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The Altkristallin of the Kreuzeck Mountains, SE Tauern Window, Eastern Alps – Basement Crust in a Convergent Plate Boundary Zone

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With 48 Text-Figures, 9 Tables and 5 Plates

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Kärnten Kreuzeck Gruppe Tauern-Gebiet Austroalpin Altkristallin

Kretazische Metamorphose Variszische Metamorphose

Extensive Deformationszone

Geochronologische Altersdaten

Zusammenfassung

In dieser Arbeit werden detailierte Struktur-, Metamorphose und geochronologische Daten vom Altkristallin (Kreuzeck-Gruppe), SE Tauern Gebiet präsentiert. Die Untersuchungen zeigen, daß sich das Altkristallin während der Alpinen Orogenese nicht als einheitliche, steife Deckeneinheit verhalten hat, sondern eine Krustenverdickung erfahren hat, die zu einer Versenkungsmetamorphose in der mittleren Kreide führte, gefolgt von einer Krustenverdünnung, während der die kretazische Metamorphose-Abfolge stark ausgedünnt wurde.

Im Untersuchungsgebiet, in der nördlichen Kreuzeck-Gruppe, treten zwei Einheiten auf. Sie unterscheiden sich in Gesteinszusammensetzung, struktureller und metamorpher Entwicklung und unterschiedlichen geochronologischen Altersdaten. In der nördlichen Polinik-Einheit hat eine Phase der Kru-(D₂-Deformation) stenverdickung zu einer Versenkungsmetamorphose geführt, die maximale Temperaturen von 620±60°C bei 6,25±1,25 kb Druck um ca. 100 Ma, erfahren hat. In der südlich gelegenen Strieden-Einheit hat die kretazische Metamorphose die Bedingungen der Grünschieferfazies nicht überschritten und überprägt eine variszische Hochtemperatur-Metamorphose. Eine von hohen Strain-Raten und dem Vorkommen von Myloniten gekennzeichnete Deformationszone (Dp-Deformation) trennt die beiden Einheiten, und befindet sich an der Basis der Strieden-Einheit. Bewegungsanzeiger, Metamorphose-Sachverhalte und geochronologische Daten deuten darauf hin, daß die Strieden-Einheit die Polinik-Einheit in NW-Richtung überfahren hat, während die beiden Einheiten von der kretazischen Metamorphose abkühlten (90-60 Ma). Dabei wurden 2-8 km der kretazischen Metamorphoseabfolge herausgeschnitten.

Diese Ergebnisse werden präsentiert und im größeren Rahmen der alpinen geologischen Entwicklung in den Ostalpen diskutiert.

Abstract

This study presents detailed structural, metamorphic and geochronological data from the Altkristallin (Kreuzeck Mountains) in the SE Tauern area. Evidence is presented which shows that the Altkristallin did not behave as a rigid body during Alpine convergence in the Cretaceous, but has experienced crustal thickening resulting in mid Cretaceous burial metamorphism, followed by post-metamorphic crustal thinning, which substantially thinned the Cretaceous metamorphic sequence.

In the study area two units can be distinguished, which differ greatly in lithologies, structural and metamorphic history and geochronological age data. In the north the Polinik Unit was affected by crustal thickening (D₂ deformation) prior to the burial metamorphism which reached peak temperatures of 620±60°C at 6.25±1.25 kb pressure at ca. 100 Ma. In the south, in the Strieden Unit, the Cretaceous metamorphism did not exceed lower greenschist facies conditions, and overprints a Variscan high temperature metamorphism. A discrete zone of high strain deformation (D_p deformation), characterised by the wide-spread occurrence of mylonites, which formed under greenschist facies conditions (main mylonite zone - MMZ), separates the two units and forms the base of the Strieden Unit. Kinematic indicators, metamorphic evidence and geochronological data, suggest that the Strieden Unit moved over the Polinik Unit in a NW direction, cutting out 2-8 km of the Cretacous metamorphic sequence. This occurred at and following the time when the two units passed through the K/Ar mica blocking temperatures at 90-80 Ma, and continued until 60 Ma.

These results are presented and discussed within the wider context of the Alpine orogeny.

1. Introduction

This publication presents a detailed study of part of a large allochthonous crystalline basement thrust sheet, referred to as the Altkristallin, in the Eastern Alps of Austria. The structural and metamorphic history of this substantial tectonic unit has important implications for our understanding of the behaviour of deeper crustal levels in zones of continental collision.

The Eastern Alps form part of a belt of intense deformation between the European Plate and the Adriatic region (Adria). It is not clear whether Adria acted as an



Text-Fig. 1.

Generalized geological map of the European Alps showing extent of the Austroalpine and Pennine units and major structural features. TW = Tauern Window; PL = Periadriatic Lineament; S = Sesia Zone; RG = Rhine Graben; BM = Bohemian Massif.



Text-Fig. 2. Generalized geological map of the Eastern Alps, showing extent of the Altkristallin and its Palaeozoic and Mesozoic cover rock sequences.

independent microplate or as a promontary of the African plate (PLATT et al., 1989; DEWEY et al., 1973, 1989). From the Permian to early-mid Cretaceous, extension in the Tethyan region produced a broad zone of oceanic and thinned continental crust, referred to as the Pennine domain, between the two major plates (FRISCH, 1979). During subsequent convergence between Adria and Europe, the basement and its sedimentary cover were shortened, with the emplacement of the southern continental margin (Austroalpine nappes including the Altkristallin) onto the oceanic Pennine domain (Fig. 1). Continental collision between Adria and the European continental margin (Bohemian massif), marked by the onset of deformation in sediments on the stable European foreland, occurred in the Tertiary (OBERHAUSER, 1980; PLATT, 1987).

Allochthonous and para-allochthonous units can be studied best in tectonic windows, such as the Tauern Window, in the axial zone of the Eastern Alps. The largest of these allochthonous sheets is the Altkristal-lin, consisting of pre-Mesozoic crystalline rocks, and forming a more or less continuous E-W trending belt, ca. 500 km long and 50–130 km wide, extending the entire length of the Eastern Alps into Switzerland (Fig. 2). Mesozoic metasedimentary rocks extend beneath the Altkristallin for at least 120 km in a N–S direction.

2. The study area

The study area lies in the SE corner of the Tauern Window and is entirely within the Altkristallin rocks of the Kreuzeck mountains (Fig. 3). In the north, a major subvertical NW trending fault, which follows part of the Möll valley (Mölltal fault zone), separates the Altkristallin from the Pennine Units, exposed in the Tauern Window. South (15-20 km) of the edge of the Tauern Window, the Altkristallin is overlain by Palaeozoic phyllites, which in turn are unconformable overlain by the Permian and Mesozoic sediments of the Drauzug.

The previous work in the SE corner of the Tauern Window has been mainly geochronological studies, both within the Altkristallin and also across the margin of the Tauern Window (OXBURGH et. al., 1966; BREWER, 1969, 1970; LAMBERT, 1970; HAWKESWORTH, 1976; WA-TERS, 1976; DEUTSCH, 1984). The pattern of geochronological ages will be described in detail in Part 5.

Major structural subdivision of the Study area

Lithological layering, prominent schistosity, major fault and fold structures trend E to SE. The prominent foliation generally dips towards the south, so that on a km-scale the northern parts of the area are at structurally deeper levels, with respect to the foliation, than the southern parts.



Text-Fig. 3. Generalized geological map of the south-east Tauern Window, showing location of the study area.



An important zone of deformation within the Altkristallin, characterised by the widespread occurrence of mylonites (Fig. 4), trends WNW and marks the boundary between two distinct units which differ greatly in lithology, deformation and thermal history, with different K/Ar ages. Concordant K/Ar mica ages of c. 80 Ma are found north of the mylonites, whereas within them and further south, ages become progressively older towards the south, ranging between 100- 320 Ma (Fig. 41). The northern unit is called the Polinik Unit, named after the prominent mountain, Mt. Polinik (2784 m). The southern unit is called the Strieden Unit after Mt. Strieden (2682 m). The boundary between the two units is at the northern limit of the mylonites. This is marked in the study area by subvertical and easttrending, predominantly brittle, fault zones, which rework the mylonites along their length. These are referred to as the Ragga - Teuchl fault zone in the east and the Strugenkopf fault zone in the west (Fig. 4).

The organisation of this paper is based on the structural division described above. The geology of the Polinik and Strieden Units is described in sections 3 and 4 respectively. In a concluding section, this is discussed within the wider context of the geological history of the Eastern Alps.

3. The Polinik Unit, its Structural and Metamorphic History

3.1. Lithologies in the Polinik Unit

Medium grained (1-5 mm) quartzo-feldspathic schists and gneisses, referred to as the Polinik Schists, predominate. They contain discontinuous bodies of amphibolite, augengneiss and pegmatite. Garnet, staurolite and kyanite-bearing schists (aluminous metapelites) are rare, found only in the Mt.Polinik and Griessbergalm areas.

Augengneisses

Biotite augengneisses contain large (up to 5 cm long) and numerous K-feldspar augen, aligned within the foliation. They have been found in only one locality, about 500 m south of the Weissenstein Alm, forming tens of metres of exposure along a forest road, and apparently are concordant with the main schistosity.

Amphibolites

Amphibolites can be subdivided into two groups, depending on whether they contain clinopyroxene (cpx). Cpx-rich amphibolites contain abundant garnet, which are enclosed in pale green symplectites after clinopyroxene, with relict omphacite. These rocks have been derived from eclogites with the original mineral assemblage omph - gt - rut - qz (see appendix A for mineral abbreviations), preserved as relicts in the cores of amphibolite pods, where the rock is massive and contains randomly oriented pale green cm-sized columnar pseudomorphs of omphacite, surrounded by dark greeen amphibole, plagioclase and irregularly shaped garnet (0.2-1 cm in diametre). Margins of amphibolite pods are rich in biotite and oligoclase blasts which overgrow the contact with the surrounding schists and gneisses.

Unusual amphibolite pods are found along the Mörnigköpfe ridge, containing the texturally equilibrated mineral assemblage qt - cpx (di) - hb - $qz \pm plaq$.

Amphibolites, which do not contain clinopyroxene, contain garnet and have relict textures (plagioclase-hornblende symplectites after clinopyroxene and mineral zoning), suggesting a derivation from eclogite amphibolites.

Pegmatites

Pegmatites are rich in muscovite and plagioclase and occur sporadically throughout the Polinik Unit. They are intensely deformed, forming sheet-like bodies or irregular-shaped pods (up to 5 m wide), and may be locally concordant with the schistosity of the host rock, though cross-cutting on a large scale (Plate 1 – Fig. 6). In the core of metre thick pegmatite bodies, muscovite books, up to 5 cm across and 0.5 cm thick, are embedded in a coarse-grained plagioclase and quartz matrix which shows no shape fabric. Towards the margins, the grainsize progressively decreases and minerals have a strong planar fabric.

3.2. Deformation History of the Polinik Unit

Detailed studies in the Polinik Unit have made it possible to relate metamorphic and microstructural data to large scale structures and the tectonic evolution of the area.

Sets of structures have been correlated, using criteria such as orientation, style and position in the locally established deformation sequence.

Five major groups of regional structures can be distinguished ($D_1 - D_5$). F_1 , F_3 and F_4 regional fold phases correspond with the sequence F_1 , F_2 , F_3 , established





Sketch of structural elements in quartzo-feldspathic schist, in the Raggaalm area (sample 7), showing refolding of D_1 fabrics during D_2 and D_3 deformation.

An early quartzo-feldspathic and mica-rich foliation (S₁) is crenulated by F_2 folds. F_1 quartz fold hinges define a lineation I_2 which is subparallel to the F_2 crenulation lineation. Subsequent crenulation occurred during D₃, and micas have not recrystallised in F_3 fold hinges.

by OXBURGH (1966) and WATERS (1976) in rock units equivalent to the Polinik Unit. The deformation history and major structures are summarised in Table 1 and in a cross-section (Fig. 9).

 D_1 deformation resulted in the earliest regional structures, though evidence for earlier deformation has been recognised locally. Variscan pegmatites were intruded after D_1 deformation and are intensely deformed by D_2 . D_2 deformation is post-dated by a Cretaceous amphibolite facies burial metamorphism, which was followed soon after by D_3 . During D_3 , the Polinik Unit was deformed into a km-scale recumbent fold, closing to the south. It was during or after the final stages of D_3 that the Polinik Unit was juxtaposed with the Pennine tectonic units. However, Tertiary movement along the Mölltal fault zone has obscured the original contact relationships. Younger deformation (D_4-D_5) resulted in regional warping and brittle faulting.





