. Oxburgh and Turcotte, 1974). At one extreme, therefore, the results of this thesis would be consistent with a model in which Alpine metamorphism took place by thermal relaxation below an overthrust sheet without a transient high heat flow. At the other extreme, the results would be consistent with a model in which metamorphism was accompanied by an abnormally high mantle heat flow.

England (1978) also calculated the present-day surface heat flow assuming an exponentially decaying erosion rate with a time constant of 60 my, and the results are also shown in Fig. 10.2. Observed surface heat flow values in the Tauern Window (ca. 1.6 to 2.0 hfu) are considerably lower than those predicted by the preferred range of depth and Unit 3 heat production solutions. However, if most of the erosion occurred immediately after the thermal peak, as suggested in part A, calculated values of present-day surface heat flow are up to 0.6 hfu lower than initially predicted, and more consistent with the observations.

The existence of a preserved M_a isothermal pressure gradient in the Mölltal Schieferhülle (Chapter 6) implies that a single one-dimensional thermal model is insufficient to explain the observed range in metamorphic conditions. The apparent increase in M_a pressure towards the south-east of the Mölltal Schieferhülle might have arisen in one of the following ways:

(i) Variable horizontal distribution of heat producing elements in the basement complex: Hawkesworth (1974) points out that there is a distinct correlation between lithology and heat production in the basement complex of the south-east Tauern. In particular, Inner Schieferhülle rocks tend to be less radiogenic than Zentralgneis lithologies, and some varieties of the latter (e.g. syenite gneiss) have very high values of heat production. If the basement structurally beneath the north-west Mölltal was significantly enriched in the more radiogenic lithologies compared with the basement elsewhere to depths of 10 km or more, the heat flow through the base of Unit 2 may hage been abnormally high in that area. Assuming a regionally uniform combined thickness of Units 1 and 2, mantle heat flow contribution and erosion history, this would have had the effect of generating an isobaric thermal gradient during metamorphism.

(ii) Variation in the pre-M_a peak uplift rate: England (1978) found that thermal models which allow 5 to 8 km of erosion between 65 and 35 mybp show no appreciable difference in the temperature at the base of Unit 2 at 35 mybp from those with no erosion. This implies that a difference in uplift rate alone could account for a singificant (ca 1 kb) pressure difference between points on a 35 mybp isotherm near the basement-cover interface.

The steepness of the negative slope of the pre-M_a peak trajectory deduced by Holland (1977) for the central Tauern as compared with those possible for the eastern Tauern (Fig. 7.22) might well be due to a difference in uplift

rate. If this were so, however, the observed 50°C difference in peak metamorphic temperature at the basement-cover interface between the two areas would appear to argue against the validity of England's findings (preceding paragraph).

Although there is some evidence that the Altkristallin sheet becomes thinner towards the north, it is unlikely that this would have had the effect of generating an isothermal pressure gradient because, provided that the heat flux through the base of Unit 2 and the conductivities of the overlying units were uniform, the thermal gradient in Units 1 and 2 would also have been uniform laterally as well as vertically in a steady-state situation.

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APPENDIX I : MINERAL ANALYSES

A. TECHNIQUES

Mineral analyses were performed on the Microscan 9 electron microprobe at the Department of Geology and Mineralogy, Oxford. Specimen preparation techniques were similar to those described by Sweatman and Long (1969). The following standards were used:

element	standard	counting time (secs)
Na	jadeite	30
Mg	periclase	15
Al	corundum	5
Si	wollastonite	25
K	orthoclase	15
Ca	wollastonite	15
Ti	rutile	10
Mn	fowlerite	15
Fe	haematite	10

Analyses were performed at 20 kV with a specimen current of 1.36×10^{-8} amps. Counting times were generally varied according to the element analysed, and are listed above. A few analyses, however, were performed with a specimen current of 3.72×10^{-8} amps and a counting time of 10 seconds for all elements. A raster with a scanning area of 3 x 3 microns was used for all minerals.

All analysed data were processed automatically using the on-line M-9 ZAF correction program. For silicates it was assumed that oxygen was the only

element that had not been measured, but for carbonates both oxygen and carbon were calculated by difference assuming an atomic ratio of 3:1.

B. RECALCULATION OF MINERAL ANALYSES

All mineral analyses were recalculated using the CALCULATE program written by R.A. Exley and D. Walker. All hydrous minerals were recalculated according to the number of oxygens in the formula unit, assuming all iron to be Fe^{2+}, except epidotes where it was assumed to be Fe^{3+}. The Fe^{3+} contents of anhydrous ferromagnesian minerals such as garnet and pyroxene were calculated and the resultant analyses recast according to the number of oxygens in the formula unit. The program uses a charge balance equation to calculate Fe^{3+}, as described in Ryburn, Råheim and Green (1976).

C. SHORT NOTES ON ANALYSED ROCKS

GD 136a Eclogitic amphibolite:

Locality : Alt 1680 m in Dösen Bach gully.

Assemblage : gt + cpx + hbl + ep + plag + qz + rut + ore.

Texture : Fresh eclogite patch composed of tiny M_e garnets enclosed in large random cpx prisms. Petrogressed area composed of large M_v garnets, sometimes rimming relict cpx, hornblende and hornblende+albite symplectite (Figs. 2.5, 2.6, 4.2).

Minerals analysed :

M_e garnets	: GTA 1-5, 8, 9, 16-20.
M_e pyroxenes	: PXA 1-18.
dark M_v garnet cores	: GTA 11, 12, 14.
M_v garnet rims	: GTA 6, 7, 10, 13, 15.
hornblende	: HBA 1-4.
plagioclase	: PLA 1.

GD 122 Eclogitic amphibolite:

Locality : Alt 1740 m on path, Dösener Tal.

Assemblage : gt + cpx + hbl + ep + plag + qz + sph + ore + ap.

Texture : Fine-grained, granulose except for hornblende which is poikiloblastic. Quartz+plagioclase total about 50%. Very small ragged ?M_v pyroxene is breaking down to hornblende.

Minerals analysed :

garnet	: GTB 1.
pyroxenes	: PXB 1, 2.
hornblende	: HBB 1
plagioclase	: PLB 1-4.
epidote	: EPB 1-5.

GD 22b Epidote-amphibolite

Locality : On road 400 m south-west of basement-cover interface, Zwenberger Tal.

Assemblage : gt + hbl + ep + biot + plag + qz + dol + cc + ore.

Texture : Large spongy hornblendes and zoned epidotes with a good parallelism. Abundant random biotite, some forming at the expense of garnet. Interstitial plagioclase. Numerous small polygonal calcite grains surround large iron-stained dolomites.

Minerals analysed : calcite : CCC 1-15.
dolomite : DLC 1-4.

GD 551 Greenschist

Locality : 50 m south-east of summit of Lonza Köpfl.

Assemblage : act + hbl + ep + chl + plag + qz + cc + sph.

Texture : Fine grained actinolite and chlorite aligned in S_a^1. Small slightly zoned epidotes. Large rotated feldspar porphyroblasts. Thin hornblende rims round actinolites (Fig. 4.8).

Minerals analysed : actinolite : ACD 1-6.
hornblende : HBD 1-4
epidote : EPD 1-6.
chlorite : CHD 1-7.
plagioclase cores : PLD 3, 4.
plagioclase rims : PLD 1, 2.

GD 193 Epidote-amphibolite

Locality : Alt 1500 m due south of Sicker Kopf, north flank of Kaponigtal.

Assemblage : hbl + ep + chl + biot + plag + qz + rut + ap.

Texture : Random euhedral poikiloblasts of hornblende overgrowing

aligned epidotes and cutting across chlorites aligned in S_a^2. Interstitial plagioclase and quartz (Fig. 2.35).

Minerals analysed : hornblende : HBE 1-11.
epidote : EPE 1-8.
chlorite : CHE 1-7.
plagioclase : PLE 1-9.

GD 160b Siliceous dolomite

Locality : Alt 1690 m, 350 m west of summit of Labiges Köpfl.

Assemblage : cc + di + trem + ep + plag + biot + qz + sph + pyr + ap.

Texture : Euhedral epidote granules locally enclosed by prismatic diopside altering to tremolite. Irregular patches of plagioclase. Quartz is rare.

Minerals analysed : diopside : DIF 1.
tremolite : TRF 1, 2.
epidote : EPF 1, 2.
plagioclase : PLF 1-4.

GD 992b Calc-schist

Locality : Alt 1665 m, 1.1 km south-east of Jamnig Alm.

Assemblage : cc + dol + biot + ep + musc + plag + qz + rut + pyr + tour + ap.

Texture : Iron stained dolomite concentrated in certain bands contacts calcite. Muscovite has good S_a^1? schistosity. Tiny granules of rutile generally associated with biotite. Rare epidotes are zoned out to clinozoisite rims. Plagioclases are zoned with rims replacing muscovite.

Minerals analysed : calcite : CCG 1-25.
dolomite : DLG 1-20.

GD 145a Garnetiferous feldspathic paragneiss

Locality : Alt 1570 m, 100 m north-east of basement-cover interface, Kaponig Tal.

Assemblage : gt + hbl + ep + biot + musc + plag + qz + mt + tour.

Texture : Small euhedral M_a garnets and tiny euhedral hornblendes cut across S_a^2 muscovite schistosity. Epidote is very abundant and forms polycrystalline patches elongated in S_a^2. Plagioclase tends to be poikiloblastic.

Minerals analysed : garnet : GTH 1-16.

hornblende : HBH 1-6.

epidote : EPH 1-5.

plagioclase : PLH 1-8.

GD 209a Inner Schieferhülle aluminous pelite

Locality : Alt 1775 m, due south of 2682 m spot height, Kaponigtal.

Assemblage : (a) st + ky + parag + plag + qz

(b) gt + st + ky + biot + chl + musc + parag + plag + ilm + ap.

Texture : Banded. Monomineralic, fine-grained paragonite aligned in S_a^2 is cut across by euhedral staurolite and kyanite and partially replaced by oligoclase in assemblage (a) (Figs. 2.21, 2.22). Elsewhere, micas are both decussate and aligned in S_a^2. Two generations of garnet occur: M_a garnets, forming isolated "pin prick" grains and rims round larger M_v garnets. Euhedral M_a staurolite and kyanite. Late porphyroblasts of plagioclase and chlorite replacing micas (Figs. 2.18, 2.19).

Minerals analysed : M_a garnet : GTI 1, 2.

staurolite : STI 1-3.

biotite : BII 1,2.

muscovite : MUI 1, 2.

schistose paragonite in (a) : PAI 1-4.

chlorite : CHI 1.

plagioclase : PLI 1.

GD 209b Inner Schieferhülle aluminous pelite

Locality : As for GD 109a.

Assemblage : gt + st + ky + biot + chl + musc + qz + plag + ilm + tour.

Texture : Large inclusion-filled M_v garnets and "pin prick" M_a garnets both present. Muscovite mostly parallel to S_a^2. Random cross-cutting M_a biotite, staurolite and kyanite. Late porphyroblasts of chlorite and oligoclase replace micas.

Minerals analysed : M_v garnets : GTJ 10-15.

M_a garnets : GTJ 1-9.

staurolite : STJ 1-12.

kyanite : KYJ 1-4.

biotite : BIJ 1-9.

chlorite : CHJ 1.

muscovite : MUJ 1.

GD 208e Inner Schieferhülle aluminous quartzite

Locality : Alt 1760 m, due south of 2682 spot height, Kaponigtal.

Assemblage : gt + musc + parag + plag + qz + ilm.

Texture : Fine grained. All micas aligned in S_a^2. Garnets very rare $<1\%$. Quartz and plagioclase as flattened polygonal grains elongated in S_a^2. The latter is in equilibrium with the micas.

Minerals analysed : garnet : GTK 1-9.

muscovite : MUK 1-17.

paragonite : PAK 1-14.

plagioclase : PLK 1-14.

GD 523 Inner Schieferhülle mica-schist

Locality : Alt 1555 m, 1 km east of Schwussner Hütte, Seebach Tal.

Assemblage : gt + biot + chl + musc + plag + qz + ore + tour + ap.

Texture : Small rounded M_a garnets cutting across biotite and muscovite schistosity. Polygonal plagioclase appears to be in equilibrium with micas. Large sprays of chlorite are probably not in equilibrium with other phases.

Minerals analysed :

garnet	: GTL 1-4.
muscovite	: MUL 1, 2.
biotite	: BIL 1, 2.
chlorite	: CHL 1, 2.
plagioclase	: PLL 1-5.

GD 465b Inner Schieferhülle aluminous pelite

Locality : Alt 2030 m, 700 m west-south-west of old Giessener Hütte site, Gössgraben.

Assemblage : (a) gt + st + biot + chl + musc + qz + il.

(b) gt + st + ky + biot + chl + musc + parag + plag.

Texture : Assemblages (a) and (b) are interbanded. The band with assemblage (a) is very rich in staurolite, biotite and sprays of chlorite. The band with assemblage (b) is composed largely of very fine-grained decussate muscovite and paragonite. All the garnets are the M_a "pin prick" type. Some are enclosed by staurolite.

Minerals analysed :

garnet in (a)	: GTM 1-7.
staurolite	: STM 1-8.
biotite in (a)	: BIM 1-5.
muscovite	: MUM 1-24.
paragonite	: PAM 1-13.
chlorite	: CHM 1-4.

GD 80b Inner Schieferhülle mica-schist

Locality : 200 m from Schmidt Haus on south shore of Dösener See.

Assemblage : biot + chl + ep + musc + plag + qz.

Texture : Medium to coarse grained schist composed of mimetically recrystallised pale orange biotite, muscovite and almost colourless chlorite in roughly equal proportions. Plagioclase forms ragged poikiloblasts enclosing micas.

Minerals analysed : biotite : BIN 1-3.
chlorite : CHN 1-3.
muscovite : MUN 1-4.
plagioclase : PLN 1, 2.

C 523 Inner Schieferhülle mica-schist

Locality : Kaponigtörl.

Assemblage : ky + biot + chl + musc + plag + qz + sph

Texture : A dark, coarse-grained schist. 1-2 cm blades of M_v kyanite are bent round an F_a^2 microfold and have been partially altered to M_a colourless chlorite, muscovite and plagioclase. Abundant pale orange biotite has mimetically recrystallised round the fold. Some forms inclusions in kyanite. Some plagioclase appears to have replaced muscovite.

Minerals analysed : biotite : BIO 1-7.
chlorite : CHO 1-3.
muscovite : MUO 1-5.
plagioclase : PLO 1.

GD 168 Peripheral Schieferhülle pelitic mica-schist

Locality : 50 m north-east of summit of Labiges Köpfl.

Assemblage : ctd + biot + musc + parag + qz + ep + sph + ore + ap.

Texture : Very fine-grained muscovite±paragonite form a schistose matrix. Tight F_a^2 microfolds bend these micas. Two generations of chloritoid: 3 mm pre-F_a^2 plates, and smaller cross-cutting flakes. Biotite is exclusively post-F_a^2, and generally associated with smaller chloritoids.

Mineral analysed :

pre-F_a^2 chloritoid	: CTP 7-10.
post-F_a^2 chloritoid	: CTP 1-6, 11, 12.
biotite	: BIP 1-8.
muscovite	: MUP 1-14.

GD 168b Peripheral Schieferhülle pelitic mica-schist

Locality : As for GD 168.

Assemblage : gt + ctd + biot + chl + musc + qz + ore.

Texture : Large post-S_a^2 garnet poikiloblasts enclose quartz stringers. Chloritoid and biotite generally cut across S_a^2, and appear to be in equilibrium (Fig. 6.4). Some chlorite is retrogressive, but some may be primary.

Minerals analysed :

garnet (complete profile $\perp S_a^2$)	: GTQ 1-19.
garnet (profile average)	: GTQ 20.
chloritoid	: CTQ 1-5.
biotite	: BIQ 1-5.

GD 18 Peripheral Schieferhülle pelitic mica-schist

Locality : From 10 cm thick band in calc-schist, 100 m south-west of basement-cover interface Zwenberger Tal.

Assemblage : gt + st + ky + biot + ctd + chl + musc + parag + qz + plag + ep + ilm + ap.

Texture : Fine-grained paragonite±muscovite aligned in S_a^1 and subsequently bent by F_a^2 microfolds. Some muscovite parallel to

S_a^2 strain-slip schistosity, the rest cross-cutting. Large garnet porphyroblasts, with euhedral post-F_a^2 rims, enclose chloritoid, epidote, chlorite and staurolite. Euhedral post-F_a^2 staurolite and kyanite, often close to garnet. Biotite aligned in S_a^2 or cross-cutting, some altering to chlorite, Some kyanite altering to plagioclase (Figs. 4.19, 6.13).

Minerals analysed :

garnet (half zoning profile)	: GTR 10-18.
garnet (adjacent to chloritoid, etc., inclusions)	: GTR 1-3.
garnet rims (adjacent to biotite, etc.):	GTR 4-9, 19, 20.
chloritoid	: CTR 1, 2.
staurolite inclusion	: STR 1, 7.
staurolite in matrix	: STR 2-6.
chlorite inclusions	: CHR 1, 2.
biotite	: BIR 1-6.
plagioclase	: PLR 1.
paragonite	: PAR 1-4.
muscovite	: MUR 1-4.

GD 234 b Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1406 m on road up west side of Mühldorfer Graben.

Assemblage : ctd + chl + musc + qz + mt.

Texture : Strongly pleochroic blocky chloritoid tablets lie generally parallel to S_a^2, as defined by muscovite and chlorite.

Minerals analysed :

chloritoid	: CTS 1-4.
chlorite	: CHS 1-4.
muscovite	: MUS 1-4.

GD 308c Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1675 m on track 2.1 km south-south-west of summit of Kampleck.

Assemblage : gt + st + ctd + chl + musc + qz + ep + ore + tour + ap.

Texture : Rare flat post-S_a^2 garnets. Abundant blocky chloritoid tablets and chlorite and muscovite flakes aligned in S_a^2. Tiny euhedral staurolites and chlorite growing at the expense of chloritoid (Fig. 6.7).

Minerals analysed :

garnet (complete profile $\parallel S_a^2$)	: GTT 1-13.
garnet (complete profile $\perp S_a^2$)	: GTT 20, 19, 18, 6, 14-17.
staurolite	: STT 1-6.
chloritoid	: CTT 1-4.
chlorite	: CHT 1-5.
muscovite	: MUT 1, 2.

GD 43a Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1360 m on track 800 m south-west of basement-cover interface, Zwenger Tal.

Assemblage : gt + chl + biot + ep + musc + plag + qz + mt + tour.

Texture : Epidote and plagioclase are more abundant than in most other pelites (ca. 30%) and apparently in equilibrium with other phases. Chlorite and biotite are in equilibrium (Fig. 4.22). Subhedral ?post-S_a^2 garnets.

Minerals analysed :

epidote	: EPU 1-4.
plagioclase	: PLU 1-6.
biotite	: BIU 1.
chlorite	: CHU 1.
muscovite	: MUU 1, 2.

GD 280b Peripheral Schieferhülle pelitic mica-schist

Locality : 80 m north-east of summit of Sicker Kopf.

Assemblage : ctd + biot + chl + musc + qz + ep + plag + ore.

Texture : Uniformly fine-grained. Chloritoid and chlorite mostly aligned in S_a^2. Biotite mostly cross-cutting. Rare ragged plagioclases replace muscovite.

Minerals analysed :
- chloritoid : CTV 1, 2.
- chlorite : CHV 1-3.
- biotite : BIV 1, 2.
- plagioclase : PLV 1, 2.
- muscovite : MUV 1, 2.

GD 821d Coarse-grained chloritoid-schist

Locality : Alt 2030 m, Tauern Bach.

Assemblage : gt + ctd + biot + chl + ab + musc + mt + ap.

Texture : Large (1 cm+) M_a porphyroblasts of albite, chloritoid and garnet (which appear to have grown in that order) are all post-S_a^1 as defined by the muscovite schistosity. Together with garnet and chloritoid, ragged cross-cutting flakes of biotite are altering to bright green chlorite, intermeshed with random muscovite flakes (Fig. 2.42).

Minerals analysed :
- garnet rims : GTW 1-3.
- chloritoid rims : CTW 1-3.
- albite rims : PLW 1-4.
- S_a^1 muscovites : MUW 4, 5.
- cross-cutting muscovites: MUW 1-3, 6.

GD 480 Hornblende-garbenschiefer

Locality : At contact between calc-schist and pelitic mica-schist, 200 m north-east of Wasenle Kopf.

Assemblage : gt + hbl + ctd + ep + biot + chl + marg + musc + plag + qz + ore + ap.

Texture : Radiating sheaves of hornblende needles and garnet porphyroblasts are conspicuous in hand specimen (Fig. 2.44). In thin section fine-grained schistose margarite ϟ is very abundant, locally replaced by ragged plagioclase porphyroblasts. Muscovite is also very abundant. Garnet, hornblende and chloritoid porphyroblasts cut across schistosity, while epidote needles lie within it. Chlorite is retrogressive.

Minerals analysed : garnet rims : GTX 1-4.
hornblende : HBX 1-6.
margarite : MAX 1-6.
muscovite : MUX 1-3.
epidote : EPX 1-4.
plagioclase : PLX 1, 2.

GD 1233 Syenite-gneiss

Locality : Alt 2360 m, on footpath 300 m east of summit of Romate Spitze.

Assemblage : cpx + biot + ep + hbl + ksp + plag + qz + cc + sph + ap.

Texture : A coarse, virtually unfoliated rock carrying distinct pseudomorphs after euhedral? hornblende (Fig. 2.30) composed of green clinopyroxene, biotite and epidote in a well recrystallised mosaic (Fig. 2.31). Matrix composed predominantly of large blocky orthoclase and "gefüllte" plagioclase crystals. Quartz and calcite are subordinate. Some fine-grained recrystallised plagioclase is also present.

Minerals analysed : clinopyroxene : PXY 1-8.

epidote : EPY 1-3.

biotite : BIY 1-3.

recrystallised plagioclase : PLY 1-6.

GD 1234b Mafic lens in syenite-gneiss

Locality : As for GD 1233.

Assemblage : cpx + biot + hbl + ksp + plag + qz + sph + mt + ap.

Texture : Dark, medium-grained schist composed of well-recrystallised subequant granules of pyroxene, plagioclase and quartz and plates of biotite. Hornblende may be an alteration product of pyroxene. K-feldspar is rare.

Minerals analysed : clinopyroxene : PXZ 1-6.

plagioclase : PLZ 1-6.

K-feldspar : KFZ 1-4.

GD 468 Leucogranite

Locality : Alt 2195 m, 700 m west-north-west of old Giessener Hütte site, Gössgraben.

Assemblage : gt + biot + musc + ksp + plag + qz + chl + ap.

Texture : An unfoliated rock containing purplish quartz and white feldspar. Difficult to say to what extent texture is igneous or metamorphic. K-feldspar occurs as both large grains of microcline and smaller polygonal untwinned grains. Plagioclase also occurs in two sizes. Abundant myrmekitic intergrowths of plagioclase+quartz are replacing microcline (Fig. 2.25). Garnet forms minute irregular grains along polygonal-feldspar grain-boundaries. The two micas occur in equal proportions.

Minerals analysed	:	garnet	: GT& 1-4.
		biotite	: BI& 1-5.
		muscovite	: MU& 1-5.
		microcline	: KF& 4-7.
		polygonal K-feldspar	: KF& 1-3.
		myrmekitic plagioclase	: PL& 4, 6, 7.
		polygonal plagioclase	: PL& 1-3.
		large plagioclase core	: PL& 5.

GD 526 ?Peripheral Schieferhülle pelitic mica-schist

Locality : Loose block from footpath at Alt 2450 m south of Ebeneck.

Assemblage : chl + ilm + ep + musc + parag + plag.

Texture : Very fine-grained mesh of muscovite and paragonite makes up 90% of rock. Large ilmenite plates are abundant and have been bent. Small euhedral epidotes have sharply demarcated clinozoisite rims. Plagioclases are ragged and replace white micas.

Minerals analysed	:	muscovite	: MU& 1-25.

COMPOSITION IN OXIDES

	GTA 1	GTA 2	GTA 3	GTA 4	GTA 5	GTA 6	GTA 7	GTA 8	GTA 9	GTA10
SiO2	39.42	39.24	38.92	39.83	39.64	39.40	39.67	39.56	39.00	39.53
Al2O3	22.26	20.71	21.73	22.08	21.36	22.26	22.37	22.14	21.85	22.14
TiO2	0.14	0.14	0.14	0.16	0.14	0.18	0.18	0.14	0.16	0.14
Fe2O3	0.48	1.41	0.30	0.00	1.14	0.98	0.61	1.35	0.01	1.06
MgO	3.16	3.22	3.04	3.15	3.23	2.87	2.78	3.31	2.43	2.85
FeO	28.81	27.93	28.32	29.02	27.96	28.37	28.62	27.66	29.55	27.73
MnO	1.37	1.82	0.92	1.03	1.60	0.45	0.67	1.17	0.49	0.55
CaO	8.83	8.98	9.12	8.85	8.98	10.24	9.98	9.44	10.26	10.45
Na2O	0.02	0.00	0.06	0.00	0.14	0.04	0.12	0.12	0.04	0.12
K2O	0.01	0.01	0.01	0.00	0.01	0.01	0.00	0.01	0.01	0.01
	104.50	103.46	102.56	104.12	104.20	104.80	105.00	104.91	104.60	104.64

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTA 1	GTA 2	GTA 3	GTA 4	GTA 5	GTA 6	GTA 7	GTA 8	GTA 9	GTA10
Si	2.986	3.014	3.001	3.020	3.013	2.975	2.988	2.980	3.018	2.996
Al	0.014	0.000	0.000	0.000	0.000	0.025	0.012	0.020	0.000	0.014
	3.000	3.014	3.001	3.020	3.013	3.000	3.000	3.000	3.018	3.010
Al	1.974	1.875	1.975	1.973	1.914	1.956	1.974	1.946	1.953	1.956
Ti	0.008	0.008	0.008	0.009	0.008	0.010	0.010	0.008	0.009	0.008
Fe3	0.027	0.082	0.017	0.000	0.065	0.056	0.035	0.077	0.001	0.061
Mg	0.357	0.369	0.349	0.356	0.366	0.323	0.312	0.372	0.275	0.321
Fe2	1.825	1.794	1.826	1.840	1.778	1.791	1.803	1.743	1.873	1.755
Mn	0.088	0.118	0.060	0.066	0.103	0.029	0.043	0.075	0.031	0.035
Ca	0.717	0.739	0.753	0.719	0.731	0.828	0.805	0.762	0.833	0.946
Na	0.003	0.000	0.009	0.000	0.021	0.006	0.018	0.018	0.006	0.018
K	0.001	0.001	0.001	0.000	0.001	0.001	0.000	0.001	0.001	0.001
	5.000	4.986	4.999	4.964	4.987	5.000	5.000	5.000	4.982	5.000

AL% IN Z	0.45	0.00	0.00	0.00	0.00	0.84	0.39	0.66	0.00	0.48
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	11.84	11.89	11.62	11.94	12.03	10.67	10.41	12.28	9.11	10.63
Fe	64.38	64.29	63.32	63.94	63.94	61.96	62.72	62.56	63.23	61.34
Ca	23.78	23.83	25.06	24.12	24.03	27.37	26.87	25.17	27.65	28.03

Ti-ANDRT	0.40	0.00	0.00	0.00	0.00	0.52	0.40	0.40	0.00	0.40
ANDRADTE	1.38	4.05	0.87	0.00	3.29	2.81	1.75	3.99	0.04	3.07
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	22.22	20.42	24.34	24.12	21.27	24.55	25.04	21.52	27.62	25.13
SPESSRTN	2.94	3.92	2.01	2.22	3.46	0.97	1.44	2.53	1.04	1.19
ALMNDINE	61.11	59.40	61.10	61.72	59.69	60.28	60.84	59.06	62.13	59.35
PYROPE	11.95	12.21	11.69	11.94	12.29	10.87	10.53	12.59	9.11	10.85

COMPOSITION IN OXIDES

	GTA11	GTA12	GTA13	GTA14	GTA15	GTA16	GTA17	GTA18	GTA19	GTA20
SiO2	38.26	38.31	38.30	38.28	38.74	38.75	38.68	38.64	38.87	38.70
Al2O3	19.07	19.63	20.96	20.27	21.09	20.25	21.16	20.73	20.69	20.31
TiO2	0.35	0.35	0.14	0.19	0.14	0.14	0.14	0.14	0.12	0.18
Fe2O3	1.97	0.69	0.39	0.40	0.00	0.39	0.30	0.28	0.03	1.01
MgO	2.72	2.98	2.82	2.71	2.86	3.06	3.00	3.01	2.78	3.02
FeO	26.16	28.29	28.41	28.54	28.23	28.73	28.51	28.41	28.25	28.48
MnO	1.14	0.12	0.41	1.37	0.63	1.39	0.77	1.00	1.24	1.36
CaO	10.56	9.39	9.35	8.35	9.45	8.41	9.09	8.72	9.32	8.67
Na2O	0.08	0.08	0.02	0.04	0.06	0.02	0.02	0.08	0.04	0.02
	100.31	99.84	100.80	100.45	101.20	101.14	101.67	101.01	101.33	101.75

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTA11	GTA12	GTA13	GTA14	GTA15	GTA16	GTA17	GTA18	GTA19	GTA20
Si	3.035	3.045	3.011	3.033	3.027	3.044	3.013	3.031	3.041	3.025
	3.035	3.045	3.011	3.033	3.027	3.044	3.013	3.031	3.041	3.025
Al	1.783	1.839	1.942	1.893	1.942	1.875	1.943	1.917	1.907	1.871
Ti	0.021	0.021	0.008	0.011	0.008	0.008	0.008	0.008	0.007	0.011
Fe3	0.117	0.041	0.023	0.024	0.000	0.023	0.017	0.016	0.002	0.060
Mg	0.322	0.353	0.330	0.320	0.333	0.358	0.348	0.352	0.324	0.352
Fe2	1.736	1.881	1.868	1.911	1.845	1.888	1.858	1.864	1.849	1.862
Mn	0.077	0.008	0.027	0.092	0.042	0.092	0.051	0.066	0.082	0.090
Ca	0.898	0.800	0.788	0.709	0.791	0.708	0.759	0.733	0.781	0.726
Na	0.012	0.012	0.003	0.006	0.009	0.003	0.003	0.012	0.006	0.003
	4.965	4.955	4.989	4.967	4.971	4.956	4.987	4.969	4.959	4.975

	GTA11	GTA12	GTA13	GTA14	GTA15	GTA16	GTA17	GTA18	GTA19	GTA20
Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

	GTA11	GTA12	GTA13	GTA14	GTA15	GTA16	GTA17	GTA18	GTA19	GTA20
Mg	10.21	11.45	10.88	10.47	11.06	11.67	11.49	11.61	10.67	11.39
Fe	61.28	62.60	63.18	66.33	62.66	65.26	63.50	64.21	63.61	65.11
Ca	28.51	25.94	25.94	23.20	26.28	23.06	25.02	24.18	25.72	23.50

	GTA11	GTA12	GTA13	GTA14	GTA15	GTA16	GTA17	GTA18	GTA19	GTA20
Ti-ANDRT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ANDRADTE	5.81	2.03	1.15	1.17	0.00	1.13	0.87	0.82	0.10	2.96
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	23.80	24.26	24.99	22.21	26.28	22.11	24.29	23.49	25.64	21.01
SPESSRTN	2.53	0.27	0.91	3.03	1.38	3.04	1.68	2.20	2.71	2.97
ALMNDINE	57.26	61.83	61.99	63.03	61.27	61.97	61.60	61.82	60.88	61.45
PYROPE	10.61	11.61	10.97	10.56	11.06	11.76	11.55	11.67	10.68	11.61

COMPOSITION IN OXIDES

	GTB 1	GTH 1	GTH 2	GTH 3	GTH 4	GTH 5	GTH 6	GTH 7	GTH 8	GTH 9
SiO2	37.66	38.65	37.05	37.36	37.33	38.36	37.27	37.58	38.15	37.46
Al2O3	21.01	21.47	19.87	20.27	20.24	21.35	20.05	21.20	21.54	21.41
TiO2	0.18	0.09	0.09	0.11	0.16	0.12	0.07	0.07	0.07	0.12
Fe2O3	1.22	0.00	1.28	0.81	0.48	0.28	1.37	1.09	0.48	0.94
MgO	1.84	1.29	0.44	0.48	0.39	0.69	0.58	1.93	0.87	1.12
FeO	27.72	27.70	24.89	24.26	25.63	26.15	26.00	30.75	27.11	26.79
MnO	0.39	3.56	5.07	4.93	3.42	2.19	2.12	2.28	1.99	1.72
CaO	10.55	9.57	10.50	11.43	11.68	12.74	12.03	6.49	11.64	11.22
Na2O	0.06	0.00	0.02	0.02	0.00	0.00	0.00	0.04	0.02	0.00
K2O	0.00	0.03	0.01	0.01	0.01	0.02	0.01	0.00	0.00	0.00
	100.63	102.36	99.22	99.58	99.34	101.90	99.51	101.43	101.87	100.73

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTB 1	GTH 1	GTH 2	GTH 3	GTH 4	GTH 5	GTH 6	GTH 7	GTH 8	GTH 9
Si	2.978	3.016	3.007	3.009	3.013	3.001	3.003	2.977	2.989	2.966
Al	0.022	0.000	0.000	0.000	0.000	0.000	0.000	0.023	0.011	0.034
	3.000	3.016	3.007	3.009	3.013	3.001	3.003	3.000	3.000	3.000
Al	1.937	1.975	1.901	1.924	1.926	1.969	1.904	1.956	1.978	1.970
Ti	0.011	0.005	0.005	0.007	0.010	0.007	0.004	0.004	0.004	0.007
Fe3	0.072	0.000	0.078	0.049	0.029	0.017	0.083	0.065	0.028	0.050
Mg	0.217	0.150	0.053	0.058	0.047	0.080	0.070	0.228	0.102	0.132
Fe2	1.834	1.808	1.689	1.634	1.730	1.711	1.752	2.037	1.776	1.774
Mn	0.026	0.235	0.349	0.329	0.234	0.145	0.145	0.153	0.132	0.115
Ca	0.894	0.800	0.913	0.986	1.010	1.068	1.038	0.551	0.977	0.952
Na	0.009	0.000	0.003	0.003	0.000	0.000	0.000	0.006	0.003	0.000
K	0.000	0.003	0.001	0.001	0.001	0.002	0.001	0.000	0.000	0.000
	5.000	4.976	4.993	4.991	4.987	4.999	4.997	5.000	5.000	5.000

Al% IN Z	0.72	0.00	0.00	0.00	0.00	0.00	0.00	0.77	0.37	1.13
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	7.13	5.01	1.73	1.89	1.54	2.66	2.26	7.51	3.37	4.37
Fe	63.50	68.26	68.65	65.85	65.34	61.98	64.12	74.33	64.22	64.14
Ca	29.38	26.73	29.62	32.27	33.12	35.35	33.63	18.16	32.41	31.49

Ti-ANDRT	0.54	0.00	0.00	0.00	0.00	0.00	0.00	0.21	0.21	0.36
ANDRADTE	3.66	0.00	3.90	2.46	1.45	0.83	4.15	3.28	1.42	2.51
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	25.89	26.73	26.49	30.34	31.99	34.72	30.40	15.06	31.09	29.14
SPESSRTN	0.88	7.86	11.60	10.96	7.74	4.83	4.81	5.15	4.42	3.38
ALMNDINE	61.73	60.39	56.23	54.33	57.27	56.94	58.31	68.62	59.47	59.66
PYROPE	7.30	5.01	1.77	1.92	1.55	2.68	2.32	7.68	3.40	4.45

COMPOSITION IN OXIDES

	GTH10	GTH11	GTH12	GTH13	GTH14	GTH15	GTH16	GTI 1	GTI 2	GTJ 1
SiO2	37.49	37.77	37.68	38.02	37.93	38.05	37.78	38.55	39.88	37.84
Al2O3	21.26	21.22	21.31	21.08	21.19	21.26	21.29	21.96	22.37	21.50
TiO2	0.07	0.12	0.04	0.58	0.09	0.11	0.14	0.07	0.05	0.09
Fe2O3	1.07	0.73	0.74	0.00	0.75	0.44	0.70	0.00	0.00	0.65
MgO	2.06	0.91	1.45	1.49	0.82	1.00	1.29	3.32	5.14	4.32
FeO	31.67	26.23	28.82	28.66	26.53	26.14	28.06	28.94	29.38	34.30
MnO	1.82	1.96	2.34	2.12	2.42	2.31	2.11	6.39	0.85	1.89
CaO	5.94	12.05	8.83	9.66	11.56	11.90	9.92	3.12	5.72	0.89
Na2O	0.02	0.00	0.00	0.02	0.04	0.02	0.02	0.02	0.02	0.06
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00
	101.41	100.98	101.21	101.63	101.32	101.22	101.31	102.38	103.42	101.54

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTH10	GTH11	GTH12	GTH13	GTH14	GTH15	GTH16	GTI 1	GTI 2	GTJ 1
Si	2.972	2.984	2.982	2.991	2.991	2.996	2.982	3.005	3.020	2.982
Al	0.028	0.016	0.018	0.009	0.009	0.004	0.018	0.000	0.000	0.018
	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.005	3.020	3.000
Al	1.959	1.959	1.969	1.946	1.960	1.968	1.963	2.017	1.996	1.978
Ti	0.004	0.007	0.002	0.034	0.005	0.007	0.008	0.004	0.003	0.005
Fe3	0.064	0.043	0.044	0.000	0.044	0.026	0.041	0.000	0.000	0.038
Mg	0.243	0.107	0.171	0.175	0.096	0.117	0.152	0.386	0.580	0.507
Fe2	2.100	1.733	1.903	1.886	1.749	1.721	1.853	1.886	1.860	2.260
Mn	0.122	0.131	0.157	0.141	0.162	0.154	0.141	0.422	0.055	0.126
Ca	0.505	1.020	0.749	0.814	0.977	1.004	0.839	0.261	0.464	0.075
Na	0.003	0.000	0.000	0.003	0.006	0.003	0.003	0.003	0.003	0.009
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.001	0.000
	5.000	5.000	5.000	4.999	5.000	5.000	5.000	4.980	4.962	5.000

Al% IN Z	0.93	0.55	0.61	0.30	0.30	0.14	0.59	0.00	0.00	0.61
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	8.02	3.53	5.65	5.79	3.18	3.88	5.02	13.05	19.60	16.87
Fe	75.35	62.85	69.63	67.21	64.57	62.90	67.26	78.13	64.71	80.63
Ca	16.63	33.62	24.72	27.00	32.25	33.22	27.73	8.82	15.68	2.50

Ti-ANDRT	0.21	0.36	0.12	0.30	0.27	0.14	0.42	0.00	0.00	0.27
ANDRADTE	3.24	2.17	2.21	0.00	2.23	1.30	2.08	0.00	0.00	1.95
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	13.54	31.58	22.76	26.70	30.23	32.06	25.61	8.82	15.68	0.32
SPESSRTN	4.11	4.38	5.26	4.68	5.42	5.14	4.73	14.28	1.84	4.25
ALMNDINE	70.70	57.93	63.92	62.52	58.62	57.44	62.08	63.65	62.87	76.13
PYROPE	8.20	3.58	5.73	5.79	3.23	3.92	5.09	13.05	19.60	17.09

COMPOSITION IN OXIDES

	GTJ 2	GTJ 3	GTJ 4	GTJ 5	GTJ 6	GTJ 7	GTJ 8	GTJ 9	GTJ10	GTJ11
SiO2	38.64	38.11	38.27	38.27	37.74	38.45	37.95	37.84	37.99	36.76
Al2O3	21.80	21.61	21.45	21.48	21.67	21.85	21.71	21.81	21.41	21.33
TiO2	0.03	0.02	0.05	0.00	0.02	0.16	0.09	0.05	0.09	0.02
Fe2O3	0.84	1.12	0.30	0.42	0.75	0.59	0.57	1.15	0.78	0.10
MgO	4.99	4.54	4.31	5.07	5.10	4.58	4.51	4.74	5.23	5.14
FeO	31.28	30.74	32.09	33.30	32.23	31.13	32.50	32.99	30.61	31.80
MnO	1.12	1.48	1.54	1.60	1.53	1.21	1.60	1.65	1.69	1.79
CaO	3.77	4.10	3.42	1.41	1.78	4.37	2.58	1.70	2.94	2.80
Na2O	0.02	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.02	0.00
	102.49	101.72	101.45	101.55	100.82	102.34	101.51	101.94	100.77	101.74

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTJ 2	GTJ 3	GTJ 4	GTJ 5	GTJ 6	GTJ 7	GTJ 8	GTJ 9	GTJ10	GTJ11
Si	2.983	2.973	2.999	2.996	2.971	2.977	2.975	2.958	2.993	3.017
Al	0.017	0.027	0.001	0.004	0.029	0.023	0.025	0.042	0.017	0.000
	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.017
Al	1.967	1.959	1.980	1.979	1.982	1.970	1.981	1.968	1.964	1.957
Ti	0.002	0.001	0.003	0.000	0.001	0.009	0.005	0.003	0.005	0.001
Fe3	0.049	0.066	0.018	0.025	0.044	0.034	0.034	0.068	0.046	0.006
Mg	0.574	0.528	0.503	0.592	0.598	0.529	0.527	0.552	0.612	0.596
Fe2	2.020	2.006	2.103	2.180	2.122	2.016	2.130	2.157	2.010	2.071
Mn	0.073	0.098	0.102	0.106	0.102	0.079	0.106	0.109	0.112	0.118
Ca	0.312	0.343	0.287	0.118	0.150	0.362	0.217	0.142	0.247	0.234
Na	0.003	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.003	0.000
	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	4.933

AL% IN Z	0.56	0.91	0.03	0.12	0.96	0.77	0.84	1.39	0.58	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	18.96	17.37	16.71	19.58	19.84	17.50	17.48	18.24	20.21	19.72
Fe	70.74	71.36	73.77	76.50	75.19	70.50	75.33	77.06	71.62	72.56
Ca	10.30	11.27	9.53	3.91	4.98	12.00	7.19	4.70	8.17	7.72

Ti-ANDRT	0.09	0.06	0.03	0.00	0.06	0.47	0.27	0.15	0.27	0.00
ANDRADTE	2.47	3.31	0.88	1.25	2.24	1.71	1.70	3.43	2.33	0.28
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	7.91	8.15	8.67	2.70	2.75	9.96	5.31	1.23	5.70	7.46
SPESSRTN	2.46	3.29	3.41	3.54	3.43	2.66	3.56	3.69	3.77	3.91
ALMNDINE	67.80	67.44	70.20	72.76	71.38	67.50	71.48	72.35	67.41	68.59
PYROPE	19.28	17.75	16.80	19.75	20.13	17.70	17.68	18.65	20.53	19.76

COMPOSITION IN OXIDES

	GTJ12	GTJ13	GTJ14	GTJ15	GTK 1	GTK 2	GTK 3	GTK 4	GTK 5	GTK 6
SiO2	38.77	38.31	38.66	38.28	37.89	37.84	37.90	37.75	38.36	37.85
Al2O3	21.51	21.64	21.77	21.81	21.18	21.18	21.45	20.96	21.32	21.38
TiO2	0.03	0.05	0.02	0.02	0.02	0.03	0.03	0.05	0.02	0.02
Fe2O3	0.00	0.01	0.00	1.17	0.00	0.64	0.00	0.00	0.03	0.90
MgO	5.23	4.35	4.71	4.78	3.48	3.83	3.76	3.27	4.08	3.97
FeO	33.56	33.28	33.13	33.58	27.55	29.12	30.04	28.61	30.37	29.37
MnO	1.73	1.85	1.78	2.04	5.44	3.23	3.43	3.75	2.71	2.94
CaO	1.10	2.30	2.08	1.20	4.29	4.51	3.93	5.02	4.29	4.57
Na2O	0.06	0.00	0.04	0.02	0.12	0.06	0.00	0.00	0.00	0.00
	102.01	101.79	102.19	102.90	99.97	100.44	100.54	99.41	101.18	101.00

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTJ12	GTJ13	GTJ14	GTJ15	GTK 1	GTK 2	GTK 3	GTK 4	GTK 5	GTK 6
Si	3.017	2.999	3.006	2.969	3.015	2.996	3.000	3.020	3.012	2.980
Al	0.000	0.001	0.000	0.031	0.000	0.004	0.000	0.000	0.000	0.020
	3.017	3.000	3.006	3.000	3.015	3.000	3.000	3.020	3.012	3.000
Al	1.973	1.995	1.995	1.963	1.986	1.972	2.001	1.976	1.973	1.964
Ti	0.002	0.003	0.001	0.001	0.001	0.002	0.002	0.003	0.001	0.001
Fe3	0.000	0.001	0.000	0.068	0.000	0.038	0.000	0.000	0.002	0.053
Mg	0.607	0.507	0.546	0.553	0.413	0.452	0.444	0.390	0.477	0.466
Fe2	2.186	2.179	2.154	2.178	1.833	1.928	1.988	1.914	1.994	1.934
Mn	0.114	0.123	0.117	0.134	0.367	0.217	0.230	0.254	0.180	0.196
Ca	0.092	0.193	0.173	0.100	0.366	0.383	0.333	0.430	0.361	0.386
Na	0.009	0.000	0.006	0.003	0.019	0.009	0.000	0.000	0.000	0.000
	4.982	5.000	4.993	5.000	4.985	5.000	4.998	4.968	4.988	5.000

AL% IN Z	0.00	0.05	0.00	1.03	0.00	0.15	0.01	0.00	0.00	0.66
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	20.24	16.90	18.25	18.22	13.86	14.98	14.81	13.05	15.84	15.35
Fe	76.70	76.67	75.95	78.49	73.86	72.34	74.06	72.55	72.19	71.94
Ca	3.06	6.42	5.79	3.29	12.28	12.68	11.13	14.40	11.97	12.70

Ti-ANDRT	0.00	0.05	0.00	0.06	0.00	0.09	0.01	0.00	0.00	0.06
ANDRADTE	0.00	0.03	0.00	3.31	0.00	1.93	0.00	0.00	0.08	2.69
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	3.06	6.35	5.79	0.00	12.28	10.82	11.12	14.40	11.89	10.18
SPESSRTN	3.80	4.09	3.92	4.52	12.31	7.27	7.68	8.50	5.98	6.55
ALMNDINE	72.90	72.58	72.03	73.47	61.55	64.72	66.39	64.05	66.19	64.86
PYROPE	20.24	16.91	18.25	18.64	13.86	15.17	14.81	13.05	15.85	15.63

COMPOSITION IN OXIDES

	GTK 7	GTK 8	GTK 9	GTL 1	GTL 2	GTL 3	GTL 4	GTM 1	GTM 2	GTM 3
SiO2	38.77	38.36	38.14	37.25	37.65	37.84	38.86	38.10	38.23	37.21
Al2O3	20.98	21.10	21.26	20.54	20.80	21.65	22.34	21.08	21.40	20.66
TiO2	0.05	0.05	0.03	0.02	0.10	0.03	0.05	0.05	0.09	1.23
Fe2O3	0.04	0.00	0.00	0.60	0.00	0.00	0.00	0.00	0.00	0.00
MgO	3.77	3.63	3.98	2.94	2.84	3.27	2.56	3.38	3.98	3.31
FeO	30.16	29.25	29.68	29.06	31.13	30.94	31.73	34.09	34.33	34.37
MnO	3.15	2.76	2.96	0.91	0.92	0.74	1.24	1.75	1.53	1.96
CaO	4.29	5.61	4.20	6.81	5.85	5.32	5.25	2.64	1.69	1.73
Na2O	0.18	0.00	0.06	0.12	0.06	0.04	0.10	0.04	0.04	0.10
K2O	0.00	0.00	0.00	0.02	0.01	0.01	0.01	0.00	0.00	0.00
	101.39	100.76	100.31	98.27	99.36	99.84	102.14	101.13	101.29	101.07

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTK 7	GTK 8	GTK 9	GTL 1	GTL 2	GTL 3	GTL 4	GTM 1	GTM 2	GTM 3
Si	3.040	3.022	3.016	3.012	3.020	3.005	3.020	3.019	3.014	2.967
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.033
	3.040	3.022	3.016	3.012	3.020	3.005	3.020	3.019	3.014	3.000
Al	1.939	1.959	1.982	1.957	1.966	2.026	2.046	1.969	1.988	1.908
Ti	0.003	0.003	0.002	0.001	0.006	0.002	0.003	0.003	0.005	0.074
Fe3	0.002	0.000	0.000	0.037	0.000	0.000	0.000	0.000	0.000	0.000
Mg	0.441	0.426	0.469	0.354	0.340	0.387	0.297	0.399	0.468	0.393
Fe2	1.978	1.927	1.963	1.965	2.088	2.055	2.062	2.259	2.263	2.325
Mn	0.209	0.184	0.198	0.062	0.063	0.050	0.082	0.117	0.102	0.132
Ca	0.360	0.474	0.356	0.590	0.503	0.453	0.437	0.224	0.143	0.148
Na	0.027	0.000	0.009	0.019	0.009	0.006	0.015	0.006	0.006	0.015
K	0.000	0.000	0.000	0.002	0.001	0.001	0.001	0.000	0.000	0.000
	4.960	4.973	4.979	4.988	4.976	4.979	4.942	4.978	4.976	4.996

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.10
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	14.73	14.16	15.71	11.78	11.34	13.15	10.31	13.31	15.72	13.12
Fe	73.22	70.12	72.37	68.61	71.86	71.48	74.50	79.22	79.49	81.95
Ca	12.05	15.73	11.92	19.61	16.80	15.37	15.19	7.47	4.80	4.93

Ti-ANDRT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.10
ANDRADTE	0.12	0.00	0.00	1.86	0.00	0.00	0.00	0.00	0.00	0.00
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	11.94	15.73	11.92	18.00	16.80	15.37	15.19	7.47	4.80	3.82
SPESSRTN	7.00	6.12	6.64	2.10	2.09	1.69	2.84	3.92	3.43	4.41
ALMNDINE	66.19	64.00	65.73	66.12	69.77	69.79	71.67	75.31	76.06	77.54
PYROPE	14.75	14.16	15.71	11.92	11.34	13.15	10.31	13.31	15.72	13.12

COMPOSITION IN OXIDES

	GTM 4	GTM 5	GTM 6	GTM 7	GTQ 1	GTQ 2	GTQ 3	GTQ 4	GTQ 5	GTQ 6
SiO2	37.75	38.12	37.98	37.76	37.77	37.44	37.73	37.75	37.53	37.59
Al2O3	21.07	20.78	21.03	21.04	21.04	21.01	20.89	20.94	20.80	20.95
TiO2	0.03	0.07	0.07	0.21	0.14	0.15	0.15	0.10	0.09	0.09
Fe2O3	0.00	0.11	0.00	0.39	0.00	0.94	0.00	0.00	0.00	0.00
MgO	3.41	3.68	3.38	4.15	1.78	1.57	1.57	1.66	1.28	1.21
FeO	35.53	33.64	34.24	34.04	34.01	33.02	33.66	33.53	33.51	33.53
MnO	2.00	1.73	1.75	1.56	3.51	4.23	4.23	3.84	4.79	4.94
CaO	0.93	2.81	2.49	1.67	3.18	3.53	3.25	3.61	3.19	3.37
Na2O	0.02	0.02	0.04	0.04	0.02	0.06	0.02	0.04	0.02	0.00
	100.74	100.96	100.98	100.86	101.45	101.95	101.50	101.47	101.21	101.68

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTM 4	GTM 5	GTM 6	GTM 7	GTQ 1	GTQ 2	GTQ 3	GTQ 4	GTQ 5	GTQ 6
Si	3.014	3.023	3.016	2.995	3.013	2.981	3.015	3.013	3.015	3.007
Al	0.000	0.000	0.000	0.005	0.000	0.019	0.000	0.000	0.000	0.000
	3.014	3.023	3.016	3.000	3.013	3.000	3.015	3.013	3.015	3.007
Al	1.983	1.942	1.968	1.962	1.978	1.953	1.967	1.970	1.969	1.975
Ti	0.002	0.004	0.004	0.013	0.008	0.009	0.009	0.006	0.005	0.005
Fe3	0.000	0.007	0.000	0.023	0.000	0.057	0.000	0.000	0.000	0.000
Mg	0.406	0.435	0.400	0.491	0.212	0.186	0.187	0.197	0.153	0.144
Fe2	2.372	2.231	2.274	2.258	2.269	2.199	2.249	2.238	2.251	2.243
Mn	0.135	0.116	0.118	0.105	0.237	0.285	0.286	0.260	0.326	0.335
Ca	0.080	0.239	0.212	0.142	0.272	0.301	0.278	0.309	0.275	0.289
Na	0.003	0.003	0.006	0.006	0.003	0.009	0.003	0.006	0.003	0.006
	4.980	4.977	4.982	5.000	4.979	5.000	4.980	4.986	4.982	4.992

Al% IN Z	0.00	0.00	0.00	0.15	0.00	0.62	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	13.56	14.37	13.32	16.25	7.08	6.15	6.23	6.57	5.10	4.79
Fe	83.78	77.75	79.63	79.04	83.83	83.90	84.50	83.15	85.76	85.62
Ca	2.66	7.89	7.05	4.70	9.09	9.95	9.27	10.28	9.14	9.59

Ti-ANDRT	0.00	0.00	0.00	0.16	0.00	0.45	0.00	0.00	0.00	0.00
ANDRADTE	0.00	0.33	0.00	1.17	0.00	2.86	0.00	0.00	0.00	0.00
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	2.66	7.58	7.05	3.42	9.09	6.83	9.27	10.28	9.14	9.59
SPESSRTN	4.52	3.85	3.92	3.50	7.93	9.60	9.54	8.64	10.85	11.12
ALMNDINE	79.26	73.85	75.71	75.38	75.90	73.99	74.96	74.51	74.92	74.50
PYROPE	13.56	14.40	13.32	16.38	7.08	6.27	6.23	6.57	5.10	4.79

COMPOSITION IN OXIDES

	GTQ 7	GTQ 8	GTQ 9	GTQ10	GTQ11	GTQ12	GTQ13	GTQ14	GTQ15	GTQ16
SiO2	37.70	37.66	37.60	37.38	37.19	36.92	37.50	37.46	37.70	37.57
Al2O3	21.04	20.72	20.81	20.94	20.77	20.60	20.86	21.01	21.18	20.89
TiO2	0.10	0.12	0.09	0.14	0.12	0.12	0.10	0.12	0.10	0.07
Fe2O3	0.05	0.00	0.03	0.00	0.25	1.10	0.00	0.10	0.00	0.21
MgO	1.26	1.28	1.31	1.30	1.10	1.10	1.35	1.40	1.70	2.13
FeO	33.45	32.99	33.37	32.90	32.82	32.23	32.86	32.41	32.23	35.56
MnO	4.80	4.49	4.76	4.92	5.57	5.41	5.40	5.66	5.17	1.45
CaO	3.53	3.92	3.38	3.59	3.10	3.36	3.19	3.33	3.36	3.25
Na2O	0.02	0.00	0.04	0.00	0.04	0.06	0.02	0.00	0.04	0.00
	101.94	101.18	101.39	101.07	100.95	100.90	101.28	101.49	101.48	101.13

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTQ 7	GTQ 8	GTQ 9	GTQ10	GTQ11	GTQ12	GTQ13	GTQ14	GTQ15	GTQ16
Si	3.006	3.020	3.014	3.003	3.001	2.983	3.009	2.999	3.006	3.005
Al	0.000	0.000	0.000	0.000	0.000	0.017	0.000	0.001	0.000	0.000
	3.006	3.020	3.014	3.003	3.001	3.000	3.009	3.000	3.006	3.005
Al	1.977	1.958	1.966	1.983	1.975	1.945	1.973	1.981	1.990	1.969
Ti	0.006	0.007	0.005	0.008	0.007	0.007	0.006	0.007	0.006	0.004
Fe3	0.003	0.000	0.002	0.000	0.015	0.067	0.000	0.006	0.000	0.013
Mg	0.150	0.153	0.157	0.156	0.132	0.132	0.161	0.167	0.202	0.254
Fe2	2.230	2.213	2.237	2.210	2.215	2.178	2.205	2.169	2.149	2.378
Mn	0.324	0.305	0.323	0.328	0.381	0.370	0.367	0.384	0.349	0.098
Ca	0.302	0.337	0.290	0.309	0.268	0.291	0.274	0.286	0.287	0.278
Na	0.005	0.000	0.006	0.000	0.006	0.009	0.003	0.000	0.006	0.000
	4.996	4.973	4.986	4.994	4.999	5.000	4.990	5.000	4.990	4.995

ALX IN Z	0.00	0.00	0.00	0.00	0.00	0.56	0.00	0.05	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	4.98	5.09	5.20	5.18	4.39	4.36	5.37	5.55	6.76	8.40
Fe	85.00	83.71	85.15	84.53	86.70	86.07	85.51	84.97	83.63	82.38
Ca	10.02	11.20	9.65	10.29	8.90	9.57	9.12	9.48	9.61	9.22

Ti-ANDRT	0.00	0.00	0.00	0.00	0.00	0.37	0.00	0.05	0.00	0.00
ANDRADTE	0.14	0.00	0.10	0.00	0.75	3.37	0.00	0.31	0.00	0.64
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	9.89	11.20	9.55	10.29	8.20	6.05	9.12	9.14	9.61	8.62
SPESSRTN	10.78	10.14	10.75	10.92	12.71	12.46	12.20	12.77	11.69	3.26
ALMNDINE	74.20	73.57	74.39	73.60	73.93	73.29	73.31	72.17	71.94	79.04
PYROPE	4.98	5.09	5.21	5.18	4.42	4.46	5.37	5.56	6.76	8.44

COMPOSITION IN OXIDES

	GTQ17	GTQ18	GTQ19	GTQ20	GTR 1	GTR 2	GTR 3	GTR 4	GTR 5	GTR 6
SiO2	37.78	37.53	37.71	37.55	37.83	37.30	38.07	37.89	36.95	38.73
Al2O3	20.85	20.83	20.87	20.89	20.04	19.42	19.98	20.46	19.26	20.78
TiO2	0.10	0.07	0.05	0.10	0.15	0.22	0.03	0.09	0.05	0.05
Fe2O3	0.00	0.00	0.00	0.01	0.00	0.14	0.32	0.17	1.47	0.35
MgO	1.82	2.08	2.13	1.52	1.49	1.51	1.82	1.93	1.14	2.56
FeO	34.20	35.13	35.80	33.65	28.43	28.04	28.42	30.77	27.74	31.58
MnO	3.08	1.67	1.48	4.19	7.04	6.69	6.90	3.62	5.89	2.99
CaO	3.22	3.43	3.01	3.36	5.21	5.33	5.24	5.79	6.46	5.46
Na2O	0.02	0.00	0.02	0.02	0.10	0.10	0.04	0.02	0.04	0.04
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
	101.07	100.74	101.07	101.28	100.29	98.75	100.82	100.74	99.01	102.55

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTQ17	GTQ18	GTQ19	GTQ20	GTR 1	GTR 2	GTR 3	GTR 4	GTR 5	GTR 6
Si	3.023	3.011	3.017	3.009	3.048	3.053	3.049	3.028	3.026	3.031
	3.023	3.011	3.017	3.009	3.048	3.053	3.049	3.028	3.026	3.031
Al	1.966	1.970	1.966	1.973	1.903	1.874	1.886	1.927	1.859	1.917
Ti	0.006	0.004	0.003	0.006	0.009	0.014	0.002	0.005	0.003	0.003
Fe3	0.000	0.000	0.000	0.001	0.000	0.009	0.020	0.010	0.091	0.021
Mg	0.217	0.249	0.254	0.182	0.179	0.184	0.217	0.230	0.139	0.299
Fe2	2.289	2.357	2.395	2.255	1.915	1.919	1.903	2.056	1.900	2.067
Mn	0.209	0.113	0.100	0.284	0.480	0.464	0.468	0.245	0.409	0.198
Ca	0.276	0.295	0.258	0.288	0.450	0.467	0.450	0.496	0.567	0.458
Na	0.003	0.000	0.003	0.003	0.016	0.016	0.006	0.003	0.006	0.006
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000
	4.966	4.988	4.981	4.991	4.952	4.947	4.951	4.972	4.974	4.969

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	7.26	8.25	8.45	6.03	5.92	6.05	7.10	4.57	4.48	9.82
Fe	83.51	81.97	82.98	84.38	79.22	78.59	78.19	76.11	77.26	75.14
Ca	9.23	9.78	8.58	9.59	14.87	15.36	14.70	16.32	18.26	15.05

TI-ANDRT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ANDRADTE	0.00	0.00	0.00	0.03	0.00	0.44	0.97	0.49	4.50	1.02
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	9.23	9.78	8.58	9.55	14.87	14.96	13.83	15.88	14.30	14.13
SPESSRTN	6.98	3.77	3.33	9.43	15.88	15.28	15.41	8.09	13.55	6.56
ALMNDINE	76.53	78.20	79.64	76.95	63.33	63.24	62.65	67.93	63.02	68.41
PYROPE	7.26	8.25	8.45	6.03	5.92	6.07	7.15	7.59	4.62	9.86

COMPOSITION IN OXIDES

	GTR 7	GTR 8	GTR 9	GTR10	GTR11	GTR12	GTR13	GTR14	GTR15	GTR16
SiO2	39.22	37.33	37.66	37.42	37.68	37.51	37.45	37.56	37.6°	37.47
Al2O3	20.83	20.55	20.82	20.26	20.42	20.56	20.36	20.42	20.00	20.25
TiO2	0.07	0.07	0.02	0.15	0.09	0.12	0.10	0.07	0.09	0.10
Fe2O3	0.00	1.98	1.32	1.04	0.86	1.42	1.42	1.25	1.52	2.09
MgO	3.04	2.74	3.03	1.72	1.63	1.65	1.54	1.27	1.16	1.04
FeO	31.86	30.51	30.83	26.96	27.07	27.04	27.11	27.12	27.25	27.06
MnO	2.06	2.56	2.30	7.45	7.52	7.27	8.77	7.18	6.52	6.20
CaO	5.51	5.17	4.99	5.46	5.65	5.80	4.69	6.35	7.12	7.50
Na2O	0.04	0.02	0.02	0.06	0.06	0.04	0.02	0.04	0.02	0.02
K2O	0.00	0.00	0.00	0.01	0.01	0.00	0.01	0.00	0.00	0.00
	102.63	100.93	100.99	100.53	100.99	101.41	101.47	101.25	101.36	101.73

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTR 7	GTR 8	GTR 9	GTR10	GTR11	GTR12	GTR13	GTR14	GTR15	GTR16
Si	3.051	2.973	2.988	3.006	3.012	2.988	2.994	3.000	3.009	2.983
Al	0.000	0.027	0.012	0.000	0.000	0.012	0.006	0.000	0.000	0.017
	3.051	3.000	3.000	3.006	3.012	3.000	3.000	3.000	3.009	3.000
Al	1.910	1.903	1.934	1.918	1.924	1.918	1.913	1.922	1.882	1.883
Ti	0.004	0.004	0.001	0.009	0.005	0.007	0.006	0.004	0.005	0.006
Fe3	0.000	0.119	0.079	0.063	0.052	0.085	0.085	0.075	0.091	0.125
Mg	0.353	0.325	0.358	0.206	0.194	0.196	0.184	0.151	0.138	0.123
Fe2	2.073	2.032	2.046	1.811	1.810	1.801	1.813	1.812	1.820	1.802
Mn	0.136	0.173	0.155	0.507	0.509	0.491	0.594	0.486	0.441	0.418
Ca	0.459	0.441	0.424	0.470	0.484	0.495	0.402	0.543	0.609	0.640
Na	0.006	0.003	0.003	0.009	0.009	0.006	0.003	0.006	0.003	0.003
K	0.000	0.000	0.000	0.001	0.001	0.000	0.001	0.000	0.000	0.000
	4.941	5.000	5.000	4.994	4.988	5.000	5.000	5.000	4.991	5.000

AlZ IN Z	0.00	0.89	0.41	0.00	0.00	0.40	0.20	0.00	0.00	0.57
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	11.67	10.53	11.70	6.74	6.37	6.39	5.96	4.93	4.45	3.97
Fe	73.12	75.19	74.44	77.89	77.76	77.48	80.98	77.35	75.89	75.45
Ca	15.21	14.28	13.85	15.37	15.87	16.13	13.05	17.72	19.65	20.58

Ti-ANDRT	0.00	0.21	0.06	0.00	0.00	0.36	0.20	0.00	0.00	0.30
ANDRADTE	0.00	5.99	3.96	3.15	2.58	4.29	4.29	3.75	4.56	6.29
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	15.21	8.64	10.20	12.55	13.56	11.95	8.95	14.41	15.69	14.85
SPESSRTN	4.49	5.81	5.15	16.93	16.99	16.44	19.85	16.24	14.66	14.02
ALMNDINE	68.63	68.39	68.59	60.50	60.38	60.39	60.59	60.55	60.50	60.40
PYROPE	11.67	10.95	12.01	6.88	6.48	6.57	6.13	5.05	4.59	4.14

COMPOSITION IN OXIDES

	GTR17	GTR18	GTR19	GTR20	GTT 1	GTT 2	GTT 3	GTT 4	GTT 5	GTT 6
SiO2	37.64	37.50	37.67	36.85	37.47	37.63	37.60	37.57	37.36	37.20
Al2O3	20.48	20.62	20.57	20.51	20.69	20.69	20.91	20.69	20.69	20.66
TiO2	0.10	0.05	0.07	0.05	0.15	0.12	0.09	0.21	0.12	0.14
Fe2O3	1.33	1.70	0.79	2.02	0.00	0.14	0.24	0.00	0.36	0.64
MgO	1.04	2.69	2.99	3.49	2.07	2.23	2.18	1.95	2.07	1.95
FeO	28.24	30.53	30.94	30.54	22.64	24.29	24.55	24.31	24.23	23.11
MnO	6.50	2.72	2.36	1.92	13.53	10.77	10.69	10.85	10.51	10.90
CaO	6.50	5.24	4.73	4.12	3.74	4.63	4.51	4.81	4.85	4.99
Na2O	0.02	0.02	0.02	0.02	0.00	0.00	0.00	0.02	0.00	0.00
K2O	0.00	0.00	0.03	0.01	0.00	0.00	0.00	0.00	0.00	0.00
	101.85	101.07	100.37	99.53	100.29	100.50	100.77	100.41	100.19	100.20

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTR17	GTR18	GTR19	GTR20	GTT 1	GTT 2	GTT 3	GTT 4	GTT 5	GTT 6
Si	2.995	2.982	3.007	2.965	3.013	3.012	3.003	3.013	3.002	2.993
Al	0.005	0.018	0.000	0.035	0.000	0.000	0.000	0.000	0.000	0.007
	3.000	3.000	3.007	3.000	3.013	3.012	3.003	3.013	3.002	3.000
Al	1.916	1.914	1.936	1.910	1.961	1.952	1.968	1.956	1.959	1.952
Ti	0.006	0.003	0.004	0.003	0.009	0.007	0.005	0.013	0.007	0.008
Fe3	0.080	0.102	0.047	0.122	0.000	0.009	0.015	0.000	0.022	0.039
Mg	0.123	0.319	0.356	0.419	0.248	0.266	0.260	0.233	0.248	0.234
Fe2	1.679	2.030	2.066	2.055	1.523	1.626	1.640	1.630	1.629	1.622
Mn	0.438	0.183	0.173	0.131	0.922	0.730	0.723	0.737	0.715	0.743
Ca	0.554	0.446	0.405	0.355	0.322	0.397	0.386	0.413	0.413	0.422
Na	0.003	0.003	0.003	0.003	0.000	0.000	0.000	0.003	0.000	0.000
K	0.000	0.000	0.003	0.001	0.000	0.000	0.000	0.000	0.000	0.000
	5.000	5.000	4.993	5.000	4.984	4.988	4.997	4.985	4.995	5.000

Al% IN Z	0.16	0.61	0.00	1.16	0.00	0.00	0.00	0.00	0.00	0.24
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	4.01	10.35	11.68	13.50	8.23	8.79	8.58	7.75	8.19	7.69
Fe	77.97	75.16	75.04	74.90	81.08	78.10	78.65	78.55	78.05	78.41
Ca	18.02	14.49	13.28	11.52	10.69	13.11	12.77	13.71	13.78	13.89

Ti-ANDRT	0.16	0.15	0.00	0.15	0.00	0.00	0.00	0.00	0.00	0.24
ANDRADTE	4.00	5.12	2.37	6.20	0.00	0.43	0.73	0.00	1.09	1.97
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	14.35	9.71	11.12	5.65	10.69	12.72	12.10	13.71	12.78	11.90
SPESSRTN	14.63	6.15	5.77	4.42	30.57	24.19	24.04	24.45	23.77	24.75
ALMNDINE	62.75	68.16	68.88	69.44	50.51	53.85	54.51	54.10	54.12	53.39
PYROPE	4.12	10.70	11.36	14.14	8.23	8.81	8.67	7.73	8.24	7.79

COMPOSITION IN OXICES

	GTT 7	GTT 8	GTT 9	GTT10	GTT11	GTT12	GTT13	GTT14	GTT15	GTT16
SiO2	37.31	37.01	37.14	37.25	37.59	37.03	36.88	37.05	37.15	37.02
Al2O3	20.58	20.73	20.70	20.61	20.57	20.82	20.68	20.71	20.45	20.63
TiO2	0.14	0.14	0.12	0.12	0.10	0.14	0.10	0.14	0.10	0.12
Fe2O3	0.58	1.01	0.00	0.66	0.00	1.00	0.89	0.87	0.89	0.22
MgO	1.93	1.88	2.15	2.09	2.22	2.34	2.35	1.95	2.14	2.04
FeO	23.88	23.57	24.38	23.84	24.40	23.14	22.91	23.71	23.93	24.09
MnO	10.89	10.87	9.92	10.54	10.22	11.06	11.85	9.93	10.34	10.49
CaO	4.99	4.96	4.84	4.93	4.79	4.60	3.97	5.54	4.84	4.70
Na2O	0.00	0.02	0.00	0.02	0.00	0.00	0.00	0.02	0.02	0.00
	100.30	100.19	99.25	100.08	99.89	100.13	99.62	99.92	99.86	99.31

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTT 7	GTT 8	GTT 9	GTT10	GTT11	GTT12	GTT13	GTT14	GTT15	GTT16
Si	2.999	2.979	3.007	2.997	3.023	2.975	2.982	2.984	2.997	3.001
Al	0.001	0.021	0.000	0.003	0.000	0.025	0.018	0.016	0.003	0.000
	3.000	3.000	3.007	3.000	3.023	3.000	3.000	3.000	3.000	3.001
Al	1.949	1.946	1.975	1.951	1.950	1.947	1.953	1.950	1.941	1.971
Ti	0.006	0.008	0.007	0.007	0.006	0.008	0.006	0.008	0.006	0.007
Fe3	0.035	0.061	0.000	0.041	0.000	0.061	0.053	0.053	0.054	0.013
Mg	0.231	0.226	0.259	0.251	0.266	0.280	0.283	0.234	0.257	0.246
Fe2	1.605	1.587	1.651	1.604	1.641	1.555	1.549	1.597	1.614	1.633
Mn	0.741	0.741	0.680	0.718	0.696	0.753	0.812	0.677	0.706	0.720
Ca	0.430	0.428	0.420	0.425	0.413	0.396	0.344	0.478	0.418	0.428
Na	0.000	0.003	0.000	0.003	0.000	0.000	0.000	0.003	0.003	0.000
	5.000	5.000	4.992	5.000	4.972	5.000	5.000	5.000	5.000	4.999

Al% IN Z	0.03	0.69	0.00	0.11	0.00	0.82	0.60	0.54	0.12	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	7.60	7.41	8.62	8.25	8.82	9.21	9.31	7.70	8.44	8.16
Fe	78.25	78.53	77.43	77.77	77.49	77.79	79.38	76.57	77.85	78.33
Ca	14.12	14.06	13.95	13.98	13.68	13.01	11.31	15.73	13.71	13.51

Ti-ANDRT	0.03	0.43	0.00	0.11	0.00	0.43	0.31	0.43	0.12	0.00
ANDRADTE	1.75	3.07	0.00	2.05	0.00	3.04	2.67	2.66	2.70	0.67
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	12.51	10.85	13.95	12.01	13.68	9.80	8.53	12.93	11.15	12.90
SPESSRTN	24.65	24.86	22.60	23.96	23.08	25.23	27.16	22.68	23.53	23.94
ALMNDINE	53.37	53.23	54.84	53.51	54.41	52.11	51.85	53.47	53.88	54.29
PYROPE	7.69	7.57	8.62	8.36	8.32	9.39	9.43	7.84	8.59	8.19

COMPOSITION IN OXIDES

	GTT17	GTT18	GTT19	GTT20	GTT21	GTT22	GTT23	GTT24	GTT25	GTW 1
SiO2	37.31	36.20	36.64	37.19	36.54	36.70	37.22	38.09	37.34	37.56
Al2O3	20.82	20.40	20.41	20.52	20.69	20.78	20.99	20.70	20.72	20.98
TiO2	0.14	0.21	0.24	0.17	0.12	0.10	0.05	0.10	0.15	0.07
Fe2O3	0.95	1.29	1.51	0.70	1.51	1.56	0.25	0.36	0.37	0.00
MgO	2.07	1.75	1.82	1.93	2.04	2.37	2.23	2.18	2.16	0.70
FeO	23.56	22.79	22.79	23.91	23.03	23.12	24.33	24.35	23.96	39.14
MnO	10.50	11.14	11.11	10.75	10.45	10.17	10.34	10.73	10.64	0.01
CaO	4.63	4.83	4.90	4.91	4.82	4.87	4.51	4.52	4.84	3.41
Na2O	0.20	0.02	0.10	0.02	0.08	0.02	0.00	0.16	0.00	0.02
	100.18	98.63	99.52	99.70	99.28	99.69	99.91	101.19	100.18	101.39

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTT17	GTT18	GTT19	GTT20	GTT21	GTT22	GTT23	GTT24	GTT25	GTW 1
Si	2.994	2.965	2.972	3.005	2.964	2.960	2.994	3.026	2.999	3.009
Al	0.006	0.035	0.028	0.000	0.036	0.040	0.006	0.000	0.001	0.000
	3.000	3.000	3.000	3.005	3.000	3.000	3.000	3.026	3.000	3.009
Al	1.963	1.933	1.923	1.954	1.942	1.936	1.985	1.938	1.960	1.981
Ti	0.008	0.013	0.015	0.010	0.007	0.006	0.003	0.006	0.009	0.004
Fe3	0.057	0.079	0.092	0.018	0.092	0.095	0.015	0.021	0.022	0.000
Mg	0.246	0.214	0.220	0.232	0.247	0.285	0.267	0.258	0.259	0.084
Fe2	1.581	1.561	1.546	1.616	1.563	1.560	1.637	1.618	1.609	2.623
Mn	0.714	0.773	0.763	0.736	0.718	0.695	0.705	0.722	0.724	0.001
Ca	0.398	0.424	0.426	0.425	0.419	0.421	0.389	0.385	0.417	0.293
Na	0.031	0.003	0.016	0.003	0.013	0.003	0.000	0.025	0.000	0.003
	5.000	5.000	5.000	4.995	5.000	5.000	5.000	4.974	5.000	4.998

AL% IN Z	0.20	1.18	0.95	0.00	1.20	1.32	0.19	0.00	0.03	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	8.26	7.00	7.22	7.68	8.12	9.33	8.88	8.59	8.53	2.79
Fe	78.46	79.10	75.31	78.28	78.09	76.89	78.22	78.60	77.72	87.45
Ca	13.28	13.89	13.97	14.04	13.79	13.78	12.90	12.81	13.74	9.76

Ti-ANDRT	0.21	0.65	0.74	0.00	0.37	0.31	0.15	0.00	0.03	0.00
ANDRADTE	2.92	4.00	4.68	0.90	4.68	4.79	0.76	1.07	1.11	0.00
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	10.41	9.61	8.99	13.22	9.16	9.12	12.06	11.83	12.70	9.76
SPESSRTN	24.27	26.01	25.83	24.45	24.37	23.47	23.51	24.21	24.06	0.02
ALMNDINE	53.77	52.54	52.31	53.70	53.04	52.68	54.60	54.24	53.49	87.43
PYROPE	8.42	7.19	7.45	7.72	8.37	9.63	8.92	8.65	8.60	2.79

COMPOSITION IN OXIDES

	GTW 2	GTW 3	GTX 1	GTX 2	GTX 3	GTX 4	GT& 1	GT& 2	GT& 3	GT& 4
SiO2	37.55	37.21	35.26	35.51	35.33	38.01	36.92	36.77	37.10	36.88
Al2O3	20.99	20.65	20.66	20.59	20.39	20.43	20.39	20.29	20.19	20.69
TiO2	0.05	0.03	0.10	0.12	0.10	0.09	0.03	0.02	0.03	0.02
Fe2O3	0.00	0.16	4.27	3.82	3.95	1.26	0.00	0.00	0.00	0.00
MgO	0.89	0.89	1.42	1.35	1.38	1.35	0.38	0.38	0.42	0.42
FeO	38.20	38.30	28.62	28.48	29.19	32.02	25.11	24.93	25.30	24.38
MnO	0.09	0.07	1.66	1.40	1.42	1.90	11.26	11.12	11.13	11.33
CaO	3.10	3.10	7.21	7.94	7.22	7.09	4.80	4.92	5.00	4.94
Na2O	0.02	0.02	0.02	0.02	0.00	0.02	0.08	0.04	0.06	0.02
	100.87	100.93	99.29	99.23	98.98	102.20	98.97	98.47	99.23	98.87

STRUCT. FORM. BASED ON 12.0 OXYGENS

	GTW 2	GTW 3	GTX 1	GTX 2	GTX 3	GTX 4	GTX 1	GTX 2	GTX 3	GTX 4
Si	3.024	3.010	2.873	2.890	2.890	3.006	3.029	3.031	3.038	3.029
Al	0.000	0.000	0.127	0.110	0.110	0.000	0.000	0.000	0.000	0.000
	3.024	3.010	3.000	3.000	3.000	3.006	3.029	3.031	3.038	3.029
Al	1.993	1.969	1.857	1.865	1.855	1.904	1.972	1.971	1.948	1.993
Ti	0.003	0.002	0.006	0.007	0.006	0.005	0.002	0.001	0.002	0.001
Fe3	0.000	0.010	0.262	0.234	0.243	0.076	0.000	0.000	0.000	0.000
Mg	0.107	0.107	0.172	0.164	0.168	0.159	0.046	0.047	0.051	0.051
Fe2	2.574	2.625	1.950	1.938	1.996	2.118	1.723	1.718	1.732	1.692
Mn	0.006	0.005	0.115	0.097	0.098	0.127	0.783	0.776	0.772	0.787
Ca	0.268	0.269	0.635	0.692	0.633	0.601	0.422	0.435	0.439	0.425
Na	0.003	0.003	0.003	0.003	0.000	0.003	0.013	0.006	0.010	0.003
	4.954	4.990	5.000	5.000	5.000	4.994	4.960	4.955	4.954	4.956

Al% IN Z	0.00	0.00	4.24	3.67	3.68	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

IRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	3.62	3.56	5.50	5.24	5.36	5.16	1.56	1.57	1.71	1.74
Fe	87.33	87.53	74.22	72.61	74.48	75.34	84.25	83.53	83.64	83.87
Ca	9.06	8.91	20.27	22.15	20.16	19.49	14.19	14.60	14.65	14.39

Ti-ANDRT	0.00	0.00	0.32	0.38	0.32	0.00	0.00	0.00	0.00	0.00
ANDRADTE	0.00	0.48	13.66	12.15	12.59	3.81	0.00	0.00	0.00	0.00
UVAROVTE	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
KNORRNGT	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
GROSSULR	9.06	8.45	8.14	11.42	8.95	16.18	14.19	14.60	14.65	14.39
SPESSRTN	0.21	0.16	3.99	3.34	3.40	4.24	26.31	26.09	25.78	26.62
ALMNDINE	87.12	87.33	67.89	67.05	68.94	72.48	57.94	57.74	57.86	57.26
PYROPE	3.62	3.57	6.00	5.66	5.81	5.30	1.56	1.57	1.71	1.74