- a) Sketch of a tight, m-scale F₂ fold, defined by a deformed quartzo-feldspathic layer. Mica schists contain an axial planar cleavage (S₂), which is crenulated (F₃).
- b) Sketch of crenulated mica schist in the cores of minor F_3 folds. Quartz layers on F_3 fold limbs are attenuated and are not crenulated.
- c) Thin section of similar style F_3 fold, showing mm-scale F_3 crenulation of mica schist.

Text-Fig. 7.

- a) Block diagram of a deformed pegmatite, intruding mica schist in the Polinik Unit, NNE of Bodensee. A pronounced foliation in the pegmatite (S_2), subparallel to the pegmatite margins, is folded by F_3 cascading subhorizontal folds. See Plate 1 Fig. 6.
- b) Equal area lower hemisphere stereographic plot of poles to S₂ foliation in the deformed pegmatite and the enclosing schists, and lineations. Poles to S₂ foliation lie on great circle with pole (X) consistent with F₃ fold axes.

\mathbf{D}_1

 D_1 fabrics are defined by a metamorphic segregation into mica and/or amphibole and quartzo-feldspathic layers (S₁), which are deformed by D_2 structures. A strong linear fabric (l₁), best developed in gneisses, is defined by the preferred orientation of stretched and broken garnets, quartzo-feldspathic rods and quartzribbons. Dismembered fold hinges (rods) also define a strong linear anisotropy (l₁ or older, Fig. 5).

\mathbf{D}_2

 D_2 caused a crenulation and substantial modification of S_1 (Figs. 5, 6). On a metre to 10's of metre-scale, S_1 is folded throughout the Polinik Unit into conspicuous 'M', 'Z' and 'S'-shaped tight folds (F₂), (Fig. 8, Plate 1 – Figs. 3–5), with subhorizontal fold axes which are nearly parallel to l_1 (within $\pm 20^\circ$, Fig. 9B). However, no large scale D_2 folds have been found, though the change in minor fold asymmetry suggests a series of km-scale folds with subhorizontal E-trending fold axes. A variably developed crenulation cleavage, S_2 , locally transposes the S_1 foliation and becomes the dominant foliation in less competent mica-rich lithologies. Eclogite amphibolite pods are aligned parallel to the S_2 foliation.

A pronounced lineation (I₂), which is either a crenulation lineation or the intersection of S₂ with S₁, parallels F₂ fold axes. The S₂ foliation is found in pegmatites, particularly at their margins (Fig. 7A, Plate 1 - Fig. 6). This is the earliest deformation fabric in the pegmatites, and therefore the pegmatites must have been emplaced after D₁, but prior to D₂. Rb/Sr two-mineral isochron ages from the coarse-grained centre of these pegmatites give ages of ca. 254 Ma (CLIFF, pers. com.). These are assumed to be emplacement ages and therefore post-date D₁ deformation, but pre-date D₂ deformation.

Inter $D_2 - D_3$ amphibolite facies metamorphism

Extensive growth of staurolite, kyanite, plagioclase and locally garnet in metapelites post-dates D_2 and pre-dates D_3 . Staurolite, kyanite and individual mica laths cross-cut the S_2 foliation and plagioclase helicitically overgrows the earlier fabrics (Plate 2 – Figs. 1–3). K/Ar radiometric ages suggests that this mineral growth occurred in the Cretaceous (ca. 100 Ma). The P/T conditions during metamorphism are discussed in section 3.3.

D₃

Subhorizontal cascading asymmetrical folds (F₃) occur throughout the Polinik Unit (Fig. 9, Plate 1 – Figs. 1–5), trending E–NE, parallel to I₁ and I₂, and forming coaxial interference folds with F₂ (Fig. 7B, 9B). A distinct variation in fold asymmetry suggests the existence of a km- scale recumbent fold, closing to the south (Fig. 9). Crenulation of S₂ defines a crenulation lineation (I₃), which is parallel to F₃ axes. In crenulation hinges, staurolite, kyanite and micas are bent and broken (Plate 2 – Figs. 4–6) and have been mechanically rotated and transposed into the foliation.

In areas of intense D_3 deformation, kyanite and staurolite are partly altered to fine-grained aggregates of sericite, margarite and chlorite; garnets are partly replaced by chlorite. This alteration may have occurred during or after D_3 deformation.

D₄ and younger deformation

During D_4 , the whole area was warped by upright, open, E-trending folds (Fig. 9), forming open antiforms and tightly pinched synforms, developed on a metre to a kilometre scale. Steep fault zones, which have accommodated several kilometres of both strike slip and dip slip displacement, cut D_4 and all earlier structures. These faults resulted in both N–S-shortening and E–Wextension. Pseudotachylite was generated along some fault surfaces (HOKE, in prep.).

3.3. Inter D₂-D₃ Cretaceous Amphibolite Facies Metamorphism

P/T conditions during the inter D_2-D_3 metamorphism have been estimated from microprobe chemical and textural analyses of staurolite, garnet and kyanite-bearing schists (aluminous metapelites). Samples were collected from a section between Mt. Polinik and the Polinik Alpine Hut (samples 68, 83, 81.3, 81.1, 81, 8, 77; Tables A,B).

3.3.1. Mineral phases

Garnets

Three generations of garnets have been recognised (Fig. 10, Plate 2 – Fig. 1) and all garnets may show some degree of chemical zoning.

- In some garnets, cores (garnet I) can be optically distinguished from their rims (garnet II), (Fig. 10). Garnet I tends to be full of quartz and tiny highly birefringent inclusions, but generally has a uniform chemical composition: almandine 73-75 mol.-%; spessartine 1.5 -3.6 mol.-%; pyrope 14-20 mol.-%; grossular 2-6 mol.-%. A chemical zonation is only developed near garnet I rims, where there is a slight decrease in Mg, and increase in Fe, while Ca and Mn show no significant variation.
- Garnet II is the most commonly found garnet generation, developed either as a clear rim around garnet I or as euhedral or spongy grains less than 1 mm in size. Garnet II often contains elongate rutile inclusions, and is found intergrown with staurolite, kyanite, biotite and muscovite.

Garnet II is characterised by higher Mg (pyrope 18–25 mol.-%) and lower Fe contents than garnet I. The Mg/Mg+Fe ratio is higher and can show a marked variation, suggesting changing equilibrium conditions between ferromagnesian minerals as a result of increasing temperature and probably pressure during garnet growth.

- Garnet III does not contain inclusions and is difficult to distinguish optically from garnet II, though it forms chemically distinctive, narrow, $100-300 \mu$ wide rims around garnet II, or more rarely occurs as swarms of tiny euhedral garnets in a quartz and mica-rich matrix. Garnet III is commonly characterised by a zonation showing a decrease in Mg and increase in Fe content (overall reduction in Mg/ Mg+Fe ratio compared to garnet II), and an increase in both Mn and Ca, towards the rim. This suggests growth during the waning stages of metamorphism (HESS, 1971; TRACY, 1979; KERR, 1981). Garnet III

Text-Fig. 8.

Sketch of m-scale, gently plunging, fold interference patterns, observed in the Polinik Schists, in the Mt. Polinik area, produced by superposition of F_3 and F_2 folds. Sketches are drawn as if viewed towards the west. a) "S"-shaped F_3 folds on "M"-shaped F_2 folds. b) "S"-shaped F_3 folds on "Z"-shaped F_2 folds. compare with Plate 1.

Ν



Text-Fig. 9.

Sketch of N-S cross-section through the Polinik Unit, showing an interpretation of the relation between structural elements.

A km-scale recumbent F_3 fold is folded into an F_4 antiformal structure. South of Mt. Polinik the structures are truncated by the Teuchl fault zone, with the formation of a subvertical shear fabric. Dismembered folds (possibly F_3) and a pronounced cleavage (S_3 or younger) occur near the base of the Altkristallin. The basal contact is not preserved in the study area, because it has been displaced by movement along the steeply dipping Mölltal fault zone. This crosssection suggests that the exposed structural thickness of the Polinik Unit, measured perpendicular to the prominent foliation, is not more than 2 km. Orientation data from Polinik Schists in the Mt. Polinik area are shown in an equal area lower hemisphere stereogram.



structures.	OTHER LITHOLOGIES	~. ◄	¢	•	<pre>Emplacement of augengneisses post-dates &1. Emplacement of pegmatites: (>254 Ma)</pre>	<pre>52 foliation in 2 pegmatites (developed best along pegmatite margins) 52 predominant foliation in</pre>	augengneisses	ŝ
rrelation of fabric elements with large-scale	MAFIC LITHOLOGIES Large scale structures Microstructures	¢•		(··		Amphibolite- facies: amphibole and plagioclase inclusions in garnet cores	Eclogite facies: garnet-omphacite- rutile-quartz No shape or cryst preferred orien- tation	Omphacite poiki- litically overgroveuhedral garnets
ion history of the Polinik Unit, showing the co	POLINIK SCHISIS (METAPELITES, QUARTZO- FELDSPATHIC SCHISTS) Large scale structures Microstructures	pre-S ₁ : local occurrence in Mittagsspitz area	51 Segregated quartzofeldspathic and mica-rich layers contain iso- clinally folded and stretched quartzo-feldspathic rods (f1 or earlier)	<pre>&1 mineral elongation Quartz-ribbon lineation (? trans- texture port direction), Garnets: Strong linear flattened and anisotropy. Quartzo-feld- spathic rods</pre>		F2 fold axes subparallel (±20°) to & rods. Folds have been modified to class 2 similar style tight to iso- cline! folds during later defor- mation. Asymmetric minor folds. folded.		? restricted occurrence of metre-size chevron folds (area 4)
diagram of the deformat	TECTONIC SIGNIFICANCE	? crystalline basement terrain	Ductile deformation. Major nappe dis- placement. Incorporation of mafic rock- suivers into para- greiss-series of Polinik Unit.	Strong mineral line- ation - possible transport direction during D ₁		Crustal shortening: Regional folding ? Underplating of Polinik Unit		
Table 1. Summary c	METAMORPHISM	^.∢	Upper green- schist facies to amphibo- lite facies			Upper green- schist facies to amphibo- lite facies	?) P-metamor- phism	-
	AGE		UT	sosijeV	riscan	nsV-teoq	?120 Ma	
	DEF OR- MAT I BN	Pre-D1	◄	<u>م</u>		D2	>	

	F ₃ folding of foliated pegmatites		area and causes a	es involve kink- nd flexural slip with faulting.
Interpreted mineral assem- blage: ioclase- b iotite	Alignment of i minerals in foliation	Hydrothermal alterations: new formation of chlorite, sphene, epidote,zoisite	rms effects the whole	mesoscopic structur folds, fracturing a folding associated
mineral assemblage to amphibolite facies mineral assemblages garnet-amphibole-plag -sphene-quartz		ب ب	tightly pinched synfor ructures. e-slip movement along- nding conjugate	·
² 2 lithologies treated in mile-lithologies Interpreted mineral assem- kyanite-staurolite-biotite- muscovite-quartz-plagioclase. Recrystallisation of quartz and mice-fabric.	F_3 recumbent folding Crystals bent and and F_2 broken in crenu- and F_2 lation hinges; quartz subgrain formation can show preferred crysta- llographic orien- tation. k_3 crenulation of S_2 Retrograde alterations of staurolite and kyanite to fine grained aggregates of sericite, chlorite and quartz- filled veins oriented purce fold axes in fold cores.	S ₃ in intensely deformed areas (area 5, south of bend in river Möll). Axial planar cleavage. Predominant foliation in area 5: phyllonitic foliation. Breakdown of garne to chlorite post- dates S ₃ formation growth ôf epidote and coisite in chlorite pseudo- morphs. Intrusion of calci	F_4 Upright folding with open antiforms and regional reorientation of the earlier st With increasing D_4 increasing degree of retrograde alterations. 2_4 in area 5 Further N-S shortening is achieved by strik parallel sets of steep ~NW-SE and NE-SW tre faults	Reactivation of faults as dip-slip faults
Prograde thermal event; thermal re-equili- bration	Recumbent folding. Possible emplacement of Pulinix Unit onto Matrei Zone and Per- ipheral Schieferhulle Possible formation of a major shear zone between the Polinik and Strieden Units.		Monoclinal structure Culmination of the Polinik Unit N-S shortening Predominant strike-slip faulting	Predominant dip- slip displacement along faults and accompanying uplift.
Thermal climax: amphibolite facies meta- morphism	- Cooling - 80-90 Ma Muscovite and biotite cooling ages Greenschist facies dis- equitorium assemblarges	Extensive hydrothermal alterations. Tauernmeta- morphism only affects	N-part lower greenschist facies uplift and cooling	
~105 Ma	₹80	>	~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~	
	۵ ^۲		0 5	

has probably not achieved equilibrium with the other garnets, since it only forms narrow rims around garnet II.