COMPOSITION IN OXIDES

	PXA 1	PXA 2	PXA 3	PXA 4	PXA 5	PXA 6	PXA 7	PXA 8	PXA 9	PXA10
SiO2	55.07	55.79	55.47	56.47	55.92	55.81	56.20	56.94	56.00	55.74
Al2O3	8.36	9.58	6.93	9.86	9.96	9.92	9.29	10.79	8.51	8.65
TiO2	0.11	0.11	0.09	0.17	0.13	0.13	0.19	0.11	0.13	0.11
Fe2O3	7.26	5.06	2.50	5.71	2.65	4.47	3.65	1.56	3.53	7.31
MgO	7.13	6.72	6.67	6.69	7.42	6.62	7.01	6.28	7.11	6.82
FeO	4.26	5.08	8.12	4.37	5.60	5.34	6.29	6.90	6.71	3.10
MnO	0.04	0.04	0.04	0.05	0.05	0.01	0.01	0.01	0.04	0.05
CaO	13.36	12.47	13.11	12.43	13.75	12.17	12.68	11.36	13.52	13.25
Na2O	6.85	7.25	6.37	7.63	6.57	7.35	6.96	7.65	6.53	7.42
K2O	0.02	0.04	0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.02
	102.46	102.14	101.31	103.39	102.06	101.83	102.29	101.61	102.09	102.50

STRUCT. FORM. BASED ON 6.0 OXYGENS

	PXA 1	PXA 2	PXA 3	PXA 4	PXA 5	PXA 6	PXA 7	PXA 8	PXA 9	PXA10
Si	1.961	1.979	1.997	1.976	1.979	1.982	1.992	2.014	1.997	1.974
Al	0.039	0.021	0.003	0.024	0.021	0.018	0.008	0.000	0.003	0.026
	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.014	2.000	2.000
Al	0.312	0.380	0.376	0.383	0.395	0.398	0.380	0.449	0.354	0.336
Ti	0.003	0.003	0.002	0.004	0.003	0.003	0.005	0.003	0.003	0.003
Fe3	0.194	0.135	0.068	0.150	0.071	0.119	0.097	0.042	0.095	0.195
Mg	0.378	0.355	0.358	0.349	0.391	0.351	0.370	0.331	0.378	0.360
Fe2	0.127	0.151	0.244	0.128	0.166	0.159	0.186	0.204	0.200	0.092
Mn	0.001	0.001	0.001	0.001	0.001	0.000	0.000	0.000	0.001	0.001
Ca	0.510	0.474	0.506	0.466	0.521	0.463	0.482	0.431	0.516	0.503
Na	0.473	0.499	0.445	0.518	0.451	0.506	0.478	0.525	0.451	0.509
K	0.001	0.002	0.000	0.000	0.000	0.000	0.001	0.001	0.001	0.001
	2.000	2.000	2.000	2.000	2.000	2.000	2.000	1.986	2.000	2.000

Al% IN Z	1.93	1.03	0.16	1.20	1.04	0.88	0.40	0.00	0.17	1.31
CHARGE	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	31.26	31.63	30.41	31.87	34.02	32.10	32.60	32.86	31.75	31.28
Fe	26.64	25.70	26.63	25.56	20.66	25.49	25.00	24.40	24.85	25.04
Ca	42.10	42.46	42.96	42.57	45.32	42.41	42.39	42.73	43.40	43.68

Na-Ti	0.29	0.29	0.24	0.45	0.35	0.35	0.51	0.00	0.35	0.29
Ti-DI	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ACMITE	19.45	13.51	6.78	15.05	7.05	11.94	9.72	4.19	9.47	19.48
Cr DIOPS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JADEITE	27.65	36.25	37.48	36.32	37.73	38.38	37.69	45.57	35.42	31.26
Ca-TS	3.57	1.76	0.08	1.95	1.73	1.40	0.28	0.00	0.00	2.34
FERRI-TS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JOHANNS	0.12	0.12	0.12	0.15	0.15	0.03	0.03	0.03	0.12	0.15
ENSTAT	18.92	17.77	17.89	17.45	19.57	17.53	18.52	16.68	18.89	18.00
FERROSIL	6.35	7.53	12.22	6.39	8.28	7.93	9.32	10.27	10.00	4.59
CaSiO3	23.64	22.76	25.18	22.25	25.13	22.44	23.92	23.27	25.76	23.99

COMPOSITION IN OXIDES

	PXA11	PXA12	PXA13	PXA14	PXA15	PXA16	PXA17	PXA18	PXB 1	PXB 2
SiO2	56.28	55.99	55.56	55.25	55.44	55.31	54.77	55.05	53.23	52.52
Al2O3	10.15	10.20	9.89	9.20	9.41	8.60	8.51	8.65	5.49	4.52
TiO2	0.11	0.07	0.11	0.15	0.13	0.11	0.15	0.11	0.19	0.17
Fe2O3	1.39	2.40	2.99	3.04	3.58	3.23	5.18	4.43	5.78	4.25
MgO	6.94	6.96	6.75	6.26	6.61	6.97	6.93	6.48	7.38	8.25
FeO	6.17	5.29	4.69	6.27	5.15	6.48	4.92	5.73	7.60	9.25
MnO	0.01	0.03	0.05	0.04	0.03	0.01	0.03	0.05	0.01	0.03
CaO	12.07	12.27	12.35	12.23	12.50	12.80	12.89	12.45	17.09	18.91
Na2O	7.20	7.24	7.32	7.13	7.21	6.67	6.86	7.04	4.55	3.20
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.01
	100.32	100.45	99.71	99.57	100.06	100.17	100.24	99.99	101.34	101.02

STRUCT. FORM. BASED ON 6.0 OXYGENS

	PXA11	PXA12	PXA13	PXA14	PXA15	PXA16	PXA17	PXA18	PXB 1	PXB 2
Si	2.014	2.002	2.002	2.008	2.000	2.004	1.985	1.999	1.959	1.952
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.015	0.001	0.041	0.048
	2.014	2.002	2.002	2.008	2.000	2.004	2.000	2.000	2.000	2.000
Al	0.428	0.430	0.420	0.394	0.400	0.367	0.348	0.369	0.197	0.150
Ti	0.003	0.002	0.003	0.004	0.004	0.003	0.004	0.003	0.005	0.005
Fe3	0.037	0.065	0.081	0.083	0.097	0.088	0.141	0.121	0.160	0.119
Mg	0.370	0.371	0.363	0.339	0.355	0.376	0.374	0.351	0.405	0.457
Fe2	0.185	0.158	0.141	0.191	0.155	0.196	0.149	0.174	0.234	0.288
Mn	0.000	0.001	0.002	0.001	0.001	0.000	0.001	0.002	0.000	0.001
Ca	0.463	0.470	0.477	0.476	0.483	0.497	0.500	0.484	0.674	0.749
Na	0.500	0.502	0.511	0.503	0.504	0.468	0.482	0.496	0.325	0.231
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000
	1.986	1.998	1.998	1.992	2.000	1.996	2.000	2.000	2.000	2.000

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.77	0.04	2.07	2.39
CHARGE	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	35.08	34.84	34.10	31.10	32.55	32.51	32.10	30.99	27.48	28.32
Fe	21.07	21.01	21.06	25.23	23.21	24.57	24.98	26.22	26.77	25.25
Ca	43.85	44.15	44.84	43.67	44.24	42.92	42.92	42.80	45.75	46.42

Na-Ti	0.00	0.00	0.00	0.00	0.00	0.00	0.41	0.08	0.53	0.48
Ti-DI	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ACMITE	3.77	6.47	8.12	8.36	9.71	8.81	14.12	12.11	15.99	11.90
Cr DIOPS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JADEITE	43.41	43.21	42.35	40.00	40.36	37.09	33.67	37.16	16.04	10.74
Ca-TS	0.00	0.00	0.00	0.00	0.00	0.00	1.13	0.00	3.62	4.30
FERRI-TS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JOHANNS	0.03	0.09	0.15	0.12	0.09	0.03	0.09	0.15	0.03	0.09
ENSTAT	18.64	18.56	18.15	17.03	17.77	18.85	18.71	17.54	20.24	22.86
FERROSIL	9.29	7.91	7.07	9.58	7.77	9.83	7.46	8.70	11.70	14.38
CaSiO3	24.85	23.75	24.16	24.92	24.29	25.39	24.41	24.25	31.86	35.26

COMPOSITION IN OXIDES

	DIF 1	PXY 1	PXY 2	PXY 3	PXY 4	PXY 5	PXY 6	PXY 7	PXY 8
SiO2	52.98	52.81	52.94	52.93	53.22	53.16	53.14	53.05	52.82
Al2O3	0.18	1.92	1.33	1.64	1.19	1.17	1.99	1.48	1.22
TiO2	0.02	0.04	0.06	0.06	0.07	0.04	0.04	0.04	0.02
Fe2O3	0.53	2.89	1.98	1.51	1.56	1.34	2.34	0.74	4.33
MgO	11.14	9.29	9.63	9.42	10.00	9.51	8.76	9.23	10.51
FeO	11.02	9.31	10.29	10.47	10.53	11.00	10.05	11.50	7.93
MnO	0.19	0.43	0.49	0.43	0.47	0.49	0.45	0.49	0.52
CaO	24.55	20.40	21.44	20.47	21.67	21.60	19.85	20.88	21.49
Na2O	0.17	2.31	1.70	2.02	1.52	1.60	2.58	1.77	1.81
K2O	0.02	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02
	100.80	99.41	99.68	98.97	100.25	99.93	99.22	99.20	100.67

STRUCT. FORM. BASED ON 6.0 OXYGENS

	DIF 1	PXY 1	PXY 2	PXY 3	PXY 4	PXY 5	PXY 6	PXY 7	PXY 8
Si	1.995	2.000	2.003	2.015	2.006	2.013	2.016	2.021	1.978
Al	0.005	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.022
	2.000	2.000	2.003	2.015	2.006	2.013	2.016	2.021	2.000
Al	0.003	0.086	0.059	0.074	0.053	0.052	0.089	0.066	0.032
Ti	0.001	0.001	0.002	0.002	0.002	0.001	0.001	0.001	0.001
Fe3	0.015	0.082	0.056	0.043	0.044	0.038	0.067	0.021	0.122
Mg	0.625	0.524	0.543	0.534	0.562	0.537	0.495	0.524	0.587
Fe2	0.347	0.295	0.325	0.333	0.332	0.348	0.319	0.367	0.248
Mn	0.006	0.014	0.016	0.014	0.015	0.016	0.014	0.016	0.016
Ca	0.990	0.828	0.869	0.835	0.875	0.876	0.807	0.852	0.962
Na	0.012	0.170	0.125	0.149	0.111	0.117	0.190	0.131	0.131
K	0.001	0.000	0.001	0.001	0.001	0.001	0.001	0.001	0.001
	2.000	2.000	1.997	1.995	1.994	1.987	1.984	1.979	2.000

Al% IN Z	0.27	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.12
CHARGE	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00	12.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	31.52	30.08	30.01	30.37	30.73	29.57	29.10	29.45	31.93
Fe	18.56	22.43	21.97	22.19	21.39	22.16	23.51	22.67	21.08
Ca	49.92	47.49	48.02	47.44	47.87	48.27	47.39	47.88	46.97

Na-Ti	0.06	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.06
Ti-DI	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ACMITE	1.28	8.22	5.66	4.37	4.42	3.85	6.74	2.14	12.20
Cr DIOPS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JADEITE	0.00	8.67	6.11	7.59	5.50	5.37	9.09	6.83	0.98
Ca-TS	0.26	0.00	0.00	0.00	0.00	0.00	0.00	0.00	2.18
FERRI-TS	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
JOHANNS	0.61	1.38	1.57	1.40	1.50	1.58	1.46	1.60	1.65
ENSTAT	31.33	26.22	27.20	26.92	28.16	27.01	24.98	26.48	29.33
FERROSIL	17.39	14.75	16.30	16.79	16.64	17.53	16.09	18.52	12.42
CaSiO3	49.08	40.75	43.16	42.94	43.77	44.66	41.66	44.43	41.19

COMPOSITION IN OXIDES

	PXZ 1	PXZ 2	PXZ 3	PXZ 4	PXZ 5	PXZ 6
SiO2	52.62	52.75	52.89	52.67	53.08	52.79
Al2O3	1.15	1.24	1.15	1.11	1.10	1.06
TiO2	0.04	0.02	0.02	0.02	0.06	0.02
Fe2O3	5.09	4.13	5.69	5.50	5.22	4.74
MgO	9.61	9.16	9.52	9.56	9.51	9.67
FeO	9.11	10.12	8.59	8.84	9.14	9.31
MnO	0.45	0.45	0.50	0.43	0.50	0.43
CaO	21.25	20.94	21.11	21.21	21.51	21.31
Na2O	1.95	2.02	2.19	2.05	2.02	1.91
	101.27	100.82	101.66	101.39	102.14	101.25

STRUCT. FORM. BASED ON 6.0 OXYGENS

	PXZ 1	PXZ 2	PXZ 3	PXZ 4	PXZ 5	PXZ 6
Si	1.972	1.987	1.973	1.972	1.974	1.979
Al	0.028	0.013	0.027	0.028	0.026	0.021
	2.000	2.000	2.000	2.000	2.000	2.000
Al	0.023	0.042	0.024	0.021	0.022	0.025
Ti	0.001	0.001	0.001	0.001	0.002	0.001
Fe3	0.144	0.117	0.160	0.155	0.146	0.134
Mg	0.537	0.514	0.529	0.533	0.527	0.540
Fe2	0.285	0.319	0.268	0.277	0.284	0.292
Mn	0.014	0.014	0.016	0.014	0.016	0.014
Ca	0.853	0.845	0.844	0.851	0.857	0.856
Na	0.142	0.148	0.153	0.149	0.146	0.139
	2.000	2.000	2.000	2.000	2.000	2.000

Al% IN Z	1.38	0.64	1.33	1.40	1.30	1.07
CHARGE	12.00	12.00	12.00	12.00	12.00	12.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	29.28	28.42	29.14	29.16	28.80	29.44
Fe	24.18	24.87	24.41	24.34	24.38	23.94
Ca	46.54	46.71	46.45	46.50	46.82	46.63

Na-Ti	0.11	0.06	0.06	0.06	0.17	0.06
Ti-DI	0.00	0.00	0.00	0.00	0.00	0.00
ACMITE	14.10	11.69	15.82	14.92	14.43	13.37
Cr DIOPS	0.00	0.00	0.00	0.00	0.00	0.00
JADEITE	0.00	3.00	0.00	0.00	0.00	0.45
Ca-TS	2.34	1.22	2.41	2.11	2.23	2.09
FERRI-TS	0.00	0.00	0.00	0.00	0.00	0.00
JOHANNS	1.43	1.44	1.53	1.37	1.58	1.37
ENSTAT	26.93	25.72	26.53	26.85	26.41	27.01
FERROSIL	14.32	15.94	13.42	13.94	14.25	14.60
CaSiO3	40.77	40.93	40.19	40.75	40.94	41.06

COMPOSITION IN OXIDES

	HBA 1	HBA 2	HBA 3	HBA 4	HBB 1	ACD 1	ACD 2	ACD 3	ACD 4	ACD 5
SiO2	44.86	45.45	43.66	43.35	43.81	54.62	55.90	54.19	52.99	55.[illegible]
Al2O3	12.59	13.06	13.59	12.62	13.00	2.69	0.91	3.64	3.97	1.98
TiO2	0.62	0.56	0.58	0.69	0.74	0.06	0.00	0.04	0.11	0.06
MgO	9.60	9.43	9.16	10.09	8.18	15.73	16.73	15.89	14.70	16.58
FeO	17.05	17.34	17.59	17.19	21.16	12.17	11.99	11.83	13.03	11.26
MnO	0.08	0.13	0.08	0.07	0.05	0.24	0.39	0.30	0.22	0.27
CaO	10.63	9.58	9.33	9.74	8.74	11.38	11.53	11.27	11.41	12.11
Na2O	3.90	4.68	3.74	3.85	2.69	0.45	0.18	0.55	0.57	0.20
K2O	0.25	0.40	0.16	0.14	0.38	0.00	0.00	0.00	0.21	0.00
	99.58	99.63	97.89	97.74	98.75	97.34	97.63	97.70	97.21	97.55

STRUCT. FORM. BASED ON 23.0 OXYGENS

	HBA 1	HBA 2	HBA 3	HBA 4	HBB 1	ACD 1	ACD 2	ACD 3	ACD 4	ACD 5
Si	6.577	6.638	6.503	6.484	6.549	7.825	7.974	7.726	7.662	7.[illegible]
Al	1.423	1.362	1.495	1.516	1.451	0.175	0.026	0.274	0.338	0.145
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	0.753	0.886	0.891	0.708	0.839	0.279	0.127	0.338	0.338	0.188
Ti	0.068	0.062	0.065	0.078	0.083	0.006	0.000	0.004	0.012	0.006
Mg	2.098	2.053	2.034	2.249	1.823	3.359	3.557	3.377	3.108	3.524
Fe2	2.091	2.118	2.192	2.150	2.645	1.458	1.430	1.411	1.576	1.343
Mn	0.010	0.016	0.010	0.009	0.006	0.029	0.047	0.036	0.027	0.033
Ca	1.670	1.343	1.489	1.561	1.400	1.747	1.762	1.722	1.768	1.850
Na	1.109	1.325	1.080	1.116	0.780	0.125	0.050	0.152	0.160	0.055
K	0.047	0.075	0.030	0.027	0.072	0.000	0.000	0.000	0.039	0.000
	7.845	7.877	7.792	7.898	7.649	7.004	6.974	7.040	7.087	6.999

ALK IN X	17.79	17.03	18.69	18.96	18.14	2.19	0.32	3.43	4.23	1.81
CHARGE	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	35.75	37.13	35.53	37.68	31.03	50.95	52.34	51.59	48.45	52.21
Fe	35.79	38.59	38.46	36.17	45.14	22.56	21.74	22.10	24.51	20.38
Ca	28.46	24.28	26.01	26.15	23.83	26.49	25.93	26.30	27.04	27.41

AMPHIBOLE SITE OCCUPANCIES

A SITE	0.845	0.877	0.792	0.898	0.649	0.004	0.000	0.040	0.087	0.000
ALK IN X	0.311	0.523	0.318	0.245	0.203	0.121	0.050	0.112	0.111	0.055
Ca IN X	1.670	1.343	1.489	1.561	1.400	1.747	1.762	1.722	1.768	1.850
Mg+Fe2 X	0.019	0.134	0.192	0.194	0.397	0.132	0.162	0.166	0.121	0.094
TOTAL X	2.000	2.000	2.000	2.000	2.000	2.000	1.974	2.000	2.000	1.999
3+ IN Y	0.821	0.947	0.956	0.736	0.923	0.286	0.127	0.342	0.350	0.195
Fe+Mg Y	4.179	4.053	4.044	4.214	4.077	4.714	4.873	4.658	4.650	4.805
Y DEFNY.	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL Y	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000

CLASSIFICATION OF ANALYSIS

AM CLASS	C-HBD	MGKAT	C-HBD.	PARG	C-HBD	TREM	TREM	TREM	TREM	TREM

COMPOSITION IN OXIDES

	ACD 6	HBD 1	HBD 2	HBD 3	HBD 4	HBE 1	HBE 2	HBE 3	HBE 4	HBE 5
SiO2	55.41	45.39	46.10	45.81	45.74	45.68	47.09	46.28	46.42	47.29
Al2O3	1.36	12.92	12.18	12.67	12.78	12.46	10.51	11.40	11.22	10.39
TiO2	0.00	0.26	0.24	0.31	0.29	0.33	0.28	0.29	0.20	0.29
MgO	16.67	10.18	10.76	10.34	10.70	12.61	13.61	13.34	13.03	13.53
FeO	11.40	14.69	15.03	14.73	14.68	12.83	12.96	12.94	13.45	13.21
MnO	0.00	0.20	0.24	0.24	0.23	0.22	0.15	0.18	0.20	0.16
CaO	11.71	11.53	11.35	11.49	11.46	11.18	11.03	11.07	11.19	11.19
Na2O	0.18	1.39	1.24	1.31	1.34	1.64	1.61	1.61	1.58	1.61
K2O	0.00	0.00	0.16	0.16	0.16	0.13	0.12	0.13	0.19	0.14
	96.73	96.56	97.30	97.06	97.38	97.08	97.36	97.24	97.39	97.70

STRUCT. FORM. BASED ON 23.0 OXYGENS

	ACD 6	HBD 1	HBD 2	HBD 3	HBD 4	HBE 1	HBE 2	HBE 3	HBE 4	HBE 5
Si	7.949	6.720	6.781	6.750	6.717	6.683	6.860	6.758	6.785	6.877
Al	0.051	1.280	1.219	1.250	1.283	1.319	1.140	1.242	1.215	1.127
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.004
Al	0.179	0.975	0.892	0.950	0.930	0.829	0.664	0.720	0.718	0.651
Ti	0.000	0.029	0.027	0.034	0.032	0.036	0.031	0.032	0.022	0.032
Mg	3.564	2.247	2.359	2.271	2.342	2.749	2.955	2.903	2.839	2.931
Fe2	1.368	1.819	1.849	1.815	1.803	1.569	1.579	1.580	1.644	1.606
Mn	0.000	0.025	0.030	0.030	0.029	0.027	0.019	0.022	0.025	0.020
Ca	1.800	1.829	1.789	1.814	1.803	1.752	1.722	1.732	1.738	1.727
Na	0.050	0.399	0.354	0.374	0.382	0.465	0.455	0.456	0.448	0.454
K	0.000	0.000	0.030	0.030	0.030	0.024	0.022	0.024	0.035	0.026
	6.961	7.323	7.329	7.318	7.350	7.453	7.446	7.469	7.468	7.446

ALK IN Z	0.64	15.99	15.24	15.63	16.03	16.48	14.25	15.53	15.19	14.09
CHARGE	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	52.95	37.95	39.14	38.30	39.19	45.08	47.10	46.55	45.45	46.65
Fe	20.32	31.15	31.17	31.11	30.64	26.18	25.46	25.69	26.72	25.87
Ca	26.74	30.90	29.68	30.59	30.17	28.73	27.44	27.76	27.83	27.48

AMPHIBOLE SITE OCCUPANCIES

A SITE	0.000	0.323	0.329	0.318	0.350	0.453	0.446	0.469	0.468	0.446
ALK IN X	0.050	0.076	0.055	0.086	0.061	0.036	0.031	0.011	0.015	0.034
Ca IN X	1.800	1.829	1.789	1.814	1.803	1.752	1.722	1.732	1.738	1.727
Mg+Fe2 X	0.111	0.095	0.156	0.100	0.135	0.212	0.247	0.257	0.247	0.239
TOTAL X	1.961	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000
3+ IN Y	0.179	1.004	0.919	0.984	0.962	0.866	0.695	0.752	0.740	0.683
Fe+Mg Y	4.821	3.996	4.081	4.016	4.038	4.134	4.305	4.248	4.260	4.317
Y DEFNY.	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL Y	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000

CLASSIFICATION OF ANALYSIS

AM CLASS TREM	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	

COMPOSITION IN OXIDES

	HBE 6	HBE 7	HBE 8	HBE 9	HBE10	HBE11	TRF 1	TRF 2	HBH 1	HBH 2
SiO2	45.87	44.97	44.75	45.56	45.71	44.82	50.45	51.74	40.41	40.04
Al2O3	11.98	13.30	12.75	12.00	11.85	13.46	2.98	2.98	16.71	17.17
TiO2	0.35	0.33	0.42	0.28	0.35	0.29	0.11	0.06	0.23	0.34
MgO	12.64	12.11	12.42	12.70	12.85	11.88	13.87	14.38	6.12	6.03
FeO	13.30	13.26	13.18	13.02	13.07	13.55	12.87	12.70	20.65	21.97
MnO	0.19	0.20	0.15	0.14	0.12	0.24	0.09	0.12	0.32	0.24
CaO	11.12	11.53	11.34	11.03	11.11	11.41	11.76	11.34	10.41	10.43
Na2O	1.72	1.62	1.53	1.58	1.68	1.59	0.50	0.55	1.77	1.96
K2O	0.19	0.20	0.12	0.14	0.15	0.19	0.14	0.12	0.22	0.30
	97.36	97.52	96.66	96.45	96.89	97.43	92.77	94.49	97.04	97.38

STRUCT. FORM. BASED ON 23.0 OXYGENS

	HBE 6	HBE 7	HBE 8	HBE 9	HBE10	HBE11	TRF 1	TRF 2	HBH 1	HBH 2
Si	6.708	6.575	6.596	6.712	6.709	6.566	7.683	7.713	6.174	6.103
Al	1.292	1.425	1.404	1.288	1.291	1.434	0.317	0.287	1.826	1.895
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	0.773	0.867	0.811	0.795	0.759	0.890	0.218	0.236	1.132	1.191
Ti	0.038	0.036	0.047	0.031	0.039	0.032	0.013	0.007	0.026	0.039
Mg	2.755	2.639	2.729	2.789	2.811	2.594	3.149	3.195	1.394	1.370
Fe2	1.627	1.621	1.625	1.604	1.604	1.660	1.639	1.583	2.664	2.674
Mn	0.024	0.025	0.019	0.017	0.015	0.030	0.012	0.015	0.041	0.031
Ca	1.742	1.806	1.791	1.741	1.747	1.791	1.919	1.891	1.704	1.704
Na	0.488	0.459	0.437	0.451	0.478	0.452	0.148	0.159	0.524	0.550
K	0.035	0.037	0.023	0.026	0.028	0.036	0.027	0.023	0.043	0.058
	7.482	7.491	7.480	7.455	7.481	7.484	7.124	7.110	7.579	7.617

AL% IN Z	16.15	17.81	17.55	16.11	16.14	17.93	3.96	3.59	22.83	23.69
CHARGE	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	44.82	43.32	44.28	45.34	45.51	42.70	46.87	47.80	24.02	23.71
Fe	26.84	27.02	26.67	26.36	26.21	27.82	24.57	23.91	46.62	46.90
Ca	28.34	29.65	29.06	28.30	28.28	29.48	28.56	28.29	29.36	29.40

AMPHIBOLE SITE OCCUPANCIES

A SITE	0.482	0.491	0.480	0.455	0.481	0.484	0.124	0.110	0.579	0.617
ALK IN X	0.041	0.005	0.000	0.023	0.025	0.003	0.051	0.072	0.000	0.000
Ca IN X	1.742	1.806	1.771	1.741	1.747	1.791	1.919	1.891	1.692	1.695
Mg+Fe2 X	0.217	0.188	0.229	0.236	0.228	0.206	0.030	0.037	0.308	0.305
TOTAL X	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000
3+ IN Y	0.812	0.903	0.857	0.826	0.797	0.922	0.231	0.243	1.209	1.230
Fe+Mg Y	4.188	4.097	4.143	4.174	4.203	4.078	4.769	4.757	3.791	3.770
Y DEFNY.	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL Y	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000

CLASSIFICATION OF ANALYSIS

AM CLASS	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	C-HBD	TREM	TREM	HAST	HAST

COMPOSITION IN OXIDES

	HBH 3	HBH 4	HBH 5	HBH 6	HBX 1	HBX 2	HBX 3	HBX 4	HBX 5	HBX 6
SiO2	40.41	40.12	40.32	40.46	39.51	39.53	40.28	40.04	39.27	39.77
Al2O3	16.38	16.00	16.99	16.86	20.86	21.33	20.81	20.40	20.67	20.65
TiO2	0.34	0.36	0.22	0.29	0.29	0.25	0.25	0.31	0.23	0.31
MgO	6.19	5.78	7.02	7.04	3.61	3.41	3.66	3.65	3.67	3.64
FeO	20.72	21.75	19.09	19.35	21.04	21.25	21.09	21.30	20.46	21.57
MnO	0.27	0.37	0.07	0.12	0.13	0.16	0.20	0.15	0.19	0.23
CaO	10.46	11.08	10.50	10.46	10.84	11.18	10.78	10.34	10.68	10.91
Na2O	1.88	1.44	1.85	2.01	1.10	0.95	1.46	1.45	1.19	1.03
K2O	0.28	0.37	0.30	0.32	0.34	0.37	0.28	0.28	0.00	0.29
	96.93	97.27	96.36	96.94	97.72	98.43	98.81	98.42	96.36	98.40

STRUCT. FORM. BASED ON 23.0 OXYGENS

	HBH 3	HBH 4	HBH 5	HBH 6	HBX 1	HBX 2	HBX 3	HBX 4	HBX 5	HBX 6
Si	6.185	6.165	6.152	6.150	5.966	5.932	6.011	6.016	5.991	5.977
Al	1.815	1.835	1.848	1.850	2.034	2.068	1.989	1.984	2.009	2.023
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	1.139	1.062	1.208	1.170	1.078	1.704	1.671	1.629	1.707	1.634
Ti	0.039	0.042	0.025	0.033	0.033	0.028	0.028	0.035	0.026	0.035
Mg	1.412	1.324	1.597	1.595	0.812	0.763	0.814	0.817	0.835	0.815
Fe2	2.652	2.795	2.436	2.463	2.657	2.667	2.632	2.739	2.610	2.711
Mn	0.035	0.048	0.009	0.015	0.017	0.020	0.025	0.019	0.025	0.029
Ca	1.715	1.824	1.717	1.703	1.754	1.798	1.724	1.665	1.746	1.757
Na	0.558	0.429	0.547	0.592	0.322	0.276	0.422	0.422	0.352	0.300
K	0.055	0.073	0.058	0.062	0.065	0.071	0.053	0.054	0.000	0.056
	7.605	7.596	7.597	7.634	7.339	7.327	7.369	7.380	7.301	7.337

AL% IN Z	22.69	22.94	23.09	23.13	25.43	25.85	24.87	24.80	25.12	25.29
CHARGE	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00	46.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	24.29	22.10	27.73	27.61	15.51	14.54	15.67	15.60	16.00	15.35
Fe	46.21	47.46	42.46	42.91	51.02	51.21	51.15	52.64	50.52	51.59
Ca	29.50	30.45	29.81	29.49	33.47	34.26	33.19	31.76	33.47	33.06

AMPHIBOLE SITE OCCUPANCIES

A SITE	0.605	0.596	0.597	0.634	0.339	0.327	0.369	0.380	0.301	0.337
ALK IN X	0.007	0.000	0.008	0.020	0.049	0.020	0.107	0.096	0.051	0.018
Ca IN X	1.715	1.730	1.717	1.703	1.754	1.798	1.724	1.665	1.746	1.757
Mg+Fe2 X	0.277	0.270	0.275	0.277	0.197	0.182	0.170	0.240	0.203	0.225
TOTAL X	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000
3+ IN Y	1.178	1.104	1.233	1.203	1.711	1.733	1.699	1.664	1.734	1.669
Fe+Mg Y	3.822	3.896	3.767	3.797	3.289	3.267	3.301	3.336	3.266	3.331
Y DEFNY.	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
TOTAL Y	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000

CLASSIFICATION OF ANALYSIS

AM CLASS	HAST	HAST	HAST	HAST	FTSCH	FTSCH	FTSCH	FTSCH	FTSCH	FTSCH

COMPOSITION IN OXIDES

	EPB 1	EPB 2	EPB 3	EPB 4	EPB 5	EPD 1	EPD 2	EPD 3	EPD 4	EPD 5
SiO2	39.50	38.63	39.16	38.75	39.22	39.04	39.34	38.68	38.90	39.31
Al2O3	25.90	26.63	26.21	26.14	25.86	29.65	30.37	29.06	27.64	29.45
TiO2	0.19	0.17	0.09	0.15	0.19	0.09	0.06	0.09	0.11	0.06
Fe2O3	10.72	9.75	9.91	10.15	10.70	4.80	4.66	5.50	7.28	5.27
MgO	0.08	0.08	0.02	0.04	0.06	0.04	0.04	0.10	0.02	0.00
MnO	0.00	0.03	0.00	0.04	0.00	0.14	0.20	0.25	0.22	0.19
CaO	23.54	23.61	23.86	23.52	23.53	24.06	24.13	23.89	23.45	24.01
Na2O	0.03	0.02	0.00	0.02	0.00	0.03	0.00	0.02	0.03	0.05
K2O	0.01	0.02	0.02	0.02	0.01	0.00	0.00	0.00	0.00	0.00
	98.97	98.94	99.27	98.83	99.57	97.85	98.80	97.79	97.65	98.34

STRUCT. FORM. BASED ON 25.0 OXYGENS

	EPB 1	EPB 2	EPB 3	EPB 4	EPB 5	EPD 1	EPD 2	EPD 3	EPD 4	EPD 5
Si	5.995	5.996	6.059	6.028	6.059	6.023	6.004	6.020	6.062	6.042
Al	0.005	0.004	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	6.000	6.000	6.059	6.028	6.059	6.023	6.004	6.020	6.062	6.042
Al	4.748	4.867	4.779	4.792	4.708	5.391	5.463	5.303	5.077	5.335
Ti	0.022	0.020	0.010	0.018	0.022	0.010	0.007	0.010	0.013	0.007
Fe3	1.257	1.138	1.154	1.188	1.244	0.557	0.535	0.641	0.854	0.619
Mg	0.019	0.019	0.005	0.009	0.014	0.009	0.009	0.023	0.005	0.000
Mn	0.000	0.004	0.000	0.005	0.000	0.018	0.026	0.033	0.029	0.025
Ca	3.927	3.926	3.955	3.920	3.895	3.977	3.946	3.963	3.916	3.954
Na	0.009	0.006	0.000	0.006	0.000	0.009	0.000	0.006	0.009	0.015
K	0.002	0.004	0.004	0.004	0.002	0.000	0.000	0.000	0.000	0.000
	9.984	9.984	9.908	9.942	9.885	9.973	9.986	9.980	9.902	9.945

Al% IN Z	0.09	0.07	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.36	0.36	0.09	0.18	0.27	0.20	0.20	0.50	0.10	0.00
Fe	24.15	22.46	22.57	23.29	24.15	12.62	12.42	14.46	18.38	13.32
Ca	75.49	77.18	77.34	76.53	75.59	87.18	87.38	85.05	81.52	86.68

ENDMEMBERS

EPIDOTE	62.69	56.76	59.70	59.98	63.72	28.60	26.77	32.97	44.22	31.34
ZOISITE	37.31	43.24	40.30	40.02	36.28	71.40	73.23	67.03	55.78	68.66

SUM IONS	15.96	15.98	15.97	15.97	15.94	16.00	15.99	16.00	15.96	15.99
Si (VI)	0.000	0.000	0.059	0.028	0.059	0.023	0.004	0.020	0.062	0.042

COMPOSITION IN OXIDES

	EPD 6	EPE 1	EPE 2	EPE 3	EPE 4	EPE 5	EPE 6	EPE 7	EPE 8	EPF 1
SiO2	39.37	38.40	39.06	38.20	38.10	38.11	37.93	38.24	38.22	36.54
Al2O3	30.28	23.32	23.68	23.57	23.40	23.58	23.31	23.39	23.41	26.26
TiO2	0.06	0.04	0.04	0.04	0.06	0.04	0.04	0.06	0.06	0.06
Fe2O3	3.92	12.94	12.78	12.96	12.86	12.98	13.06	12.85	13.36	6.91
MgO	0.04	0.00	0.02	0.04	0.02	0.02	0.00	0.04	0.04	0.08
MnO	0.10	0.22	0.20	0.31	0.24	0.41	0.23	0.24	0.19	0.08
CaO	24.12	23.22	23.14	23.17	23.10	23.27	23.26	23.38	23.27	23.71
Na2O	0.00	0.00	0.02	0.00	0.03	0.02	0.02	0.02	0.00	0.03
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03
	97.89	98.14	97.94	98.29	97.81	98.43	97.85	98.22	98.55	95.70

STRUCT. FORM. BASED ON 25.0 OXYGENS

	EPD 6	EPE 1	EPE 2	EPE 3	EPE 4	EPE 5	EPE 6	EPE 7	EPE 8	EPF 1
Si	6.047	6.085	6.042	6.048	6.060	6.032	6.040	6.060	6.042	6.136
	6.047	6.085	6.042	6.048	6.060	6.032	6.040	6.060	6.042	6.136
Al	5.481	4.356	4.431	4.398	4.386	4.399	4.375	4.369	4.362	4.927
Ti	0.007	0.005	0.005	0.005	0.007	0.005	0.005	0.007	0.007	0.007
Fe3	0.453	1.543	1.527	1.544	1.539	1.546	1.565	1.532	1.589	0.826
Mg	0.009	0.000	0.005	0.009	0.005	0.005	0.000	0.009	0.009	0.019
Mn	0.013	0.030	0.027	0.042	0.032	0.055	0.031	0.032	0.025	0.011
Ca	3.969	3.943	3.936	3.930	3.936	3.946	3.968	3.970	3.941	4.044
Na	0.000	0.000	0.006	0.000	0.009	0.006	0.006	0.006	0.000	0.009
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.006
	9.933	9.875	9.936	9.928	9.915	9.962	9.949	9.925	9.934	9.852

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.21	0.00	0.09	0.17	0.09	0.08	0.00	0.17	0.17	0.39
Fe	10.49	28.51	28.28	28.69	28.50	28.84	28.68	28.22	29.01	17.11
Ca	89.30	71.49	71.64	71.14	71.41	71.08	71.32	71.61	70.82	82.50

ENDMEMBERS

EPIDOTE	23.44	81.27	78.00	79.50	79.93	79.50	80.69	80.60	81.47	47.18
ZOISITE	76.56	18.73	22.00	20.50	20.07	20.50	19.31	19.40	18.53	52.82

SUM IONS	15.98	15.96	15.98	15.98	15.98	15.99	15.99	15.99	15.98	15.99
Si (VI)	0.047	0.085	0.042	0.048	0.060	0.032	0.040	0.060	0.042	0.136

COMPOSITION IN OXIDES

	EPD 6	EPE 1	EPE 2	EPE 3	EPE 4	EPE 5	EPE 6	EPE 7	EPE 8	EPF 1
SiO2	39.37	38.40	39.06	38.20	38.10	38.11	37.93	38.24	38.22	36.54
Al2O3	30.28	23.32	23.68	23.57	23.40	23.58	23.31	23.39	23.41	26.26
TiO2	0.06	0.04	0.04	0.04	0.06	0.04	0.04	0.06	0.06	0.06
Fe2O3	3.92	12.94	12.78	12.96	12.86	12.98	13.06	12.85	13.36	6.91
MgO	0.04	0.00	0.02	0.04	0.02	0.02	0.00	0.04	0.04	0.08
MnO	0.10	0.22	0.20	0.31	0.24	0.41	0.23	0.24	0.19	0.08
CaO	24.12	23.22	23.14	23.17	23.10	23.27	23.26	23.38	23.27	23.71
Na2O	0.00	0.00	0.02	0.00	0.03	0.02	0.02	0.02	0.00	0.03
K2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.03
	97.89	98.14	97.94	98.29	97.81	98.43	97.85	98.22	98.55	95.70

STRUCT. FORM. BASED ON 25.0 OXYGENS

	EPD 6	EPE 1	EPE 2	EPE 3	EPE 4	EPE 5	EPE 6	EPE 7	EPE 8	EPF 1
Si	6.047	6.085	6.042	6.048	6.060	6.032	6.040	6.060	6.042	6.136
	6.047	6.085	6.042	6.048	6.060	6.032	6.040	6.060	6.042	6.136
Al	5.481	4.356	4.431	4.398	4.386	4.399	4.375	4.369	4.362	4.927
Ti	0.007	0.005	0.005	0.005	0.007	0.005	0.005	0.007	0.007	0.007
Fe3	0.453	1.543	1.527	1.544	1.539	1.546	1.565	1.532	1.589	0.828
Mg	0.009	0.000	0.005	0.009	0.005	0.005	0.000	0.009	0.009	0.019
Mn	0.013	0.030	0.027	0.042	0.032	0.055	0.031	0.032	0.025	0.011
Ca	3.969	3.943	3.936	3.930	3.936	3.946	3.968	3.970	3.941	4.044
Na	0.000	0.000	0.006	0.000	0.009	0.006	0.006	0.006	0.000	0.009
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.004
	9.933	9.875	9.936	9.928	9.915	9.962	9.949	9.925	9.934	9.852

AL% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.21	0.00	0.09	0.17	0.09	0.08	0.00	0.17	0.17	0.39
Fe	10.49	28.51	28.28	28.69	28.50	28.84	28.68	28.22	29.01	17.11
Ca	89.30	71.49	71.64	71.14	71.41	71.08	71.32	71.61	70.82	82.50

ENDMEMBERS

EPIDOTE	23.44	81.27	78.00	79.50	79.93	79.50	80.69	80.60	81.47	47.19
ZOISITE	76.56	18.73	22.00	20.50	20.07	20.50	19.31	19.40	18.53	52.82

SUM IONS	15.98	15.96	15.98	15.98	15.98	15.99	15.99	15.99	15.98	15.99
Si (VI)	0.047	0.085	0.042	0.048	0.060	0.032	0.040	0.060	0.042	0.136