Staurolite

Zoning patterns in staurolites are related to replacement of staurolite by white mica shimmer aggregates. Zn, released by the breakdown of staurolite in this retrograde reaction, was incorporated into the remaining staurolite to form a Zn-rich rim (up to 2.5 % ZnO), similar to the retrograde Mn-rich rims in garnets. Otherwise, staurolites appear to be unzoned.

Biotite

Individual biotite grains are always of uniform composition, however, biotites adjacent to garnets or



Text-Fig. 10.

Microprobe profiles of chemical composition across garnets in the Polinik schists (samples 61, 81, 83, 77; see Appendices B, C, D), showing zoning and compositional variations in different garnet generations (gt I, gt II, gt III).

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staurolites (with Mg-poor rims) can be richer in Mg than those away from other Fe-Mg bearing phases. The garnets have well defined rims, marked by a change in chemical composition, presumably as a consequence of ionic exchange with biotite. For instance, in sample 81, Mg/Mg+Fe ratios vary between 47-49 in biotites next to garnets, and are between 43.5-44 away from other Fe-Mg bearing phases.

All biotites contain significantly more Al than phlogopite and annite. This is largely due to the tschermak substitution $AI^{VI}AI^{VV} = (MgFe)^{VI}SI^{VV}$ (DEER et al., 1966).

White mica

Three generations can be distinguished (Fig. 11):

- Coarse, aligned platy muscovite I, which defines the foliation S₂.
- Coarse, platy muscovite II which cuts across S₂ and, together with staurolite, kyanite, garnet II, biotite and quartz, forms a texturally equilibrated amphibolite facies mineral assemblage.
- White mica shimmer aggregates (muscovite III) which are retrograde and variably replace staurolite and kyanite and also occur in the kinked zones of coarse mica laths.

Fig. 11A illustrates the extent of phengite substitution in the different mica generations (phengite substitution:

$A^{3+[V]} \cdot A^{3+[V]} = S^{3+[V]} \cdot (Fe,Mg)^{[V]}$.

There is a negative correlation between Fe+Mg and Al in the octahedral position, with a higher phengite

content in muscovite I compared to muscovite II and III.

The paragonite content also systematically differs in the 3 muscovite generations (Fig. 11B,C). Muscovite I shows the smallest range in content of paragonite, varying between 4.8-15 mol.-%, which is generally lower than that of muscovite II (10-29.5 mol.-%), though partly overlapping with it. No free paragonite was detected in white mica. Muscovite III shows a bimodal distribution, occurring with 0-4 mol-% (almost pure muscovite) and 21-23 mol-% paragonite.

Most muscovites are deficient in alkalies, with Na+K+Ca totalling between 1.4–1.7 atoms per 22 oxygens. This phenomenon is generally attributed to H_3O^+ ions substituting for K⁺ and Na⁺ in the interlayer sites (BUTLER, 1967; CIPRIANI et al., 1971).

Chlorite

Chlorite variably replaces biotite and garnet. All analysed chlorites are of ripidolite composition (classification of HEY, 1954).

Plagioclase

Three generations of plagioclase have been distinguished:

- Plagioclase I is polysynthetically twinned and tends to be clouded with minute inclusions of probably white mica and/or zoisite. Its composition varies between 10-14 % anorthite with a slight decrease of anorthite towards the rim.
- Plagioclase II often forms clear rims around plagioclase I. It is rarely twinned, however it variably



shows "gefüllte" textures, containing white mica microliths. It is the most commonly found plagioclase in metapelites and occurs closly associated with kyanite, staurolite, garnet II, muscovite II and biotite. It overgrows the S₂ foliation helicitically. It either shows a homogeneous composition or is inversely zoned with more albite rich cores and more anorthite rich rims. Compositions vary between 6.48-5.4 mol.-% anorthite and <1 mol.-% orthoclase, with a mean composition of 14 mol.-% anorthite.

Equilibria which might have controlled the composition of plagioclase II include:

$$gros + 2 ky + qz = 3 an$$

 $pg + qz = ab + ky + H_2O$

Increasing temperature displaces both these equilibria in favour of the right hand side.

 Plagioclase III is almost of pure albite composition and is found in strongly altered samples associated with white mica shimmer aggregates.

3.3.2. P/T Sensitive Equilibria

The phases quartz – garnet II – biotite – muscovite II – kyanite – plagioclase II – staurolite – rutile – ilmenite form the most common mineral assemblage in Polinik Unit metapelites (Plate 2 – Fig. 2).

Mineral compatibility and phase relations in quartz and muscovite bearing metapelites can be represented in the AFM projection (Thompson 1975), projected from muscovite, provided that H_2O behaved as a perfectly mobile component during metamorphism (Fig. 12).

Internally consistant thermodynamic datasets (POW-ELL & HOLLAND, 1985; HOLLAND & POWELL, 1985) and published calibrations of appropriate geothermometers and geobarometers (see below) are used to evaluate the P-T conditions for the amphibolite facies equilibrium mineral assemblages.

The P/T conditions (Fig. 14) can be bracketed by the kyanite stability limits, and by the temperature sensitive breakdown of mica in the equilibrium:

 $pg + qz = ky + ab + H_2O$

and by the pressure sensitve equilibrium

 $3 \text{ iIm} + \text{Al}_2 \text{SiO}_5 + 2 \text{ qz} = \text{aIm} + 3 \text{ rut}$

Kyanite stability limits

Kyanite is the only Al_2SiO_5 polymorph observed in the Polinik Unit and is present in most aluminous metapelites. Therefore, peak metamorphic conditions throughout the Polinik Unit were within the kyanite field. Phase equilibria between the minerals kyanite, andalusite and sillimanite have been studied experimentally by RICHARDSON et al. (1969) and HOLDAWAY (1971) whose results are in good agreement. The position of the kyanite = sillimanite equilibrium puts a lower pressure limit for the peak metamorphism.

The equilibrium

paragonite + quartz = kyanite + albite + H_2O

in the system $Na_2O - CaO - Al_2O_3 - SiO_2 - H_2O$ This equilibrium was investigated, using the internally consistant data set of POWELL & HOLLAND (1985). This data set has been derived by optimising enthalpies over the whole data base simultaneously, with the re-

sult that there is very good agreement between most

$$pa+q = ab+ky+H_2O$$



Text-Fig. 13.

Temperature - Xco_2 graphs for the reaction $pa + qz = ab + ky + H_2O$ for different activities of albite and paragonite from Polinik Unit schists. Reaction lines were plotted using HOLLAND's program Thermo (HOLLAND et al., 1986).

calculated equilibrium curves and their corresponding experimental uncertainties. It also takes account of the compressibility and thermal expansion of minerals. The activity-composition relations used to calibrate the equilibrium constants are listed in Table 2.

Equilibrium curves were calculated for each sample using mean mineral compositions (representative probe analyses in Appendix D). Only minerals which are thought to still contain their peak metamorphic compositions (e.g. garnet II, white mica II, plagioclase II) have been used for the calibrations of equilibrium constants (Table 2).

The calculated equilibrium curves are subject to errors. A description of the common errors and how they propagate through the calculation has been outlined by POWELL (1985).

The position of the mica dehydration equilibrium

is relatively insensitive to errors in mineral composition. Temperatures calculated for this equilibrium are shown in Fig. 14. Fig. 13 shows the $T - X_{CO_2}$ curves for a range of activities at 6 kbars. It can be seen that temperature is fairly insensitive to vapour compositions. Assuming that the diluent is CO_2 , only low values of XH₂O are required to reduce the temperature substantially. For example, XH₂O of less than 0.2 is needed to bring the equilibrium temperature below 500°C.

	Table 2. Equilibrium constants u:	sed for thermodynamic calculations.		
Component	Formula	Activity (a)	Activity coefficient (Y)	Reference
anorthite (an)	CaA&_Si ₂ 0 ₈	0.25X _{an} (1+X _{an}) ² Y	exp {(1-Xan)2(2050+9392 Xan)} RT	Newton et al.(1980)
albite (ab)	NaA&Si ₃ 0 ₈	(X _{ab}) ² (2-X _{ab})۲	exp $\left\{ \frac{(1-X_{ab})^2 (6746-9442 \ X_{ab})}{R1} \right\}$	Newton et al.(1980)
grossular (gros)	Ca ₃ A&2Si3012	$(x_{ca}\gamma)^{3}(x_{A\&})^{2}$	exp $\left\{ \frac{(3300-1.51)(X_{py}^{2}+X_{py}X_{alm})}{RI} \right\}^{*}$	Newton and Haselton (1981)
almandine (alm)	Fe ₃ A&2Si ₂ 012	(X _{Fe} Y) ³ (X _A R) ²	exp {(-3300+1.51)(X _{py} X _{gr})} RT	Ganguly and Kennedy (1974) Newton and Haselton (1981)
paragonite (pa)	NaA&2 ^{M1} A& ^T Si ^T 30 ₁₀ ^(OH) 2	$9.48 \left(X_{NB} \gamma \right) \left(X_{A\mathfrak{A}}^{M1} \right)^2 \left(X_{\Box}^{M2} \right) \left(X_{A\mathfrak{A}}^{I} \right) \left(X_{S\mathfrak{i}}^{I} \right)^3$	$\exp\left\{\frac{(1-N^2(3250.5+0.1671)+(2342+0.4541)N^*)}{RI}\right\}$	Eugster et al.(1972) Powell (1978)
			N* = Na/(Na+K)	
Pyrope (py)	M9 ₃ A1 ₂ Si ₂ 0 ₁₂	(X _{Mg} Y) ³ (X _{A1}) ²	exp (3300-1.5T)(X ² r+X _{gr} X _{alm}) RT	Newton (1986)
R = 1.9872 cal de	eg-1mo1-1 * ;	X py = X _{Mg} = Mg/Mg+Fe+Ca+Mn, etc.		



Text-Fig. 14. Estimated metamorphic conditions (shaded region) for the Polinik Unit metapelites. Error bars (2 σ standard variations) are shown.

Table	З.
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Pressure estimates based on the equilibrium 3 ilm + Al_2SiO_5 + 2 qz = alm + 3 rut (BOHLEN et.al., 1983). Pressures have been read off the P-T-logK plot presented in BOHLEN (op. cit.; Fig. 3).

SAMPLE	^a alm	K = (1/a _{alm})	log K	₽(Kb) at 600°C
77	0.458897	2.179	0.3382	7.3
	0.5527	1.809	0.2575	8
81	0.473031	2.114	0.3251	7.4
	0.364785	2.741	0.4379	6
	0.344450	2.903	0.4629	5.8
	0.368711	2.712	0.4333	6
81.3	0.4910	2.037	0.3089	7.5
	0.5495	1.820	0.2600	7.8
	0.402404	2.485	0.3953	6.3
	0.485500	2.060	0.3138	7.4
	0.360269	2.776	0.4434	5.9
83	0.479314	2.086	0.3194	7.2
	0.399389	2.504	0.3986	6.2
	0.430532	2.323	0.3660	7.2
85.4	0.373715	2.966	0.4275	6.1
	0.358712	2.788	0.4452	5.7
	0.4385	2.281	0.3580	7.0
68	0.385067	2.593	0.4145	6.3
	0.3741	2.673	0.4270	6.1
9	0.5055	1.978	0.2963	7.6
	0.4602	2.173	0.3370	7.2

Pressure sensitive equilibrium in the system FeO – $TiO_2 - Al_2O_3 - SiO_2$

BOHLEN et al. (1983) calibrated the pressure sensitive equilibrium

3 ilmenite + Al_2SiO_3 + 2 quartz = almandine + 3 rutile.

Assemblages containing these reactants are common in the metapelites of the Polinik Unit. Ilmenite tends to be rimmed by rutile, suggesting that the reaction proceeded towards the right-hand side, and therefore estimates based on this geobarometer will give low pressure estimates. Retrogression textures have not been observed, such as rutile needles scattered in the titaniferous biotites. However rutile and ilmenite occur as inclusions in, or are intergrown with, garnet II. Therefore the assemblage garnet II, ilmenite, kyanite, rutile, and quartz can be confidently considered to be a peak metamorphic assemblage.

BOHLEN et al (1983) present a P-T-logK plot for this geobarometer. K is defined as:

$$\mathsf{K} = \frac{(\mathsf{a}_{\mathsf{ilm}})^3 \cdot (\mathsf{a}_{\mathsf{ky}})}{\mathsf{a}_{\mathsf{alm}} \cdot (\mathsf{a}_{\mathsf{rut}})^3}$$

since kyanite and rutile occur as pure phases (a = 1), and analysed ilmenites closely approximate the ideal formula (Fe²⁺Ti⁴⁺O₃) and can be treated as a pure phase (no hematite solid solution), the equilibrium constant simplifies to: K = $1/a_{alm}$ (Table 3).

The activity of almandine has been calculated using the expression given in Table 2, and varies between 0.344 and 0.485 at 600°C. Therefore logK varies between 0.4629 and 0.3138, with a mean of 0.3919 (Table 3). These values suggest pressures at 600°C of 6.2 ± 1.2 kbars. As garnets are rich in almandine and are in the RAOULT's law region, activities are quite reliable and this barometer should be good.

Fig. 14 shows all the equilibria discussed above. The metamorphic conditions deduced from Polinik Unit metapelite are in the range of $620\pm60^{\circ}$ C and 6.25 ± 1.25 kbars.

4. The Strieden Unit

4.1. Introduction

Schists in the Strieden Unit show a varied mineralogy and a distinct metamorphic zonation, indicating increasing temperatures with increasing structural depth. They range from sillimanite-bearing metapelites, with evidence for partial melting in the most northerly outcrops, at the base of the Strieden Unit, to low-grade phyllites at structurally higher levels, further south, immediately underlying the Mesozoic Drauzug sequence. Marble and amphibolite occur throughout the Strieden Unit, forming concordant discontinuous bodies and bands of varying thickness. The amphibolites, in contrast to those in the Polinik Unit, show no textural evidence for a former eclogite assemblage.

Radiometric age data (BREWER, 1969, 1970, discussed in section 5) suggest that the high temperature mineral assemblages are Variscian in age. Alpine metamorphism did not exceed lowermost greenschist facies, which is in contrast to the mid Cretaceous amphibolite facies metamorphism in the Polinik Unit.

The Strieden Unit is separated from the Polinik Unit by subvertival fault zones. Immediately south of these faults, a 400–700 m wide zone of intense deformation forms the base of the Strieden Unit (Fig. 4). This is an important discrete tectonic contact within the Altkristallin sheet, referred to as the main mylonite zone (MMZ). Radiometric age data suggest it is late Cretaceous to early Tertiary in age (section 5). It generally dips towards the south $(50-80^\circ)$, trending WNW and can be traced across the entire width of the field area. It is comprised of mylonites, which formed during lower greenschist facies metamorphic conditions, derived from a variety of high grade rocks such as diopside-bearing marbles, cummingtonite-bearing amphibolites, augengneisses and pegmatites and sillimanite and andalusite-bearing metapelites.

Subvertical and undeformed N to NNW trending porphyritic, fine- grained, biotite and amphibole lamprophyre dykes (dated at 30–40 Ma, DEUTSCH, 1984) have been found throughout the Strieden Unit. They cut across the mylonites and suddenly terminate at the Ragge – Teuchl fault zone, and are not found in the Polinik Unit.

4.2. Deformation History of the Strieden Unit

The deformation history is complex and not easy to correlate with that in the Polinik Unit. Deformation events are labelled D_l-D_q , and at least four of these events (D_l-D_o) are probably Variscan or older (section 5). D_p deformation is the dominant deformation at the base of the Strieden Unit, in the MMZ, and occurred in the late Cretaceous-early Tertiary (section 5), post-dated by D_q . Brittle faulting and the emplacement of dykes cut the D_q features and all earlier structures. The deformation history is summarised in Table 4.

4.2.1. D_1 , D_m , D_n fabrics

In schists, D_I is defined by a segregation layering (S_I) consisting of alternating mm-thick quartzitic and micaceous (muscovite with graphite and illmenite) layers, and straight inclusion trails of ilmenite and graphite in garnet cores (garnet I, Fig. 15, Plate 3 – Figs. 2–4). In amphibolites, S_I is defined by layers rich in plagioclase, quartz, amphibole with rare epidote. In marbles, S_I consists of mm- thick, more weather-resistant, dolomite layers, or layers rich in quartz clasts or other impurities such as mica-laths, graphite trains, oxides and amphiboles.

 D_m structures are the earliest to occur regionally. S_I is isoclinally folded (F_m), (Fig. 16). In schists, F_m is preserved as dismembered and isoclinal fold hinges, which form quartz rods (I_m) aligned within the foliation (S_m). Garnets (gtl) have rotated and have quartz pressure shadows parallel to S_m (Fig. 15). In amphibolites and marbles, S_I defines similar style, tight to isoclinal folds (F_m), with dm to m-scale wavelengths.

 $F_{\rm n}$ folds form the most prominent folds above the MMZ, defined by the repetition of amphibolite, marble and metapelite zones on a 100-200 m scale, as well as the variation in the orientation of $S_{\rm m}$ on a 10-30 m scale.

On a regional scale, Fn forms upright to slightly Nverging open to tight folds with up to 100 m



Text-Fig. 15.

Sketch of typical $D_{\mbox{\tiny II}}$, $D_{\mbox{\tiny m}}$ and $D_{\mbox{\tiny n}}$ fabrics in schists above the MMZ, in the Strieden Unit:

Superposition of fabrics shown in (d), suggests the following sequence of fabric development.

- a) Garnet overgrows a foliation (S_i or older).
- b) Garnets rotate during D_m deformation with the development of pressure shadow zones.
- c) Pressure shadow zones are crenulated during D_n deformation.
- d) Growth of garnet and staurolite occurs after D_n deformation.

wavelengths. Fold axes vary from subhorizontal to plunging gently E-SE (Fig. 17). Both Fm and Fn fold axes are nearly parallel, resulting in co-axial interference patterns on a dm- to 10's of metres-scale (Fig. 16). Minor dm- to m-scale parasitic folds, occurring in all lithologies, are associated with the large scale Fn structures, and show systematic changes in asymmetry. S_m is crenulated in the minor fold cores.

Further rotation of garnets (gtl) within the enclosing mica- and quartz-rich fabric occurred during D_n , and D_m pressure shadow zones were crenulated (Fig. 15). L_m quartz rods are deformed by D_n crenulations, and the angle between both crenulation lineations (F_n/I_n and F_m/I_m) is less than 30° (Figs. 16, 17).



Text-Fig. 16.

Field sketch , showing D_m and D_n structural elements in Strieden Unit quartzo-feldspathic schists, 100 m north of Mt. Schneestellkopf. An earlier quartzo-feldspathic foliation (Si) is folded isoclinally (F_m). F_m axial planes are subsequently deformed by open F_n folds, with 10's cm wavelengths. D_n fold axes are markedly oblique to D_m fold axes.



Text-Fig. 17.

Equal area stereogram showing relation between poles to the D_m foliation (S_m) and D_m and D_n fold axes $(F_m,\ F_n)$ in the Mt. Schneestellkopf area. F_n fold axes approximately lie on the pole to the great circle formed by poles to $S_m.$

4.2.2. Variscan High-Temperature Metamorphism and $D_{\rm o}$

 D_n deformation was followed by a static high-temperature metamorphism (section 4.3). In schists, the S_n foliation is overgrown by staurolite, garnet II (Fig. 15D), andalusite (Fig. 18) and sillimanite (Plate 3 - Fig. 1-6), and by diopside in marbles. The distinct metamorphic zonation in Strieden Unit schists indicates increasing



Text-Fig. 18.

Field sketch showing andalusite-bearing quartz veins which cut across ${\sf F}_{\sf n}$ folds in aluminous metapelites in the Mt. Strieden area.

Andalusite, staurolite and garnet (garnet II) overgrow F_n crenulations and therefore post-date D_n deformation.

temperatures towards the base of the Strieden Unit during metamorphism (section 4.3.1.). Mineral assemblages, which overgrow the steeply dipping (60°SSE) S_m or S_n foliation, define S-SSE dipping isograd surfaces (dipping 0–20°S). Subsequent intense D_p deformation occurred at the base of the Strieden Unit under greenschist facies conditions.

There is evidence for localised deformation, referred to as D₀, in sillimanite bearing schists, with textural evidence for partial melting (Fig. 28, Plate 3 Figs. 7,8). D₀ was either synchronous with or postdates the formation of the high temperature minerals, such as sillimanite, but pre-dates the emplacement of pegmatites. Sillimanite- bearing gneisses, quartzitic biotite-rich schists, and rare dm-thick layers of cummingtonite amphibolite contain a well-developed foliation (D_{01}), defined by the alignment of the high-temperature minerals (sillimanite, biotite, quartz, feldspar, amphibole), which is folded (D_{02}) into ptygmatic folds. These structures are cross-cut by pegmatites which were emplaced at ca. 265 Ma (CLIFF, pers. comm.). The localised occurrence of D₀ may be related to the thermal softening of rocks in and near a zone of partial melting (see Section 4.3).

4.2.3. $D_p - D_q$ Fabrics

The MMZ is the product of high strain and plastic deformation under lower greenschist facies conditions. A suite of progressively formed structures (D_p-D_q) can be distinguished:

- The main phase structures (D_n), consisting of the mylonitic foliation (S_p) and a stretching lineation (l_p) and associated folds (F_p).
- 2) Sets of microfaults and shear bands which formed at a late stage of D_p and caused extension parallel to the mylonitic foliation and lineation.
- A set of open folds (F_q) and faults (with a thrustsense displacement) which deform the mylonites, resulting in shortening approximately perpendicular to the mylonitic lineation.

Both boundaries of the MMZ are shown in Fig. 19, 20. The northern contact has been reworked cataclastically by movement along subvertical fault zones. The southern contact is either faulted, with an abrupt change in deformation style above the MMZ (e.g. in the Latisch area where undeformed and mylonitized pegmatites are juxtaposed with each other), or the mylonitic deformation decreases gradually with no sharp boundary to the MMZ (Poliniksee area, Fig. 20). In the latter case, the F_n fold geometry is progressively modified by D_p. Folds have been tightened and flattened and reorientated within the mylonitic foliation (So) (Fig. 21). Inter $D_n - D_p$ mineral assemblages are aligned in the transposed foliation. Fn fold axes have been rotated towards parallelism with the mylonitic stretching lineation (lp).

\boldsymbol{S}_p and \boldsymbol{F}_p in different lithologies

The substantial ductile strain during D_p has transposed the layering and schistosity into a new orientation, axial planar to F_p (S_p) (Fig. 23). However, the degree of transposition varies considerably in the wide range of lithologies found in the MMZ, such as augengneisses, pegmatites, marbles, amphibolites and quartz and mica-rich schists. In general, quartz is the only mineral which shows plastic deformation, deformed by slip along certain crystallographic planes, resulting in crystallographic preferred orientation patterns (Plate 4). Some dynamic recrystallisation occurred in micas. Feldspars behaved as resistant clasts, affected only by brittle deformation, which locally resulted in granulation of the clasts (Plate 5 – Figs. 3,7).

Quartzitic mylonites

A good mesoscopically visible foliation (S_p) is rarely developed in mylonites derived from quartzitic schists and quartzites . Instead, the mica fabric is a relict foliation (S_n) which is folded around F_p fold hinges (Fig. 23). S_p is only apparent on the microscopic scale, where it is defined by the preferred orientation of elongate quartz grains or quartz ribbons, which are aligned within F_p axial surfaces, with their elongation direction often perpendicular to or at a high angle to mica-rich layers (Fig. 23, Plate 4). Individual quartz ribbons have aspect ratios of 30 : 1 in large grains (up to 6 mm long) and 10:1 in smaller grains (0.2 mm long). Ribbon grains are variably recrystallised along grain margins or along subgrain boundaries to form grains less than 0.2 mm in diameter (matrix grains). Ribbon grains commonly have subgrain boundaries which make a constant angle with the grain elongation direction. In thin sections cut perpendicular to both Sp and Ip, quartz











Text-Fig. 21. Cross-sections through the base of the Strieden Unit (cross- sections 1 and 2 in Fig. 20). Note the structural position of the MMZ (at the base of the Strieden Unit) and the increase in deformation, observed in pegmatites and from F_p fold structures, at progressively lower structural levels.