COMPOSITION IN OXIDES

	EPF 2	EPH 1	EPH 2	EPH 3	EPH 4	EPH 5	EPU 1	EPU 2	EPU 3	EPU 4
SiO2	38.67	38.08	38.50	38.01	38.19	38.29	39.70	38.46	39.29	38.61
Al2O3	27.21	25.15	26.67	26.43	26.56	25.90	23.41	22.73	23.48	22.82
TiO2	0.32	0.07	0.15	0.11	0.09	0.11	0.04	0.07	0.04	0.04
Fe2O3	6.38	11.18	9.06	9.55	9.49	10.37	12.18	12.38	11.32	12.10
MgO	0.04	0.02	0.02	0.02	0.02	0.06	0.04	0.04	0.02	0.03
MnO	0.08	0.28	0.15	0.11	0.16	0.09	0.47	0.66	0.65	0.40
CaO	23.79	23.26	23.53	23.46	23.64	23.20	23.41	21.88	21.64	21.31
Na2O	0.02	0.00	0.00	0.02	0.00	0.08	0.00	0.00	0.02	0.00
K2O	0.02	0.00	0.00	0.00	0.00	0.01	0.01	0.02	0.01	0.01
	96.57	98.04	98.08	97.71	98.15	98.11	99.26	96.24	96.46	95.87

STRUCT. FORM. BASED ON 25.0 OXYGENS

	EPF 2	EPH 1	EPH 2	EPH 3	EPH 4	EPH 5	EPU 1	EPU 2	EPU 3	EPU 4
Si	6.087	6.005	6.017	5.978	5.979	6.007	6.198	6.193	6.267	6.223
Al	0.000	0.000	0.000	0.022	0.021	0.000	0.000	0.000	0.000	0.000
	6.087	6.005	6.017	6.000	6.000	6.007	6.198	6.193	6.267	6.223
Al	5.048	4.674	4.912	4.877	4.880	4.789	4.307	4.314	4.415	4.335
Ti	0.038	0.008	0.018	0.013	0.011	0.013	0.005	0.008	0.005	0.005
Fe3	0.756	1.327	1.065	1.130	1.118	1.224	1.431	1.500	1.360	1.468
Mg	0.019	0.005	0.005	0.005	0.005	0.014	0.009	0.010	0.005	0.019
Mn	0.011	0.037	0.020	0.015	0.021	0.012	0.062	0.090	0.089	0.055
Ca	4.012	3.930	3.940	3.953	3.966	3.900	3.916	3.775	3.699	3.766
Na	0.006	0.000	0.000	0.006	0.000	0.024	0.000	0.000	0.006	0.000
K	0.004	0.000	0.000	0.000	0.000	0.002	0.002	0.004	0.002	0.002
	9.893	9.981	9.960	9.998	10.001	9.979	9.732	9.701	9.579	9.649

Al IV Z	0.00	0.00	0.00	0.37	0.35	0.00	0.00	0.00	0.00	0.00
CHARGE	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00	50.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.39	0.09	0.09	0.09	0.09	0.27	0.17	0.18	0.09	0.36
Fe	15.97	25.74	21.57	22.43	22.30	24.00	27.56	29.59	28.10	28.49
Ca	83.64	74.17	78.33	77.48	77.61	75.73	72.27	70.23	71.81	71.15

ENDMEMBERS

EPIDOTE	41.90	66.30	53.86	56.31	55.95	60.80	82.32	82.71	76.62	81.43
ZOISITE	58.10	33.70	46.14	43.69	44.05	39.20	17.68	17.29	23.38	18.57

SUM IONS	15.98	15.99	15.98	16.00	16.00	15.99	15.93	15.89	15.85	15.87
Si (VI)	0.087	0.005	0.017	0.000	0.000	0.007	0.198	0.193	0.267	0.223

COMPOSITION IN OXIDES

	EPX 1	EPX 2	EPX 3	EPX 4	EPY 1	EPY 2	EPY 3
SiO2	39.24	40.03	39.56	38.83	38.09	38.07	38.07
Al2O3	28.10	28.49	28.09	27.95	23.31	22.85	23.07
TiO2	0.17	0.21	0.11	0.13	0.11	0.07	0.09
Fe2O3	6.88	6.69	6.45	6.60	13.85	14.17	14.01
MgO	0.04	0.06	0.02	0.04	0.06	0.04	0.04
MnO	0.18	0.14	0.14	0.16	0.22	0.14	0.18
CaO	23.52	23.79	23.47	23.43	22.16	22.21	22.19
Na2O	0.02	0.02	0.02	0.02	0.00	0.00	0.00
K2O	0.02	0.03	0.02	0.02	0.02	0.02	0.02
	98.17	99.46	97.88	97.58	97.82	97.57	97.66

STRUCT. FORM. BASED ON 25.0 OXYGENS

	EPX 1	EPX 2	EPX 3	EPX 4	EPY 1	EPY 2	EPY 3
Si	6.070	6.101	6.122	6.049	6.057	6.077	6.067
	6.070	6.101	6.122	6.049	6.057	6.077	6.067
Al	5.123	5.118	5.123	5.132	4.369	4.299	4.333
Ti	0.020	0.024	0.013	0.015	0.013	0.008	0.011
Fe3	0.801	0.767	0.751	0.774	1.657	1.702	1.681
Mg	0.009	0.014	0.005	0.009	0.014	0.010	0.010
Mn	0.024	0.018	0.018	0.021	0.030	0.019	0.024
Ca	3.898	3.885	3.891	3.978	3.776	3.798	3.787
Na	0.006	0.006	0.006	0.006	0.000	0.000	0.000
K	0.004	0.006	0.004	0.004	0.004	0.004	0.004
	9.884	9.837	9.811	9.939	9.862	9.840	9.850

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	50.00	50.00	50.00	50.00	50.00	50.00	50.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.19	0.29	0.10	0.19	0.26	0.17	0.17
Fe	17.42	16.77	16.48	16.63	30.80	31.13	30.99
Ca	82.39	82.94	83.42	83.18	68.94	68.70	68.84

ENDMEMBERS

EPIDOTE	41.63	40.71	40.06	40.61	81.80	85.07	83.45
ZOISITE	58.37	59.29	59.94	59.39	18.20	14.93	16.55

SUM IONS	15.95	15.94	15.93	15.99	15.92	15.92	15.92
Si (VI)	0.070	0.101	0.122	0.049	0.057	0.077	0.067

COMPOSITION IN OXIDES

	PLA 1	PLB 1	PLB 2	PLB 3	PLB 4	PLD 1	PLD 2	PLD 3	PLD 4	PLE 1
SiO2	68.89	68.36	70.08	69.96	67.92	63.32	62.29	69.93	69.73	69.13
Al2O3	19.89	19.95	19.56	19.47	19.51	23.84	24.92	19.86	19.88	19.59
TiO2	0.02	0.00	0.00	0.00	0.02	0.02	0.02	0.02	0.02	0.00
MgO	0.02	0.02	0.02	0.04	0.02	0.00	0.00	0.00	0.02	0.02
FeO	0.19	0.09	0.07	0.12	0.07	0.06	0.09	0.07	0.00	0.01
MnO	0.01	0.00	0.00	0.04	0.00	0.03	0.00	0.00	0.01	0.00
CaO	0.34	0.16	0.16	0.23	0.26	4.53	5.77	0.23	0.23	0.13
Na2O	11.67	11.62	12.00	11.75	12.13	8.83	8.34	11.54	11.54	11.87
K2O	0.05	0.05	0.04	0.05	0.04	0.10	0.10	0.07	0.10	0.07
	101.08	99.28	101.93	101.66	100.02	100.73	101.53	101.72	101.53	100.82

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLA 1	PLB 1	PLB 2	PLB 3	PLB 4	PLD 1	PLD 2	PLD 3	PLD 4	PLE 1
Si	11.924	12.029	12.014	12.023	11.905	11.108	10.888	11.996	11.986	11.982
Al	4.057	3.936	3.952	3.944	4.030	4.392	5.112	4.004	4.014	4.002
Ti	0.003	0.000	0.000	0.000	0.003	0.000	0.000	0.000	0.000	0.000
	15.984	15.966	15.966	15.967	15.938	16.000	16.000	16.000	16.000	15.984
Al	0.000	0.000	0.000	0.000	0.000	0.037	0.022	0.012	0.013	0.000
Ti	0.000	0.000	0.000	0.000	0.000	0.003	0.003	0.003	0.003	0.000
Mg	0.005	0.005	0.005	0.010	0.005	0.000	0.000	0.000	0.005	0.005
Fe2	0.028	0.013	0.010	0.017	0.013	0.009	0.013	0.010	0.000	0.001
Mn	0.001	0.000	0.000	0.006	0.000	0.004	0.000	0.000	0.001	0.000
Ca	0.063	0.030	0.029	0.042	0.049	0.851	1.081	0.042	0.042	0.024
Na	3.916	3.965	3.989	3.915	4.139	3.020	2.826	3.838	3.846	3.999
K	0.011	0.011	0.009	0.011	0.009	0.022	0.022	0.015	0.022	0.015
	4.025	4.024	4.042	4.002	4.213	3.947	3.967	3.920	3.932	4.045

Al% IN Z	25.36	24.66	24.75	24.70	25.29	30.58	31.95	25.02	25.09	25.04
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	5.31	10.79	11.48	15.54	8.12	0.00	0.00	0.00	10.47	16.30
Fe	29.81	27.22	22.53	30.49	15.95	1.53	1.20	19.20	2.99	4.71
Ca	64.88	62.00	65.99	55.97	75.92	98.47	98.80	80.80	86.55	78.99

TRIANGULAR DIAGRAM Na - Ca - K

Ca	1.58	0.75	0.73	1.07	1.16	21.66	27.50	1.09	1.07	0.60
Na	98.14	98.97	99.05	98.66	98.62	77.56	71.93	98.52	98.36	99.02
K	0.28	0.28	0.22	0.28	0.21	0.57	0.57	0.39	0.56	0.38

COMPOSITION IN OXIDES

	PLE 2	PLE 3	PLE 4	PLE 5	PLE 6	PLE 7	PLE 8	PLE 9	PLE 1	PLE 2
SiO2	61.03	63.05	69.76	66.70	61.69	60.70	68.46	67.93	69.34	66.47
Al2O3	24.75	23.30	19.33	19.36	24.02	24.33	19.55	19.74	19.15	19.28
TiO2	0.02	0.02	0.04	0.04	0.02	0.02	0.00	0.02	0.02	0.02
MgO	0.04	0.00	0.04	0.02	0.02	0.06	0.00	0.02	0.02	0.02
FeO	0.07	0.09	0.10	0.06	0.09	0.10	0.07	0.10	0.03	0.11
MnO	0.03	0.00	0.00	0.00	0.00	0.03	0.01	0.01	0.01	0.01
CaO	6.26	4.53	0.13	0.17	5.73	6.13	0.10	0.64	1.07	0.77
Na2O	8.04	9.11	11.89	11.81	8.30	8.21	11.84	11.45	11.34	11.50
K2O	0.10	0.10	0.06	0.06	0.10	0.06	0.07	0.11	0.13	0.17
	100.34	100.20	100.35	100.22	99.97	99.66	100.10	99.92	101.11	100.41

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLE 2	PLE 3	PLE 4	PLE 5	PLE 6	PLE 7	PLE 8	PLE 9	PLE 1	PLE 2
Si	10.816	11.138	11.984	11.985	10.953	10.838	11.959	11.989	12.004	11.971
Al	5.173	4.351	3.917	3.923	5.126	5.120	4.025	4.078	3.907	3.861
Ti	0.003	0.003	0.005	0.005	0.003	0.003	0.000	0.003	0.003	0.003
	15.993	15.992	15.960	15.970	15.982	15.961	15.985	15.970	15.913	15.835
Mg	0.011	0.000	0.010	0.005	0.006	0.016	0.000	0.005	0.005	0.005
Fe2	0.010	0.013	0.015	0.009	0.013	0.015	0.011	0.015	0.004	0.001
Mn	0.005	0.000	0.000	0.000	0.000	0.005	0.001	0.001	0.001	0.001
Ca	1.189	0.857	0.024	0.032	1.090	1.173	0.019	0.120	0.193	0.144
Na	2.765	3.120	4.018	3.995	2.857	2.842	4.010	3.891	3.806	3.857
K	0.023	0.023	0.013	0.013	0.023	0.014	0.016	0.025	0.029	0.029
	4.000	4.013	4.081	4.054	3.989	4.069	4.056	4.057	4.044	4.036

Al% IN Z	32.33	30.33	24.68	24.92	31.45	32.08	25.19	25.54	24.55	24.36
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.87	0.00	21.10	11.37	0.45	1.32	0.00	3.69	2.46	3.42
Fe	1.23	1.53	29.60	19.14	1.21	1.61	38.48	11.40	2.77	1.93
Ca	97.90	98.47	49.30	69.49	98.32	97.07	61.52	84.91	94.76	94.65

TRIANGULAR DIAGRAM Na - Ca - K

Ca	29.91	21.43	0.60	0.79	27.46	29.08	0.46	2.98	4.92	3.54
Na	69.52	78.00	99.07	98.88	71.97	70.47	99.15	96.41	94.37	95.75
K	0.57	0.56	0.33	0.33	0.57	0.45	0.39	0.61	0.71	0.71

COMPOSITION IN OXIDES

	PLF 3	PLF 4	PLH 1	PLH 2	PLH 3	PLH 4	PLH 5	PLH 6	PLH 7	PLH 8
SiO2	69.23	68.69	63.04	66.95	62.47	63.67	61.34	61.96	61.49	62.96
Al2O3	19.27	18.43	23.52	20.82	23.45	23.00	24.23	23.74	24.03	23.40
TiO2	0.02	0.02	0.02	0.00	0.00	0.02	0.00	0.04	0.04	0.02
MgO	0.02	0.02	0.02	0.04	0.02	0.02	0.00	0.02	0.02	0.02
FeO	0.01	0.03	0.12	0.16	0.06	0.10	0.38	0.23	0.00	0.19
MnO	0.00	0.00	0.01	0.00	0.01	0.03	0.36	0.01	0.29	0.00
CaO	0.98	0.73	4.89	1.92	4.83	4.26	5.68	5.34	5.80	4.57
Na2O	11.94	11.49	8.87	10.69	9.00	9.64	8.13	8.40	8.54	9.16
K2O	0.15	0.12	0.13	0.10	0.11	0.17	0.08	0.13	0.13	0.10
	100.62	99.53	100.62	100.58	99.95	100.91	100.20	99.87	100.33	100.42

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLF 3	PLF 4	PLH 1	PLH 2	PLH 3	PLH 4	PLH 5	PLH 6	PLH 7	PLH 8
Si	11.910	12.074	11.099	11.689	11.078	11.163	10.895	11.007	10.908	11.110
Al	3.965	3.813	4.861	4.284	4.901	4.761	5.072	4.971	5.025	4.956
Ti	0.003	0.003	0.003	0.000	0.000	0.003	0.000	0.005	0.005	0.003
	15.878	15.895	15.962	15.974	15.979	15.947	15.968	15.983	15.939	15.979
Mg	0.005	0.005	0.005	0.010	0.005	0.005	0.000	0.005	0.005	0.005
Fe2	0.001	0.004	0.014	0.023	0.009	0.015	0.056	0.034	0.000	0.028
Mn	0.000	0.000	0.001	0.000	0.002	0.004	0.054	0.002	0.044	0.000
Ca	0.183	0.137	0.922	0.340	0.918	0.802	1.081	1.016	1.103	0.864
Na	4.041	3.916	3.028	3.619	3.094	3.283	2.800	2.893	2.938	3.134
K	0.033	0.027	0.029	0.022	0.025	0.038	0.018	0.029	0.029	0.023
	4.265	4.090	4.004	4.015	4.053	4.147	4.010	3.980	4.119	4.054

Al% IN Z	24.97	24.02	30.54	26.82	30.67	29.86	31.77	31.10	31.53	30.46
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	2.74	3.56	0.55	2.78	0.57	0.63	0.00	0.50	0.46	0.59
Fe	0.77	3.00	2.02	6.24	1.11	2.32	9.23	3.37	3.78	3.12
Ca	96.49	93.44	97.42	90.98	98.32	97.05	90.72	96.13	95.76	96.29

TRIANGULAR DIAGRAM Na - Ca - K

Ca	4.30	3.37	23.18	8.55	22.73	19.45	27.72	25.80	27.09	21.49
Na	94.91	95.97	76.09	90.89	76.65	79.63	71.81	73.45	72.19	77.95
K	0.78	0.66	0.73	0.56	0.62	0.92	0.46	0.75	0.72	0.56

COMPOSITION IN OXIDES

	PLI 1	PLK 1	PLK 2	PLK 3	PLK 4	PLK 5	PLK 6	PLK 7	PLK 8	PLK 9
SiO2	61.28	62.25	62.33	61.65	63.78	60.41	62.12	61.58	61.98	62.54
Al2O3	24.82	22.44	23.72	23.33	21.81	24.34	24.49	25.07	24.81	24.09
TiO2	0.00	0.02	0.02	0.02	0.02	0.02	0.00	0.02	0.02	0.00
MgO	0.02	0.00	0.02	0.04	0.02	0.00	0.00	0.00	0.00	0.06
FeO	0.01	0.06	0.06	0.03	0.01	0.03	0.03	0.04	0.01	0.04
MnO	0.01	0.03	0.00	0.04	0.05	0.03	0.01	0.01	0.00	0.00
CaO	6.40	5.46	5.60	5.79	4.30	5.64	5.77	6.08	5.87	5.73
Na2O	7.60	8.77	8.55	8.33	9.02	8.49	8.77	8.60	8.57	8.49
K2O	0.08	0.12	0.09	0.08	0.10	0.08	0.11	0.08	0.00	0.11
	100.22	99.15	100.68	99.31	99.11	99.04	101.30	101.48	101.26	102.00

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLI 1	PLK 1	PLK 2	PLK 3	PLK 4	PLK 5	PLK 6	PLK 7	PLK 8	PLK 9
Si	10.846	11.147	10.999	11.020	11.361	10.845	10.903	10.800	10.870	10.884
Al	5.152	4.736	4.933	4.915	4.579	5.150	5.066	5.102	5.120	5.116
Ti	0.000	0.003	0.003	0.003	0.003	0.003	0.000	0.003	0.001	0.000
	16.000	15.885	15.935	15.937	15.943	15.997	15.969	15.905	16.000	16.000
Al	0.027	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.010
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000
Mg	0.005	0.000	0.005	0.011	0.005	0.000	0.000	0.000	0.000	0.016
Fe2	0.001	0.009	0.009	0.004	0.001	0.005	0.004	0.006	0.001	0.006
Mn	0.001	0.005	0.000	0.006	0.008	0.005	0.001	0.001	0.000	0.000
Ca	1.214	1.048	1.059	1.109	0.821	1.085	1.085	1.143	1.103	1.078
Na	2.609	3.045	3.028	2.887	3.115	2.955	2.984	2.924	2.914	2.865
K	0.018	0.027	0.019	0.018	0.023	0.018	0.025	0.018	0.000	0.022
	3.876	4.133	4.119	4.035	3.973	4.067	4.100	4.092	4.020	3.996

Al% IN Z	32.20	29.81	30.96	30.84	28.72	32.19	31.72	32.42	32.05	31.97
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.43	0.00	0.49	0.94	0.64	0.00	0.00	0.00	0.00	1.42
Fe	0.24	1.25	0.83	0.93	1.09	0.83	0.54	0.64	0.13	0.53
Ca	99.32	98.72	98.68	98.12	98.28	99.17	99.46	99.36	99.87	98.05

TRIANGULAR DIAGRAM Na - Ca - K

Ca	31.61	25.43	25.79	27.62	20.73	26.73	26.50	27.97	27.46	27.13
Na	67.92	73.91	73.77	71.92	78.69	72.82	72.90	71.59	72.54	72.26
K	0.47	0.67	0.44	0.45	0.57	0.45	0.60	0.44	0.00	0.56

COMPOSITION IN OXIDES

	PLK10	PLK11	PLK12	PLK13	PLK14	PLL 1	PLL 2	PLL 3	PLL 4	PLL 5
SiO2	61.66	61.42	61.62	60.83	63.09	58.57	62.77	65.58	60.72	61.93
Al2O3	24.34	24.36	24.24	24.82	24.01	25.55	22.88	22.23	23.49	24.13
TiO2	0.02	0.02	0.02	0.02	0.00	0.00	0.00	0.00	0.02	0.00
MgO	0.04	0.02	0.02	0.02	0.00	0.00	0.00	0.00	0.00	0.00
FeO	0.14	0.03	0.10	0.12	0.03	0.04	0.10	0.06	0.10	0.03
MnO	0.03	0.00	0.04	0.03	0.03	0.01	0.03	0.00	0.00	0.00
CaO	5.51	5.67	5.48	6.25	4.95	7.90	5.02	3.88	5.01	6.00
Na2O	8.46	8.67	8.69	8.46	9.32	7.06	8.02	9.27	8.89	8.16
K2O	0.10	0.10	0.07	0.08	0.08	0.11	0.12	0.11	0.10	0.11
	100.30	100.29	100.28	100.63	101.51	99.24	98.94	101.13	98.33	100.36

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLK10	PLK11	PLK12	PLK13	PLK14	PLL 1	PLL 2	PLL 3	PLL 4	PLL 5
Si	10.916	10.887	10.918	10.774	11.028	10.545	11.201	11.422	10.970	10.952
Al	5.079	5.089	5.062	5.181	4.946	5.422	4.799	4.563	5.001	5.029
Ti	0.003	0.003	0.003	0.003	0.000	0.000	0.000	0.000	0.003	0.000
	15.998	15.978	15.982	15.958	15.975	15.967	16.000	15.986	15.974	15.981
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.013	0.000	0.000	0.000
Mg	0.011	0.005	0.005	0.005	0.000	0.000	0.000	0.000	0.000	0.000
Fe2	0.021	0.004	0.015	0.018	0.004	0.006	0.015	0.009	0.015	0.004
Mn	0.004	0.000	0.006	0.005	0.004	0.002	0.005	0.000	0.000	0.000
Ca	1.045	1.077	1.040	1.186	0.927	1.524	0.960	0.724	0.970	1.137
Na	2.904	2.980	2.985	2.905	3.159	2.465	2.775	3.131	3.114	2.798
K	0.023	0.023	0.016	0.018	0.018	0.025	0.027	0.024	0.023	0.025
	4.007	4.089	4.067	4.137	4.112	4.021	3.794	3.888	4.122	3.964

Al% IN Z	31.75	31.85	31.67	32.47	30.96	33.96	29.99	28.55	31.31	31.47
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.98	0.49	0.50	0.44	0.00	0.00	0.00	0.00	0.00	0.00
Fe	2.33	0.41	1.95	1.84	0.94	0.49	1.99	1.19	1.53	0.39
Ca	96.69	99.10	97.55	97.73	99.06	99.51	98.01	98.81	98.47	99.61

TRIANGULAR DIAGRAM Na - Ca - K

Ca	26.32	26.40	25.74	28.86	22.59	37.97	25.51	18.67	23.61	28.71
Na	73.12	73.05	73.87	70.70	76.97	61.40	73.76	80.70	75.83	70.66
K	0.57	0.55	0.39	0.44	0.43	0.63	0.73	0.63	0.56	0.63

COMPOSITION IN OXIDES

	PLN 1	PLN 2	PLO 1	PLR 1	PLU 1	PLU 2	PLU 3	PLU 4	PLU 5	PLU 6
SiO2	59.25	59.64	61.84	63.21	65.63	63.69	68.19	70.20	64.82	64.80
Al2O3	25.41	25.40	23.82	22.65	22.69	23.23	19.11	19.39	23.11	21.30
TiO2	0.00	0.00	0.02	0.00	0.00	0.02	0.00	0.00	0.04	0.00
MgO	0.00	0.02	0.02	0.02	0.04	0.00	0.00	0.02	0.02	0.02
FeO	0.00	0.06	0.00	0.00	0.35	0.23	0.09	0.06	0.07	0.19
MnO	0.00	0.00	0.00	0.00	0.11	0.03	0.00	0.00	0.01	0.01
CaO	7.21	7.51	6.31	4.82	4.09	4.61	0.48	0.35	5.02	4.20
Na2O	7.01	6.98	8.63	9.05	8.66	8.56	10.96	11.95	8.81	9.34
K2O	0.11	0.11	0.08	0.07	0.01	0.07	0.06	0.00	0.08	0.08
	98.99	99.72	100.72	99.82	101.58	100.44	98.89	101.97	101.98	99.94

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLN 1	PLN 2	PLO 1	PLR 1	PLU 1	PLU 2	PLU 3	PLU 4	PLU 5	PLU 6
Si	10.651	10.653	10.931	11.208	11.377	11.200	12.028	12.031	11.235	11.453
Al	5.349	5.347	4.963	4.733	4.623	4.800	3.972	3.917	4.771	4.437
Ti	0.000	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.005	0.000
	16.000	16.000	15.897	15.941	16.000	16.000	16.000	15.947	15.961	15.889
Al	0.034	0.000	0.000	0.000	0.012	0.014	0.001	0.000	0.000	0.000
Ti	0.000	0.000	0.000	0.000	0.000	0.003	0.000	0.000	0.000	0.000
Mg	0.000	0.005	0.005	0.005	0.010	0.000	0.000	0.005	0.005	0.005
Fe2	0.000	0.009	0.000	0.000	0.051	0.034	0.013	0.009	0.010	0.028
Mn	0.000	0.000	0.000	0.000	0.016	0.004	0.000	0.000	0.001	0.001
Ca	1.389	1.437	1.195	0.916	0.760	0.869	0.091	0.064	0.932	0.795
Na	2.443	2.417	2.958	3.111	2.911	2.918	3.748	3.971	2.961	3.201
K	0.025	0.025	0.018	0.016	0.002	0.016	0.014	0.000	0.018	0.018
	3.892	3.894	4.176	4.048	3.762	3.858	3.867	4.049	3.927	4.049

Al% IN Z	33.43	33.42	31.22	29.69	28.90	30.00	24.83	24.56	29.58	27.92
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.00	0.37	0.44	0.57	1.23	0.00	0.00	6.55	0.54	0.63
Fe	0.00	0.62	0.00	0.00	7.99	4.22	12.77	11.03	1.22	3.56
Ca	100.00	99.02	99.56	99.43	90.77	95.78	87.23	82.42	98.23	95.80

TRIANGULAR DIAGRAM Na - Ca - K

Ca	36.00	37.05	28.65	22.65	20.68	22.84	2.35	1.59	23.84	19.81
Na	63.34	62.31	70.91	76.96	79.26	76.75	97.29	98.41	75.71	79.74
K	0.65	0.65	0.43	0.39	0.06	0.41	0.35	0.00	0.45	0.45

COMPOSITION IN OXIDES

	PLV 1	PLV 2	PLW 1	PLW 2	PLW 3	PLW 4	PLX 1	PLX 2	PLY 1	PLY 2
SiO2	59.69	61.95	69.44	68.84	68.22	68.14	57.38	51.31	69.27	69.28
Al2O3	24.28	23.47	19.46	19.60	19.91	19.83	27.34	30.66	19.56	19.81
TiO2	0.02	0.00	0.00	0.02	0.00	0.04	0.02	0.02	0.02	0.02
MgO	0.02	0.04	0.00	0.02	0.00	0.02	0.04	0.00	0.02	0.02
FeO	0.09	0.10	0.03	0.04	0.06	0.16	0.16	0.12	0.01	0.04
MnO	0.03	0.01	0.00	0.03	0.03	0.00	0.01	0.00	0.01	0.03
CaO	5.77	4.82	0.10	0.45	0.58	0.82	9.10	13.61	0.16	0.23
Na2O	7.80	8.36	12.14	11.58	11.30	11.40	6.15	3.74	10.89	10.71
K2O	0.07	0.07	0.09	0.07	0.07	0.09	0.10	0.07	0.11	0.18
	97.77	98.82	101.26	100.65	100.17	100.50	100.30	99.53	100.05	100.32

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLV 1	PLV 2	PLW 1	PLW 2	PLW 3	PLW 4	PLX 1	PLX 2	PLY 1	PLY 2
Si	10.837	11.084	11.996	11.959	11.907	11.880	10.252	9.371	12.067	12.020
Al	5.163	4.916	3.962	4.013	4.093	4.075	5.748	6.600	3.953	3.980
Ti	0.000	0.000	0.000	0.003	0.000	0.005	0.000	0.003	0.000	0.000
	16.000	16.000	15.958	15.975	16.000	15.959	16.000	15.974	16.000	16.000
Al	0.033	0.033	0.000	0.000	0.003	0.000	0.009	0.000	0.057	0.070
Ti	0.003	0.000	0.000	0.000	0.000	0.000	0.003	0.000	0.003	0.003
Mg	0.005	0.011	0.000	0.005	0.000	0.005	0.011	0.000	0.005	0.005
Fe2	0.014	0.015	0.004	0.006	0.009	0.023	0.024	0.018	0.001	0.006
Mn	0.005	0.002	0.000	0.004	0.004	0.000	0.002	0.000	0.001	0.004
Ca	1.122	0.924	0.019	0.084	0.108	0.153	1.742	2.663	0.030	0.043
Na	2.746	2.900	4.066	3.900	3.824	3.854	2.130	1.324	3.672	3.603
K	0.016	0.016	0.020	0.016	0.016	0.020	0.023	0.016	0.024	0.040
	3.943	3.900	4.109	4.015	3.965	4.055	3.943	4.022	3.794	3.774

Al% IN Z	32.27	30.73	24.83	25.12	25.58	25.53	35.92	41.32	24.70	24.88
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.47	1.12	0.00	5.22	0.00	2.86	0.60	0.00	13.67	8.90
Fe	1.59	1.73	18.97	10.31	10.84	12.84	1.43	0.68	7.72	17.57
Ca	97.93	97.15	81.03	84.47	89.16	84.30	97.97	99.32	78.61	73.54

TRIANGULAR DIAGRAM Na - Ca - K

Ca	28.90	24.06	0.45	2.09	2.75	3.80	44.72	66.52	0.80	1.16
Na	70.69	75.52	99.07	97.52	96.86	95.70	54.69	33.08	98.55	97.76
K	0.42	0.42	0.48	0.39	0.39	0.50	0.59	0.41	0.65	1.08

COMPOSITION IN OXIDES

	PLY 3	PLY 4	PLY 5	PLY 6	PLZ 1	PLZ 2	PLZ 3	PLZ 4	PLZ 5	PLZ 6
SiO2	69.12	68.66	68.84	69.04	69.98	69.22	68.73	68.82	69.05	68.52
Al2O3	19.47	19.43	19.54	19.57	19.12	19.24	19.41	19.26	19.39	19.21
TiO2	0.00	0.02	0.00	0.02	0.02	0.00	0.00	0.02	0.02	0.00
MgO	0.00	0.02	0.00	0.02	0.02	0.00	0.00	0.02	0.02	0.00
FeO	0.04	0.07	0.10	0.04	0.07	0.12	0.07	0.07	0.14	0.17
MnO	0.00	0.01	0.03	0.01	0.00	0.03	0.00	0.00	0.01	0.00
CaO	0.06	0.13	0.07	0.13	0.07	0.09	0.28	0.12	0.09	0.10
Na2O	10.93	10.69	10.81	10.80	11.77	11.77	11.46	11.84	11.69	11.73
K2O	0.20	0.15	0.15	0.16	0.20	0.20	0.21	0.26	0.20	0.20
	99.82	99.18	99.54	99.79	101.25	100.67	100.16	100.41	100.61	99.93

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PLY 3	PLY 4	PLY 5	PLY 6	PLZ 1	PLZ 2	PLZ 3	PLZ 4	PLZ 5	PLZ 6
Si	12.055	12.046	12.039	12.040	12.072	12.024	11.994	11.996	12.001	11.998
Al	3.945	3.954	3.961	3.960	3.888	3.939	3.992	3.957	3.972	3.964
Ti	0.000	0.000	0.000	0.000	0.003	0.000	0.000	0.003	0.003	0.000
	16.000	16.000	16.000	16.000	15.963	15.963	15.986	15.955	15.975	15.963
Al	0.057	0.064	0.066	0.063	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.000	0.003	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.000
Mg	0.000	0.005	0.000	0.005	0.005	0.000	0.000	0.005	0.005	0.000
Fe2	0.006	0.010	0.015	0.006	0.010	0.017	0.010	0.010	0.020	0.025
Mn	0.000	0.001	0.004	0.001	0.000	0.004	0.000	0.000	0.001	0.000
Ca	0.011	0.024	0.013	0.024	0.013	0.017	0.052	0.022	0.017	0.019
Na	3.696	3.636	3.665	3.652	3.937	3.964	3.877	4.002	3.939	3.982
K	0.044	0.034	0.033	0.036	0.044	0.044	0.047	0.058	0.044	0.045
	3.814	3.778	3.797	3.790	4.009	4.047	3.987	4.097	4.027	4.071

AL% IN Z	24.66	24.71	24.76	24.75	24.35	24.68	24.97	24.80	24.86	24.84
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.00	12.63	0.00	14.13	18.25	0.00	0.00	13.74	11.84	0.00
Fe	34.23	28.38	59.25	19.87	35.84	56.60	16.33	26.99	49.36	57.02
Ca	65.77	58.99	40.75	66.01	45.91	43.40	83.67	59.27	38.30	42.98

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.30	0.66	0.35	0.65	0.32	0.42	1.32	0.55	0.42	0.46
Na	98.52	98.43	98.75	98.39	98.57	98.48	97.51	98.03	98.47	98.43
K	1.19	0.91	0.90	0.96	1.10	1.10	1.18	1.42	1.11	1.10

COMPOSITION IN OXIDES

	PL& 1	PL& 2	PL& 3	PL& 4	PL& 5	PL& 6	PL& 7	KFZ 1	KFZ 2	KFZ 3
SiO2	65.93	68.52	68.31	66.50	65.86	65.73	66.24	65.05	64.66	64.34
Al2O3	21.05	19.43	19.88	21.13	20.93	21.10	20.99	18.25	18.31	18.33
TiO2	0.00	0.02	0.02	0.02	0.02	0.02	0.00	0.04	0.02	0.02
MgO	0.02	0.00	0.00	0.00	0.00	0.02	0.02	0.00	0.00	0.00
FeO	0.01	0.01	0.04	0.06	0.04	0.01	0.04	0.03	0.03	0.29
MnO	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.03	0.01	0.00
CaO	2.50	0.63	1.21	2.55	2.47	2.52	2.30	0.00	0.00	0.00
Na2O	10.15	11.21	10.78	10.20	10.14	10.20	9.99	1.01	1.07	1.22
K2O	0.21	0.13	0.15	0.29	0.27	0.27	0.21	15.71	15.36	15.48
	99.87	99.95	100.40	100.75	99.74	99.88	99.80	100.12	99.46	99.68

STRUCT. FORM. BASED ON 32.0 OXYGENS

	PL& 1	PL& 2	PL& 3	PL& 4	PL& 5	PL& 6	PL& 7	KFZ 1	KFZ 2	KFZ 3
Si	11.607	11.978	11.901	11.613	11.615	11.582	11.650	12.000	11.989	11.941
Al	4.368	4.003	4.082	4.349	4.351	4.382	4.350	3.968	4.001	4.010
Ti	0.000	0.003	0.003	0.003	0.003	0.003	0.000	0.006	0.003	0.003
	15.975	15.984	15.985	15.965	15.969	15.967	16.000	15.973	15.993	15.954
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Mg	0.005	0.000	0.000	0.000	0.000	0.005	0.005	0.000	0.000	0.000
Fe2	0.001	0.001	0.006	0.009	0.006	0.001	0.006	0.005	0.005	0.045
Mn	0.000	0.000	0.001	0.000	0.001	0.001	0.001	0.005	0.002	0.000
Ca	0.472	0.118	0.226	0.477	0.467	0.476	0.433	0.000	0.000	0.000
Na	3.465	3.800	3.641	3.454	3.467	3.485	3.407	0.361	0.385	0.439
K	0.047	0.029	0.033	0.065	0.061	0.061	0.047	3.697	3.633	3.665
	3.990	3.948	3.908	4.004	4.002	4.030	3.901	4.067	4.024	4.149

Al% IN Z	27.34	25.05	25.54	27.24	27.24	27.44	27.19	24.84	25.02	25.13
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	1.10	0.00	0.00	0.00	0.00	1.09	1.18	0.00	0.00	0.00
Fe	0.31	1.22	3.13	1.80	1.56	0.61	1.65	100.00	100.00	100.00
Ca	98.59	98.78	96.87	98.20	98.44	98.30	97.17	0.00	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	11.84	2.99	5.79	11.94	11.68	11.83	11.15	0.00	0.00	0.00
Na	86.98	96.28	93.35	86.44	86.80	86.66	87.64	8.90	9.57	10.70
K	1.18	0.73	0.85	1.62	1.52	1.51	1.21	91.10	90.43	89.30

COMPOSITION IN OXIDES

	KFZ 4	KF& 1	KF& 2	KF& 3	KF& 4	KF& 5	KF& 6	KF& 7
SiO2	64.45	65.51	65.06	65.52	65.55	66.82	65.29	65.36
Al2O3	18.16	18.02	18.25	18.14	18.14	18.72	18.03	18.11
TiO2	0.04	0.02	0.02	0.04	0.02	0.02	0.02	0.02
MgO	0.00	0.00	0.00	0.09	0.00	0.00	0.00	0.00
FeO	0.03	0.01	0.01	0.00	0.03	0.03	0.00	0.03
MnO	0.04	0.01	0.03	0.03	0.01	0.03	0.01	0.01
CaO	0.00	0.00	0.00	0.06	0.00	0.02	0.00	0.00
Na2O	1.07	1.28	0.98	0.95	1.21	1.19	1.33	1.20
K2O	15.48	15.37	15.51	15.78	15.32	15.85	15.19	15.09
	99.27	100.22	99.86	100.61	100.28	102.68	99.87	99.82

STRUCT. FORM. BASED ON 32.0 OXYGENS

	KFZ 4	KF& 1	KF& 2	KF& 3	KF& 4	KF& 5	KF& 6	KF& 7
Si	11.988	12.050	12.014	12.023	12.043	12.006	12.044	12.049
Al	3.981	3.907	3.972	3.923	3.928	3.964	3.920	3.935
Ti	0.006	0.003	0.003	0.006	0.003	0.003	0.003	0.003
	15.975	15.959	15.989	15.952	15.974	15.973	15.967	15.986
Mg	0.000	0.000	0.000	0.025	0.000	0.000	0.000	0.000
Fe2	0.005	0.002	0.002	0.000	0.005	0.005	0.000	0.005
Mn	0.006	0.002	0.005	0.005	0.002	0.005	0.002	0.002
Ca	0.000	0.000	0.000	0.012	0.000	0.004	0.000	0.000
Na	0.386	0.457	0.351	0.338	0.431	0.415	0.476	0.429
K	3.673	3.606	3.654	3.694	3.590	3.633	3.574	3.549
	4.070	4.066	4.011	4.073	4.028	4.060	4.052	3.984

Al% IN Z	24.92	24.48	24.84	24.59	24.59	24.82	24.55	24.61
CHARGE	64.00	64.00	64.00	64.00	64.00	64.00	64.00	64.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	0.00	0.00	0.00	59.93	0.00	0.00	0.00	0.00
Fe	100.00	100.00	100.00	11.35	100.00	70.21	100.00	100.00
Ca	0.00	0.00	0.00	28.72	0.00	29.79	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.00	0.00	0.29	0.00	0.10	0.00	0.00
Na	9.51	11.24	8.76	8.36	10.72	10.23	11.75	10.78
K	90.49	88.76	91.24	91.35	89.28	89.67	88.25	89.22

COMPOSITION IN OXIDES

	MAX 1	MAX 2	MAX 3	MAX 4	MAX 5	MAX 6	PAI 1	PAI 2	PAI 3	PAI 4
SiO2	32.94	32.9[illegible]	33.17	34.10	33.29	33.37	47.12	47.35	46.33	47.22
Al2O3	48.65	48.95	48.61	48.13	47.34	48.10	40.40	39.99	39.42	39.36
TiO2	0.11	0.09	0.09	0.13	0.13	0.13	0.13	0.09	0.17	0.19
MgO	0.11	0.09	0.11	0.07	0.13	0.11	0.09	0.11	0.04	0.08
FeO	0.46	0.45	0.39	0.56	0.76	0.59	0.52	0.52	0.61	0.61
MnO	0.04	0.00	0.01	0.03	0.01	0.01	0.00	0.03	0.01	0.00
CaO	10.31	10.90	10.51	10.08	10.26	10.36	0.80	0.45	0.49	0.47
Na2O	1.39	1.31	1.42	1.42	1.15	1.22	6.17	6.00	6.19	6.48
K2O	0.34	0.25	0.00	0.78	0.26	0.79	1.09	1.30	1.39	1.37
	94.37	95.02	94.31	95.30	93.35	94.69	96.32	95.84	94.55	95.72