Text-Fig. 23.

Sketch showing the relation between D_p and $\mathsf{pre-D}_\mathsf{p}$ fabrics in a sheared quartzitic schist.

A pre-D_p foliation (S_n) has been folded and crenulated in the core of a F_p fold, while a folded quartz band contains an axial planar fabric defining the D_p foliation (S_p). C-axes in quartz subgrains have a constant preferred orientation throughout the F_p fold structure.

bands are tightly folded on a mm-scale, with axial planes parallel to S_p and fold axes oriented subparallel to the stretching lineation I_p (Plate 4 – Figs. 3,4,5). The preferred crystallographic orientation of quartz subgrains is unaffected by these F_p folds.

Mylonitic amphibolites

 $\boldsymbol{S}_{\boldsymbol{p}}$ is parallel to a compositional layering and consists of 0.1-1 mm wide bands of pure quartz alternating with layers rich in porphyroclasts of green hornblende, plagioclase (20-30 % anorthite) and sphene aligned within a black, fine banded matrix of comminuted material (ultramylonite) derived from the parent rock (amph - plag - qz - sph - ore). With increasing deformation, the percentage of clasts and their grainsize decreases until, in extreme cases, cmthick bands of ultramylonite, devoid of clasts, alternate with layers which are rich in aligned porphyroclasts. A mineral lineation (I_p) is defined by the preferred orientation of the porphyroclast grains which form rods in the well developed mylonitic foliation and define a stretching lineation (I_p).

Mylonitic marbles

 $S_{\rm p}$ is apparent in thin section only, defined by the alignment of elongate calcite grains and occasional mica, chlorite and amphibole grains. $S_{\rm p}$ is either parallel to or cuts the compositional layering. The latter is

defined by layers of impurities and concentrations of more weather resistant dolomite grains. Where marbles are intruded by pegmatites, D_p penetrative deformation appears to have been accommodated preferentially in the marbles. The pegmatites are boudinaged and contain the S_p foliation only near their margins. Thin amphibolite layers within marbles are also boudinaged and have been stretched, within S_p , parallel to I_p . The orientation of individual boudins is oblique to S_p , which is deflected around the boudins. The sense of shear, inferred from the sense of obliquity with respect to S_p , agrees with other shear sense indicators in the mylonites (see below).

Mylonitised pegmatites and augengneisses (Plate 5)

These have a well developed mesoscopic S_p which is defined by the preferred orientation of elongate and planar feldspar, tourmaline (restricted to pegmatites) and mica porphyroclasts, embedded in a rock matrix of fine-grained recrystallised quartz, granular feldspar and sericite. Rods of porphyroclasts of tourmaline and feldspar define an extensional lineation (I_p), (Plate 5 – Figs. 1,2,4).

Geometry of D_p fabric

Sp is axial planar to tight to isoclinal asymmetric folds, recognised on all scales, plunging E to ESE at 30-60° (Fig. 19). These refold Fn, forming complex interference patterns (co-axial refolding and doubly plunging geometries) ; $\mathbf{F}_{\mathbf{p}}$ axes make angles of up to 85° with F_n , but are within 10° of I_p stretching lineation. On a cm-scale, Sp is itself locally deformed into asymmetric tight to isoclinal similar folds (Plate 4 - Figs. 3, 4, 5). These have variable vergences, with axial surfaces at a low angle or parallel to the mylonitic foliation. Fold axes are parallel to the mylonitic lineation. However, the quartz fabric shows a well developed crystallographic preferred orientation which is constant throughout the fold. These folds therefore probably nucleated after a strong foliation had already formed, possibly as a result of flow instability in a highly anisotropic material (PLATT et al., 1980) and were progressively amplified, flattened and rotated during continued flow in the mylonites. Fold axes may also have been rotated in opposite directions towards the long axis (xdirection) of the finite strain ellipsoid. This process leads to the formation of sheath folds. Sheath folds have been observed in the Kelluckerlkopf augengneiss, where m-scale elliptical traces of Sp are found in sections perpendicular to lp, and in marbles south of Poliniksee.

Late phase D_p structures in mylonites

 S_p is cut by microfaults (cm- to mm-scale). If the sense of movement along these faults is described relative to S_p , normal faults and thrust faults can be distinguished. Sets of faults with a normal fault sense (Plate 5 – Fig. 5) are observed in sections perpendicular to S_p and parallel to I_p . They tend to have a listric geometry and cut S_p at angles less than 37°. They intersect S_p at a high angle to I_p . Movement on such faults has resulted in extension along I_p in the plane of S_p . PLATT and BEHRMANN (1986) describe similar features, which they call extensional crenulation cleavage. Extensional crenulation is very common in mica-rich mylonites, where it forms single or conjugate sets with opposite senses of displacement. The cleavage zones contain fine aggregates of white mica (50–200 μ) de-



Text-Fig. 24.

Block diagram showing the mylonitic fabric observed in three orthogonal sections of a Strieden Unit quartzitic schist. The sections are described relative to three orthogonal axes x, y and z. The x-axis is parallel to the mylonitic lineation and the z axis is perpendicular to the mylonitic foliation. Typical shear sense indicators (sinistral) are illustrated: A,B and C show various styles of microfaulting in porphyroclasts; D, E show extensional crenulation cleavages in a micaceous fabrica.

rived, but differing chemically, from coarser older grains. New grains are richer in iron, which suggests that they have formed during a phase of non-isochemical recrystallisation, which probably occurred during the progressive D_p deformation, when the fluid composition changed and the temperature decreased.

Both extensional faults and extensional crenulations probably formed in the last phases of D_p . Conjugate sets of faults, symmetrically oriented with respect to the mylonitic foliation, suggest that the deformation during at least the late phases of D_p may have deviated from non coaxial deformation (shear deformation) and may in places have resulted in coaxial flattening of S_p .

Faults with a thrust movement sense are commonly observed in thin sections perpendicular to both S_p and I_p . They too have a listric geometry and have accommodated shortening in a direction normal to I_p . Asymmetric mm- to cm-scale crenulations of S_p , with axes approximately parallel to I_p , appear to have been caused by drag on these faults.

Kinematic indicators

 D_p occurs in a distinct zone of intense deformation within the Altkristallin, which is likely to have accommodated a large amount of relative motion of its boundaries. However, no displaced marker horizons have been found, and its boundaries cannot be precisely defined. Therefore, features of the internal deformation must be used to determine the sense of shear, defined as the sense of rotation of the vorticity (spin) of the strain field which acted during deformation in the MMZ. Knowledge of this provides considerable constraints on tectonic models for the development of the Altkristallin.

Microstructures and crystallographic fabrics in naturally deformed rocks have proved particularly useful in assessing the kinematics of deformation events (LISTER & WILLIAMS, 1979; SCHMID et al., 1986; BOUCHEZ et al., 1983; SIMPSON & SCHMID, 1983; BEHRMANN et al., 1982). A variety of fabrics, such as the orientation of minor shear bands, the preferred orientation of quartz c-axes



Text-Fig. 25.

Diagram showing the assumed relation between a single quartz c-axis girdle distribution and the mylonitic foliation (S_p) , for the case of sinistral simple shear.

The c-axis girdle tends to be perpendicular to the shear zone walls (flow direction) and hence is rotated anticlockwise of the mylonitic foliation (S_p).

(CPO), and the asymmetry of pressure shadows and sigmoidal micas, have been used to deduce a sense of shear (Fig. 24, 25).

Quartz crystallographic preferred orientation

Various models (LISTER & WILLIAMS, 1979) suggest that the greatest concentration of quartz c-axes lie in a plane normal to the flow or shear direction and that the foliation marks the xy plane of the finite strain ellipsoid (BOUCHEZ, 1978). Therefore, the relative orientation of the flow direction, which is parallel to the shear zone walls in simple shear, and the mylonitic foliation can be used to deduce the sense of shear. If the former is anticlockwise of the latter, then the shear is sinistral (Fig. 25).

Using a Universal Stage microscope, C-axis orientations have been measured from dynamically recrystallised quartz ribbons in several orientated mylonite samples. The preferred orientation pattern of the crystallographic c-axes can show:



 A single c-axes maximum, which forms angles of 35-45° with the y-axis, inclined 12-15° with respect to the yz plane (Fig. 26A).

The sense of shear deduced from this data agrees with that derived from other indicators: the orientation of muscovite fish, which, in pegmatite and augengneiss mylonites, form displaced and broken grains, displaced along (0001) planes and orientated oblique to S_p (Plate 5); shear bands, which consistantly make an angle of 30–40° with the mylonitic foliation (Plate 5 – Figs. 1,2,6) and have accommodated shear with associated drag of adjacent quartz layers; and the asymmetry of granulated tails of feldspar clasts, which can be traced for several centimetres along the mylonitic foliation before they taper out (Plate 5 – Fig. 7).

$\mathbf{D}_{\mathbf{q}}$ deformation

On a regional scale, the orientation of S_p varies from gently to steeply dipping. The orientations of D_p fold axes and lineations are shown in Fig. 19 and in Plate 4 - Fig. 2. In the Kelluckerlkopf area, asymmetric folds (F_q), which deform the mylonitic foliation on a cm- to m-scale, can be correlated with large scale variations in the oientation of the mylonitic foliation (Fig. 22). Areas where S_p is steeply dipping or flat lying correspond to F_q fold limbs or hinges respectively.

The poles of the mylonitic foliation crudely lie on a great circle in the stereographic plot, and the pole to this great circle has a similar orientation to that of small scale F_q asymmetric folds. This suggests that the mylonitic foliation (S_p) has been folded on a large scale by a series of moderately plunging folds. The outcrop pattern of the Kehlluckerlkopf augengneiss, with the repetition of lenses of augengneiss and schists, marbles and amphibolites, may be the product of D_q .



Text-Fig. 26.

Equal area lower hemisphere stereograms showing quartz c-axis distribution for quartz tectonites.

 a) C-axis orientations in sample 329, measured with an optical microscope and sensitive tint plate, in two orthogonal sections parallel to the xz and yz sections (see Text. Fig. 24).

b) C-axis orientations in sample 382, measured on an universal stage...

 D_q deformation is restricted to the MMZ. This may be explained in a number of ways:

- D_q occurred in a zone of weakness controlled by lithology, rock fabric or temperature.
- D_q occurs in the core of a large km-scale upright, ESE to SE trending, antiformal structure, which has been subsequently truncated by the Ragga – Teuchl fault zone.

In the latter case, D_q may correlate with the D_4 antiformal structures within the Polinik Unit. The correlation of D_q with deformation in the Polinik Unit is discussed in section 5.

 D_q structures have been modified by movement along the Ragga-Teuchl fault zone. Away from the fault, the trend of F_q fold axes is markedly oblique to that of the fault, whereas, close to the fault, they trend more nearly parallel to the fault, as well as plunging more gently (Fig. 19). This suggests that F_q fold axes have been rotated about horizontal and vertical axes as a consequence of movement along the fault, and that movement along the fault zone contained a component of strike slip.

4.2.4. Original Orientation of S_p and I_p

The following evidence suggests that the mylonitic foliation originally formed as a sub-horizontal planar fabric:

- In the structurally higher parts of the Strieden Unit, above the MMZ, thin mylonite zones (for instance the quartz mylonite near Mt. Strieden, Fig. 19), show no evidence for D_q, and hence may be near their original orientation, dipping at 15° or less.
- 2) After removing the effects of D_q on the mylonitic foliation (without correction of the fold plunge) in the MMZ, S_p dips gently (less than 30°) though in



Unfolding the I_p stretching lineation, taking account of F_q folding: the lower hemisphere equal area stereogram shows the I_p stretching lineations in the MMZ.

Lineations are scattered and do not lie on a single small circle about an average F_q axis. Dashed curve is the largest small circle about the average F_q axis wich encompasses most of the data. If the F_q axis is rotated to the horizontal (about a horizontal axis), the intersection of the circle with the edge of the stereogram gives an estimate of the range in unfolded lineation trends, assuming that the mylonitic foliation was horizontal initially. The median trend (and scatter, (M)), as well as that for the unfolded Strieden quartz mylonite (SM), are marked by arrows.

variable directions. This apparent variation in the orientation of S_p orientation, manifested in the variation (trend and plunge) of F_q fold axes, may be the result of subsequent warping and twisting of the deformed mylonite zone as a consequence of movement on the Ragga – Teuchl fault zone.



Text-Fig. 28.