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MAX 1	MAX 2	MAX 3	MAX 4	MAX 5	MAX 6	PAI 1	PAI 2	PAI 3	PAI 4
Si	4.385	4.364	4.408	4.499	4.477	4.438	5.959	6.011	5.977	6.019
Al	3.615	3.636	3.592	3.501	3.523	3.562	2.041	1.989	2.023	1.981
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	4.018	3.999	4.022	3.993	3.980	3.977	3.981	3.995	3.971	3.942
Ti	0.011	0.009	0.009	0.013	0.013	0.013	0.012	0.009	0.016	0.018
Mg	0.022	0.018	0.022	0.014	0.026	0.022	0.017	0.021	0.008	0.015
Fe2	0.053	0.050	0.043	0.062	0.088	0.066	0.055	0.055	0.066	0.065
Mn	0.005	0.000	0.001	0.003	0.001	0.001	0.000	0.003	0.001	0.000
Ca	1.471	1.545	1.496	1.425	1.478	1.476	0.108	0.061	0.068	0.064
Na	0.359	0.336	0.366	0.363	0.300	0.315	1.513	1.477	1.548	1.601
K	0.058	0.042	0.000	0.131	0.045	0.134	0.176	0.211	0.212	0.223
	5.996	5.999	5.959	5.994	5.931	6.004	5.862	5.832	5.890	5.919
AL% IN Z	45.19	45.45	44.90	43.77	44.04	44.53	25.51	24.86	25.29	24.76
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	1.41	1.10	1.39	0.92	1.64	1.39	9.41	14.82	5.40	10.52
Fe	3.74	3.09	2.65	4.33	5.58	4.27	30.49	41.60	47.01	45.03
Ca	94.85	95.81	95.76	94.75	92.79	94.34	60.10	43.58	47.59	44.45

TRIANGULAR DIAGRAM Na - Ca - K

Ca	77.93	80.33	80.35	74.24	81.10	76.69	6.03	3.50	3.70	3.40
Na	19.01	17.47	19.65	18.92	16.45	16.34	84.18	84.46	84.69	84.80
K	3.06	2.19	0.00	6.84	2.45	6.96	9.78	12.04	11.61	11.80

COMPOSITION IN OXIDES

	PAR 1	PAR 2	PAR 3	PAR 4	PAK 1	PAK 2	PAK 3	PAK 4	PAK 5	PAK 6
SiO2	45.06	47.82	47.57	47.34	46.86	46.61	47.76	47.14	47.27	47.76
Al2O3	38.90	38.45	38.43	38.76	39.04	38.53	37.70	38.87	39.66	38.99
TiO2	0.04	0.09	0.07	0.06	0.07	0.13	0.11	0.15	0.13	0.17
MgO	0.08	0.13	0.10	0.10	0.15	0.10	0.13	0.13	0.09	0.15
FeO	0.53	0.52	0.53	0.69	0.88	0.89	0.94	1.01	1.01	1.15
MnO	0.04	0.01	0.00	0.01	0.01	0.00	0.03	0.04	0.00	0.01
CaO	0.33	0.36	0.39	0.32	0.35	0.33	0.77	0.35	0.38	0.29
Na2O	7.09	6.34	6.77	6.90	6.88	7.15	7.17	5.88	6.38	6.20
K2O	0.78	1.20	0.98	0.77	0.80	1.06	1.13	1.36	1.19	1.22
	92.85	94.92	94.84	94.95	95.04	94.80	95.74	94.93	96.11	95.94

STRUCT. FORM. BASED ON 22.0 OXYGENS

	PAR 1	PAR 2	PAR 3	PAR 4	PAK 1	PAK 2	PAK 3	PAK 4	PAK 5	PAK 6
Si	5.926	6.128	6.106	6.071	6.016	6.018	6.115	6.056	6.005	6.074
Al	2.074	1.872	1.894	1.929	1.984	1.982	1.885	1.944	1.995	1.926
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.956	3.935	3.921	3.929	3.923	3.882	3.803	3.942	3.943	3.917
Ti	0.004	0.009	0.007	0.006	0.007	0.013	0.011	0.014	0.012	0.016
Mg	0.016	0.025	0.019	0.019	0.029	0.019	0.025	0.025	0.017	0.028
Fe2	0.058	0.056	0.057	0.074	0.094	0.096	0.101	0.109	0.107	0.122
Mn	0.004	0.001	0.000	0.001	0.001	0.000	0.003	0.004	0.000	0.001
Ca	0.047	0.049	0.054	0.044	0.048	0.046	0.106	0.048	0.052	0.040
Na	1.808	1.575	1.685	1.716	1.713	1.790	1.780	1.465	1.571	1.529
K	0.131	0.196	0.160	0.126	0.131	0.175	0.185	0.223	0.193	0.198
	6.024	5.846	5.902	5.915	5.946	6.020	6.013	5.830	5.896	5.852

Al% IN Z	25.92	23.40	23.67	24.11	24.80	24.77	23.57	24.30	24.94	24.08
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	12.55	18.94	14.76	13.83	16.65	11.95	10.59	13.39	9.68	14.86
Fe	50.23	43.35	43.88	54.34	55.43	59.69	44.34	60.70	60.94	64.49
Ca	37.22	37.71	41.37	31.82	27.92	28.36	45.07	25.91	29.38	20.65

TRIANGULAR DIAGRAM Na - Ca - K

Ca	2.34	2.71	2.82	2.33	2.54	2.27	5.10	2.78	2.85	2.24
Na	91.07	86.51	88.73	90.99	90.53	89.04	85.98	84.38	86.53	86.56
K	6.59	10.77	8.45	6.68	6.93	8.69	8.92	12.84	10.62	11.21

COMPOSITION IN OXIDES

	PAK 7	PAK 8	PAK 9	PAK10	PAK11	PAK12	PAK13	PAK14	PAM 1	PAM 2
SiO2	46.70	45.37	47.12	46.28	47.06	45.33	46.20	46.36	46.09	47.63
Al2O3	38.69	38.92	38.69	38.66	39.02	39.66	39.72	39.75	38.91	38.97
TiO2	0.17	0.11	0.15	0.13	0.13	0.09	0.15	0.07	0.17	0.19
MgO	0.11	0.15	0.17	0.19	0.25	0.13	0.08	0.21	0.11	0.13
FeO	0.94	0.92	1.02	1.31	0.91	0.92	0.88	0.98	0.85	0.89
MnO	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00
CaO	0.42	0.46	0.45	0.48	0.33	0.87	0.49	0.71	0.41	0.32
Na2O	5.63	6.45	5.60	5.62	6.65	6.47	6.43	5.48	6.84	6.42
K2O	1.52	1.18	1.70	1.86	1.30	1.21	1.36	1.18	1.54	1.75
	94.19	93.56	94.90	94.53	95.65	94.69	95.31	94.74	94.92	96.30

STRUCT. FORM. BASED ON 22.0 OXYGENS

	PAK 7	PAK 8	PAK 9	PAK10	PAK11	PAK12	PAK13	PAK14	PAM 1	PAM 2
Si	6.047	5.934	6.063	6.002	6.017	5.871	5.933	5.962	5.959	6.055
Al	1.953	2.066	1.937	1.998	1.983	2.129	2.067	2.038	2.041	1.945
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.952	3.934	3.931	3.912	3.897	3.925	3.944	3.988	3.889	3.894
Ti	0.017	0.011	0.015	0.013	0.012	0.009	0.014	0.007	0.017	0.018
Mg	0.021	0.029	0.033	0.037	0.048	0.025	0.015	0.040	0.021	0.025
Fe2	0.102	0.101	0.110	0.142	0.097	0.100	0.095	0.105	0.092	0.095
Mn	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000
Ca	0.058	0.064	0.062	0.067	0.045	0.121	0.067	0.098	0.057	0.044
Na	1.414	1.636	1.397	1.413	1.649	1.625	1.601	1.367	1.715	1.582
K	0.251	0.197	0.279	0.308	0.212	0.200	0.223	0.194	0.254	0.284
	5.816	5.972	5.826	5.891	5.961	6.005	5.959	5.798	6.044	5.941

AL% IN Z	24.41	25.82	24.21	24.97	24.79	26.61	25.84	25.47	25.51	24.31
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	11.64	15.05	15.95	14.96	25.06	10.18	8.64	16.53	12.48	15.13
Fe	56.41	51.78	53.70	57.87	51.17	40.86	53.32	43.29	54.09	58.11
Ca	31.95	33.17	30.35	27.17	23.77	48.96	38.04	40.18	33.43	26.77

TRIANGULAR DIAGRAM Na - Ca - K

Ca	3.38	3.40	3.57	3.73	2.37	6.21	3.56	5.90	2.80	2.28
Na	82.04	86.22	80.38	79.05	86.50	83.52	84.65	82.42	84.66	82.86
K	14.57	10.38	16.05	17.21	11.13	10.28	11.78	11.68	12.54	14.86

COMPOSITION IN OXIDES

	PAM 3	PAM 4	PAM 5	PAM 6	PAM 7	PAM 8	PAM 9	PAM10	PAM11	PAM12
SiO2	46.55	46.74	46.40	47.52	46.72	46.84	46.44	46.74	47.31	47.44
Al2O3	39.25	38.97	39.30	39.18	38.58	38.88	38.76	39.07	39.20	39.07
TiO2	0.19	0.19	0.17	0.20	0.17	0.19	0.19	0.15	0.20	0.20
MgO	0.11	0.11	0.08	0.09	0.08	0.08	0.10	0.08	0.08	0.11
FeO	0.92	0.94	0.75	0.88	0.81	0.85	0.84	0.74	0.75	0.76
MnO	0.00	0.00	0.01	0.03	0.00	0.00	0.00	0.01	0.03	0.01
CaO	0.48	0.35	0.48	0.28	0.30	0.33	0.30	0.39	0.33	0.26
Na2O	6.41	6.84	6.86	6.17	6.41	6.40	6.36	6.70	6.29	6.28
K2O	1.35	1.28	1.30	1.87	1.71	1.70	1.85	1.19	1.76	1.75
	95.26	95.42	95.35	96.22	94.78	95.27	94.84	95.07	95.95	95.88

STRUCT. FORM. BASED ON 22.0 OXYGENS

	PAM 3	PAM 4	PAM 5	PAM 6	PAM 7	PAM 8	PAM 9	PAM10	PAM11	PAM12
Si	5.978	5.998	5.960	6.044	6.035	6.020	6.003	6.006	6.031	6.049
Al	2.022	2.002	2.040	1.956	1.965	1.980	1.997	1.994	1.969	1.951
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.920	3.892	3.909	3.917	3.908	3.910	3.907	3.924	3.921	3.920
Ti	0.018	0.018	0.016	0.019	0.017	0.018	0.018	0.014	0.019	0.019
Mg	0.021	0.021	0.015	0.017	0.015	0.015	0.019	0.015	0.015	0.021
Fe2	0.099	0.101	0.081	0.094	0.087	0.091	0.091	0.080	0.080	0.081
Mn	0.000	0.000	0.001	0.003	0.000	0.000	0.000	0.001	0.003	0.001
Ca	0.066	0.048	0.066	0.038	0.042	0.045	0.042	0.054	0.045	0.036
Na	1.596	1.702	1.708	1.522	1.605	1.595	1.594	1.669	1.555	1.553
K	0.221	0.210	0.213	0.303	0.282	0.279	0.305	0.195	0.286	0.285
	5.941	5.992	6.010	5.913	5.956	5.954	5.976	5.953	5.925	5.915

Al% IN Z	25.27	25.02	25.50	24.45	24.57	24.75	24.97	24.92	24.61	24.39
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	11.33	12.37	9.39	11.22	10.66	10.07	12.71	10.24	10.60	15.09
Fe	53.15	59.33	50.09	63.68	60.59	60.06	59.89	53.87	57.09	59.27
Ca	35.53	28.30	40.52	25.09	28.75	29.87	27.40	35.89	31.42	25.64

TRIANGULAR DIAGRAM Na - Ca - K

Ca	3.51	2.46	3.32	2.05	2.15	2.37	2.14	2.80	2.39	1.90
Na	84.75	86.85	85.96	81.67	83.24	83.11	82.14	87.03	82.43	82.90
K	11.74	10.69	10.72	16.28	14.61	14.52	15.72	10.17	15.18	15.20

COMPOSITION IN OXIDES

	PAM13	MUI 1	MUI 2	MUJ 1	MUK 1	MUK 2	MUK 3	MUK 4	MUK 5	MUK 6
SiO2	46.76	46.14	48.24	48.59	47.06	48.32	47.20	47.38	45.73	47.02
Al2O3	39.48	36.35	33.18	31.61	33.81	31.52	33.01	31.74	33.05	33.97
TiO2	0.19	0.30	0.56	0.62	0.22	0.32	0.28	0.32	0.30	0.37
MgO	0.08	0.56	1.68	2.01	0.74	1.22	1.02	1.33	0.84	0.83
FeO	0.75	1.36	1.86	2.09	2.79	3.21	2.87	2.88	2.43	3.04
MnO	0.00	0.01	0.03	0.01	0.01	0.00	0.00	0.00	0.01	0.00
CaO	0.48	0.00	0.00	0.00	0.00	0.06	0.04	0.03	0.06	0.07
Na2O	6.92	1.80	1.46	1.10	2.12	1.78	1.94	1.83	1.76	1.71
K2O	1.07	8.42	8.95	9.06	8.28	8.50	8.66	8.78	8.31	8.51
	95.73	94.94	95.96	95.09	95.03	94.93	95.02	94.29	92.49	95.52

STRUCT. FORM. BASED ON 22.0 OXYGENS

	PAM13	MUI 1	MUI 2	MUJ 1	MUK 1	MUK 2	MUK 3	MUK 4	MUK 5	MUK 6
Si	5.972	6.112	6.348	6.455	6.275	6.460	6.311	6.389	6.250	6.247
Al	2.028	1.888	1.652	1.545	1.725	1.540	1.689	1.611	1.740	1.753
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.915	3.786	3.494	3.405	3.587	3.426	3.512	3.433	3.593	3.566
Ti	0.018	0.030	0.055	0.062	0.022	0.032	0.028	0.032	0.031	0.037
Mg	0.015	0.111	0.330	0.398	0.147	0.243	0.203	0.267	0.171	0.164
Fe2	0.080	0.151	0.205	0.232	0.311	0.359	0.321	0.325	0.278	0.338
Mn	0.000	0.001	0.003	0.001	0.001	0.000	0.000	0.000	0.001	0.000
Ca	0.066	0.000	0.000	0.000	0.000	0.009	0.006	0.004	0.009	0.010
Na	1.714	0.462	0.372	0.283	0.548	0.461	0.503	0.478	0.467	0.440
K	0.174	1.423	1.502	1.535	1.408	1.450	1.477	1.510	1.451	1.442
	5.982	5.964	5.961	5.917	6.025	5.980	6.050	6.051	6.002	5.998

Al% IN Z	25.35	23.60	20.65	19.31	21.57	19.25	21.12	20.14	21.75	21.91
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	9.46	42.14	61.30	63.04	32.02	39.81	38.36	44.82	37.30	32.10
Fe	49.75	57.86	38.70	36.96	67.98	58.78	60.56	54.45	60.79	65.96
Ca	40.79	0.00	0.00	0.00	0.00	1.41	1.08	0.73	1.91	1.95

TRIANGULAR DIAGRAM Na - Ca - K

Ca	3.36	0.00	0.00	0.00	0.00	0.45	0.29	0.22	0.46	0.53
Na	87.71	24.52	19.87	15.58	28.01	24.04	25.33	24.01	24.24	23.27
K	8.92	75.48	80.13	84.42	71.99	75.52	74.38	75.78	75.30	76.20

COMPOSITION IN OXIDES

	MUK 7	MUK 8	MUK 9	MUK10	MUK11	MUK12	MUK13	MUK14	MUK15	MUK16
SiO2	47.01	44.98	46.74	46.36	47.47	46.98	46.53	47.91	46.64	47.11
Al2O3	29.77	34.00	35.13	34.87	33.02	33.18	35.37	33.73	34.44	34.01
TiO2	0.47	0.22	0.20	0.26	0.47	0.32	0.20	0.34	0.22	0.26
MgO	1.64	0.70	0.64	0.81	1.26	1.57	0.69	1.28	0.77	0.94
FeO	3.27	2.82	2.38	2.49	2.62	2.85	2.56	3.21	2.63	2.85
MnO	0.03	0.00	0.01	0.01	0.01	0.05	0.01	0.01	0.01	0.00
CaO	0.04	0.15	0.06	0.04	0.06	0.09	0.03	0.09	0.04	0.12
Na2O	1.46	2.12	1.87	1.94	1.33	1.21	1.77	1.67	1.68	1.59
K2O	9.05	7.66	8.37	8.22	8.90	8.79	8.30	8.58	8.45	8.42
	92.74	92.65	95.40	95.00	95.14	94.84	95.46	96.82	95.08	95.30

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUK 7	MUK 8	MUK 9	MUK10	MUK11	MUK12	MUK13	MUK14	MUK15	MUK16
Si	6.472	6.150	6.192	6.173	6.324	6.285	6.163	6.283	6.215	6.260
Al	1.528	1.850	1.808	1.827	1.676	1.715	1.837	1.717	1.785	1.740
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.303	3.629	3.678	3.645	3.509	3.517	3.684	3.497	3.623	3.586
Ti	0.049	0.023	0.020	0.026	0.047	0.032	0.020	0.034	0.022	0.026
Mg	0.337	0.143	0.126	0.161	0.250	0.273	0.136	0.250	0.153	0.186
Fe2	0.377	0.322	0.264	0.277	0.292	0.319	0.284	0.352	0.293	0.317
Mn	0.003	0.000	0.001	0.001	0.001	0.006	0.001	0.001	0.001	0.000
Ca	0.006	0.022	0.009	0.006	0.009	0.013	0.004	0.013	0.006	0.017
Na	0.390	0.562	0.480	0.501	0.344	0.314	0.455	0.425	0.486	0.410
K	1.589	1.336	1.415	1.396	1.513	1.500	1.402	1.435	1.436	1.427
	6.053	6.037	5.992	6.013	5.964	5.974	5.985	6.006	6.020	5.969

Al% IN Z	19.10	23.12	22.59	22.84	20.95	21.43	22.97	21.46	22.32	21.75
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Na - Fe2+Fe3+Mn - Ca

Mg	46.58	29.29	31.62	36.14	45.34	44.74	32.04	40.62	33.77	35.81
Fe	52.60	66.20	66.25	62.58	53.11	53.15	66.96	57.33	64.97	60.91
Ca	0.82	4.51	2.13	1.28	1.55	2.11	1.00	2.05	1.26	3.29

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.30	1.14	0.45	0.30	0.46	0.71	0.23	0.68	0.30	0.92
Na	19.63	29.27	25.24	26.32	18.42	17.18	24.42	22.68	25.20	22.10
K	80.07	69.58	74.32	73.38	81.12	82.11	75.35	76.65	74.51	76.98

COMPOSITION IN OXIDES

	MUK17	MUL 1	MUL 2	MUM 1	MUM 2	MUM 3	MUM 4	MUM 5	MUM 6	MUM 7
SiO2	46.64	50.56	50.45	45.86	46.72	46.83	46.09	46.34	47.60	47.02
Al2O3	33.43	32.66	33.68	35.19	33.17	33.60	33.78	33.24	33.59	33.36
TiO2	0.22	0.45	0.39	0.35	0.54	0.45	0.52	0.50	0.48	0.45
MgO	0.89	1.03	0.64	0.51	1.09	0.79	0.88	0.94	0.94	0.88
FeO	2.71	1.01	0.71	2.03	2.25	2.00	2.32	2.32	2.25	2.23
MnO	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.00
CaO	0.06	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na2O	1.77	1.06	1.16	2.29	1.72	1.67	1.93	1.96	1.57	1.73
K2O	8.29	9.47	9.38	8.16	8.75	8.90	8.61	8.78	8.93	8.94
	94.01	96.24	96.42	94.39	94.24	94.25	94.14	94.09	95.37	94.61

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUK17	MUL 1	MUL 2	MUM 1	MUM 2	MUM 3	MUM 4	MUM 5	MUM 6	MUM 7
Si	6.281	6.582	6.541	6.139	6.281	6.286	6.209	6.253	6.316	6.298
Al	1.719	1.418	1.459	1.861	1.719	1.714	1.791	1.747	1.684	1.702
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.586	3.594	3.687	3.691	3.536	3.602	3.573	3.539	3.568	3.564
Ti	0.022	0.044	0.038	0.035	0.055	0.045	0.053	0.051	0.048	0.045
Mg	0.179	0.200	0.124	0.102	0.218	0.158	0.177	0.189	0.186	0.176
Fe2	0.305	0.110	0.077	0.227	0.253	0.225	0.261	0.262	0.250	0.250
Mn	0.000	0.000	0.001	0.000	0.000	0.001	0.001	0.001	0.001	0.000
Ca	0.009	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na	0.462	0.268	0.292	0.594	0.448	0.435	0.504	0.513	0.404	0.449
K	1.424	1.573	1.551	1.393	1.501	1.524	1.480	1.511	1.511	1.528
	5.987	5.788	5.770	6.043	6.011	5.990	6.048	6.066	5.968	6.012

AL% IN Z	21.49	17.72	18.24	23.26	21.49	21.42	22.39	21.84	21.06	21.28
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	36.27	64.51	61.30	30.93	46.34	41.19	40.23	41.83	42.57	41.29
Fe	61.97	35.49	38.70	69.07	53.66	58.81	59.77	58.17	57.43	58.71
Ca	1.76	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.46	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na	24.39	14.54	15.82	29.90	23.00	22.19	25.41	25.33	21.09	22.73
K	75.15	85.46	84.18	70.10	77.00	77.81	74.59	74.67	78.91	77.27

COMPOSITION IN OXIDES

	MUM 8	MUM 9	MUM10	MUM11	MUM12	MUM13	MUM14	MUM15	MUM16	MUM17
SiO2	47.13	47.16	47.44	45.96	46.30	46.14	45.86	45.40	46.40	47.09
Al2O3	33.25	33.96	32.82	34.61	34.49	34.95	34.84	33.77	33.45	32.20
TiO2	0.60	0.50	0.60	0.41	0.43	0.30	0.43	0.35	0.45	0.65
MgO	1.07	0.84	1.15	0.60	0.62	0.51	0.60	1.02	0.77	1.24
FeO	2.28	2.12	2.55	2.07	2.16	2.17	2.16	3.15	2.19	2.53
MnO	0.01	0.01	0.03	0.01	0.00	0.00	0.00	0.01	0.01	0.00
CaO	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00
Na2O	1.54	1.89	1.58	2.21	2.34	2.32	2.05	1.86	1.87	1.46
K2O	9.10	8.52	8.96	8.24	8.22	8.28	8.46	8.64	8.73	9.11
	94.98	95.01	95.13	94.11	94.56	94.68	94.40	94.21	93.87	94.28

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUM 8	MUM 9	MUM10	MUM11	MUM12	MUM13	MUM14	MUM15	MUM16	MUM17
Si	6.293	6.272	6.328	6.175	6.194	6.166	6.150	6.146	6.263	6.345
Al	1.707	1.728	1.672	1.825	1.806	1.834	1.850	1.854	1.737	1.655
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.526	3.595	3.488	3.655	3.631	3.670	3.657	3.534	3.585	3.459
Ti	0.060	0.050	0.060	0.041	0.043	0.030	0.043	0.036	0.046	0.066
Mg	0.213	0.167	0.229	0.120	0.124	0.102	0.120	0.206	0.155	0.249
Fe2	0.255	0.236	0.284	0.233	0.242	0.243	0.242	0.357	0.247	0.285
Mn	0.001	0.001	0.003	0.001	0.000	0.000	0.000	0.001	0.001	0.000
Ca	0.000	0.001	0.000	0.000	0.000	0.001	0.000	0.001	0.000	0.000
Na	0.399	0.487	0.409	0.576	0.607	0.601	0.533	0.488	0.489	0.381
K	1.550	1.445	1.525	1.412	1.403	1.411	1.447	1.492	1.503	1.566
	6.004	5.983	5.998	6.038	6.049	6.058	6.043	6.115	6.027	6.006

Al% IN Z	21.33	21.60	20.90	22.82	22.58	22.93	23.12	23.18	21.71	20.69
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	45.44	41.13	44.27	33.95	33.84	29.40	33.11	36.43	38.41	46.62
Fe	54.56	58.52	55.73	66.05	66.16	70.19	66.89	63.32	61.59	53.38
Ca	0.00	0.35	0.00	0.00	0.00	0.41	0.00	0.26	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.07	0.00	0.00	0.00	0.07	0.00	0.07	0.00	0.00
Na	20.46	25.20	21.14	28.96	30.20	29.85	26.92	24.64	24.56	19.59
K	79.54	74.73	78.86	71.04	69.80	70.08	73.08	75.29	75.44	80.41

COMPOSITION IN OXIDES

	MUM18	MUM19	MUM20	MUM21	MUM22	MUM23	MUM24	MUN 1	MUN 2	MUN 3
SiO2	47.41	45.33	45.95	46.22	46.32	46.07	46.42	47.05	49.74	48.22
Al2O3	33.33	33.67	31.42	33.93	33.20	34.71	33.86	34.81	35.12	34.17
TiO2	0.52	0.35	0.63	0.50	0.52	0.39	0.48	0.36	0.19	0.28
MgO	0.90	0.71	1.77	0.79	0.92	0.58	0.85	0.89	1.03	0.83
FeO	2.29	3.97	3.75	2.35	2.28	1.94	2.35	0.71	0.76	0.59
MnO	0.00	0.20	0.01	0.04	0.00	0.01	0.01	0.00	0.00	0.01
CaO	0.00	0.19	0.00	0.00	0.00	0.00	0.01	0.00	0.13	0.00
Na2O	1.79	1.92	1.32	2.10	1.86	1.96	2.07	1.26	1.49	1.31
K2O	8.90	7.75	9.16	8.48	8.74	8.60	8.64	9.18	8.76	9.22
	95.14	94.09	94.01	94.41	93.84	94.26	94.69	94.26	97.22	94.63

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUM18	MUM19	MUM20	MUM21	MUM22	MUM23	MUM24	MUN 1	MUN 2	MUN 3
Si	6.314	6.146	6.267	6.208	6.261	6.181	6.220	6.267	6.392	6.384
Al	1.686	1.854	1.733	1.792	1.739	1.819	1.780	1.733	1.608	1.616
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.545	3.526	3.318	3.579	3.549	3.670	3.567	3.731	3.711	3.717
Ti	0.052	0.036	0.065	0.051	0.053	0.039	0.048	0.036	0.018	0.028
Mg	0.179	0.143	0.360	0.158	0.185	0.116	0.170	0.177	0.197	0.164
Fe2	0.255	0.450	0.428	0.264	0.258	0.218	0.263	0.079	0.082	0.065
Mn	0.000	0.023	0.001	0.005	0.000	0.001	0.001	0.000	0.000	0.001
Ca	0.000	0.028	0.000	0.000	0.000	0.000	0.001	0.000	0.018	0.000
Na	0.462	0.505	0.349	0.547	0.487	0.510	0.538	0.325	0.371	0.336
K	1.512	1.340	1.594	1.453	1.507	1.472	1.477	1.560	1.436	1.557
	6.005	6.051	6.114	6.056	6.039	6.026	6.066	5.908	5.834	5.868

AL% IN Z	21.08	23.18	21.66	22.40	21.74	22.73	22.25	21.67	20.10	20.19
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	41.19	22.27	45.62	37.07	41.83	34.64	38.97	69.08	66.46	71.14
Fe	58.81	73.44	54.38	62.93	58.17	65.36	60.70	30.92	27.51	28.86
Ca	0.00	4.28	0.00	0.00	0.00	0.00	0.33	0.00	6.03	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	1.47	0.00	0.00	0.00	0.00	0.07	0.00	0.98	0.00
Na	23.41	26.95	17.97	27.35	24.44	25.73	26.68	17.26	20.34	17.76
K	76.59	71.57	82.03	72.65	75.56	74.27	73.25	82.74	78.68	82.24

COMPOSITION IN OXIDES

	MUN 4	MUO 1	MUO 2	MUO 3	MUO 4	MUO 5	MUP 1	MUP 2	MUP 3	MUP 4
SiO2	48.98	45.70	46.03	45.95	46.10	45.41	47.80	47.79	47.55	49.12
Al2O3	35.17	34.79	34.64	35.35	35.06	33.92	35.23	34.52	33.79	32.09
TiO2	0.19	0.45	0.47	0.39	0.49	0.45	0.22	0.32	0.26	0.17
MgO	0.55	0.86	0.79	0.75	0.86	0.98	0.76	0.91	1.21	1.67
FeO	0.63	0.74	0.79	0.75	0.75	0.77	1.30	1.67	1.37	1.52
MnO	0.01	0.01	0.01	0.03	0.00	0.00	0.00	0.01	0.00	0.03
CaO	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.04	0.01	0.01
Na2O	1.09	2.06	2.11	2.04	1.91	2.06	1.44	1.42	1.18	0.89
K2O	9.46	8.33	8.42	8.58	8.51	8.33	8.58	8.60	9.25	9.30
	96.08	92.95	93.26	93.84	93.68	91.92	95.34	95.28	94.62	94.80

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUN 4	MUO 1	MUO 2	MUO 3	MUO 4	MUO 5	MUP 1	MUP 2	MUP 3	MUP 4
Si	6.381	6.175	6.202	6.156	6.180	6.210	6.285	6.305	6.331	6.517
Al	1.619	1.825	1.798	1.844	1.820	1.790	1.715	1.695	1.669	1.483
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.781	3.716	3.704	3.739	3.720	3.677	3.745	3.673	3.634	3.535
Ti	0.019	0.046	0.048	0.039	0.049	0.046	0.022	0.032	0.026	0.017
Mg	0.107	0.173	0.159	0.150	0.172	0.200	0.149	0.179	0.240	0.330
Fe2	0.069	0.084	0.089	0.084	0.084	0.088	0.143	0.184	0.153	0.169
Mn	0.001	0.001	0.001	0.003	0.000	0.000	0.000	0.001	0.000	0.003
Ca	0.000	0.001	0.000	0.000	0.000	0.000	0.001	0.006	0.001	0.001
Na	0.275	0.540	0.551	0.530	0.496	0.546	0.367	0.363	0.305	0.229
K	1.572	1.436	1.447	1.466	1.455	1.453	1.439	1.447	1.571	1.574
	5.824	5.997	5.999	6.011	5.977	6.010	5.866	5.885	5.929	5.859

Al% IN Z	20.24	22.81	22.47	23.04	22.75	22.38	21.43	21.19	20.86	18.54
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	60.50	66.77	63.76	63.14	67.15	69.40	50.78	48.37	60.93	65.57
Fe	39.50	32.68	36.24	36.86	32.85	30.60	48.74	50.10	38.71	34.15
Ca	0.00	0.56	0.00	0.00	0.00	0.00	0.48	1.53	0.36	0.28

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.07	0.00	0.00	0.00	0.00	0.08	0.31	0.08	0.08
Na	14.90	27.30	27.58	26.55	25.44	27.32	20.31	20.00	16.23	12.69
K	85.10	72.63	72.42	73.45	74.56	72.68	79.61	79.69	83.70	87.23

COMPOSITION IN OXIDES

	MUP 5	MUP 6	MUP 7	MUP 8	MUP 9	MUP10	MUP11	MUP12	MUP13	MUP14
SiO2	47.82	47.25	48.82	46.53	46.76	48.98	47.47	48.75	49.77	48.13
Al2O3	33.05	32.74	32.29	35.49	33.97	32.09	33.36	31.82	30.54	33.01
TiO2	0.26	0.28	0.24	0.21	0.22	0.26	0.34	0.21	0.17	0.21
MgO	1.19	1.29	1.19	0.54	0.60	1.54	1.10	1.52	1.75	1.19
FeO	2.12	1.83	1.66	1.28	1.73	1.47	2.06	1.73	1.59	1.25
MnO	0.04	0.00	0.03	0.03	0.00	0.00	0.01	0.00	0.00	0.01
CaO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.06
Na2O	1.04	1.12	0.91	1.54	1.33	0.92	1.17	0.98	1.35	1.74
K2O	9.34	9.23	9.31	8.78	9.16	9.34	9.34	9.30	8.21	7.85
	94.86	93.75	94.45	94.40	93.97	94.60	94.85	94.31	93.41	93.45

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUP 5	MUP 6	MUP 7	MUP 8	MUP 9	MUP10	MUP11	MUP12	MUP13	MUP14
Si	6.376	6.368	6.506	6.200	6.285	6.513	6.334	6.513	6.664	6.430
Al	1.624	1.632	1.494	1.800	1.715	1.487	1.666	1.487	1.336	1.564
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.569	3.568	3.577	3.773	3.666	3.542	3.579	3.524	3.484	3.639
Ti	0.026	0.028	0.024	0.021	0.022	0.026	0.034	0.021	0.017	0.021
Mg	0.236	0.259	0.236	0.107	0.160	0.305	0.219	0.303	0.349	0.237
Fe2	0.236	0.206	0.185	0.143	0.194	0.163	0.230	0.193	0.178	0.140
Mn	0.005	0.000	0.003	0.003	0.000	0.000	0.001	0.000	0.000	0.001
Ca	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.004	0.009
Na	0.269	0.293	0.235	0.396	0.347	0.237	0.303	0.254	0.350	0.451
K	1.589	1.587	1.583	1.492	1.571	1.584	1.590	1.585	1.402	1.339
	5.930	5.943	5.843	5.938	5.961	5.858	5.956	5.880	5.785	5.837

Al% IN Z	20.30	20.40	18.68	22.50	21.44	18.59	20.83	18.58	16.70	19.55
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	49.54	55.51	55.65	42.35	45.18	65.12	48.64	61.03	65.70	61.34
Fe	50.46	44.18	44.35	57.65	54.82	34.88	51.36	38.97	33.49	36.44
Ca	0.00	0.31	0.00	0.00	0.00	0.00	0.00	0.00	0.81	2.22

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.00	0.24	0.48
Na	14.47	15.56	12.93	21.05	18.08	13.02	15.99	13.81	19.95	25.08
K	85.53	84.36	87.07	78.95	81.92	86.98	84.01	86.19	79.81	74.44

COMPOSITION IN OXIDES

	MUR 1	MUR 2	MUR 3	MUR 4	MUS 1	MUS 2	MUS 3	MUS 4	MUT 1	MUT 2
SiO2	47.56	44.88	45.99	47.40	46.27	44.63	46.68	45.31	46.84	46.84
Al2O3	31.83	32.71	34.70	32.00	32.04	32.71	33.14	33.77	33.74	35.25
TiO2	0.24	0.30	0.32	0.32	0.26	0.22	0.28	0.26	0.22	0.21
MgO	1.29	0.95	0.57	1.10	1.10	0.89	0.95	0.70	0.81	0.69
FeO	3.33	2.91	2.16	2.64	3.28	2.71	3.34	2.94	2.55	2.46
MnO	0.00	0.03	0.00	0.01	0.01	0.00	0.04	0.00	0.01	0.00
CaO	0.03	0.01	0.01	0.04	0.00	0.01	0.01	0.01	0.00	0.00
Na2O	1.26	1.67	1.83	1.46	1.13	1.28	1.29	1.44	1.70	1.47
K2O	9.01	8.48	7.90	8.34	9.15	8.75	8.83	8.73	9.04	9.04
	94.55	91.94	93.48	93.31	93.24	91.20	94.56	93.16	94.91	95.96

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUR 1	MUR 2	MUR 3	MUR 4	MUS 1	MUS 2	MUS 3	MUS 4	MUT 1	MUT 2
Si	6.401	6.212	6.196	6.418	6.327	6.220	6.281	6.184	6.267	6.186
Al	1.599	1.788	1.804	1.582	1.673	1.780	1.719	1.816	1.733	1.814
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.450	3.547	3.707	3.524	3.490	3.594	3.536	3.616	3.588	3.672
Ti	0.024	0.031	0.032	0.033	0.027	0.023	0.028	0.027	0.022	0.021
Mg	0.259	0.196	0.114	0.222	0.224	0.185	0.191	0.142	0.162	0.136
Fe2	0.375	0.337	0.243	0.299	0.375	0.316	0.376	0.336	0.285	0.272
Mn	0.000	0.004	0.000	0.001	0.001	0.000	0.005	0.000	0.001	0.000
Ca	0.004	0.001	0.001	0.006	0.000	0.001	0.001	0.001	0.000	0.000
Na	0.329	0.448	0.478	0.383	0.300	0.346	0.337	0.381	0.441	0.376
K	1.547	1.497	1.358	1.440	1.596	1.556	1.516	1.520	1.543	1.523
	5.988	6.062	5.934	5.908	6.013	6.021	5.989	6.024	6.042	6.000

Al% IN Z	19.99	22.35	22.55	19.78	20.92	22.24	21.49	22.70	21.66	22.68
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	40.57	36.44	31.56	42.05	37.34	36.81	33.29	29.70	36.06	33.33
Fe	58.76	63.28	67.74	56.85	62.66	62.89	66.46	69.99	63.94	66.67
Ca	0.68	0.28	0.40	1.10	0.00	0.30	0.25	0.31	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.23	0.08	0.08	0.32	0.00	0.08	0.08	0.08	0.00	0.00
Na	17.49	23.02	26.02	20.95	15.80	18.18	18.16	20.03	22.23	19.82
K	82.28	76.90	73.90	78.73	84.20	81.75	81.77	79.89	77.77	80.18

COMPOSITION IN OXIDES

	MUU 1	MUU 2	MUV 1	MUV 2	MUW 1	MUW 2	MUW 3	MUW 4	MUW 5	MUW 6
SiO2	50.72	48.17	46.28	47.60	46.82	46.15	46.33	45.60	45.64	46.71
Al2O3	28.26	28.86	34.22	31.76	31.88	33.09	31.92	34.58	34.51	30.88
TiO2	0.21	0.13	0.22	0.19	0.22	0.20	0.15	0.19	0.19	0.19
MgO	2.20	2.05	0.79	1.33	0.74	0.62	0.65	0.34	0.39	0.63
FeO	3.86	4.23	1.43	1.93	3.41	2.98	4.25	2.32	2.61	4.78
MnO	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00
CaO	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01
Na2O	0.86	0.84	1.35	0.95	0.74	0.89	0.86	1.31	1.45	0.85
K2O	9.44	9.52	8.50	8.58	9.69	9.96	9.54	9.65	9.00	9.51
	95.57	93.81	92.80	92.36	93.51	93.89	93.70	93.99	93.79	93.56

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUU 1	MUU 2	MUV 1	MUV 2	MUW 1	MUW 2	MUW 3	MUW 4	MUW 5	MUW 6
Si	6.759	6.580	6.266	6.478	6.390	6.278	6.340	6.174	6.130	6.417
Al	1.241	1.420	1.734	1.522	1.610	1.722	1.660	1.826	1.870	1.583
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.197	3.227	3.726	3.573	3.518	3.583	3.489	3.692	3.687	3.417
Ti	0.021	0.013	0.022	0.019	0.023	0.020	0.015	0.019	0.019	0.020
Mg	0.437	0.417	0.159	0.270	0.151	0.126	0.133	0.069	0.079	0.129
Fe2	0.430	0.483	0.162	0.220	0.389	0.339	0.486	0.263	0.296	0.549
Mn	0.001	0.001	0.001	0.001	0.001	0.000	0.000	0.000	0.000	0.000
Ca	0.001	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.001
Na	0.222	0.222	0.354	0.251	0.196	0.235	0.228	0.344	0.381	0.226
K	1.605	1.659	1.468	1.490	1.687	1.728	1.665	1.667	1.554	1.667
	5.915	6.024	5.893	5.825	5.965	6.031	6.017	6.053	6.015	6.010

Al% IN Z	15.52	17.74	21.68	19.02	20.12	21.53	20.75	22.83	22.76	19.78
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	50.24	46.29	49.44	54.83	27.83	27.05	21.42	20.71	21.03	18.98
Fe	49.59	53.71	50.56	44.88	72.17	72.95	78.58	79.29	78.97	80.80
Ca	0.16	0.00	0.00	0.30	0.00	0.00	0.00	0.00	0.00	0.22