- a) Opposite page.
- Topographic map of part of the Strieden Unit in the Latischalm and Striedenalm area, showing localities of samples (mineral assemblages of samples listed in Table 5) and isograds (cross-sections are shown in Fig. 28b). b) This page.
- Two cross-sections (see Fig. 28a), showing the geometry of the andalusite/ sillimanite and staurolite-in isograd surfaces.



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eral	Emplacement of pegmatites as sheets and irregular-shaped bodies, cutting across D _n and D _o structures.	F sheath-folds in Kehlluckerlkopf adgengneiss. S in augengneisses defined by plastically deformed quartz grains asymetrically aligned mica-fish, granulated feldspars. In pegmatites tour(often define L (pegmatite mylonites), boudinage of pegmatite bodies. Marble-tectonites (brittle def. processes dominate, cc and dol porphyroclasts are aligned in S. S in marble tectonites cuts across m			s fault zone .
Tight Fo2 folds, deforming sill-bearing schists ^{O2} (? plastic deformation of hot rocks). lyneatin. S defined by the alignment of sill-bi-rich dômains.		D restricted to MMZ and thin mylonites. F socilinal folding L stretching lineation S mylonitic foliation Earlier fabric elements rotated into parallelism with the mylonitic foliätion S and the lineation lp. T defined by quartz recryst. grains and sub-grains which contain recryst. grains and sub-grains which show a cryst. preferred orien- tation. S deforms amph.facies mineral assemblages which are retrogressed to greenschist facies mineral assemblages. (sericite, chl, marg, ab). Grain-size reduction processes dominate over recrystallisation (seen only in qz and the new-formation of sericite). Extensional microfaulting of clasts. further stretching within S further stretching within S to 1.	F folds(cm-100m) Smal-scale thrust faults cause 1q crenulations shortening of the mylonitic foliation perpendicular to 1. Quartz preferred cPyst. orientation deformed by crenulations and microthrusts.	Dykes cut across F _q	ong the Ragga-Teuchl fault zone. No dykes found north of thi
Restricted to lower sill-zone Cooling ages (K/Ar). Greenschist facies meta- morphism	Emplacement of pegmatites - in structural lower part of the Strieden Unit.	Lowermost green- schist facies met. MMZ : subhorizontal MMZ : subhorizontal main mylonite zone and minor thin mylonites, transport direction towards the NW URIt and the Strieden Unit together	r Greenschist facies Restricted to MMZ NE directed shortening. Steepening of S tewards the Ragga- Teuchl fault zone.	Emplacement of subvertical dykes (N-S trending) contemporaneously with develop- ment of a conjugate set of sub vertical faults, N-S shortening and W-E extension.	Movement (strike-slip and dip-slip) al
300 Ma	265 Ma	90-60Ma	less than 60 Ma ago	30-40 Ma	less than 30 Ma ago
D ₀ (1+2		ے۔۔۔۔۔ م ^م	۵		

Post- D_q deformation may also explain the scatter in I_p lineations, which might otherwise be expected to fall, on a stereogram, on a small circle about the F_q axis (assuming a mainly flexural flow mechanism of folding, accommodated along the mylonitic foliation), (Fig. 27). Lineations, which fall on the largest small circle about the average F_q axis, have been unfolded by

- rotating ${\rm F_q}$ to the horizontal (about a horizontal axis); then
- rotating (in both directions) the lineation along the rotated small circle to the horizontal.

The resultant two possible trends of the unfolded lineation give some estimate of the uncertainty in the original orientation of I_p , assuming that S_p was horizontal.

In general, microstructures suggest that the mylonites have developed in the following way:

- 1) Simple shear and plastic deformation in quartz bands.
- 2) Fracturing and extension of porphyroclasts (e.g. feldspar, tourmaline) parallel to the mylonitic foliation.
- Shortening of the mylonitic foliation in a direction within the foliation and perpendicular to the lineation, accommodated along mesoscopic low-angle thrust faults and asymmetric folds.
- 4) Extension of the mylonitic foliation in a direction within the mylonitic foliation and perpendicular to that in (3), accommodated by meso- and microscopic normal faults with a listric fault geometry.

Evidence discussed earlier suggests that the mylonitic foliation (S_p) was originally subhorizontal. Assuming both this and that movement was parallel to the "mylonitic lineation, microstructures suggest that, after removing the effect of D_q (discussed earlier), the original upper plate to the mylonitic shear zone moved NW (transport direction $300\pm40^\circ$), (Fig. 27), with a preferred direction of ca. 315°, relative to the lower plate.

4.3. Metamorphism of the Strieden Unit

4.3.1. Distribution of Minerals (Figs. 28,29)

Schists in the Strieden Unit show a distinct high temperature metamorphic zonation. A high temperature (upper) sillimanite zone, with local evidence for partial melting, can be distinguished from a lower sillimanite zone, where sillimanite coexists with staurolite. A zone of andalusite-bearing schists can also be distinguished. The boundary between the sillimanite and andalusite zones defines a surface where sillimanite is in equilibrium with andalusite. Another surface, which marks the limit of staurolite occurrence, is subparallel to the sillimanite-andalusite transition zone. The intersections of these surfaces with the topography suggest that they dip gently (0-20° S-SE), but steepen up both near the MMZ (ca. 50° S) and the Strieden shear zone (ca. 90°) (Fig. 28B). The steepening near the MMZ may be due to D_{q} deformation, which folds the mylonitic foliation (S_n) (section 4). Furthermore, these surfaces are oblique to the MMZ. The MMZ cuts across the isograd surfaces at a shallow angle, and appears to

cut down into higher grade rocks in the transport direction, towards NW. Sillimanite and andalusite-bearing schists are juxtaposed and deformed within the MMZ. In the following sections, the sillimanite-andalusite and staurolite surfaces are used as reference surfaces for determining the structural position of the metamorphic assemblages (Table 5). Therefore, the present maximum structural separation of samples discussed in this section is estimated to be about 1.5 km (ca. 0.5 kbar pressure difference).

4.3.2. Mineral Chemistry

Garnets

Garnets in andalusite-bearing metapelites (sample 554, Fig. 30A), show characteristic zoning patterns, and cores (gt I) can be texturally and chemically distinguished from rims (gt II). Garnets are generally euhedral and typically contain inclusions of quartz, ilmenite and graphite, in some cases showing evidence of rotation during growth (garnet I, Fig. 15; Plate 3 -Fig. 3). 'Normal' chemical zoning (HOLLISTER, 1966), characterised by bell-shaped patterns (Fig. 30A) for Mn and Ca, are typical, whereas Fe and Mg increase from the core towards the rim. The Mg zoning profile shows symmetrical maxima near the garnet rim, caused by a drop in the Mg/Fe ratio at the edges. At these Mg maxima, the mean garnet composition is: pyrope 8.8 mol-%; grossular 1 mol.-%; spessartine 3 mol.-%; almandine 87.2 mol.-%. In the cores of garnets, the composition is: pyrope 4.6 mol-%; grossular 5.6 mol.-%; spessartine 7.3 mol.-%; almandine 82.5 mol.-%.

The garnet rims (garnet II) are texturally different from the rest of the garnet (garnet I); they form clear, narrow, $100-500 \mu$ wide, often discontinuous bands, often intergrown with small euhedral staurolite and biotite which crosscut the foliation S_n (Figs. 15, 18). It is therefore likely that the change in the garnet zoning pattern towards the garnet rim marks the crystallisation of new Fe- and Mg-bearing phases, such as staurolite and biotite; the resulting equilibria requiring the garnet to become less Mg-rich.

Within the sillimanite zone, garnets are variably replaced by biotite (e.g. samples 516.2, 136, Fig. 30B,C). In general, a modal rise of sillimanite and biotite is observed with increasing structural depth, and is accompanied by a decrease in staurolite, muscovite and garnet.

Corroded garnets form atolls, where garnet cores are replaced by coarse, randomly oriented biotite laths which contain bundles of fibrolitic sillimanite (Plate 3 – Fig. 1). The end-product of garnet breakdown is a matted knot of fibrolite intergrown with biotite, and still preserving the shape of the original garnet.

All analysed garnets within sillimanite-bearing rocks are characterised by an unusually high Mn or Ca content. In sample 516.2, which contains sillimanite-biotite pseudomorphs after garnet and rare garnet, the mean garnet core composition is: pyrope 8.2 mol.-%; grossular 3.2 mol.-%; spessartine 18.9 mol.-%; almandine 69.3 mol.-%; andradite 0.4 mol.-%. In this case, the high spessartine content (18.9 mol%), or in sample 136 the high grossular content (18.7 mol.-%), may have prevented garnet from further break down (Appendix D).

The Mn enrichment of garnet can be explained by either the result of:



- replacement phenomena, with Mn concentrating in the remaining garnet as it breaks down; or
- the breakdown of ilmenite, releasing Mn, which was incorporated into garnet, while biotite accomodated the Ti.

Textural evidence suggests that ilmenite, which occurs as inclusions in garnets and also in the surrounding micas, recrystallised into coarse clots during garnet breakdown. Recrystallised ilmenite is typically more Mn-rich than primary ilmenite in the matrix (e.g. $0.9-1.2 \ \% \ TiO_2$ in the original ilmenite, $4-5 \ \% \ TiO_2$ in



recrystallised ilmenite clots). Presumably, Mn in this ilmenite was derived from the breakdown of garnet, while the Ti and Fe came from the garnet and its original ilmenite inclusions.

Biotite

Sillimanite-free schists contain two biotite generations (samples 195,554):

- The older biotite generation (biotite I), which is aligned within the S_m foliation and also occurs as inclusions in garnet I, usually has a higher TiO₂ and Fe_{tot}O content than
- a second generation (biotite II), which occurs as individual blades which cross-cut the S_m foliation. Biotite II is unzoned, with a homogeneous chemical composition throughout the thin section.

Biotites in sillimanite-bearing schists: In general, with increasing structural depth, a rise in the modal amount of biotite and sillimanite is accompanied by a decrease in the amounts of staurolite, muscovite and garnet. Biotites which occur as pseudomorphs after garnet, intergrown with sillimanite and also in the matrix, have a homogeneous composition on a thin section scale. However, on a regional scale, there is a trend towards more Fe-rich compositions at higher structural levels (Mg/Mg+Fe = 35-43), compared with those at deeper structural levels (Mg/Mg+Fe = 43-60). This trend reflects the relation between temperature of crystallisation and the composition of biotite end-members ($T_{Mq} > T_{Fe} > T_{Mn}$).

In the upper sillimanite zone, biotite, with inclusions of rutile needles, occurs intergrown with sillimanite fibrolite and shows ilmenite exsolution, particularly in kinked zones. This suggests that the original Ti content of biotites, during peak metamorphic conditions, may have been higher than their present content (1.9-2.6TiO₂) and was exsolved during cooling.

White mica

(Fig. 31)

Two white mica generations can be distinguished:

- Coarse platy muscovite laths (mus I) aligned parallel to the S_m or S_n foliation;
- and
 fine-grained white mica aggregates (mus II) or shimmer aggregates which variably replace andalusite, sillimanite and staurolite.

All white micas are deficient in alkalis (Fig. 32C). The ratio of Na/Na+K in coarse grained muscovites (musc I) systematically varies and decreases with increasing structural depth: the mean value is 0.085 in sillimanite free schists, whereas it is 0.057 in sillimanite-bearing rocks. This K-enrichment in mus I is associated with a modal decrease of muscovite and a modal rise in sillimanite, which can be described by the continuous reaction:

Na-rich mus + qz = sill + ab + K-rich mus + H_2O

The occurrence of more sodic rims around plagioclase in sillimanite-bearing rocks also suggests this reaction, though these rims could also be related to exsolution of albite in plagioclase during cooling.

In any one sample, coarse grained platy muscovites (mus I) tend to be more sodic than the retrogressive muscovite (II). The latter co- exists with Na and Ca-rich muscovite (up to 7 % margarit), chlorite and albite, in shimmer aggregates replacing andalusite, sillimanite and staurolite.