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.08	0.00	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.08
Na	12.15	11.83	19.45	14.39	10.40	11.96	12.05	17.10	19.67	11.95
K	87.77	88.17	80.55	85.52	89.60	88.04	87.95	82.90	80.33	87.97

COMPOSITION IN OXIDES

	MUX 1	MUX 2	MUX 3	MUS 1	MUS 2	MUS 3	MUS 4	MUS 5	MUS 6	MUS 7
SiO2	46.21	46.56	46.71	47.75	46.16	46.99	47.23	46.43	47.71	47.42
Al2O3	35.92	35.28	35.69	30.95	33.41	33.66	33.45	33.55	31.79	32.64
TiO2	0.21	0.17	0.19	0.22	0.24	0.20	0.22	0.22	0.19	0.24
MgO	0.41	0.56	0.50	1.49	0.83	0.90	0.86	0.83	1.48	1.16
FeO	1.41	1.46	1.46	3.78	2.68	2.38	2.52	2.71	2.78	2.69
MnO	0.03	0.01	0.01	0.03	0.00	0.00	0.01	0.01	0.01	0.00
CaO	0.00	0.00	0.00	0.00	0.00	0.10	0.00	0.03	0.00	0.01
Na2O	0.83	1.27	0.93	0.99	1.23	1.72	1.22	1.23	1.07	1.07
K2O	9.92	9.33	9.76	9.69	9.48	8.51	9.51	9.38	9.51	9.53
	94.94	94.64	95.25	94.90	94.03	94.46	95.02	94.39	94.54	94.76

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUX 1	MUX 2	MUX 3	MUS 1	MUS 2	MUS 3	MUS 4	MUS 5	MUS 6	MUS 7
Si	6.154	6.208	6.194	6.441	6.251	6.291	6.314	6.258	6.418	6.361
Al	1.846	1.792	1.806	1.559	1.749	1.709	1.686	1.742	1.582	1.639
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.793	3.753	3.772	3.362	3.583	3.603	3.584	3.588	3.458	3.521
Ti	0.021	0.017	0.019	0.022	0.024	0.020	0.022	0.022	0.019	0.024
Mg	0.081	0.111	0.099	0.300	0.168	0.180	0.171	0.167	0.297	0.232
Fe2	0.157	0.163	0.162	0.426	0.304	0.266	0.282	0.305	0.313	0.302
Mn	0.003	0.001	0.001	0.003	0.000	0.000	0.001	0.001	0.001	0.000
Ca	0.000	0.000	0.000	0.000	0.000	0.014	0.000	0.004	0.000	0.001
Na	0.214	0.328	0.239	0.259	0.323	0.446	0.316	0.321	0.279	0.278
K	1.685	1.587	1.651	1.667	1.638	1.453	1.622	1.613	1.632	1.631
	5.955	5.960	5.943	6.040	6.039	5.983	5.998	6.022	5.998	5.989

Al% IN Z	23.07	22.40	22.57	19.49	21.87	21.36	21.08	21.77	19.78	20.49
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	33.66	40.44	37.74	41.07	35.57	39.01	37.73	34.91	48.60	43.34
Fe	66.34	59.56	62.26	58.93	64.43	57.88	62.27	64.19	51.40	56.39
Ca	0.00	0.00	0.00	0.00	0.00	3.12	0.00	0.91	0.00	0.27

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.00	0.00	0.00	0.00	0.75	0.00	0.22	0.00	0.08
Na	11.28	17.14	12.65	13.44	16.47	23.32	16.32	16.58	14.60	14.57
K	88.72	82.86	87.35	86.56	83.53	75.93	83.68	83.20	85.40	85.36

COMPOSITION IN OXIDES

	MUS 8	MUS 9	MUS10	MUS11	MUS12	MUS13	MUS14	MUS15	MUS16	MUS17
SiO2	47.32	48.92	47.61	47.96	49.04	48.58	47.19	47.43	47.84	46.97
Al2O3	35.04	33.18	34.81	35.13	33.62	32.67	35.44	34.96	33.88	35.45
TiO2	0.30	0.19	0.26	0.26	0.13	0.20	0.20	0.26	0.17	0.24
MgO	1.50	1.92	1.39	1.28	1.98	1.88	1.29	1.25	1.71	1.25
FeO	3.88	3.42	3.65	2.82	3.44	4.01	4.02	3.11	4.24	3.41
MnO	0.01	0.01	0.00	0.04	0.04	0.00	0.05	0.01	0.01	0.01
CaO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00
Na2O	1.02	0.74	1.06	1.01	0.78	0.71	0.96	1.06	0.76	1.08
K2O	9.91	9.70	9.79	10.14	9.96	10.53	9.97	9.79	10.26	9.66
	98.98	98.09	98.57	98.64	98.99	98.58	99.12	97.88	98.87	98.07

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MUS 8	MUS 9	MUS10	MUS11	MUS12	MUS13	MUS14	MUS15	MUS16	MUS17
Si	6.130	6.352	6.178	6.201	6.321	6.332	6.109	6.181	6.220	6.118
Al	1.870	1.648	1.822	1.799	1.679	1.668	1.891	1.819	1.780	1.882
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.479	3.430	3.501	3.554	3.428	3.351	3.516	3.551	3.411	3.560
Ti	0.029	0.019	0.025	0.025	0.013	0.020	0.019	0.025	0.017	0.024
Mg	0.290	0.372	0.269	0.247	0.380	0.365	0.249	0.243	0.331	0.243
Fe2	0.420	0.371	0.396	0.305	0.371	0.437	0.435	0.339	0.461	0.371
Mn	0.001	0.001	0.000	0.004	0.004	0.000	0.005	0.001	0.001	0.001
Ca	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Na	0.256	0.186	0.267	0.253	0.195	0.179	0.241	0.268	0.192	0.273
K	1.638	1.607	1.620	1.672	1.638	1.751	1.646	1.628	1.702	1.605
	6.113	5.987	6.079	6.060	6.029	6.104	6.112	6.056	6.114	6.076

AL% IN Z	23.38	20.60	22.78	22.49	20.99	20.85	23.64	22.73	22.25	23.53
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	40.73	49.85	40.43	44.37	50.35	45.52	36.09	41.56	41.76	39.45
Fe	59.27	49.97	59.57	55.63	49.65	54.48	63.91	58.20	58.24	60.55
Ca	0.00	0.19	0.00	0.00	0.00	0.00	0.00	0.24	0.00	0.00

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.07	0.00	0.00
Na	13.53	10.38	14.13	13.15	10.64	9.30	12.77	14.12	10.12	14.52
K	86.47	89.54	85.87	86.85	89.36	90.70	87.23	85.81	89.88	85.48

COMPOSITION IN OXIDES

	MU$18	MU$19	MU$20	MU$21	MU$22	MU$23	MU$24	MU$25	MU& 1	MU& 2
SiO2	47.29	47.60	47.61	47.62	47.72	47.88	46.59	47.78	48.22	48.66
Al2O3	34.85	34.90	36.32	37.09	36.24	32.94	37.71	36.03	28.94	27.63
TiO2	0.22	0.24	0.26	0.28	0.24	0.19	0.24	0.24	0.60	0.56
MgO	1.28	1.59	1.25	0.91	1.36	1.79	0.76	1.19	1.33	1.50
FeO	4.27	3.35	3.49	3.16	3.39	4.11	3.18	3.22	3.80	4.24
MnO	0.03	0.00	0.01	0.04	0.01	0.00	0.00	0.00	0.04	0.04
CaO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.32	0.26
Na2O	0.97	0.94	1.10	1.58	1.04	0.72	1.33	0.95	0.30	0.17
K2O	10.00	9.62	9.84	9.26	9.69	10.16	9.64	9.31	10.27	10.20
	98.91	98.25	99.88	99.94	99.69	97.79	99.45	98.73	93.82	93.26

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MU$18	MU$19	MU$20	MU$21	MU$22	MU$23	MU$24	MU$25	MU& 1	MU& 2
Si	6.144	6.179	6.091	6.066	6.106	6.286	5.979	6.149	6.594	6.700
Al	1.856	1.821	1.909	1.934	1.894	1.714	2.021	1.851	1.406	1.300
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	3.480	3.518	3.568	3.635	3.572	3.384	3.682	3.614	3.258	3.184
Ti	0.021	0.023	0.025	0.027	0.023	0.019	0.023	0.023	0.062	0.058
Mg	0.248	0.308	0.238	0.173	0.259	0.350	0.145	0.228	0.271	0.308
Fe2	0.464	0.364	0.373	0.337	0.363	0.451	0.341	0.347	0.435	0.488
Mn	0.003	0.000	0.001	0.004	0.001	0.000	0.000	0.000	0.005	0.005
Ca	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.047	0.038
Na	0.244	0.237	0.273	0.390	0.258	0.183	0.331	0.237	0.080	0.045
K	1.657	1.593	1.606	1.505	1.582	1.702	1.578	1.528	1.791	1.792
	6.118	6.043	6.085	6.070	6.058	6.089	6.101	5.978	5.948	5.918

AlX IN Z	23.21	22.77	23.86	24.17	23.67	21.42	25.27	23.14	17.58	16.25
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	34.66	45.73	38.89	33.63	41.62	43.70	29.87	39.62	35.80	36.69
Fe	65.34	54.06	61.11	66.37	58.38	56.30	70.13	60.14	58.01	58.74
Ca	0.00	0.21	0.00	0.00	0.00	0.00	0.00	0.24	6.19	4.57

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.08	2.44	2.05
Na	12.85	12.92	14.52	20.59	14.03	9.72	17.33	13.42	4.15	2.42
K	87.15	87.00	85.48	79.41	85.97	90.28	82.67	86.51	93.41	95.53

COMPOSITION IN OXIDES

	MU& 3	MU& 4	MU& 5
SiO2	47.84	47.88	48.19
Al2O3	27.90	28.25	28.31
TiO2	0.54	0.58	0.54
MgO	1.62	1.41	1.42
FeO	4.31	4.01	4.18
MnO	0.07	0.12	0.04
CaO	0.01	0.07	0.04
Na2O	0.24	0.18	0.26
K2O	10.65	10.57	10.74
	93.18	93.07	93.72

STRUCT. FORM. BASED ON 22.0 OXYGENS

	MU& 3	MU& 4	MU& 5
Si	6.622	6.620	6.625
Al	1.378	1.380	1.375
	8.000	8.000	8.000
Al	3.174	3.223	3.212
Ti	0.056	0.060	0.056
Mg	0.334	0.291	0.291
Fe2	0.499	0.464	0.481
Mn	0.008	0.014	0.005
Ca	0.001	0.010	0.006
Na	0.064	0.048	0.069
K	1.881	1.864	1.883
	6.018	5.974	6.002

Al% IN Z	17.22	17.25	17.19
CHARGE	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	39.66	37.32	37.20
Fe	60.17	61.35	62.04
Ca	0.18	1.33	0.75

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.08	0.54	0.30
Na	3.31	2.51	3.54
K	96.61	96.95	96.16

COMPOSITION IN OXIDES

	BII 1	BII 2	BIJ 1	BIJ 2	BIJ 3	BIJ 4	BIJ 5	BIJ 6	BIJ 7	BIJ 8
SiO2	36.30	37.69	37.75	37.73	37.48	37.35	37.46	37.72	37.5[illegible]	36.4[illegible]
AL2O3	19.67	19.58	19.80	20.42	19.23	18.83	19.13	18.09	18.61	19.56
TiO2	1.40	1.62	1.55	1.50	1.59	1.70	1.42	1.53	1.57	1.[illegible]4
MgO	10.17	11.90	12.16	11.98	11.81	11.15	11.76	11.64	11.59	11.[illegible]8
FeO	17.49	16.87	16.5[illegible]	15.84	15.91	16.27	16.93	17.45	16.64	16.[illegible]7
MnO	0.09	0.11	0.08	0.04	0.08	0.12	0.05	0.05	0.08	0.08
CaO	0.01	0.01	0.01	0.01	0.00	0.01	0.00	0.00	0.01	0.04
Na2O	1.05	0.36	0.18	0.20	0.43	0.57	0.50	0.28	0.43	0.34
K2O	[illegible].70	[illegible].[illegible]3	[illegible].[illegible]5	[illegible].80	[illegible].64	[illegible].51	[illegible].56	[illegible].5[illegible]	[illegible].55	[illegible].4[illegible]
	94.8[illegible]	96.97	96.78	96.52	95.17	94.51	95.87	95.25	95.1[illegible]	94.[illegible]2

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BII 1	BII 2	BIJ 1	BIJ 2	BIJ 3	BIJ 4	BIJ 5	BIJ 6	BIJ 7	BIJ [illegible]
Si	5.473	5.515	5.516	5.505	5.562	5.595	5.544	5.631	5.584	5.445
Al	2.527	2.485	2.484	2.495	2.438	2.405	2.456	2.369	2.416	2.555
	8.000	[illegible].000	[illegible].000	[illegible].[illegible]00	8.000	8.000	[illegible].000	8.000	8.000	[illegible].[illegible]00
Al	0.96[illegible]	0.[illegible]91	0.926	1.017	0.925	0.919	0.881	0.[illegible]11	[illegible].[illegible]84	0.[illegible]15
Ti	0.159	0.178	0.170	0.165	0.177	0.192	0.165	0.172	0.176	0.[illegible]06
Mg	2.285	2.595	2.648	2.606	2.612	2.490	2.594	2.590	2.572	2.[illegible]67
Fe2	2.205	2.064	2.002	1.933	1.974	2.038	2.096	2.178	2.072	1.[illegible]97
Mn	0.011	0.014	0.010	0.005	0.010	0.[illegible]15	[illegible].[illegible]16	0.[illegible]06	0.010	0.[illegible]10
Ca	0.002	0.002	0.002	0.002	0.000	0.002	0.000	0.000	0.002	0.[illegible]06
Na	0.307	0.102	0.051	0.[illegible]57	0.124	0.166	0.143	0.081	0.124	0.[illegible]9[illegible]
K	1.673	1.648	1.650	1.633	1.636	1.526	1.616	1.619	1.624	1.614
	7.611	7.[illegible]94	7.459	7.421	7.459	7.447	7.502	7.457	7.484	7.[illegible]85

AL% IN Z	31.59	31.07	31.05	31.1[illegible]	30.[illegible]8	30.06	30.69	29.62	30.21	31.93
CHARGE	44.00	44.0[illegible]	44.0[illegible]	44.[illegible]0	44.00	44.0[illegible]	44.[illegible]0	44.[illegible]0	44.[illegible]	44.[illegible]

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	50.74	55.52	56.[illegible]1	57.[illegible]3	56.83	54.78	55.24	56.24	55.25	5[illegible].[illegible]9
Fe	49.22	44.45	43.1[illegible]	42.64	43.17	45.1[illegible]	44.76	43.[illegible]6	44.[illegible]2	4[illegible].[illegible]
Ca	0.04	0.03	0.03	0.03	0.00	0.04	0.00	0.00	0.[illegible]3	0.14

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.08	0.09	0.09	0.09	0.00	0.09	0.00	0.00	0.09	0.[illegible]7
Na	15.49	5.[illegible]3	3.0[illegible]	3.34	7.0[illegible]	9.23	[illegible].15	4.77	7.[illegible]9	5.[illegible]2
K	84.43	94.08	96.91	96.57	92.97	90.68	91.[illegible]5	95.23	92.[illegible]1	93.91

COMPOSITION IN OXIDES

	BIJ 9	BIL 1	BIL 2	BIM 1	BIM 2	BIM 3	BIM 4	BIM 5	BIN 1	BIN 2
SiO2	36.60	39.8.	39.61	36.42	36.62	36.11	36.29	35.95	42.42	42.64
Al2O3	18.71	18.15	17.76	19.00	17.83	17.98	18.16	18.10	17.32	17.74
TiO2	1.9>	1.52	1.43	1.62	1.44	1.67	1.27	1.47	0.94	0.9
MgO	11.65	10.06	10.25	10.25	11.02	10.34	10.00	10.08	16.47	16.05
FeO	17.01	17.06	16.75	1[illegible].35	19.17	1[illegible].97	19.94	19.41	9.30	9.14
MnO	0.08	0.05	0.04	0.05	0.05	0.05	0.09	0.11	0.04	0.12
CaO	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.0
Na2O	0.54	0.21	0.20	0.19	0.21	0.27	0.23	0.16	0.00	0.65
K2O	9.51	9.19	9.[illegible]	9.35	9.06	9.29	9.27	9.40	9.24	9.57
	95.09	95.[illegible]3	94.[illegible]7	9[illegible].23	95.42	94.8[illegible]	95.24	94.[illegible]2	96.16	9[illegible].7

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BIJ 9	BIL 1	BIL 2	BIM 1	BIM 2	BIM 3	BIM 4	BIM 5	BIM 6	BIN 2
Si	5.482	5.875	5.890	5.500	5.540	5.504	5.529	5.496	6.041	5.994
Al	2.51[illegible]	2.124	2.11[illegible]	2.5[illegible]0	2.460	2.496	2.471	2.504	1.9[illegible]4	2.006
	8.000	[illegible].000	[illegible].000	8.000	8.000	[illegible].000	[illegible].000	[illegible].000	[illegible].000	[illegible].000
Al	0.785	1.027	1.0[illegible]2	[illegible].[illegible]2	0.719	0.735	[illegible].79[illegible]	0.757	0.92[illegible]	[illegible].9[illegible]
Ti	0.225	0.168	0.160	0.184	0.164	0.214	0.146	0.192	0.100	0.1[illegible]4
Mg	2.6[illegible]5	2.20[illegible]	2.27[illegible]	2.3[illegible]7	2.4[illegible]5	2.349	2.271	2.297	3.463	3.[illegible]3
Fe2	2.131	2.102	2.083	2.[illegible]17	2.425	2.418	2.541	2.482	1.097	1.[illegible]77
Mn	[illegible].01[illegible]	[illegible].00[illegible]	[illegible].[illegible]05	[illegible].006	[illegible].[illegible]6	[illegible].[illegible]6	[illegible].012	[illegible].[illegible]4	[illegible].[illegible]6	[illegible].[illegible]4
Ca	0.000	0.002	0.002	0.000	0.000	0.000	0.000	0.000	0.002	[illegible].0[illegible]
Na	0.157	0.060	0.05[illegible]	0.05[illegible]	[illegible].05[illegible]	0.0[illegible]0	[illegible].06[illegible]	0.047	[illegible].0[illegible]	[illegible].1[illegible]
K	1.626	1.671	1.68[illegible]	1.8[illegible]1	1.752	1.806	1.8[illegible]2	1.833	1.666	1.7[illegible]9
	7.535	7.246	7.265	7.554	7.614	7.609	7.63[illegible]	7.622	7.253	7.[illegible]8

Al% IN Z	31.47	26.55	26.39	31.25	30.75	31.20	30.89	31.30	24.49	25.[illegible]8
CHARGE	44.00	44.0[illegible]	44.[illegible]0	44.[illegible]0	44.[illegible]0	44.[illegible]0	44.[illegible]0	44.00	44.00	44.[illegible]0

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	54.87	51.15	52.09	49.82	50.54	49.21	47.08	47.97	75.[illegible]4	75.50
Fe	45.13	48.[illegible]1	47.[illegible]7	50.1[illegible]	49.4[illegible]	50.7[illegible]	52.92	52.[illegible]1	24.[illegible]5	24.[illegible]0
Ca	0.00	0.04	0.04	0.00	0.00	0.00	0.00	0.00	0.[illegible]3	0.[illegible]0

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.0[illegible]	0.0[illegible]	0.00	0.00	0.00	0.00	0.00	0.0[illegible]	[illegible].0[illegible]
Na	[illegible].8[illegible]	3.4[illegible]	3.[illegible]1	3.00	[illegible].4[illegible]	4.23	[illegible].6[illegible]	2.52	0.[illegible]	9.[illegible]9
K	91.2[illegible]	96.45	96.[illegible]0	97.00	96.[illegible]0	95.77	96.37	97.48	99.91	90.[illegible]1

COMPOSITION IN OXIDES

	BIN 3	BIO 1	BIO 2	BIO 3	BIO 4	BIO 5	BIO 6	BIO 7	BIP 1	BIP 2
SiO2	41.56	40.25	41.65	41.59	40.12	40.15	41.02	40.63	35.97	35.65
Al2O3	18.01	18.43	17.76	17.77	18.80	18.44	17.48	17.94	19.41	19.53
TiO2	0.90	1.00	0.89	0.91	1.02	0.95	0.91	0.99	1.43	1.43
MgO	15.80	19.11	20.03	19.39	18.94	19.47	20.28	19.40	9.49	9.36
FeO	9.12	6.25	6.22	6.32	6.40	6.31	6.41	6.39	19.93	20.11
MnO	0.09	0.05	0.01	0.08	0.03	0.07	0.09	0.03	0.13	0.16
CaO	0.00	0.05	0.04	0.16	0.00	0.00	0.11	0.01	0.14	0.14
Na2O	0.00	0.41	0.32	0.47	0.43	0.44	0.40	0.35	0.38	0.31
K2O	9.64	8.96	8.12	8.21	8.42	8.29	8.36	8.32	8.47	8.47
	95.12	95.08	95.75	95.50	95.24	95.04	95.45	95.15	95.11	94.54

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BIN 3	BIO 1	BIO 2	BIO 3	BIO 4	BIO 5	BIO 6	BIO 7	BIP 1	BIP 2
Si	5.948	5.774	5.932	5.849	5.690	5.705	5.784	5.766	5.456	5.448
Al	2.052	2.226	2.068	2.151	2.310	2.295	2.216	2.234	2.544	2.552
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	0.935	0.844	0.763	0.795	0.833	0.791	0.859	0.766	0.920	0.929
Ti	0.097	0.106	0.094	0.096	0.109	0.091	0.096	0.105	0.163	0.153
Mg	3.371	4.026	4.191	4.065	4.004	4.121	4.262	4.103	2.146	2.132
Fe2	1.092	0.759	0.728	0.743	0.769	0.749	0.756	0.758	2.516	2.563
Mn	0.011	0.006	0.001	0.010	0.004	0.008	0.010	0.004	0.017	0.021
Ca	0.000	0.000	0.000	0.009	0.000	0.000	0.002	0.002	0.007	0.007
Na	0.000	0.112	0.097	0.128	0.116	0.121	0.109	0.096	0.112	0.092
K	1.760	1.616	1.575	1.599	1.704	1.681	1.594	1.687	1.631	1.613
	7.316	7.449	7.449	7.445	7.541	7.562	7.519	7.521	7.517	7.514

ALZ IN Z	25.65	27.83	27.10	26.83	28.87	28.69	27.70	27.93	31.20	31.91
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	75.35	84.39	85.17	84.21	83.83	84.47	84.75	84.31	45.80	46.14
Fe	24.65	15.61	14.83	15.60	16.17	15.53	15.22	15.69	54.16	54.72
Ca	0.00	0.00	0.00	0.19	0.00	0.00	0.03	0.03	0.14	0.14

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.00	0.00	0.00	0.52	0.00	0.00	0.09	0.09	0.37	0.38
Na	0.00	6.50	5.23	7.34	6.49	6.72	6.42	5.60	6.34	5.15
K	100.00	93.50	94.77	92.10	93.51	93.28	93.50	94.52	93.24	94.27

COMPOSITION IN OXIDES

	BIP 3	BIP 4	BIP 5	BIP 6	BIP 7	BIP 8	BIG 1	BIG 2	BIG 3	BIG 4
SiO2	35.[illegible]	34.80	35.90	35.56	35.78	35.63	36.92	3[illegible].65	37.40	3[illegible].01
Al2O3	19.20	19.25	19.16	19.77	20.04	19.44	18.14	18.75	17.48	17.39
TiO2	1.47	1.4[illegible]	1.34	1.78	0.[illegible]2	1.38	1.29	[illegible].9[illegible]	1.09	1.[illegible]9
MgO	9.54	[illegible].74	9.04	9.62	9.09	9.26	[illegible].57	9.61	9.32	9.[illegible]4
FeO	19.95	20.[illegible]6	21.72	19.[illegible]4	20.[illegible]0	20.0[illegible]	19.[illegible]7	21.[illegible]0	20.[illegible]4	19.[illegible]7
MnO	0.09	0.12	0.17	0.11	0.19	0.13	0.01	0.01	0.03	0.[illegible]5
CaO	0.03	0.01	0.[illegible]3	[illegible].1	0.19	[illegible].04	[illegible].1	[illegible].4	.16	.1
Na2O	0.32	0.29	0.33	0.41	0.46	0.34	0.31	0.23	0.24	0.27
K2O	[illegible].34	[illegible].31	[illegible].40	[illegible].29	7.71	[illegible].25	[illegible].37	[illegible].51	7.[illegible]7	[illegible].5[illegible]
	97.[illegible]2	93.32	[illegible]4.[illegible]7	94.[illegible]9	9[illegible].57	94.55	9[illegible].39	[illegible]7.7[illegible]	[illegible]3.[illegible]5	94.[illegible]

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BIP 3	BIP 4	BIP 5	BIP 6	BIP 7	BIP 8	BIG 1	BIG 2	BIG 3	BIG 4
Si	5.461	5.408	5.464	5.393	5.449	5.442	5.618	5.6[illegible]7	5.7[illegible]5	5.664
Al	2.53[illegible]	2.592	2.53[illegible]	2.[illegible]07	2.551	2.55[illegible]	2.[illegible]82	2.[illegible]13	2.2[illegible]5	2.[illegible]46
	8.000	.000	[illegible].000	[illegible].000	[illegible].100	[illegible].000	[illegible].000	[illegible].[illegible]0	[illegible].[illegible]	[illegible].
Al	0.904	0.934	0.911	0.929	1.046	0.[illegible]41	0.[illegible]60	0.[illegible]39	0.[illegible]95	[illegible].1[illegible]
Ti	0.168	0.173	0.154	0.203	0.094	0.159	0.148	0.10[illegible]	0.126	0.1[illegible]4
Mg	2.164	2.[illegible]24	2.057	2.175	2.[illegible]63	2.1[illegible]0	2.176	2.1[illegible]8	2.13[illegible]	2.1[illegible]7
Fe2	2.539	2.646	2.645	2.453	2.585	2.565	2.535	2.[illegible]84	2.57[illegible]	2.564
Mn	0.012	0.[illegible]16	0.[illegible]22	0.[illegible]14	[illegible].025	[illegible].017	[illegible].[illegible]01	0.[illegible]01	0.[illegible]04	0.[illegible]7
Ca	0.005	0.002	0.005	0.002	0.031	0.007	0.002	0.006	0.026	0.[illegible]2
Na	0.094	[illegible].075	0.09[illegible]	[illegible].121	0.1[illegible]3	[illegible].1[illegible]1	[illegible].092	[illegible].[illegible]6	0.[illegible]71	[illegible].[illegible]
K	1.61[illegible]	1.647	1.635	1.[illegible]04	1.498	1.607	1.62[illegible]	1.597	1.545	1.672
	7.506	7.517	7.[illegible]26	7.499	7.474	7.504	[illegible].46[illegible]	7.410	7.[illegible]8[illegible]	7.[illegible]5[illegible]

	BIP 3	BIP 4	BIP 5	BIP 6	BIP 7	BIP 8	BIG 1	BIG 2	BIG 3	BIG 4
Al% IN Z	31.74	32.40	31.[illegible]0	32.5[illegible]	31.[illegible]9	31.98	29.[illegible]8	2[illegible].[illegible]1	2[illegible].31	29.2[illegible]
CHARGE	44.00	44.0[illegible]	[illegible].0[illegible]	[illegible].0[illegible]	[illegible].[illegible]0	[illegible].00	[illegible].0[illegible]	44.0[illegible]	[illegible].[illegible]0	[illegible].[illegible]0

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+ Mn - Ca

	BIP 3	BIP 4	BIP 5	BIP 6	BIP 7	BIP 8	BIG 1	BIG 2	BIG 3	BIG 4
Mg	45.8[illegible]	43.1[illegible]	43.5[illegible]	46.[illegible]3	43.[illegible]6	44.89	46.16	44.[illegible]5	45.[illegible]3	45.[illegible]9
Fe	54.[illegible]4	[illegible].[illegible]	[illegible].4[illegible]	[illegible].1[illegible]	[illegible].[illegible]	[illegible].97	[illegible].[illegible]	[illegible].[illegible]2	[illegible].41	[illegible].
Ca	0.10	0.04	0.10	0.03	0.66	0.14	0.03	0.13	0.56	0.[illegible]3

TRIANGULAR DIAGRAM Na - Ca - K

	BIP 3	BIP 4	BIP 5	BIP 6	BIP 7	BIP 8	BIG 1	BIG 2	BIG 3	BIG 4
Ca	0.2[illegible]	0.10	0.2[illegible]	0.09	1.[illegible]7	0.3[illegible]	0.09	0.3[illegible]	1.6[illegible]	0.[illegible]9
Na	[illegible].4[illegible]	4.[illegible]7	5.6[illegible]	6.[illegible]8	[illegible].0[illegible]	5.[illegible]7	5.32	[illegible].93	4.34	[illegible].[illegible]
K	[illegible]4.22	95.54	[illegible]4.10	92.92	[illegible]0.14	93.75	[illegible]4.5[illegible]	95.69	[illegible]4.[illegible]6	95.[illegible]7

COMPOSITION IN OXIDES

	BIQ 5	BIR 1	BIR 2	BIR 3	BIR 4	BIR 5	BIR 6	BIU 1	BIV 1	BIV 2
SiO2	37.89	35.91	36.6[illegible]	36.75	35.90	34.76	34.[illegible]2	3[illegible].[illegible]0	33.9[illegible]	35.77
Al2O3	17.53	18.18	18.54	18.23	17.18	20.49	18.28	17.87	18.51	18.[illegible]0
TiO2	1.55	1.20	1.47	1.29	1.47	[illegible].94	[illegible].75	1.56	1.11	1.[illegible]1
MgO	9.00	10.23	9.85	10.22	9.79	9.[illegible]8	8.87	13.01	9.2[illegible]	9.52
FeO	20.97	18.84	20.56	20.60	19.86	22.9[illegible]	21.91	14.7[illegible]	19.87	19.1[illegible]
MnO	0.07	0.08	0.05	0.12	0.04	0.15	0.15	0.15	0.11	0.[illegible]7
CaO	0.01	0.06	0.10	0.11	0.17	0.16	0.23	0.[illegible]0	0.13	0.[illegible]3
Na2O	0.29	0.36	0.36	0.29	0.36	0.16	0.48	0.19	0.29	0.22
K2O	8.60	8.65	8.47	8.56	8.55	6.42	7.[illegible]5	9.32	8.[illegible]9	8.61
	95.91	93.51	96.0[illegible]	96.17	93.32	93.92	93.34	94.[illegible]9	91.[illegible]5	93.42

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BIQ 5	BIR 1	BIR 2	BIR 3	BIR 4	BIR 5	BIR 6	BIU 1	BIV 1	BIV 2
Si	5.714	5.534	5.522	5.534	5.578	5.312	5.453	5.651	5.3[illegible]7	5.518
Al	2.286	2.466	2.478	2.466	2.422	2.69[illegible]	2.547	2.349	2.61[illegible]	2.482
	8.000	8.000	[illegible].00[illegible]	8.000	[illegible].000	[illegible].000	[illegible].[illegible]00	[illegible].000	[illegible].	[illegible]
Al	0.829	0.835	0.[illegible]12	0.770	0.723	1.002	0.827	0.783	0.[illegible]4[illegible]	0.[illegible]9[illegible]
Ti	0.176	0.139	0.166	0.146	0.172	0.108	0.088	0.174	0.132	0.152
Mg	2.023	2.350	2.210	2.294	2.267	2.250	2.071	2.[illegible]84	2.188	2.1[illegible]3
Fe2	2.645	2.428	2.589	2.594	2.530	2.679	2.870	1.838	2.635	2.476
Mn	0.009	0.010	0.006	0.015	0.005	[illegible].019	0.020	0.[illegible]19	0.015	0.[illegible]09
Ca	0.002	0.010	0.016	0.018	0.028	0.026	0.039	0.000	0.022	0.[illegible]05
Na	0.085	0.103	0.105	[illegible].085	[illegible].108	[illegible].047	[illegible].146	[illegible].[illegible]55	[illegible].089	[illegible].
K	1.654	1.700	1.627	1.644	1.695	1.252	1.568	1.76[illegible]	1.737	1.694
	7.422	7.580	7.532	7.566	7.579	7.384	7.628	7.520	7.664	7.51[illegible]

	BIQ 5	BIR 1	BIR 2	BIR 3	BIR 4	BIR 5	BIR 6	BIU 1	BIV 1	BIV 2
AlZ IN Z	28.58	30.83	30.97	30.82	30.28	33.60	31.83	29.37	32.66	31.02
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.[illegible]	44.

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	43.24	48.97	45.84	46.61	46.45	45.24	41.42	60.83	45.03	46.7[illegible]
Fe	56.72	50.[illegible]2	53.[illegible]2	53.[illegible]3	52.97	54.24	57.[illegible]1	39.17	54.[illegible]2	[illegible].11
Ca	0.03	0.21	0.33	0.36	0.58	0.53	0.77	0.00	0.45	0.11

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.09	0.54	0.92	1.02	1.55	1.98	2.20	0.00	1.19	0.[illegible]8
Na	4.87	5.92	6.01	4.[illegible]5	5.92	3.59	3.32	3.01	4.[illegible]2	5.73
K	95.04	93.54	93.07	94.14	92.53	94.45	89.48	96.99	93.9[illegible]	95.[illegible]9

COMPOSITION IN OXIDES

	BIY 1	BIY 2	BIY 3	BI& 1	BI& 2	BI& 3	BI& 4	BI& 5
SiO2	39.01	39.33	39.17	35.61	35.54	35.13	33.69	36.07
Al2O3	13.26	13.43	13.35	17.59	17.45	18.11	17.21	17.73
TiO2	1.16	1.17	1.17	2.41	2.43	2.44	2.72	2.99
MgO	13.09	14.46	13.78	3.54	3.29	3.38	3.72	3.76
FeO	17.62	17.62	17.62	24.59	24.27	25.60	27.21	22.66
MnO	0.25	0.32	0.29	0.45	0.41	0.40	0.53	0.44
CaO	0.06	0.09	0.07	0.14	0.23	0.21	0.21	0.64
Na2O	0.09	0.07	0.09	0.27	0.06	0.17	0.20	0.11
K2O	9.76	9.79	9.78	8.93	8.70	8.93	7.57	8.43
	94.30	96.28	95.32	93.53	92.43	94.37	93.05	93.43

STRUCT. FORM. BASED ON 22.0 OXYGENS

	BIY 1	BIY 2	BIY 3	BI& 1	BI& 2	BI& 3	BI& 4	BI& 5
Si	5.958	5.881	5.918	5.666	5.705	5.567	5.453	5.751
Al	2.042	2.119	2.082	2.334	2.295	2.433	2.547	2.249
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	0.345	0.248	0.295	0.964	1.006	0.950	0.738	1.028
Ti	0.133	0.132	0.133	0.288	0.299	0.291	0.331	0.353
Mg	2.980	3.223	3.103	0.840	0.787	0.798	0.896	0.879
Fe2	2.251	2.203	2.226	3.272	3.258	3.393	3.655	2.972
Mn	0.032	0.041	0.037	0.061	0.056	0.054	0.073	0.058
Ca	0.010	0.014	0.011	0.024	0.040	0.036	0.036	0.108
Na	0.027	0.020	0.026	0.083	0.019	0.052	0.063	0.033
K	1.902	1.867	1.885	1.812	1.781	1.805	1.564	1.687
	7.679	7.748	7.716	7.344	7.245	7.379	7.337	7.118

AL% IN Z	25.52	26.49	26.03	29.18	28.69	30.41	31.83	28.11
CHARGE	44.00	44.00	44.00	44.00	44.00	44.00	44.00	44.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	56.52	58.80	57.70	20.01	19.01	18.65	19.14	21.88
Fe	43.30	40.94	42.09	79.42	80.03	80.52	80.09	75.44
Ca	0.19	0.26	0.21	0.57	0.96	0.83	0.78	2.68

TRIANGULAR DIAGRAM Na - Ca - K

Ca	0.51	0.76	0.59	1.24	2.15	1.88	2.19	5.88
Na	1.38	1.07	1.37	4.34	1.02	2.76	3.78	1.83
K	98.12	98.18	98.04	94.42	96.83	95.36	94.03	92.29

COMPOSITION IN OXIDES

	CHD 1	CHD 2	CHD 3	CHD 4	CHD 5	CHD 6	CHD 7	CHE 1	CHE 2	CHE 3
SiO2	27.35	26.78	26.40	26.66	25.73	26.29	26.54	27.07	27.15	27.19
Al2O3	20.44	21.13	21.67	22.04	21.51	21.94	21.45	21.66	21.41	21.24
TiO2	0.07	0.05	0.07	0.05	0.02	0.09	0.05	0.07	0.07	0.07
MgO	19.50	18.74	18.67	18.88	18.34	18.70	18.90	23.77	23.71	24.28
FeO	19.70	20.11	19.93	19.98	19.07	19.98	19.81	12.63	13.14	12.79
MnO	0.21	0.20	0.17	0.25	0.23	0.17	0.20	0.07	0.19	0.13
CaO	0.06	0.08	0.07	0.03	0.06	0.04	0.06	0.06	0.07	0.08
Na2O	0.04	0.04	0.02	0.05	0.05	0.00	0.02	0.02	0.02	0.03
	87.37	87.13	87.05	87.94	85.01	87.21	86.93	85.35	85.76	85.81

STRUCT. FORM. BASED ON 23.0 OXYGENS

	CHD 1	CHD 2	CHD 3	CHD 4	CHD 5	CHD 6	CHD 7	CHE 1	CHE 2	CHE 3
Si	5.600	5.512	5.436	5.429	5.412	5.402	5.467	5.466	5.473	5.471
Al	2.400	2.488	2.564	2.571	2.588	2.598	2.533	2.534	2.526	2.529
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	2.532	2.636	2.694	2.718	2.745	2.715	2.675	2.620	2.564	2.509
Ti	0.011	0.008	0.011	0.008	0.003	0.014	0.008	0.011	0.011	0.011
Mg	5.951	5.749	5.730	5.730	5.750	5.727	5.773	7.154	7.127	7.282
Fe2	3.373	3.462	3.440	3.403	3.355	3.433	3.413	2.133	2.216	2.152
Mn	0.036	0.035	0.030	0.043	0.041	0.030	0.035	0.012	0.032	0.022
Ca	0.013	0.018	0.015	0.007	0.014	0.009	0.013	0.013	0.015	0.017
Na	0.016	0.016	0.008	0.020	0.020	0.000	0.008	0.008	0.009	0.012
	11.932	11.925	11.928	11.929	11.928	11.928	11.925	11.950	11.973	12.005

AL% IN Z	30.01	31.10	32.05	32.14	32.34	32.48	31.66	31.68	31.56	31.61
CHARGE	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	63.49	62.06	62.18	62.41	62.78	62.26	62.52	76.83	75.89	76.97
Fe	36.37	37.75	37.66	37.52	37.07	37.65	37.34	23.03	23.94	22.95
Ca	0.14	0.19	0.17	0.07	0.15	0.10	0.14	0.14	0.16	0.18

COMPOSITION IN OXIDES

	CHE 4	CHE 5	CHE 6	CHE 7	CHI 1	CHJ 1	CHL 1	CHL 2	CHM 1	CHM 2
SiO2	27.00	26.96	27.25	27.10	25.72	25.29	23.93	24.20	24.13	24.44
Al2O3	21.26	21.47	21.69	21.46	23.62	23.75	22.00	22.05	22.34	22.31
TiO2	0.07	0.05	0.05	0.07	0.09	0.07	0.09	0.07	0.09	0.09
MgO	24.05	23.81	24.09	23.95	17.48	17.85	15.50	15.41	16.33	15.99
FeO	12.65	12.75	13.12	12.85	21.75	21.47	22.62	21.70	22.36	23.47
MnO	0.11	0.13	0.11	0.12	0.11	0.11	0.07	0.11	0.09	0.09
CaO	0.06	0.06	0.04	0.06	0.01	0.06	0.04	0.07	0.01	0.01
Na2O	0.05	0.05	0.00	0.02	0.02	0.00	0.00	0.04	0.00	0.00
K2O	0.00	0.00	0.00	0.00	0.01	0.02	0.04	0.01	0.00	0.00
	85.31	85.28	86.35	85.63	88.81	89.02	84.29	83.66	85.35	86.40