Plagioclase

Three feldspar generations can be distinguished:

- The oldest plagioclase generation (plag I) occurs as inclusions in garnet I, where it is rarely observed. Here, the anorthite content varies between 20-41 mol.-%. The extreme anorthite content of 40 mol.-% is probably due to local Ca exchange between garnet and plagioclase.
- The second generation (plag II) predominates and occurs as coarse and often twinned grains, which overgrow S_n and typically contain trails of graphite and muscovite as inclusions, forming remnants of an older foliation. Plagioclase II is interpreted as one of the phases of the peak metamorphic mineral assemblage. Its composition is homogeneous or shows a slight inverse zoning in any one grain. On a regional scale, there is a progressive increase in the anorthite content with structural depth. The mean anorthite content in metapelites from andalusitebearing schists and in the lower sillimanite zone is 13.7 mol-%, between varying 9.8-20 mol.-%, whereas in the upper sillimanite zone it is 26.9 mol.-% (varying between 22.4 and 30.42 mol.-%) respectively. The orthoclase content does not exceed 2 mol.-%.

In the lower sillimanite zone, oligoclase is rimmed by more sodic plagioclase which possibly formed in the reaction

Na-rich mus + qz = sill + ab + K-rich mus + H_2O during peak metamorphic conditions.

Another reaction which possibly has controlled the plagoclase composition is:

$$gros + 2 Al_2SiO_5 + qz = 3 an.$$

 Pure albite compositions (plag III) are rare (< 2 mol.-% anorthite) and have only been found in shimmer aggregates, together with fine grained white mica and chlorite, which replace andalusite and staurolite. They are attributed to late stage alteration.

The sillimanite and andalusite Al_2SiO_5 polymorphs The formation of both sillimanite and andalusite post-dates the D_n deformation.

Andalusite occurs as thin slender crystals in schists, largely replacing the mica-rich matrix and overgrowing garnet II, plagioclase II and staurolite (Plate 3 – Figs. 3,6; samples 554,195). It also occurs as euhedral, intergrown cm-sized crystals in quartz veins. Sillimanite predominantly occurs as fibrolite, nucleating and replacing biotite (Plate 3 – Fig. 1) and often intimately associated with ilmenite. It rarely forms prismatic crystals.

The chemical composition of both sillimanite and andalusite is fairly constant and approximates Al_2SiO_5 . The only other ion detected during microprobe analyses is Fe_{tot}, which varies between 0–0.032 formula units, with a mean value of 0.006. It probably occurs as Fe³⁺, replacing Al in the structure (Appendix D).

Staurolite

Two generations of staurolite can be distinguished (Plate 3 - Fig. 7), both of which seem to post-date D_n . They cross-cut each other and only differ chemically in their Zn content, with the second generation characterised by a higher ZnO content (up to 2.2 % ZnO) than


the first generation (ca. 0.3 % ZnO). Overall, the modal proportion of staurolite decreases with structural depth and individual staurolite crystals show an increased Zn-content. This can be described by the following continuous reaction:

low-Zn st + mus + qz = sill + bi + Zn-rich st + H_2O in which the modal rise of sillimanite is ascribed to the continuous breakdown of staurolite.

Chloritoid

Chloritoid has only been found in shimmer aggregates, replacing andalusite. It forms fine-grained laths which overgrow the white mica fabric and occur associated with albite and chlorite.

Mg/Mg+Fe ratios in chloritoid range from 13-17, and the Mn content does not exceed 2 mol.-%. A possible chloritoid-forming reaction, which accounts for the localised mineral assemblage in which chloritoid is found, is:

 $5 \text{ pg} + \text{chl} + 3 \text{ qz} = 7 \text{ ctd} + 5 \text{ ab} + 4 \text{ H}_2\text{O}.$

Chlorite

It occurs as a retrograde phase associated with the breakdown of garnet, staurolite and biotite, and also in shimmer aggregates. In the latter it can form crosscutting blades.

All analysed chlorites are of ripidolite composition.

4.3.3. Phase relations

The different mineral assemblages within quartz – rich metapelites can be described in the system K_2O – FeO – MgO – SiO₂ – H₂O. They are shown on an AFM diagram, projected from the main potassium-bearing phase (muscovite or K-feldspar in muscovite free metapelites, Fig. 32), assuming that H₂O behaved as a mobile component during metamorphism, and ignoring the minor components TiO₂, Fe₂O₃, MnO, CaO, Na₂O.

Andalusite-bearing metapelites

(Fig. 32A)

In andalusite-bearing metapelites, the presence of four texturally equilibrated phases (st - gt - bi - and) suggests an univariant equilibrium. However, this is unlikely, because of the widespread occurrence of this assemblage. It is likely that the equilibrium assemblage buffered water with staurolite-producing and water consuming reactions such as:

 $7 \text{ gt} + 5 \text{ mus} + 6 \text{ H}_2\text{O} = 6 \text{ st} + 5 \text{ bi} + 9 \text{ qz}.$

Textural evidence supports this reaction, and the presence of graphite in all andalusite-bearing metapelites suggests that the activity of water was less than

 a) Andalusite-bearing schists; the phases staurolite, garnet, biotite and andalusite are shown.

b) Sillimanite-bearing schists from the lower sillimanite zone; the phases sillimanite, staurolite, garnet, biotite are shown.

c) Sillimanite-bearing schists from the upper sillimanite zone; the phases sillimanite and biotite are shown.



Text-Fig. 32.

Phase relations in quartz-rich metapelites of the Strieden Unit, described in the system $K_2O - FeO - MgO - SiO_2 - H_2O$ on an AFM phase diagram. Phases are projected from the main potassium-bearing phase (muscovite or K-feldspar in muscovite-free metapelites), assuming that H_2O behaved as a mobile component during metamorphism, and ignoring the minor components, TiO₂, Fe₂O₃, MnO, CaO, Na₂O.

one, diluted by other components such as CH_4 or CO_2 during metamorphism. A low water activity is also suggested by the absence of fluid inclusions in the andalusite-bearing metapelites. The phase relations and mineral compatibilities explain also the breakdown of a more Mg-rich garnet (gt I) to staurolite, biotite II (more Mg-rich than biotite I), and almandine-rich garnet (gt II). The formation of staurolite appears to depend on

- the composition of garnet;
- and
- the availability of H_2O .

Lower sillimanite zone

(Fig. 32B)

The co-existance of staurolite and sillimanite on a thin section scale has been used to define the lower sillimanite zone (Fig. 28, Table 5). Within this zone, sillimanite replaces andalusite, and the modal proportion of staurolite decreases towards lower structural levels, accompanied by an increase in the staurolite Zn content. Locally two generations of staurolite occur, distinguished by their different Zn contents. The decrease in the modal proportion of staurolite can be explained by the staurolite consuming reaction:

st + Na-rich mus + qz =

sill + bi + K-rich mus + ab + H_2O + gt.

The disappearance of staurolite takes place within the stability field of muscovite and sillimanite.

The lower sillimanite zone is ca. 400 m thick (Fig. 28, Table 5) and is sandwiched between overlying andalusite-bearing metapelites and the underlying upper sillimanite zone (staurolite absent).

Upper sillimanite zone

(Fig. 32C)

It is defined by the presence of sillimanite without staurolite. The modal proportion of white mica decreases at structurally lower levels, accompanied by a progressive increase in the potassium content and decrease in the paragonite content of white micas. These changes can be explained by the reaction:

Na-rich mus + qz = sill + ab + K-rich mus + qz.

Garnet is progressively replaced, towards structurally lower levels, by biotite intergrown with sillimanite, described by the reactions:

gt + mus = bi + sill + qz

gt + Na-rich mus = bi + sill + ab + qz

Phase relations are shown in Fig. 32C which illustrates the increasing Mg/(Mg+Fe) ratio in biotite at structurally lower levels.

Partial melting

(Plate 5 - Fig. 28)

About 700 m below the top of the upper sillimanite zone, primary muscovite is absent in schists (e.g. sample 516). Muscovite only occurs as a retrograde phase, forming either small grains within or around feldspars, or large ragged crystals overgrowing biotite and sillimanite in cleavage zones. The disappearance of primary muscovite in the presence of quartz and plagioclase has been experimentally determined (STORRE & KARATKE, 1971, 1972) :

mus + qz + ab + H_2O = liquid + Al_2SiO_5 .

This suggests that one might expect the products of anatectic melts in this part of the sillimanite zone, consisting of plagioclase, quartz, and Al_2SiO_5 . The water

content of such a melt would depend on that in the host rock. Biotite, which is ubiquitous, can be expected to behave as a refractory mineral (WICKHAM, 1987). There is indeed some evidence for partial melting in the lowermost part of the sillimanite zone.

Within a small area, immediately above the upper boundary of the MMZ (200 m S of the Latischalm hut), small leucogranitic pods (qz - ksp - plag - bi - sill) a few centimetres across, are found within sillimanitebiotite schists (Plate 3 - Figs. 9, 10). Within the pods, quartz, plagioclase and K-feldspar form a granular to polygonal texture with grains 0.5-2 mm across. The margins to the pods are cuspate in places, mantled by biotite (and sillimanite). The pods can be thought of as leucosomes, mantled by melanosomes or restites, which formed as a result of local melting of the surrounding schist (MEHNART, 1968) . Elsewhere, leucogranitic layers alternate with biotite-rich layers (Plate 3 -Fig. 9). These have been deformed, containing a foliation, defined by the alignment of biotite laths, which is folded. These deformation structures are cut by pegmatites.

If the leucogranitic pods and layers do indeed represent partial melts, they place some constraints on the P/T conditions during metamorphism in the sillimanite zone (Fig. 38).

4.3.4. Metamorphic conditions

General considerations

The experimentally determined stability field of staurolite, in the presence of muscovite and quartz

All examined metapelites, which contain staurolite, also contain quartz and muscovite. HOSCHEK (1969) has shown experimentally that in reactions which involve muscovite and quartz, staurolite with a MgO/ (MgO+FeO) ratio between 0.2 and 0.4 breaks down at $575\pm15^{\circ}$ C for a 2 kbar water pressure, and at $675\pm15^{\circ}$ C for a 5.5 kbar water pressure. These temperatures may be too high because the reactions have not been experimentally reversed. The reaction line is plotted in Fig. 38.

2 Stability of Al₂SiO₅ polymorphs

Andalusite and sillimanite are the main Al₂SiO₅ polymorphs in the Strieden Unit. Kyanite has only been found at one locality, about 200 m W of Saukopf (locality 397, Fig. 18A, Table 5) . However, the presence of all three polymorphs structurally close together suggests that metamorphic conditions in this part of the Strieden Unit were close to the triple point where the kyanite, sillimanite, and andalusite stability fields meet. This triple point has not been precisely determined, and the estimates of RICHARDSON et al. (1969), ALTHAUS (1967), and HOLDAWAY (1971) vary by several kilobars and up to a 100°C (see Fig. 38). SALJE (1986) shows that the thermodynamic equilibrium conditions depend very much on the structural state of the material, with the and =sill phase boundary becoming steeper in P/T space with increasing fibrolitization.

For the sake of argument, a triple point position, midway between that of RICHARDSON and HOLDAWAY, is used in this study (this is also used by HOLLAND & POW-ELL, 1985). However, as will be discussed later on, this may lead to important inconsistancies. The slopes of

Table 5. List of mineral assembalges of analysed schists in the Strieden Unit, their structural position relative to the isograds, and thickness of mineral zones.

.

SAMPLE Mineral assemblage	Mineral assemblage		- metres and /sill isograd	above (to is st-in isograd	ograds) upper boundary of MMZ	Total thickness in m
195 gt-st-bi-mi-qz-plag-zr-chl (and in adjacent rock)	-mi-qz-plag-zr-chl adjacent rock)	-jlm	200	650	1300	
554 and -gt-st-bi-mi-qz-plag-zr-	st-bi-mi-qz-plag-zr-	ap-chl-ilm	separated from all	other samples	by Streiden fault zone (Fig.8.1)	
397 sill-ky-qz-plag-tour-ap-mi-b	qz-plag-tour-ap-mi-b	i-ilm	just NE of Strieder	n fault zone		
136 sill-bi-mi-qz-plag-ap-zr-ilm 169 sill-bi-mi-qz-plag-ap-zr-ilm	gt-bi-mi-qz-plag-tou mi-qz-plag-ap-zr-ilm	r-zr-chl-ilm-gph -gph	· · · · ·	100 50	850	400
<pre>302 sill-gt-mi-bi-qz-plag-ap-zr-i 336 sill-bi-mi-qz-plag-ap-tour-ch 31 sill-bi-mi-K.fsp-plag-gt-tour</pre>	mi-bi-qz-plag-ap-zr-j mi-qz-plag-ap-tour-c ^t mi-K.fsp-plag-gt-tour	llm-gph Nl-or zr-rut-ilm	4 1 1	ţ I I	650 350 350	
408 sill-bi-mi-plag-qz-gt-ap-tour 516.1 sill-bi-mus(secondary)-qz-pla 516.2 sill-bi-mus(secondary)-qz-gt-	mi-plag-qz-gt-ap-tour mus(secondary)-qz-pla mus(secondary)-qz-gt-	-ilm-rut 9-K.fsp-zr-ap-tour-ilm-rut plag-ilm-rut			25 20 20	:
514 and -sill-mi-bi-plag-qz-chl-ru	l-mi-bi-plag-qz-chl-ru	it-sph-or(hem)	1	'		
70 sill-bi-gt-st-mi-bi-qz-plag-ch 91 sill-bi-gt-mi-qz-plag-tour-rut	gt-st-mi-bi-qz-plag-ch gt-mi-qz-plag-tour-rut	⊧l-ilm-ap-zr ilm-hem-chl	1 1	1 1		





the boundaries to the stability fields, in a P/T diagram, are taken from HOLLAND and POWELL (1985), who essentially use a mean between HOLDAWAY'S (1971) and RICHARDSON'S et al. (1969) experimental determinations.