STRUCT. FORM. BASED ON 28.0 OXYGENS

	CHE 4	CHE 5	CHE 6	CHE 7	CHI 1	CHJ 1	CHL 1	CHL 2	CHM 1	CHM 2
Si	5.473	5.457	5.453	5.464	5.230	5.161	5.193	5.260	5.157	5.134
Al	2.527	2.543	2.547	2.536	2.770	2.839	2.807	2.740	2.843	2.866
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	2.541	2.579	2.568	2.563	2.890	2.831	2.819	2.909	2.784	2.761
Ti	0.011	0.008	0.008	0.011	0.014	0.011	0.015	0.011	0.014	0.014
Mg	7.250	7.184	7.185	7.198	5.298	5.409	5.213	4.993	5.202	5.055
Fe2	2.140	2.158	2.195	2.167	3.698	3.718	4.105	3.945	3.996	4.163
Mn	0.019	0.022	0.019	0.020	0.019	0.019	0.013	0.020	0.016	0.016
Ca	0.013	0.013	0.009	0.013	0.002	0.013	0.009	0.016	0.002	0.002
Na	0.020	0.020	0.000	0.008	0.008	0.000	0.000	0.017	0.000	0.016
K	0.000	0.000	0.000	0.000	0.003	0.005	0.011	0.003	0.000	0.000
	11.992	11.984	11.982	11.980	11.932	12.006	11.985	11.914	12.015	12.013

Al% IN Z	31.59	31.78	31.84	31.70	34.63	35.74	35.09	34.25	35.54	35.20
CHARGE	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	76.95	76.61	76.37	76.59	58.75	59.05	54.85	55.64	56.44	54.73
Fe	22.91	23.25	23.54	23.27	41.23	40.80	45.05	44.18	43.54	45.25
Ca	0.14	0.14	0.09	0.14	0.02	0.14	0.10	0.18	0.02	0.02

COMPOSITION IN OXIDES

	CHM 3	CHM 4	CHN 1	CHN 2	CHN 3	CHO 1	CHO 2	CHO 3	CHR 1	CHR 2
SiO2	24.35	24.54	25.84	25.19	25.74	27.59	27.12	27.63	25.12	24.01
Al2O3	22.40	22.37	22.48	22.32	22.03	22.84	23.41	22.47	23.35	23.44
TiO2	0.09	0.07	0.04	0.04	0.04	0.07	0.04	0.04	0.03	0.07
MgO	15.81	16.41	22.79	23.48	23.78	27.99	28.36	27.51	13.45	9.95
FeO	23.35	23.24	11.69	11.48	11.54	7.68	7.96	7.72	26.56	29.66
MnO	0.09	0.09	0.15	0.16	0.12	0.08	0.08	0.11	0.29	0.73
CaO	0.01	0.00	0.01	0.03	0.03	0.01	0.01	0.03	0.10	0.11
Na2O	0.00	0.02	0.00	0.00	0.03	0.05	0.00	0.06	0.06	0.02
K2O	0.00	0.00	0.04	0.04	0.01	0.01	0.02	0.01	0.01	0.01
	86.10	86.74	83.04	82.74	83.32	86.32	87.00	85.58	88.99	88.20

STRUCT. FORM. BASED ON 28.0 OXYGENS

	CHM 3	CHM 4	CHN 1	CHN 2	CHN 3	CHO 1	CHO 2	CHO 3	CHR 1	CHR 2
Si	5.180	5.177	5.344	5.237	5.309	5.367	5.248	5.422	5.239	5.166
Al	2.820	2.823	2.656	2.763	2.691	2.633	2.752	2.578	2.761	2.834
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	2.797	2.739	2.623	2.705	2.664	2.603	2.586	2.618	2.971	3.111
Ti	0.014	0.011	0.006	0.006	0.006	0.010	0.006	0.006	0.005	0.011
Mg	5.013	5.160	7.025	7.275	7.311	8.116	8.179	8.046	4.181	3.191
Fe2	4.154	4.100	2.022	1.996	1.991	1.249	1.288	1.267	4.636	5.573
Mn	0.016	0.016	0.026	0.028	0.021	0.013	0.013	0.018	0.051	0.133
Ca	0.002	0.000	0.002	0.007	0.007	0.002	0.002	0.006	0.022	0.025
Na	0.000	0.008	0.000	0.000	0.012	0.019	0.000	0.023	0.024	0.018
K	0.000	0.000	0.011	0.011	0.003	0.002	0.005	0.003	0.003	0.003
	11.997	12.035	11.915	12.028	12.014	12.015	12.089	11.987	11.900	11.856

AL% IN Z	35.25	35.29	33.20	34.54	33.64	32.91	34.40	32.23	34.51	35.42
CHARGE	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	54.57	55.63	77.41	78.18	78.37	86.52	86.26	86.17	47.03	36.58
Fe	45.40	44.37	22.57	21.75	21.56	13.46	13.72	13.76	52.72	63.13
Ca	0.02	0.00	0.02	0.07	0.07	0.02	0.02	0.07	0.25	0.29

COMPOSITION IN OXIDES

	CHS 1	CHS 2	CHS 3	CHS 4	CHT 1	CHT 2	CHT 3	CHT 4	CHT 5	CHU 1
SiO2	24.45	24.57	25.06	24.93	25.43	25.23	25.43	25.60	25.56	27.42
Al2O3	23.03	22.95	21.89	21.50	22.93	22.84	22.11	22.82	24.36	21.50
TiO2	0.10	0.05	0.07	0.14	0.04	0.05	0.04	0.05	0.07	0.09
MgO	16.00	15.75	17.07	15.73	18.15	18.12	18.75	18.06	17.97	20.25
FeO	24.39	24.65	24.21	24.08	20.11	20.07	20.16	19.44	19.57	18.04
MnO	0.31	0.35	0.34	0.35	0.53	0.55	0.54	0.49	0.51	0.25
CaO	0.07	0.08	0.01	0.04	0.01	0.01	0.01	0.11	0.07	0.03
Na2O	0.11	0.13	0.02	0.06	0.02	0.02	0.02	0.02	0.07	0.02
K2O	0.02	0.04	0.01	0.01	0.00	0.04	0.01	0.01	0.05	0.01
	88.48	88.57	88.69	86.84	87.22	86.93	87.07	86.60	88.53	87.61

STRUCT. FORM. BASED ON 28.0 OXYGENS

	CHS 1	CHS 2	CHS 3	CHS 4	CHT 1	CHT 2	CHT 3	CHT 4	CHT 5	CHU 1
Si	5.091	5.118	5.200	5.285	5.243	5.224	5.262	5.296	5.222	5.542
Al	2.909	2.882	2.800	2.715	2.757	2.776	2.738	2.704	2.778	2.458
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	2.743	2.753	2.553	2.658	2.615	2.798	2.655	2.659	3.020	2.664
Ti	0.016	0.008	0.011	0.022	0.006	0.008	0.006	0.008	0.011	0.014
Mg	4.966	4.890	5.280	4.971	5.578	5.592	5.793	5.569	5.409	6.101
Fe2	4.247	4.294	4.201	4.270	3.467	3.475	3.489	3.363	3.305	3.050
Mn	0.055	0.062	0.060	0.063	0.093	0.096	0.095	0.086	0.087	0.043
Ca	0.016	0.018	0.002	0.009	0.002	0.002	0.002	0.024	0.015	0.006
Na	0.044	0.053	0.008	0.025	0.008	0.008	0.008	0.008	0.027	0.008
K	0.005	0.011	0.003	0.003	0.000	0.011	0.003	0.003	0.013	0.003
	12.092	12.088	12.118	12.020	11.969	11.991	12.041	11.920	11.888	11.885

Al% IN Z	36.36	36.02	35.00	33.93	34.46	34.79	34.22	33.80	34.72	30.72
CHARGE	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00	56.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	53.49	52.79	55.32	53.38	61.03	61.01	61.73	61.59	61.35	66.32
Fe	46.34	47.02	44.65	46.52	38.95	38.97	38.25	38.14	38.49	33.61
Ca	0.17	0.19	0.02	0.10	0.02	0.02	0.02	0.27	0.17	0.07

COMPOSITION IN OXIDES

	CHV 1	CHV 2	CHV 3
SiO2	24.04	24.42	24.74
Al2O3	20.26	21.93	21.34
TiO2	0.07	0.10	0.07
MgO	13.24	13.92	13.77
FeO	24.51	25.77	25.95
MnO	0.21	0.16	0.14
CaO	0.03	0.06	0.04
Na2O	0.02	0.02	0.02
K2O	0.02	0.01	0.03
	82.40	86.39	86.10

STRUCT. FORM. BASED ON 28.0 OXYGENS

	CHV 1	CHV 2	CHV 3
Si	5.408	5.248	5.339
Al	2.592	2.752	2.661
	8.000	8.000	8.000
Al	2.779	2.803	2.767
Ti	0.012	0.016	0.011
Mg	4.439	4.459	4.430
Fe2	4.611	4.632	4.684
Mn	0.040	0.029	0.026
Ca	0.007	0.014	0.009
Na	0.009	0.008	0.008
K	0.006	0.003	0.008
	11.902	11.964	11.944

Al% IN Z	32.41	34.40	33.26
CHARGE	56.00	56.00	56.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	48.80	48.82	48.42
Fe	51.12	51.03	51.48
Ca	0.08	0.15	0.10

COMPOSITION IN OXIDES

	CTP 1	CTP 2	CTP 3	CTP 4	CTP 5	CTP 6	CTP 7	CTP 8	CTP 9	CTP10
SiO2	24.32	24.40	24.17	24.27	24.17	24.22	24.22	24.02	24.01	24.13
Al2O3	40.55	40.32	40.57	40.43	40.62	40.59	40.05	40.10	40.30	40.37
TiO2	0.02	0.03	0.03	0.02	0.03	0.02	0.02	0.00	0.02	0.02
MgO	2.61	2.89	2.30	2.21	2.59	2.36	3.02	2.53	2.55	2.59
FeO	23.66	23.56	24.34	24.47	23.51	23.95	23.44	23.79	23.69	23.16
MnO	0.79	0.75	0.85	0.78	0.75	0.83	0.85	0.75	0.71	0.74
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01
Na2O	0.00	0.02	0.00	0.04	0.05	0.00	0.00	0.00	0.00	0.02
	91.95	91.97	92.26	92.27	91.73	91.97	91.41	91.19	91.28	91.53

STRUCT. FORM. BASED ON 12.0 OXYGENS

	CTP 1	CTP 2	CTP 3	CTP 4	CTP 5	CTP 6	CTP 7	CTP 8	CTP 9	CTP10
Si	2.017	2.022	2.006	2.014	2.008	2.012	2.019	2.012	2.009	2.002
	2.017	2.022	2.006	2.014	2.008	2.012	2.019	2.012	2.009	2.002
Al	3.963	3.938	3.969	3.959	3.974	3.973	3.935	3.960	3.972	3.964
Ti	0.001	0.002	0.002	0.001	0.002	0.001	0.001	0.000	0.001	0.001
Mg	0.323	0.357	0.284	0.273	0.321	0.292	0.375	0.316	0.319	0.359
Fe2	1.641	1.633	1.689	1.698	1.634	1.664	1.654	1.657	1.657	1.635
Mn	0.055	0.053	0.060	0.055	0.053	0.058	0.046	0.053	0.053	0.052
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Na	0.000	0.003	0.000	0.006	0.010	0.000	0.000	0.000	0.000	0.003
	5.983	5.986	6.003	5.994	5.997	5.989	5.993	5.995	5.998	6.014

Al% IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	15.98	17.48	13.99	13.49	15.93	14.51	18.25	15.52	15.70	17.54
Fe	84.02	82.52	86.01	86.51	84.02	85.49	81.71	84.48	84.30	82.46
Ca	0.00	0.00	0.00	0.00	0.00	0.00	0.04	0.00	0.00	0.00

COMPOSITION IN OXIDES

	CTP11	CTP12	CTQ 1	CTQ 2	CTQ 3	CTQ 4	CTQ 5	CTR 1	CTR 2	CTS 1
SiO2	23.91	24.16	24.66	24.79	24.91	25.62	25.98	23.95	23.87	24.67
Al2O3	40.71	40.40	39.49	39.18	39.05	38.65	39.34	38.39	38.75	40.01
TiO2	0.00	0.02	0.00	0.00	0.02	0.00	0.02	0.02	0.02	0.02
MgO	2.74	2.62	2.84	2.17	2.51	2.74	2.43	3.13	2.65	3.23
FeO	23.42	23.75	23.92	24.81	23.51	24.22	24.88	23.25	23.02	24.00
MnO	0.81	0.77	0.17	0.37	0.15	0.24	0.35	0.68	0.79	1.06
CaO	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00
Na2O	0.00	0.04	0.02	0.00	0.00	0.02	0.02	0.02	0.02	0.02
K2O	0.00	0.00	0.04	0.01	0.02	0.02	0.01	0.00	0.00	0.00
	91.59	91.76	91.15	91.33	90.20	91.51	93.03	89.44	89.13	93.03

STRUCT. FORM. BASED ON 12.0 OXYGENS

	CTP11	CTP12	CTQ 1	CTQ 2	CTQ 3	CTQ 4	CTQ 5	CTR 1	CTR 2	CTS 1
Si	1.990	2.010	2.061	2.078	2.098	2.133	2.131	2.046	2.043	2.028
Al	0.010	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	2.000	2.010	2.061	2.078	2.098	2.133	2.131	2.046	2.043	2.028
Al	3.985	3.962	3.890	3.871	3.877	3.792	3.803	3.864	3.910	3.877
Ti	0.000	0.001	0.000	0.000	0.001	0.000	0.001	0.001	0.001	0.001
Mg	0.340	0.325	0.354	0.271	0.315	0.340	0.297	0.398	0.334	0.402
Fe2	1.631	1.653	1.672	1.739	1.656	1.656	1.707	1.661	1.648	1.650
Mn	0.057	0.054	0.012	0.026	0.013	0.017	0.024	0.049	0.057	0.072
Ca	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.000
Na	0.000	0.006	0.003	0.000	0.000	0.003	0.003	0.003	0.003	0.003
K	0.000	0.000	0.004	0.001	0.002	0.002	0.001	0.000	0.000	0.000
	6.012	6.001	5.936	5.909	5.865	5.841	5.837	5.977	5.959	6.005

AL% IN Z	0.48	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	16.77	15.99	17.35	13.31	15.88	16.64	14.65	18.90	16.54	18.93
Fe	83.23	84.01	82.60	86.69	84.12	83.36	85.35	81.10	83.42	81.07
Ca	0.00	0.00	0.04	0.00	0.00	0.00	0.00	0.00	0.04	0.00

COMPOSITION IN OXIDES

	CTS 2	CTS 3	CTS 4	CTT 1	CTT 2	CTT 3	CTT 4	CTV 1	CTV 2	CTV 1
SiO2	24.02	24.05	23.86	24.85	25.04	25.08	25.30	23.77	23.89	23.44
Al2O3	39.65	39.36	39.53	40.48	40.74	41.09	41.12	40.15	39.26	39.66
TiO2	0.02	0.02	0.02	0.02	0.02	0.02	0.00	0.02	0.00	0.02
MgO	3.40	3.30	3.43	4.09	3.88	3.92	4.21	2.92	2.65	0.92
FeO	23.76	23.36	23.23	21.70	22.31	22.46	22.10	23.04	21.76	27.12
MnO	1.01	0.88	1.01	1.53	1.51	1.53	1.64	0.47	0.59	0.00
CaO	0.03	0.01	0.01	0.00	0.00	0.01	0.00	0.01	0.01	0.00
Na2O	0.02	0.00	0.02	0.02	0.02	0.04	0.02	0.04	0.04	0.04
K2O	0.01	0.02	0.02	0.00	0.00	0.00	0.01	0.01	0.00	0.00
	91.92	91.00	91.16	92.69	93.52	94.15	94.40	90.43	88.21	91.20

STRUCT. FORM. BASED ON 12.0 OXYGENS

	CTS 2	CTS 3	CTS 4	CTT 1	CTT 2	CTT 3	CTT 4	CTV 1	CTV 2	CTV 1
Si	2.001	2.019	2.001	2.031	2.032	2.023	2.031	2.000	2.049	2.019
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	2.001	2.019	2.001	2.031	2.032	2.023	2.031	2.000	2.049	2.019
Al	3.894	3.895	3.906	3.899	3.897	3.905	3.891	3.961	3.969	3.924
Ti	0.001	0.001	0.001	0.001	0.001	0.001	0.000	0.001	0.000	0.001
Mg	0.422	0.413	0.429	0.498	0.469	0.471	0.504	0.366	0.339	0.115
Fe2	1.656	1.640	1.632	1.483	1.514	1.515	1.484	1.621	1.562	1.953
Mn	0.071	0.063	0.072	0.106	0.104	0.105	0.112	0.033	0.042	0.000
Ca	0.003	0.001	0.001	0.000	0.000	0.001	0.000	0.001	0.001	0.000
Na	0.003	0.000	0.003	0.003	0.003	0.006	0.003	0.007	0.007	0.007
K	0.001	0.002	0.002	0.000	0.000	0.000	0.001	0.001	0.000	0.000
	6.051	6.014	6.047	5.990	5.988	6.004	5.994	6.012	5.920	6.003

AlZ IN Z	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CHARGE	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	19.62	19.51	20.09	23.87	22.49	22.53	24.00	18.11	17.43	5.70
Fe	80.25	80.45	79.87	76.13	77.51	77.43	76.00	81.84	82.53	94.30
Ca	0.12	0.04	0.04	0.00	0.00	0.04	0.00	0.04	0.05	0.00

COMPOSITION IN OXIDES

	CTW 2	CTW 3
SiO2	23.58	23.34
Al2O3	38.88	38.70
TiO2	0.02	0.00
MgO	1.05	0.98
FeO	26.69	26.40
MnO	0.00	0.05
CaO	0.00	0.01
Na2O	0.02	0.02
	90.24	89.50

STRUCT. FORM. BASED ON 12.0 OXYGENS

	CTW 2	CTW 3
Si	2.024	2.019
	2.024	2.019
Al	3.932	3.946
Ti	0.001	0.000
Mg	0.134	0.126
Fe2	1.916	1.910
Mn	0.000	0.004
Ca	0.000	0.001
Na	0.003	0.003
	5.987	5.990

AL% IN Z	0.00	0.00
CHARGE	24.00	24.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	6.55	6.19
Fe	93.45	93.76
Ca	0.00	0.05

COMPOSITION IN OXIDES

	STI 1	STI 2	STI 3	STR 1	STR 2	STR 3	STR 4	STR 5	STR 6	STR 7
SiO2	28.26	28.67	28.47	27.03	28.49	28.54	27.75	27.83	27.93	29.94
Al2O3	54.70	53.72	54.55	54.88	53.97	52.67	52.33	52.55	52.87	53.64
TiO2	0.52	0.57	0.54	0.23	0.47	0.63	0.55	0.50	0.36	0.27
MgO	1.78	1.79	1.62	1.12	1.77	1.70	1.67	1.72	1.54	1.29
FeO	13.40	13.75	13.48	13.69	14.39	14.31	14.49	14.45	14.34	13.76
MnO	0.19	0.19	0.19	0.53	0.17	0.19	0.12	0.08	0.24	0.59
CaO	0.01	0.00	0.00	0.00	0.01	0.01	0.00	0.01	0.01	0.01
Na2O	0.03	0.02	0.02	0.02	0.02	0.13	0.02	0.02	0.05	0.02
K2O	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.01
	98.90	98.72	98.88	97.51	99.30	98.20	96.95	97.17	97.35	99.53

STRUCT. FORM. BASED ON 46.0 OXYGENS

	STI 1	STI 2	STI 3	STR 1	STR 2	STR 3	STR 4	STR 5	STR 6	STR 7
Si	7.735	7.873	7.792	7.535	7.803	7.911	7.803	7.803	7.814	8.145
Al	0.265	0.127	0.208	0.465	0.197	0.089	0.197	0.197	0.186	0.000
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.145
Al	17.379	17.258	17.389	17.565	17.225	17.118	17.147	17.168	17.246	17.199
Ti	0.107	0.118	0.111	0.048	0.097	0.131	0.116	0.105	0.076	0.055
Mg	0.726	0.733	0.661	0.465	0.723	0.702	0.700	0.719	0.642	0.523
Fe2	3.067	3.158	3.086	3.191	3.296	3.317	3.408	3.388	3.355	3.131
Mn	0.044	0.044	0.044	0.125	0.039	0.045	0.029	0.019	0.057	0.136
Ca	0.003	0.000	0.000	0.000	0.003	0.003	0.000	0.003	0.003	0.003
Na	0.016	0.011	0.011	0.011	0.011	0.070	0.011	0.011	0.027	0.011
K	0.003	0.004	0.003	0.004	0.003	0.007	0.007	0.004	0.004	0.003
	21.346	21.324	21.305	21.409	21.396	21.393	21.418	21.417	21.410	21.061

Al% IN Z	3.32	1.59	2.60	5.82	2.46	1.11	2.46	2.46	2.33	0.00
CHARGE	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	18.91	18.62	17.44	12.30	17.79	17.27	16.92	17.41	15.83	13.79
Fe	81.01	81.38	82.56	87.70	82.13	82.66	83.08	82.52	84.10	86.13
Ca	0.08	0.00	0.00	0.00	0.07	0.07	0.00	0.07	0.07	0.08

COMPOSITION IN OXIDES

	STT 1	STT 2	STT 3	STT 4	STT 5	STT 6	STJ 1	STJ 2	STJ 3	STJ 4
SiO2	29.06	28.91	28.80	28.78	29.24	29.24	27.23	27.50	27.37	27.57
Al2O3	55.46	55.95	55.32	55.17	56.39	54.67	53.23	54.15	53.48	53.90
TiO2	0.29	0.18	0.25	0.34	0.36	0.02	0.63	0.70	0.72	0.59
MgO	1.70	1.64	1.65	1.79	1.86	1.91	2.06	1.81	2.06	1.68
FeO	11.50	11.49	12.28	12.45	12.17	12.30	13.61	13.34	13.79	13.27
MnO	0.95	0.87	0.92	0.97	0.92	0.95	0.17	0.19	0.16	0.17
CaO	0.00	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00
Na2O	0.02	0.03	0.00	0.02	0.08	0.02	0.03	0.03	0.05	0.02
K2O	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00
	98.99	99.09	99.23	99.54	101.04	99.12	96.96	97.72	97.63	97.20

STRUCT. FORM. BASED ON 46.0 OXYGENS

	STT 1	STT 2	STT 3	STT 4	STT 5	STT 6	STJ 1	STJ 2	STJ 3	STJ 4
Si	7.883	7.830	7.826	7.809	7.796	7.953	7.631	7.628	7.623	7.682
Al	0.117	0.170	0.174	0.191	0.204	0.047	0.369	0.372	0.377	0.318
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	17.613	17.690	17.542	17.450	17.515	17.478	17.213	17.332	17.179	17.384
Ti	0.059	0.037	0.051	0.069	0.072	0.004	0.133	0.146	0.151	0.124
Mg	0.687	0.662	0.668	0.724	0.739	0.774	0.861	0.748	0.855	0.698
Fe2	2.609	2.603	2.791	2.825	2.714	2.798	3.190	3.095	3.212	3.092
Mn	0.218	0.200	0.212	0.223	0.208	0.219	0.040	0.045	0.038	0.040
Ca	0.000	0.003	0.000	0.003	0.003	0.000	0.000	0.000	0.000	0.000
Na	0.011	0.016	0.000	0.011	0.041	0.011	0.016	0.016	0.027	0.011
K	0.003	0.003	0.003	0.003	0.003	0.003	0.000	0.000	0.000	0.000
	21.200	21.213	21.267	21.308	21.295	21.287	21.453	21.382	21.462	21.349

Al% IN Z	1.47	2.12	2.18	2.39	2.55	0.59	4.61	4.64	4.71	3.97
CHARGE	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	19.56	19.10	18.21	19.18	20.18	20.43	21.04	19.25	20.83	18.22
Fe	80.44	80.82	81.79	80.75	79.74	79.57	78.96	80.75	79.17	81.78
Ca	0.00	0.08	0.00	0.08	0.08	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	STJ 5	STJ 6	STJ 7	STJ 8	STJ 9	STJ10	STJ11	STJ12	STM 1	STM 2
SiO2	27.30	27.78	27.76	27.51	28.02	27.89	28.16	28.02	27.93	28.28
Al2O3	53.40	53.21	54.01	53.61	54.48	53.76	54.47	54.24	52.95	51.88
TiO2	0.63	0.56	0.59	0.63	0.47	0.50	0.54	0.50	0.38	0.47
MgO	1.78	1.74	1.67	1.84	1.97	1.94	1.84	1.91	1.76	2.03
FeO	13.56	13.09	13.05	13.38	13.17	13.66	12.89	13.24	13.72	14.10
MnO	0.15	0.16	0.17	0.16	0.19	0.13	0.12	0.15	0.17	0.15
CaO	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na2O	0.02	0.05	0.02	0.02	0.05	0.03	0.03	0.03	0.00	0.03
	96.85	96.59	97.28	97.15	98.35	97.91	98.05	98.09	96.91	96.94

STRUCT. FORM. BASED ON 46.0 OXYGENS

	STJ 5	STJ 6	STJ 7	STJ 8	STJ 9	STJ10	STJ11	STJ12	STM 1	STM 2
Si	7.653	7.783	7.717	7.677	7.708	7.727	7.753	7.729	7.821	7.934
Al	0.347	0.217	0.283	0.323	0.292	0.273	0.247	0.271	0.179	0.066
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	17.294	17.352	17.414	17.310	17.371	17.282	17.427	17.363	17.295	17.089
Ti	0.133	0.118	0.123	0.132	0.097	0.104	0.112	0.104	0.080	0.099
Mg	0.744	0.727	0.692	0.765	0.808	0.801	0.755	0.785	0.735	0.849
Fe2	3.179	3.067	3.034	3.123	3.030	3.165	2.968	3.054	3.213	3.308
Mn	0.036	0.038	0.040	0.038	0.044	0.031	0.028	0.035	0.040	0.036
Ca	0.003	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na	0.011	0.027	0.011	0.011	0.027	0.016	0.016	0.016	0.000	0.016
	21.399	21.328	21.317	21.379	21.377	21.399	21.306	21.358	21.362	21.397

Al% IN Z	4.34	2.72	3.53	4.03	3.65	3.41	3.09	3.38	2.24	0.82
CHARGE	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00	92.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	18.78	18.96	18.36	19.50	20.81	20.05	20.13	20.27	18.42	20.25
Fe	81.15	81.04	81.56	80.50	79.19	79.95	79.87	79.73	81.58	79.75
Ca	0.08	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	STM 3	STM 4	STM 5	STM 6	STM 7	STM 8
SiO2	28.05	27.64	28.10	27.87	28.01	27.99
Al2O3	52.67	52.84	52.71	52.51	52.96	52.65
TiO2	0.54	0.47	0.45	0.39	0.54	0.47
MgO	1.64	1.74	1.93	1.69	1.78	1.78
FeO	13.65	13.66	14.18	13.85	13.75	13.85
MnO	0.16	0.15	0.09	0.17	0.12	0.15
Na2O	0.02	0.00	0.00	0.08	0.03	0.02
	96.73	96.50	97.46	96.56	97.19	96.91

STRUCT. FORM. BASED ON 46.0 OXYGENS

	STM 3	STM 4	STM 5	STM 6	STM 7	STM 8
Si	7.866	7.774	7.840	7.842	7.821	7.844
Al	0.134	0.226	0.160	0.158	0.179	0.156
	8.000	8.000	8.000	8.000	8.000	8.000
Al	17.273	17.290	17.172	17.256	17.250	17.233
Ti	0.114	0.099	0.094	0.083	0.113	0.099
Mg	0.685	0.729	0.803	0.709	0.741	0.743
Fe2	3.201	3.213	3.309	3.259	3.211	3.246
Mn	0.038	0.036	0.021	0.041	0.028	0.036
Na	0.011	0.000	0.000	0.044	0.016	0.011
	21.322	21.368	21.399	21.390	21.359	21.368

Al% IN Z	1.68	2.82	2.00	1.97	2.24	1.95
CHARGE	92.00	92.00	92.00	92.00	92.00	92.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	17.47	18.34	19.42	17.68	18.61	18.47
Fe	82.53	81.66	80.58	82.32	81.39	81.53
Ca	0.00	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	STI 1	STI 2	STI 3	STR 1	STR 2	STR 3	STR 4	STR 5	STR 6	STR 7
SiO2	28.26	28.67	28.47	27.03	28.49	28.54	27.75	27.83	27.93	29.94
Al2O3	54.70	53.72	54.55	54.88	53.97	52.67	52.33	52.55	52.87	53.64
TiO2	0.52	0.57	0.54	0.23	0.47	0.63	0.55	0.50	0.36	0.27
MgO	1.78	1.79	1.62	1.12	1.77	1.70	1.67	1.72	1.54	1.29
FeO	13.40	13.75	13.48	13.69	14.39	14.31	14.49	14.45	14.34	13.76
MnO	0.19	0.19	0.19	0.53	0.17	0.19	0.12	0.08	0.24	0.59
CaO	0.01	0.00	0.00	0.00	0.01	0.01	0.00	0.01	0.01	0.01
Na2O	0.03	0.02	0.02	0.02	0.02	0.13	0.02	0.02	0.05	0.02
K2O	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01	0.01
	98.90	98.72	98.88	97.51	99.30	98.20	96.95	97.17	97.35	99.53

STRUCT. FORM. BASED ON 47.0 OXYGENS

	STI 1	STI 2	STI 3	STR 1	STR 2	STR 3	STR 4	STR 5	STR 6	STR 7
Si	7.903	8.044	7.962	7.699	7.973	8.083	7.973	7.972	7.984	8.322
AL	0.097	0.000	0.038	0.301	0.027	0.000	0.027	0.028	0.016	0.000
	8.000	8.044	8.000	8.000	8.000	8.083	8.000	8.000	8.000	8.322
Al	17.931	17.763	17.941	18.120	17.773	17.581	17.694	17.715	17.795	17.573
Ti	0.109	0.120	0.114	0.049	0.099	0.134	0.119	0.108	0.077	0.056
Mg	0.742	0.749	0.675	0.475	0.738	0.718	0.715	0.734	0.656	0.534
Fe2	3.134	3.226	3.153	3.261	3.368	3.389	3.482	3.462	3.428	3.199
Mn	0.045	0.045	0.045	0.128	0.040	0.046	0.029	0.019	0.058	0.139
Ca	0.003	0.000	0.000	0.000	0.003	0.003	0.000	0.003	0.003	0.003
Na	0.016	0.011	0.011	0.011	0.011	0.071	0.011	0.011	0.028	0.011
K	0.004	0.004	0.004	0.004	0.004	0.007	0.007	0.004	0.004	0.004
	21.984	21.918	21.942	22.049	22.036	21.949	22.057	22.056	22.049	21.519

AL% IN Z	1.22	0.00	0.48	3.77	0.34	0.00	0.34	0.34	0.21	0.00
CHARGE	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	18.91	18.62	17.44	12.30	17.79	17.27	16.92	17.41	15.83	13.79
Fe	81.01	81.38	82.56	87.70	82.13	82.66	83.08	82.52	84.10	86.13
Ca	0.08	0.00	0.00	0.00	0.07	0.07	0.00	0.07	0.07	0.08

COMPOSITION IN OXIDES

	STT 1	STT 2	STT 3	STT 4	STT 5	STT 6	STJ 1	STJ 2	STJ 3	STJ 4
SiO2	29.06	28.91	28.80	28.78	29.24	29.24	27.23	27.50	27.37	27.57
Al2O3	55.46	55.95	55.32	55.17	56.39	54.67	53.23	54.15	53.48	53.90
TiO2	0.29	0.18	0.25	0.34	0.36	0.02	0.63	0.70	0.72	0.59
MgO	1.70	1.64	1.65	1.79	1.86	1.91	2.06	1.81	2.06	1.68
FeO	11.50	11.49	12.28	12.45	12.17	12.30	13.61	13.34	13.79	13.27
MnO	0.95	0.87	0.92	0.97	0.92	0.95	0.17	0.19	0.16	0.17
CaO	0.00	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.00
Na2O	0.02	0.03	0.00	0.02	0.08	0.02	0.03	0.03	0.05	0.02
K2O	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00
	98.99	99.09	99.23	99.54	101.04	99.12	96.96	97.72	97.63	97.20

STRUCT. FORM. BASED ON 47.0 OXYGENS

	STT 1	STT 2	STT 3	STT 4	STT 5	STT 6	STJ 1	STJ 2	STJ 3	STJ 4
Si	8.054	8.000	7.996	7.978	7.965	8.126	7.797	7.794	7.789	7.849
Al	0.000	0.000	0.004	0.022	0.035	0.000	0.203	0.206	0.211	0.151
	8.054	8.000	8.000	8.000	8.000	8.126	8.000	8.000	8.000	8.000
Al	18.116	18.248	18.097	18.004	18.069	17.906	17.761	17.883	17.726	17.936
Ti	0.060	0.037	0.052	0.071	0.074	0.004	0.136	0.149	0.154	0.126
Mg	0.702	0.676	0.683	0.740	0.755	0.791	0.879	0.765	0.874	0.713
Fe2	2.665	2.659	2.851	2.886	2.772	2.859	3.259	3.162	3.282	3.160
Mn	0.223	0.204	0.216	0.228	0.212	0.224	0.041	0.046	0.039	0.041
Ca	0.000	0.003	0.000	0.003	0.003	0.000	0.000	0.000	0.000	0.000
Na	0.011	0.016	0.000	0.011	0.042	0.011	0.017	0.016	0.028	0.011
K	0.004	0.004	0.004	0.004	0.003	0.004	0.000	0.000	0.000	0.000
	21.781	21.848	21.903	21.945	21.932	21.798	22.093	22.021	22.102	21.987

AlZ IN Z	0.00	0.00	0.05	0.27	0.44	0.00	2.54	2.57	2.64	1.88
CHARGE	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	19.56	19.10	18.21	19.18	20.18	20.43	21.04	19.25	20.83	18.22
Fe	80.44	80.82	81.79	80.75	79.74	79.57	78.96	80.75	79.17	81.78
Ca	0.00	0.08	0.00	0.08	0.08	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	STJ 5	STJ 6	STJ 7	STJ 8	STJ 9	STJ10	STJ11	STJ12	STM 1	STM 2
SiO2	27.30	27.78	27.76	27.51	28.02	27.89	28.16	28.02	27.93	28.28
Al2O3	53.40	53.21	54.01	53.61	54.48	53.76	54.47	54.24	52.95	51.88
TiO2	0.63	0.56	0.59	0.63	0.47	0.50	0.54	0.50	0.38	0.47
MgO	1.78	1.74	1.67	1.84	1.97	1.94	1.84	1.91	1.76	2.03
FeO	13.56	13.09	13.05	13.38	13.17	13.66	12.89	13.24	13.72	14.10
MnO	0.15	0.16	0.17	0.16	0.19	0.13	0.12	0.15	0.17	0.15
CaO	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na2O	0.02	0.05	0.02	0.02	0.05	0.03	0.03	0.03	0.00	0.03
	96.85	96.59	97.28	97.15	98.35	97.91	98.05	98.09	96.91	96.94

STRUCT. FORM. BASED ON 47.0 OXYGENS

	STJ 5	STJ 6	STJ 7	STJ 8	STJ 9	STJ10	STJ11	STJ12	STM 1	STM 2
Si	7.819	7.952	7.885	7.844	7.875	7.895	7.921	7.897	7.991	8.107
Al	0.181	0.048	0.115	0.156	0.125	0.105	0.079	0.103	0.009	0.000
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.107
Al	17.844	17.903	17.966	17.861	17.923	17.832	17.980	17.915	17.845	17.528
Ti	0.136	0.121	0.126	0.135	0.099	0.106	0.114	0.106	0.082	0.101
Mg	0.760	0.742	0.707	0.782	0.825	0.819	0.771	0.802	0.751	0.867
Fe2	3.248	3.134	3.100	3.191	3.096	3.234	3.032	3.121	3.283	3.380
Mn	0.036	0.039	0.041	0.039	0.045	0.031	0.029	0.036	0.041	0.036
Ca	0.003	0.000	0.003	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na	0.011	0.028	0.011	0.011	0.027	0.016	0.016	0.016	0.000	0.017
	22.038	21.966	21.954	22.018	22.015	22.038	21.943	21.996	22.001	21.930

Al% IN Z	2.26	0.60	1.44	1.95	1.56	1.31	0.98	1.28	0.12	0.00
CHARGE	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00	94.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	18.78	18.96	18.36	19.50	20.81	20.05	20.13	20.27	18.42	20.25
Fe	81.15	81.04	81.56	80.50	79.19	79.95	79.87	79.73	81.58	79.75
Ca	0.08	0.00	0.08	0.00	0.00	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	STM 3	STM 4	STM 5	STM 6	STM 7	STM 8
SiO2	28.05	27.64	28.10	27.87	28.01	27.99
Al2O3	52.67	52.84	52.71	52.51	52.96	52.65
TiO2	0.54	0.47	0.45	0.39	0.54	0.47
MgO	1.64	1.74	1.93	1.69	1.78	1.78
FeO	13.65	13.66	14.18	13.85	13.75	13.85
MnO	0.16	0.15	0.09	0.17	0.12	0.15
Na2O	0.02	0.00	0.00	0.08	0.03	0.02
	96.73	96.50	97.46	96.56	97.19	96.91

STRUCT. FORM. BASED ON 47.0 OXYGENS

	STM 3	STM 4	STM 5	STM 6	STM 7	STM 8
Si	8.037	7.943	8.010	8.012	7.991	8.014
Al	0.000	0.057	0.000	0.000	0.009	0.000
	8.037	8.000	8.010	8.012	8.000	8.014
Al	17.786	17.840	17.709	17.792	17.799	17.767
Ti	0.116	0.102	0.096	0.084	0.116	0.101
Mg	0.700	0.745	0.820	0.724	0.757	0.760
Fe2	3.271	3.283	3.381	3.330	3.281	3.316
Mn	0.039	0.037	0.022	0.041	0.029	0.036
Na	0.011	0.000	0.000	0.045	0.017	0.011
	21.923	22.007	22.028	22.017	21.998	21.992

AL% IN Z	0.00	0.71	0.00	0.00	0.11	0.00
CHARGE	94.00	94.00	94.00	94.00	94.00	94.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	17.47	18.34	19.42	17.68	18.61	18.47
Fe	82.53	81.66	80.58	82.32	81.39	81.53
Ca	0.00	0.00	0.00	0.00	0.00	0.00

COMPOSITION IN OXIDES

	KYJ 1	KYJ 2	KYJ 3	KYJ 4
SiO2	37.27	37.51	36.77	37.13
Al2O3	63.37	63.07	62.33	62.70
TiO2	0.00	0.00	0.04	0.02
Fe2O3	0.79	0.39	0.66	0.51
MgO	0.03	0.02	0.03	0.02
MnO	0.01	0.04	0.00	0.01
CaO	0.01	0.00	0.01	0.00
Na2O	0.00	0.01	0.03	0.01
K2O	0.00	0.00	0.02	0.02
	101.48	101.04	99.89	100.42

STRUCT. FORM. BASED ON 5.0 OXYGENS

	KYJ 1	KYJ 2	KYJ 3	KYJ 4
Si	0.994	1.003	0.996	0.999
Al	0.006	0.000	0.004	0.001
	1.000	1.003	1.000	1.000
Al	1.985	1.987	1.985	1.998
Ti	0.000	0.000	0.001	0.000
Fe3	0.016	0.008	0.013	0.010
Mg	0.001	0.001	0.001	0.001
Mn	0.000	0.001	0.000	0.000
Ca	0.000	0.000	0.000	0.000
Na	0.000	0.001	0.002	0.001
K	0.000	0.000	0.001	0.001
	2.003	1.997	2.003	2.001

ALZ IN Z	0.63	0.00	0.42	0.07
CHARGE	10.00	10.00	10.00	10.00

TRIANGULAR DIAGRAM Mg - Fe2+Fe3+Mn - Ca

Mg	6.20	8.36	8.15	7.05
Fe	91.57	91.64	89.90	92.95
Ca	1.63	0.00	1.95	0.00

	DLG 17	DLG 18	DLG 19	DLG 20	CCC 1	CCC 2	CCC 3	CCC 4	CCC 5	CCC 6
CaO	28.38	28.11	28.50	28.14	50.54	51.32	51.80	51.39	51.05	51.22
MgO	15.30	16.15	15.98	15.70	1.76	1.48	2.04	1.96	2.04	2.07
MnO	0.85	0.61	0.62	0.79	-	-	-	-	-	-
FeO	7.29	5.44	5.62	5.63	2.42	2.45	2.65	2.91	2.58	2.74
CO_2	47.06	48.01	47.78	47.99	44.69	44.39	43.77	43.88	44.21	44.02
	98.88	98.32	98.50	98.25	99.41	99.64	100.27	100.14	99.88	100.04

Triangular diagram Mg - Fe - Ca

Ca	51.4	51.2	51.7	51.7	92.0	91.8	91.2	91.0	91.2	90.9
Mg	38.5	41.1	40.3	40.2	4.6	4.8	5.2	5.0	5.3	5.3
Fe	10.1	7.7	8.1	8.1	3.4	3.4	3.6	4.0	3.6	3.8

	CCC 7	CCC 8	CCC 9	CCC 10	CCC 11	CCC 12	CCC 13	CCC 14	CCC 15	CCG 1
CaO	51.74	51.32	50.81	51.41	51.51	51.26	52.28	51.42	52.08	55.78
MgO	1.91	1.91	2.04	1.67	2.07	1.72	1.78	1.67	2.00	1.04
MnO	-	-	-	-	-	-	-	-	-	0.46
FeO	2.71	2.64	2.67	2.22	2.69	2.39	2.44	2.36	2.52	0.98
CO_2	43.83	44.09	44.28	44.39	43.89	44.35	43.76	44.30	43.72	42.82
	100.19	99.96	99.80	99.69	100.16	99.72	100.25	99.76	100.32	101.08

Triangular diagram Mg - Fe - Ca

Ca	91.4	91.4	91.0	92.6	91.0	92.2	92.1	92.4	91.5	96.3
Mg	4.9	4.9	5.3	4.4	5.3	4.4	4.5	4.3	5.1	2.3
Fe	3.7	3.7	3.7	3.1	3.7	3.4	3.3	3.3	3.5	1.3

	CCG 2	CCG 3	CCG 4	CCG 5	CCG 6	CCG 7	CCG 8	CCG 9	CCG 10	CCG 11
CaO	54.78	55.99	55.28	55.13	56.66	55.51	55.25	55.79	56.37	55.49
MgO	2.34	1.97	2.15	1.67	1.99	2.12	2.02	1.97	1.41	1.91
MnO	0.61	0.46	0.50	0.57	0.59	0.63	0.56	0.57	0.53	0.59
FeO	1.87	1.79	1.86	1.53	1.60	1.70	1.79	1.70	1.33	1.63
CO_2	42.18	41.84	42.07	42.51	41.51	41.98	42.15	41.93	42.11	42.15
	101.78	102.05	101.86	101.41	102.35	101.94	101.76	101.96	101.75	101.76

Triangular diagram Mg - Fe - Ca

Ca	92.3	93.0	92.6	94.0	93.3	92.6	93.0	93.0	95.0	93.6
Mg	5.4	4.7	5.0	4.0	4.7	5.0	4.7	4.7	3.4	4.4
Fe	2.4	2.3	2.3	2.0	2.0	2.3	2.3	2.3	1.7	2.0

	CCG 12	CCG 13	CCG 14	CCG 15	CCG 16	CCG 17	CCG 18	CCG 19	CCG 20	CCG 21
CaO	55.92	56.45	53.17	52.14	52.20	50.81	50.78	51.14	51.74	51.39
MgO	1.88	1.45	1.99	1.54	1.93	1.88	1.84	1.93	1.60	1.84
MnO	0.56	0.54	0.50	0.49	0.50	0.56	0.47	0.43	0.53	0.54
FeO	1.47	1.40	1.64	1.25	1.59	1.48	1.51	1.47	1.30	1.47
CO_2	42.04	42.00	43.37	44.34	43.94	44.71	44.78	44.60	44.47	44.44
	101.87	101.85	100.67	99.76	100.15	99.45	99.38	99.57	99.64	99.69

Triangular diagram Mg - Fe - Ca

Ca	93.6	95.0	93.0	94.3	93.0	93.3	93.3	93.0	94.3	93.3
Mg	4.4	3.4	4.7	4.0	4.7	4.7	4.7	5.0	4.0	4.7
Fe	2.0	1.7	2.3	1.7	2.3	2.0	2.0	2.0	1.7	2.0

	CCG 22	CCG 23	CCG 24
CaO	51.40	50.60	49.61
MgO	1.73	1.69	1.93
MnO	0.46	0.60	0.50
FeO	1.44	1.47	1.41
CO_2	44.55	44.89	45.39
	99.58	99.25	98.84

Triangular diagram Mg - Fe - Ca

Ca	93.6	93.6	93.0
Mg	4.4	4.4	5.0
Fe	2.0	2.0	2.0

APPENDIX II : THERMODYNAMIC BACKGROUND

The aim of this section is to introduce the equations required for thermochemical calculations in a logical sequence.

For the sake of brevity, source equations are not derived here; for these the reader is referred to textbooks of classical chemical thermodynamics (e.g. Smith, 1973). Comprehensive reviews of the application of thermodynamics to metamorphic petrology are given by Powell (1973), Wood and Fraser (1976) and Holland (1977). Definitions of the terms system, phase, component, intensive variable and extensive variable are set out in Wood and Fraser and will not be repeated here.

(a) Chemical potential

The chemical potential of a component i, μ_i, is a measure of the tendency of that component to migrate. Migration is invariably from regions of high μ_i to regions of low μ_i. If two phases, A and B, each contain n moles of i, and $\mu_i^A > \mu_i^B$, then material will be transferred from A to B, i.e. the reaction:

$$\underset{\text{phase A}}{n\,i} = \underset{\text{phase B}}{n\,i}$$

is driven to the right. At equilibrium, $\mu_i^A = \mu_i^B$, or, in general ,

$$\sum_i \mu_i n_i = 0 \qquad (A.1)$$

The chemical potential is an intensive variable, i.e. independent of the amount of component present.

(b) Gibbs free energy

The free energy of a closed system at constant temperature, T, and pressure, P, (The Gibbs free energy, G) can be defined by:

$$G = U + PV - TS = H - TS \qquad (A.2)$$

where U is the internal energy, H the enthalpy, V the volume and S the entropy. Differentiating (A.2) leads to:

$$dG = dU + PdV + VdP - TdS - SdT \qquad (A.3)$$

However, from the first and second laws of thermodynamics, it can be shown that:

$$dU = TdS - PdV \qquad (A.4)$$

which, when substituted in (A.3), gives:

$$dG = VdP - SdT \qquad (A.5)$$

This expression is very useful as it relates changes in free energy to changes in the intensive variables pressure and temperature. Thus, at constant pressure:

$$\left(\frac{\partial G}{\partial T}\right)_P = -S \qquad (A.6)$$

and at constant temperature:

$$\left(\frac{\partial G}{\partial P}\right)_T = V \qquad (A.7)$$

However, free energy is an extensive property and, in an open system, varies not only with pressure and temperature but also with the amount of material. For a pure phase, the free energy is simply the chemical potential multiplied by the mole number. For a multicomponent system:

$$G = \sum_i \mu_i n_i \qquad (A.8)$$

A more general version of (A.5) is therefore:

$$dG = VdP - SdT + \sum_i \mu_i dn_i \qquad (A.9)$$

Although absolute values of chemical potential and free energy cannot be measured, it is possible to evaluate changes in these quantities associated with changes of state, i.e. chemical reactions:

$$\sum \mu \text{ (products)} - \sum \mu \text{ (reactants)} = \Delta\mu \qquad (A.10)$$

$$\sum G(\text{products}) - \sum G(\text{reactants}) = \Delta G \qquad (A.11)$$

Tabulated free energy values generally refer to ΔG for the reaction:

pure elements - pure phase.

A reaction will proceed spontaneously only if ΔG is negative. At equilibrium:

$$\Delta G = \Delta\mu = 0 \quad \text{(cf. equation (A.1))} \qquad (A.12)$$

(c) The Gibbs-Duhem equation and the phase rule

Differentiating equation (A.8) yields:

$$dG = \sum_i n_i d\mu_i + \sum \mu_i dn_i \qquad (A.13)$$

Substituting for dG in equation (A.9) results in the Gibbs-Duhem equation:

$$VdP - SdT - \sum_i n_i d\mu_i = 0 \qquad (A.14)$$

This equation is important because it relates changes in all the intenstive variables to one another.

For a system containing p phases and c components, one can write p Gibbs-Duhem equations. The total number of intensive variables is 2 + c (=P, T, and C μ's). The number of independent variables, or degrees of freedom, F, is given by the total number of variables less the number of equations that can be written between them, i.e.

$$F = C + 2 - p \tag{A.15}$$

This is the Gibbs phase rule, and has already been employed in Chapter 6.

(d) Heat capacity and the P, T dependence of G and μ

It is generally necessary to know values of G at pressures and temperatures other than those at which they have been evaluated. At any given pressure and temperature, P_1 and T_1, the free energy is given by equation (A.2):

$$G_{P_1T_1} = H_{P_1T_1} + T_1S_{P_1T_1} \tag{A.16}$$

At some higher temperature, T_2, the free energy is given by:

$$G_{P_1T_2} = H_{P_1T_2} + T_2S_{P_1T_2} \tag{A.17}$$

The relationship between $G_{P_1T_1}$ and $G_{P_1T_2}$ can be expressed as:

$$G_{P_1T_2} = G_{P_1T_1} + \int_{T_1}^{T_2} dG = H_{P_1T_1} + \int_{T_1}^{T_2} dH - T_2 \; S_{T_1P_1} + \int_{T_1}^{T_2} dS \tag{A.18}$$

$$\Rightarrow \quad G_{P_1T_2} = G_{P_1T_1} + \int_{T_1}^{T_2} dH - T_2 \int_{T_1}^{T_2} dS \tag{A.19}$$

The temperature dependence of the enthalpy and entropy are given by:

$$\left(\frac{\partial H}{\partial T}\right)_P = C_p \tag{A.20}$$

and $$\left(\frac{\partial S}{\partial T}\right)_P = \frac{C_p}{T} \tag{A.21}$$

where C_p is the heat capacity at constant pressure. Substituting (A.20) and (A.21) into (A.19) gives:

$$G_{P_1T_2} = G_{P_1T_1} + \int_{T_1}^{T_2} C_p dT - T_2 \int_{T_1}^{T_2} \frac{C_p dT}{T} \tag{A.22}$$

If the free energy is required at some higher pressure, P_2, the pressure correction at constant temperature (equation (A.7)) can be applied to give:

$$G_{P_2T_2} = G_{P_1T_1} + \int_{T_1}^{T_2} C_p dT - T_2 \int_{T_1}^{T_2} \frac{C_p dT}{T} + \int_{P_1}^{P_2} V dP \tag{A.23}$$

In order to evaluate the last term in (A.23) it is convenient to distinguish between the volume contribution of the fluid and solid parts of the system. The main reason for this is that molar volumes of solids, V_s, change very little with pressure and temperature (only 2-3% up to 10 kb and 1000 K) and the expansion due to increasing temperature tends to cancel the effects of increasing pressure. Also, the change in ΔV_s for a balanced reaction is proportionally much less than changes in individual V_s values. Consequently, for solids:

$$\int_{P_1}^{P_2} V_s dP = V_s(P_2 - P_1) \tag{A.24}$$

By comparison, fluids are extremely compressible. If the fluid behaves like an ideal gas, it obeys Boyle's law:

$$PV_g = nRT \tag{A.25}$$

Thus, for one mole of ideal gas:

$$\int_{P_1}^{P_2} V_g dP = \int_{P_1}^{P_2} \frac{RT}{P} dP = RT\ln\left(\frac{P_2}{P_1}\right) \tag{A.26}$$

However, fluids are unlikely to behave ideally at geologically realistic pressures (several kilobars) where the finite size of the molecules is no longer negligible. The fugacity, f, of a real gas is defined so that the following relationship always holds:

$$fV_g = nRT \tag{A.27}$$

Consequently:

$$\int_{P_1}^{P_2} V_g dP = RT\ln\left(\frac{f_2}{f_1}\right) \tag{A.28}$$

Fugacity is related to pressure by the fugacity coefficient, Γ:

$$f = P\Gamma \tag{A.29}$$

Substituting (A.24) and (A.28) into (A.23), the free energy of a mixed solid-fluid system at the pressure and temperature of interest is now given by:

$$G_{P_2T_2} = G_{P_1T_1} + \int_{T_1}^{T_2} C_p dT - T_2 \int_{T_1}^{T_2} \frac{C_p dT}{T} + V_s(P_2 - P_1) + RT\ln\left(\frac{f_2}{f_1}\right) \tag{A.30}$$

(e) Activity

The activity of component i, a_i, reflects the difference between the chemical potential of i at the pressure, temperature and composition of interest (μ_i) and the chemical potential of i in the standard state (μ_i^o). It is defined by the equation:

$$\mu_i = \mu_i^o + RT\ln a_i \qquad (A.31)$$

The choice of standard state is purely a matter of convenience. For solids, the standard state is normally chosen as the pure component at the pressure and temperature of interest, so that the activity is merely a factor correcting the chemical potential of component i for its dilution in solid solution. For pure solid phases, $a_i = 1$.

For pure gases the standard state is normally chosen as 1 bar and the temperature of interest (i.e. $\mu_i^o = \mu_{i,\ 1,T}$) so that:

$$\mu_{i\ P,T} = \mu_i^o + \int_1^P d\mu_i \qquad (A.32)$$

$$= \mu_i^o + \int_1^P V_g dP \qquad (A.33)$$

From (A.28):

$$\mu_{i\ P,T} = \mu_i^o + RT\ln\left(\frac{f_P}{1}\right) \qquad (A.34)$$

Thus, for a pure gas (comparing (A.31) and (A.34)),

$$a_i = \left(\frac{f_P}{1}\right) = \left(\frac{P\Gamma}{1}\right) \qquad (A.35)$$

(f) Equilibrium condition

Consider the reaction:

$$3FeAl_2SiO_5(OH)_2 = Fe_3Al_2Si_3O_{12} + 2Al_2O_3 + 3H_2O$$

chloritoid garnet corundum fluid

At equilibrium, from A.1 and A.12;

$$\Delta_r G_{P,T} = 0 = \sum_i \mu_i n_i(\text{products}) - \sum_i \mu_i n_i(\text{reactants}) \qquad (A.36)$$

$$\Rightarrow \quad \Delta_r G_{P,T} = 0 = 3\mu(Fe_3Al_2Si_3O_{12})_{P,T} + 2\mu(Al_2O_3)_{P,T} + 3\mu(H_2O)_{P,T}$$

$$- 3\mu(FeAl_2SiO_5(OH)_2)_{P,T} \qquad (.A.37)$$

Choosing a P,T standard state for all components and substituting (A.31):

$$\Delta_r G_{P,T} = 0 = \Delta G^o_{P,T} + RT\ln\left[\frac{(a^{gt}_{Fe_3Al_2Si_3O_{12}}).(a^{cor}_{Al_2O_3})^2.(a^{fluid}_{H_2O})^3}{(a^{ctd}_{FeAl_2SiO_5(OH)})^3}\right] \qquad (A.38)$$

$$= \Delta G^o_{P,T} + RT\ln K \qquad (A.39)$$

$$\Rightarrow \quad \Delta G^o_{P,T} = -RT\ln K \qquad \text{at equilibrium.} \qquad (A.40)$$

But $\Delta G^o_{P,T}$ can also be expressed in terms of $\Delta G^o_{1,298}$, the free energy change of the reaction involving pure phases at the reference state (1 bar, 298 K) to which most tabulated free energies are tied. Setting T_1 at 298 K and P_1 at 1 bar and substituting (A.30) in (A.11),

$$\Delta G^o_{P,T} = \Delta G^o_{1,T} + \Delta V_s(P - 1) + RT\ln f^n \qquad (A.41)$$

where n is the number of moles of fluid component (3 in this example) and

where:

$$\Delta G^{o}_{1,T} = \Delta G^{o}_{1,298} + \int_{298}^{T} C_p dT - T \int_{298}^{T} \frac{C_p dT}{T} \qquad \text{(cf. A.22)} \qquad \text{(A.42)}$$

Combining (A.39) and (A.41):

$$\Delta G_{P,T} = 0 = \Delta G^{o}_{1,T} + \Delta V_s (P - 1) + RT\ln(K.f^{n}) \qquad \text{(A.43)}$$

This equation forms the basis of most of the calculations in part E, below.

(g) Ideal solutions

Ideal solutions obey Roault's law:

$$P = \sum_i X_i P_i \qquad \text{(A.44)}$$

where X_i is the mole fraction of component i, and $P_i = P.X_i$.

(i) Gases

Lewis and Randall (1923) showed that real gas mixtures approximately obey the relationship:

$$f = \sum_i X_i f_i, \text{ where } f_i = f.X_i \qquad \text{(A.45)}$$

Taking the standard state as the pure component at 1 bar and the temperature of interest and applying (A.34):

$$\mu_{i\ P,T} = \mu^{o}_i + RT\ln(f_{P,T}.X_i) \qquad \text{(A.46)}$$

(h) Solids

The free energy of a solid solution, G^{SS}, can be expressed as:

$$G^{SS} = G^{o} + G^{M} \qquad (A.47)$$

where G^{M} is the free energy of mixing and G^{o} the combined free energy of pure components:

$$G^{o} = \sum_{i} X_i \mu_i^{o} \qquad (A.48)$$

From (A.2), $G^{M} = H^{M} - TS^{M}$ (A.49)

By definition, an ideal solution has no enthalpy of mixing, H^{M}, so that:

$$G^{M} = -TS^{M} \qquad (A.50)$$

A statistical treatment of the entropy of mixing, S^{M}, on n sites yields:

$$S^{M} = -nR\sum_{i} X_i \ln X_i. \quad \text{(Wood and Fraser, 1976, p. 84)} \qquad (A.51)$$

Substituting in (A.47):

$$G^{SS} = \sum_{i} X_i \mu_i^{o} + nRT\sum_{i} X_i \ln X_i \qquad (A.52)$$

From (A.8) (cf. (7.48)),

$$G^{SS} = \sum_{i} X_i \mu_i \qquad (A.53)$$

Bearing in mind that $\sum_{i} X_i = 1$, this can be rearranged to give

$$\mu_i = G^{SS} + (1 - X_i)\left(\frac{\partial G^{SS}}{\partial X_i}\right)_{P,T} \qquad (A.54)$$

Differentiating (A.52) with respect to X_i and substituting (A.54) yields:

$$\mu_i = \mu_i^{o} + nRT\ln X_i \qquad (A.55)$$

Comparing this with (A.31) it can be seen that, for a P,T standard state:

$$a_i = X_i^n \qquad \text{(A.56)}$$

Thus, the activity of a component in a multisite crystalline ideal solution, for example pyrope in garnet, is given by:

$$a^{gt}_{Mg_3Al_2Si_3O_{12}} = (X^{VIII}_{Mg})^3.(X^{VI}_{Al})^2.(X^{IV}_{Si})^3 \qquad \text{(A.57)}$$

where IV, VI and VIII refer to tetrahedral, octahedral and cubic sites, respectively.

(h) Non-ideal solutions

Many solid solutions are non-ideal and do not obey (A.56). The departure from ideality is expressed by the activity coefficient, γ, defined as:

$$\mu_i = \underbrace{\mu_i^o + RT\ln X_i}_{\text{ideal term}} + \underbrace{RT\ln\gamma_i}_{\text{excess term}} \qquad \text{(A.58)}$$

where $$X_i\gamma_i = a_i \qquad \text{(A.59)}$$

For multi-site solutions (e.g. pyrope in garnet) one can write:

$$a^{gt}_{Mg_3Al_2Si_3O_{12}} = (X^{VIII}_{Mg}.\gamma^{VIII}_{Mg})^3.(X^{VI}_{Al}.\gamma^{VI}_{Al})^2.(X^{IV}_{Si}.\gamma^{IV}_{Si})^3 \qquad \text{(A.60)}$$

Non-ideality can also be expressed in terms of an excess free-energy G^{XS}, defined as:

$$G^{SS} = G^o + (nRT\sum_i x_i \ln x_i) + G^{XS} \qquad \text{(A.61)}$$

where $G^{SS} - G^{XS}$ is the free energy of an ideal solution.

The relationship between γ and G^{XS} can therefore be expressed as:

$$RT\ln\gamma_i = \left(\frac{\partial G^{XS}}{\partial n_i}\right) \qquad (A.62)$$

(i) Regular solutions

In the regular solution model, G^{XS} is assumed to be a symmetrical function of composition. For a binary solid solution $(A, B)_nO$,

$$G^{XS} = X_A X_B W^G \qquad (A.63)$$

W^G, the interchange-free energy, is zero in ideal solutions. It is generally dependent upon pressure and temperature, and can be subdivided into interchange-enthalpy (W^H), -entropy (W^S) and -volume (W^V):

$$W^G = W^H - TW^S + (P - 1)W^V \qquad (A.64)$$

Differentiating (A.63) with respect to n_a (noting that $X_A = n_A/n_A + n_B$ and $X_B = n_B/n_A + n_B$),

$$RT\ln\gamma_A = (1 - X_A)^2 W = X_B^2 W \qquad (A.65)$$

In a ternary regular solution, it is assumed that the excess free energy is the sum of the interchange free energy terms for each pair of substituting atoms:

$$G^{XS} = (X_A X_B W_{AB}) + (X_A X_C W_{AC}) + (X_B X_C W_{BC}) \qquad (A.66)$$

Differentiating, as before:

$$RT\ln\gamma_A = X_B(1 - X_A)W_{AB} + X_C(1 - X_A)W_{AC} - X_B X_C W_{BC} \qquad (A.67)$$

(ii) Subregular solutions

The subregular solution model is more general as it allows for the possibility of excess functions being asymmetric, i.e. when the interchange energy between a pair of dissimilar atoms A, B is itself composition dependent. Thus, for a binary solution,

$$G^{XS} = X_A X_B (W_1 X_A + W_2 X_B) \qquad (A.68)$$

where W_1 and W_2 are the interchange energies of an A-B pair in pure A_nO and B_nO respectively. Differentiating, as above, leads to

$$RT\ln\gamma_A = X_B^2 (W_1 + 2X_A(W_2 - W_1)). \qquad (A.69)$$

(i) Solvi

The presence of a solvus in a binary system A_nO-B_nO must be the result of non-ideality as it implies that the curve of G^{SS} has two minima in composition-space, not one as predicted by (A.52). At equilibrium, the chemical potential of component A in two coexisting phases y and z is the same:

$$\mu_A^y = \mu_A^z \qquad (A.70)$$

Substituting (A.58) and (A.65) into (A.70):

$$\mu_A^o + RT\ln X_A^y + (1 - X_A^y)^2 W = \mu_A^o + RT\ln X_A^z + (1 - X_A^z)^2 W \qquad (A.71)$$

$$\Rightarrow W = \left(\frac{RT\ln(X_A^z / X_A^y)}{(1 - X_A^y)^2 - (1 - X_A^z)^2} \right) \qquad (A.72)$$

The temperature dependence of W can be obtained by solving (A.72) at a number of temperatures and calculating W^S and W^H from (A.64). The pressure dependence, W^V, can be calculated from:

$$W^V = \frac{V^{XS}}{X_A X_B} \qquad (A.73)$$

V^{XS}, the excess volume of mixing, is measurable and given by:

$$V^{XS} = V^{SS} - V^o \qquad \text{(cf. A.61)} \qquad (A.74)$$

where V^{SS} is the molar volume of a homogeneous phase $(A_xB_{1-x})_nO$, and V^o, the molar volume of a mechanical mixture of end-members of the same composition.

A similar procedure can be used to extract interchange energies from an asymmetric solvus, but it is necessary to solve for W_1 and W_2 simultaneously.

For an asymmetric solvus, substituting (A.58) and (A.69) into (A.70),

$$\mu_A^o + RT\ln X_A^y + (1 - X_A^y)^2 W_1 + 2X_A^y(W_2 - W_1) \qquad (A.75)$$

$$= \mu_A^o + RT\ln X_A^z + (1 - X_A^z)^2 W_2 + 2X_A^z(W_1 - W_2) \; .$$

APPENDIX III : OXYGEN ISOTOPE ANALYSES OF MINERAL SEPARATES

A. SHORT NOTES ON ANALYSED ROCKS

GD 60 Meta-tonalite

Locality : Alt 1810 m in floor of Zwenberger Tal.

Assemblage : biot + ep + ksp + plag + qz + musc + chl + ap + zirc.

Texture : Dark coarse poorly-foliated rock rich in biotite. Large blocky "gefüllte" plagioclases are abundant, but K-feldspar is rare, untwinned and embayed by myrmekitic intergrowths of plagioclase and quartz. Most of the quartz forms a fine-grained polygonal mosaic between plagioclase grains. Muscovite and epidote are only important as inclusions in "gefüllte" plagioclase.

GD 449 Meta-tonalite

Locality : Alt 2418 m, 500 m south-south-west of the Villacher Hütte, Ankogel group.

Assemblage : biot + ep + ksp + plag + qz + musc + chl + ap.

Texture : as for GD 60.

GD 417 Meta-tonalite

Locality : Alt 1970 m, 1.3 km south of Untere Moos Hütte, Riekengraben.

Assemblage and texture not examined in thin section. A dark, moderately well-foliated rock in which biotite, quartz and feldspar occur in roughly equal proportion.

GD 475 Granodiorite

Locality : Koschach quarry, Maltatal.

Assemblage and texture not examined in thin section. Virtually unfoliated.

GD 468 Leucogranite

Locality : Alt 2195 m, 700 m west-north-west of old Giessener Hütte site, Gössgraben.

Assemblage : gt + biot + musc + ksp + plag + qz + chl + ap.

Texture : see Appendix I .

GD 469 Granodiorite

Locality : Alt 2000 m, 700 m south-east of old Giessener Hütte site, Gössgraben.

Assemblage and texture not examined in thin section. Virtually unfoliated.

GD 595 Augen-gneiss

Locality : Groppensteinerfall, Raufen.

Assemblage : ep + biot + musc + ksp + plag + qz + all + chl + cc + ap.

Texture : A strongly lineated rock. Biotite and muscovite aligned in both S_a and S_a^2, occur in roughly equal proportions. K-feldspar forms orthoclase augen. Abundant rounded clear albites. Yellow epidotes, sometimes with allanitic cores. Quartz forms polycrystalline patches, elongated in $?S_a^1$, in which individual grains are strained and grain boundaries are sutured.

GD 596 Augen-gneiss

Locality : Alt 770 m on road to Kaponigtal, Neuhäusler, Obervellach.

Assemblage and texture not examined in thin section. A strongly schistose rock in which phengitic muscovite and quartz are both very abundant. Biotite is subordinate, and K-feldspar is not as abundant as in GD 595.

GD 1226 Biotite granite-gneiss

Locality : On footpath, 800 m east-north-east of summit of Romate Spitze, Mallnitzer Tauern.

Assemblage : biot + musc + ep + ksp + plag + qz + chl.

Texture : A moderately well-foliated rock in which biotite, white feldspar and purplish quartz are all abundant. In thin section, large orthoclase perthite crystals have been embayed by myrmekitic intergrowths of plagioclase and quartz. Small recrystallised grains and large "gefüllte" grains of plagioclase are abundant. Epidote and muscovite only occur as inclusions within the latter.

GD 334 Augen-gneiss

Locality : Summit of Kampleck, Reisseck group.

Assemblage and texture not examined in thin section. A well-foliated rock in which quartz forms discrete 1-2 mm thick layers. Biotite and muscovite occur in approximately equal proportions.

GD 532 Granite-pegmatite

Locality : Loose block. Alt 2300 m, 300 m south of Hochkedl, Reisseck group.

Assemblage : biot + plag + qz + musc.

Texture : Very coarse grained. 2cm+ crystals of plagioclase (An 20) and quartz and monomineralic aggregates of biotite occur in

roughly equal proportions. Muscovite is rare and appears to be secondary.

GD 642 Inner Schieferhülle mica-schist

Locality : Alt 1680 m, 40 m east of footpath, 1.1 km south-west of summit of Hindenburg Höhe, Mallnitzer Tauern.

Assemblage and texture not examined in thin-section. A coarse-grained mica-schist rich in both biotite and muscovite. Together the micas make up about 40 to 50% of the rock, the rest being formed of fine grained quartz and feldspar.

GD 1230a Inner Schieferhülle mica-schist

Locality : Alt 2425 m, 200 m south-south-west of summit of Klein Woisgen Kopf, Mallnitzer Tauern.

Assemblage : gt + biot + chl + musc + plag + qz + ore + ap.

Texture : Small (~0.5 mm) garnets have dusty ?M_v cores and clear ?M_a rims. Abundant biotite, muscovite and rare chlorite are aligned parallel to S_a^1. Quartz stringers (up to 1 mm thick) composed of unstrained polygonal grains. Plagioclase is in equilibrium with micas (Fig. 2.14a).

GD 1230b Quartz + garnet vein from Inner Schieferhülle mica-schist

Locality : As for GD 1230a.

Assemblage : qz + gt + chl + biot + musc.

Texture : Quartz and garnet are both very abundant, the sheet silicates much less so (see Fig. 2.16 for modal analysis). At least some of the garnets show evidence of having grown in a foliated rock and cannot be primary vein precipitates (Fig. 2.15). These must be M_v garnets as the vein itself has been folded by F_a^1.

GD 302 Inner Schieferhülle mica-schist

Locality : Alt 2600 m, 200 m south-east of summit of Lawogge, Reisseck group.

Assemblage : gt + chl + musc + mt + ep + plag + qz + ap.

Texture : Small euhedral M_a garnets, possibly pre-S_a^2. Quartz is rare and forms thin (<0.25 mm) stringers. Magnetite is very abundant (>5%) and forms pre-S_a^2 augen. Plagioclase is replacing muscovite.

GD 28b Peripheral Schieferhülle pelitic mica-schist

Locality : On road, 650 m south-west of basement-cover interface, Zwenberger Tal.

Assemblage : gt + chl + musc + qz + plag + ep + mt + ap.

Texture : Sporadic large pre- or syn-S_a^2 garnets. Abundant muscovite and chlorite aligned in S_a^2. Thin (<0.5 mm) stringers of recrystallised quartz. Numerous small euhedral magnetite octahedra.

GD 575 Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1560 m on road 600 m east of Jamnig Alm, Tauern Tal.

Assemblage and texture not studied in thin-section. A uniformly fine-grained schist rich in muscovite, chlorite and euhedral magnetite. Very little quartz.

GD 234a Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1406 m on road up west side of Mühldorfer Graben.

Assemblage : gt + chl + musc + qz + ep + plag + mt + tour + ap.

Texture : As for 28b.

GD 498c Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1500 m, 1.1 km south-south-west of Mallnitz.

Assemblage : gt + chl + musc + qz + ep + plag + mt + ap.

Texture : Rare 1 mm diameter euhedral garnets. Numerous < 1.5 mm thick stringers of recrystallised quartz. Muscovite and chlorite are aligned in S_a^1. Cross-cutting euhedral magnetites.

GD 573 Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 1540 m on road 800 m east of Jamnig Alm, Tauern Tal.

Assemblage and texture not examined in thin-section. Chlorite, muscovite, quartz and magnetite recognisable in hand-specimen. Quartz is largely confined to 0.5-1.5 mm thick isoclinally folded veins in S_a^1. Muscovite and chlorite are also aligned in S_a^1 but have been tightly crenulated by F_a^2.

GD 992a Peripheral Schieferhülle micaceous quartzite

Locality : Alt 1665 m, 1.1 km south-east of Jamnig Alm, Tauern Tal.

Assemblage : musc + qz + mt + ep + biot + chl + ilm + ap.

Texture : Quartz (over 50%) forms 1-2 mm thick lenses and stringers aligned in S_a^1 and parallel to muscovite flakes (over 40%). Magnetite (1.2%) forms euhedral crystals. All other minerals occur in accessory amounts.

GD 890 Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 2100 m, 1.5 km west of Jamnig Alm, Tauern Tal.

Assemblage : gt + biot + chl + musc + qz + plag + ilm + ap.

Texture : Muscovite flakes are aligned in S_a^1 but have been strongly crenulated by F_a^2. Some distinctly bent. Biotite flakes mostly aligned parallel to F_a^2 axial planes. Ilmenites

occur as small tablets mostly in S_a^1. Euhedral inclusion-free pre-F_a^2 garnets are locally altering to chlorite. Plagioclases have pre-F_a^2 albite cores and ragged post-F_a^2 oligoclase rims. Quartz forms 2 mm thick stringers parallel to S_a^1.

GD 903 Peripheral Schieferhülle pelitic mica-schist

Locality : Alt 2540 m, 100 m east of summit of Greil Kopf, Mallnitzer Tauern.

Assemblage : ep + chl + musc + qz + plag + mt + ap.

Texture : Chlorite and muscovite are aligned in S_a^1. Quartz forms 1 cm thick veins parallel to S_a^1. Magnetites are euhedral.

GD AU 1 Chlorite + albite vein

Locality : Loose block from summit of Auernig, Reisseck group.

Assemblage : chl + ab + rut + sph.

Texture : Coarse-grained. Albite, rutile and sphene all euhedral to chlorite, Sphene rims rutile.

GD AU 2 Chlorite + albite vein

Locality : As for GD AU 1.

Assemblage : chl + ab + qz + ilm.

Texture : Coarse-grained. Albite, quartz, and ilmenite all euhedral to chlorite.

B. ANALYTICAL TECHNIQUES

(a) Preparation of samples

Fresh 1 kg specimens were crushed, sieved and separated into constituent minerals using standard magnetic and heavy liquid techniques as described in Allman and Lawrence (1972). Separations were mostly performed on 40/60- and 60/80-mesh sieve fractions, but coarser fractions were used when hand-picking was necessary. Quartz and garnet separates were cleaned of feldspar, chlorite and micas by immersion for 2 to 5 mins. in concentrated hydrofluoric acid. Mica separates were cleaned of non-sheet-silicate impurities by grinding for 1 to 2 mins. in an agate mortar. No success was achieved in separating feldspars from quartz other than by hand-picking, so in general they were neglected. Purity was estimated as better than 98% for micas and ilmenite, and better than 99% for quartz, feldspar, garnet and rutile.

(b) Extraction and analysis

The oxygen was liberated from the minerals by reaction with bromine pentafluoride and converted to carbon dioxide for collection and measurement using the method of Clayton and Mayeda (1963). Garnet, ilmenite, rutile and magnetite samples were powdered to improve reaction yields and all samples were oven-dried prior to extraction. 15 to 20 mg aliquots were left to react with BrF_5 for 12 hrs. at temperatures between 550 and 600°C for quartz, feldspar and micas, and between 670 and 700°C for the rest. Reaction yields ranged from 97% to 110%, those above 103% usually from magnetites. Samples were run in duplicate and repeated until an agreement of 0.3‰ or better was obtained.

The isotopic composition of the CO_2 was measured using a 60°, 15 cm single focussing mass-spectrometer with an analytical reproducibility better than ±0.05‰.

C. ANALYTICAL DATA

Temperatures in brackets have been calculated using the method of Javoy et al. (1970). w.r. = calculated whole-rock composition.

Zentralgneis Lithologies

		δO^{18}		ΔO^{18}	T (°C)
GD 60	qz	+10.6	qz-biot	5.2	540
	biot	+ 5.4			
	w.r.	+ 8			
GD 449	qz	+11.3	qz-biot	5.8	510
	biot	+ 5.5			
	w.r.	+ 9			
GD 417	qz	+10.55	qz-biot	5.05	550
	biot	+ 5.5			
	w.r.	+ 9			
GD 475	qz	+ 8.85	qz-biot	5.9	500
	biot	+ 2.95			
	w.r.	+ 7			
GD 468	qz	+10.2	qz-biot	5.7	510
	fsp	+ 8.4	qz-fsp	1.8	
	musc	+ 7.3	qz-musc	2.9	610
	biot	+ 4.5	fsp-biot	3.9	500
	w.r.	+ 9	musc-biot	2.8	490
					(530)

		δO^{18}		ΔO^{18}		T (°C)
GD 469	qz	+10.55	qz-biot	5.05		548
	biot	+ 5.5				
	w.r.	+ 7				
GD 595	qz	+11.4	qz-biot	6.5		470
	musc	+ 8.1	qz-musc	3.3		510
	biot	+ 4.9	musc-biot	3.2		460
	w.r.	+ 9				(470)
GD 596	qz	+11.0	qz-biot	6.0		480
	musc	+ 8.0	qz-musc	3.0		580
	biot	+ 4.7	musc_biot	3.3		460
	w.r.	+ 9				(510)
GD 1226	qz	+10.0	qz-biot	6.3		480
	biot	+ 3.7				
	w.r.	+ 8				
GD 334	qz	+ 9.05	qz-biot	5.55		520
	musc	+ 7.35	qz-musc	1.7		1290
	biot	+ 3.5	musc-biot	3.85		430
	w.r.	+ 8				
GD 532	qz	+ 8.9	qz-biot	4.1		610
	plag	+ 7.8	plag-biot	3.0		550
	biot	+ 4.8	qz-plag	1.1		1220

Inner Schieferhülle Lithologies

		δO^{18}		ΔO^{18}	T (°C)
GD 642	qz	+12.4	qz-biot	6.1	490
	musc	+ 9.7	qz-musc	2.7	680
	biot	+ 6.3	musc-biot	3.4	450
	w.r.	+10			(550)
GD 1230a	qz	+13.0	qz-biot	6.4	480
	musc	+10.3	qz-musc	2.7	680
	gt	+ 8.75	qz-gt	4.25	600
	biot	+ 6.6	musc-biot	3.7	440
	w.r.	+11	gt-biot	2.15	420
					(560)
GD 1230b	qz	+11.95	qz-gt	-0.2	∞
	gt	+12.15			
	w.r.	+12			
GD 302	qz	+10.75	qz-musc	3.1	550
	musc	+ 7.65	qz-mt	10.2	560
	mt	+ 0.55	musc-mt	7.1	560
	w.r.	+ 7			(510)

Peripheral Schieferhülle Lithologies

GD 28b	qz	+12.5	qz-musc	3.0	580
	musc	+ 9.5	qz-mt	10.6	540
	mt	+ 1.9	musc-mt	7.6	530
	w.r.	+10			(510)

		δO^{18}		ΔO^{18}	T (°C)
GD 575	musc	+13.35	musc-mt	8.45	480
	mt	+ 4.9			
	w.r.	+12			
GD 234	qz	+15.2	qz-musc	2.7	680
	musc	+12.5	qz-gt	4.85	510
	gt	+10.35	qz-mt	9.6	590
	mt	+ 5.6	musc-mt	6.9	570
	w.r.	+14	gt-mt	4.75	630
					(550)
GD 498c	qz	+15.0	qz-musc	3.35	500
	musc	+11.65	qz-mt	10.3	550
	mt	+ 4.7	musc-mt	6.95	570
	w.r.	+13			(480)
GD 573	qz	+14.4	qz-musc	2.4	800
	musc	+12.0	qz-mt	9.85	570
	mt	+ 4.55	musc-mt	7.45	540
	w.r.	+12			(520, qz-mt)
GD 992	qz	+19.05	qz-musc	1.8	1200
	musc	+17.25	qz-mt	10.4	550
	mt	+ 8.65	musc-mt	8.6	470
	w.r.	+18			(490, qz-mt)

		δO^{18}	ΔO^{18}		T (°C)
GD 890	qz	+15.35	qz-musc	2.75	660
	musc	+12.6	qz-gt	5.35	450
	gt	+10.0	qz-biot	6.05	490
	biot	+ 9.3	qz-ilm	9.6	540
	ilm	+ 5.75	musc-biot	3.3	460
	w.r.	+13	musc-ilm	6.85	520
			gt-biot	0.7	520
			gt-ilm	4.25	600
					(510)
GD 903	qz	+14.25	qz-musc	2.15	940
	musc	+12.1	qz-mt	10.05	560
	mt	+ 4.2	musc-mt	7.9	510
					(510, qz-mt)

Late Stage Veins

GD AU 1	ab	+10.75	ab-rut	6.5	530
	rut	+ 4,25			
	sph	+ 7.05			
GD AU 2	qz	+12.85	qz-ab	2.65	340
	ab	+10.2	qz-ilm	10.05	520
	ilm	+ 2.8	ab-ilm	7.4	560
GD AU 1 and GD AU 2 combined using GD AU 1 albite	qz	+12.85	qz-ab	2.1	450
	ab	+10.75	qz-rut	8.6	520
	rut	+ 4.25	qz-ilm	10.05	520

		δO^{18}	ΔO^{18}		T (°C)
GD AU ≪ (Cont.)	ilm	+ 2.8	ab-rut	6.5	530
			ab-ilm	7.95	530
			rut-ilm	1.45	520