Metamorphic conditions during the emplacement of pegmatites

The primary mineral assemblage observed in pegmatites, intruding sillimanite-bearing schists, is ksp – plag – qz – sill (±tour) with muscovite only occurring as a subsolidus phase.



Text-Fig. 34.

Temperature – Xco_2 equilibrium curves at various pressures for the reaction pa + qz = ab + sill (and) + H₂O for different activities derived from the phases involved, in different schist samples of the Strieden Unit, using the program Thermo (HOLLAND et al., 1986).

The P/T field in which the pegmatites crystallised is constrained by the following:

- The absence of primary muscovite.
- The co- existence of ksp plag qz sill.

The P/T conditions are therefore likely to fall in the triangular area in Fig. 37. The maximum pressure at which the primary assemblage ksp – plag – qz – sill has crystallised (muscovite absent), is ca. 3 kb (at 665° C) and therefore at approximately 10 km depth. The possible temperature range at which crystallisation has occurred is between 730–665°C (for a pressure range of 1–3 kb). After emplacement, the pegmatites must have cooled at a constant pressure towards the regional geothermal gradient present at the time, crossing the reaction line

 $ksp + Al_2SiO_5 + H_2O = mus + qz.$

This led to the hydration reaction and breakdown of sillimanite and formation of subsolidus muscovite. This is supported by the observation that sillimanite in pegmatites is always mantled by muscovite. The abundance of muscovite, particularly at the pegmatite margins, suggests that there has been considerable interaction with water in these zones, resulting in the growth of muscovite and the obliteration of original igneous mineral textures.

The constraints on the metamorphic conditions of pegmatite emplacement suggest that the pegmatites have formed at lower pressures, and during a later separate event, than the abundent sillimanite found in the surrounding schists. Therefore the pegmatites cannot be regarded as the heat source required for the crystallisation of sillimanite in the metapelites. Geochronological date (section 5.3.2) also support this conclusion.



Pressure and temperature (P/T) calibrations

Garnet-biotite geothermometer

Compositions from garnet II and adjacent biotites have been used, applying the method outlined by FERRY and SPEAR (1978). Garnets (except for sample 31) have suitably low Ca and Mn contents (Ca+Mn)/ (Ca+Mn+Fe+Mg) ratios less than 0.2). However the Al^{IV} and Ti contents are high and the (Al^{IV}+Ti)/ (Al^{IV}+Ti+Mg+Fe) ratio is greater than the recommended 0.15 (varying between 0.1–0.27).

 K_D values (Table 6, Fig. 33) vary between 0.094 and 0.171, corresponding to a temperature range at 6 kbars of 409–575°C (±50°C).

2 Paragonite-albite geothermometer

(Figs. 34,35)

The same method is used as described in Section 3.3.2, using white mica II and plagioclase II. Paragonite activities vary between 0.36 and 0.78, and albite activities vary between 0.67 and 0.85. $T-Xco_2$ curves for 5.5 and 4 kbars (Fig. 34) show that likely temperatures are fairly insensitive to vapour composition. XH_{20} of less than 0.3 is required to bring the equilibrium temperature below 550°C. Calibrated reaction lines are plotted in Fig. 35.

Ilmenite-rutile geobarometer

The same method has been used as outlined in Section 3.3 and the results are shown in Fig. 36 and Table 7. This barometer could only be used in a few samples.

Results and discussion

The temperature and pressure constraints on samples from the Strieden Unit are summarised in Fig. 38, Table 8. This clearly shows a progessive decrease in both temperature and pressure towards structurally higher levels, as might be expected if the rocks have

> Text-Fig. 35. Temperature-pressure plots for the equilibrium pa + qz = ab + sill (and) + H_2O for different samples of Strieden Unit schists.

> Error bars show 2 sigma standard variation, using HOLLAND's et al. (1986) program Thermo.





Text-Fig. 36.

P - T - logKp equilibrium reaction lines taken from BOHLEN et al. (1983) for different values of logKp (0.2-0.6) derived from the pressure sensitive reaction 3 ilm + sill + qz = alm + 3 rut. LogKp-values (Table 7) calculated for different samples of Strieden Unit schists and the corresponding pressures at 600°C, are shown.

P kb (1) $ab+or+qz+H_2O=L$



Text-Fig. 37.

Pressure-temperature diagram showing estimated conditions during emplacement of the ksp - plag - qz - sill ± tour pegmatites (no primary muscovite) at the base of the Strieden Unit.

The pegmatites crystallised at a maximum pressure of 3 kbars (at ca. 665°C). After emplacement, the pegmatites must have cooled, probably isobarically towards the regional geothermal gradient present at the time, crossing the reaction line (3) leading to the hydration reaction and breakdown of sillimanite and formation of subsolidus muscovite.

Table 6.

Mg/Fe ratios in co-existing garnet and biotite pairs from different samples of the Strieden Unit schists. The distribution coefficients $K_D = (Mg/Fe)$ gt / (Mg/Fe) bi for different garnet/biotite pairs are shown and the equilibrium temperatures T°(K) = (12454 + 0.057 × P) / (4.662–3R × InK_D) (FERRY & SPEAR, 1978) are calculated.

		В	IOTITE				GARNE	T		ĸ	TºC 6kb	T°C 8kb
SAMPLE	Fe	Mg	Mn	ALIV	Ti	Fe	Mg	Mn	Ca	ĽD	± 50°C	± 50°C
554	2.965 3.070 3.070 2.569	1.569 1.528 1.528 1.444		0.995 0.985 0.985 0.289	0.193 0.192 0.192 0.192 0.142	2.678 2.661 2.678 2.689	0.205 0.205 0.232 0.237	0.063 0.077 0.037 0.038	0.018 0.027 0.022 0.016	0.145 0.155 0.174 0.157	517 538 575 542	525 545 583 549
195	2.571 2.458 2.458	2.028 1.965 1.965	- - -	1.000 1.122 1.122	0.159 0.165 0.165	2.488 2.488 2.485	0.277 0.277 0.284	0.153 0.153 0.143	0.04 0.04 0.038	0.141 0.139 0.143	510 507 514	517 514 521
136	2.370 2.539 2.602	1.964 1.382 1.798	- - -	1.135 1.361 1.000	0.180 0.076 0.256	2.373 2.380 2.352	0.185 0.221 0.215	0.259 0.254 0.246	0.12 0.093 0.101	0.094 0.171 0.132	409 569 492	415 576 499
31	2.263	2.299	0.016	0.979	0.197	1.911	0.258	0.673	0.104	0.133	494	500
302	2.782 2.931 2.784	1.445 1.619 1.637		1.097 0.881 1.003	0.266 0.266 0.270	2.347 2.360 2.392	0.190 0.180 0.194	0.275 0.280 0.292	0.089 0.181 0.103	0.156 0.138 0.138	540 504 504	547 511 511
408.6 408.3 408.4	2.963 2.243 2.243 2.502	1.617 2.579 2.579 2.322	0.022 - - 0.16	0.883 0.705 0.705 0.674	0.261 0.211 0.211 0.182	2.15 2.014 2.023 2.192	0.219 0.375 0.300 0.266	0.532 0279 0.343 0.272	0.032 0.306 0.280 0.253	0.187 0.162 0.129 0.131	599 552 485 489	607 559 492 496

12 - P kb





Table 7.

List of samples and pressures calculated for a fixed temperature of 600° C, using the reaction 3 ilm + sill + qz = alm + 3 rut (BOHLEN et al., 1983).

SAMPLE	^a alm	K = (1/a _{əlm})	log K	~P(Kb) at 600°C	P(Kb) at 800°C
516	0.392828	2.55	0.41	6.0	7.5
408.3	0.311118	3.21	0.51	5.0	6.5
408.6	0.392640	2.55	0.41	6.0	7.5
31	0.266581	3.75	0.57	4.5	6
8	0.269761	3.71	0.57	4.5	6

Pressures for various logK values have been taken from Fig. 3 in BOHLEN et al. (op. cit.).

undergone a burial metamorphism. However, the uncertainties are large. The samples have experienced pressures of 5.5 ± 1.5 kbars, and temperatures of $650\pm100^{\circ}$ C, roughly falling on a line in P/T space with a gradient of $25-30^{\circ}$ C/km. This is a reasonable 'normal' geothermal gradient, however the pressure differences between the structurally highest and lowest samples appears to be about 2.5 kbars, which is equivalent to about a 9 km structural height difference (Fig. 38). This is clearly incompatible with the presently observed maximum structural separation of about 1.5 km. There may be several explanations for this inconsistency:

- 1) It is an artifact of a combination of the large uncertainties in the P/T estimates, especially those based on the position of the Al_2SiO_5 triple point, and unconstrained systematic errors. If the Al_2SiO_5 triple point is in reality closer to that of RICHARDSON et al. (1969), at a higher pressure, the apparent pressure separation between the structurally lowest and highest samples is largely removed. However, even in this case P/T estimates of some samples will be out of the brackets given in Fig. 38, if they are to fall in the narrow pressure range suggested by the present structure.
- 2) Important structural deletion has occurred subsequent to the metamorphic imprint. An important mylonite zone, which shows an extensional fabric (D_p) cuts off the base of the metamorphic sequence. Other minor extensional deformation zones occur within the metamorphic sequence (section 4.2.3). However, the amount of crustal thinning is not known, though it does not seem likely that the thin mylonite zones within the metamorphic sequence have been responsible for deleting more than 7.5 kilometres of structural thickness, though they could have removed a kilometre or so.

5. Geological Synthesis

The differences between the Polinik Unit and the Strieden Unit, such as their structural separation, contrasting lithologies, different structural and metamorphic histories, make it difficult to provide a relative, let alone absolute, combined chronology for the study area, without using geochronological data, which require interpretation.

5.1. Geochronological Data from the SE Tauern Area

A large number of samples from the south-eastern Tauern area have been dated, using the K/Ar method (OXBURGH et al., 1966; BREWER, 1969, 1970; LAMBERT, 1970; WATERS, 1976; HAWKESWORTH, 1976; CLIFF et al., 1971; CLIFF, pers. com.) (Fig. 39).

5.1.1. General Considerations on the Significance of K/Ar ages

The K/Ar age of a mineral will depend to a large extent on the thermal history of the rock (Fig. 40). In the simplest case, the K/Ar age of a system gives the time elapsed since the system cooled below some critical temperature (blocking temperature) at which the rate of radiogenic argon escape from the system becomes negligible (DODSON, 1973). A combination of K/Ar ages from different minerals with different known blocking temperatures, such as muscovite and biotite, in a suite of rock samples from different structural depths, can be used to estimate uplift rates and the geothermal gradient that prevailed during cooling.

K/Ar ages are also affected by other factors:

- The grain size of the potassium-bearing mineral: Large minerals are completely outgassed at higher temperatures or over a longer period than small grains, and therefore can give different ages from those obtained from small grains.
- 2) K-bearing minerals can form diagenetically at temperatures below their blocking temperatures. In this case, the mineral will either give an age dating mineral growth, or a mixed age if the host environment contains argon.
- 3) If, even at temperatures far above the blocking temperature, the mineral is in an environment with a high argon pressure, the mineral will record a mixed age after it has cooled, as the initial argon concentration immediately after cooling will not be zero (excess argon). The effects of excess argon have been described by BREWER (1969, 1970) and WATERS (1976) from the Altkristallin, within, and in areas adjacent to the study area.

	Mineral zones	, andalusite zone	▶ sillimanite zone	uPPER sillimanite zone	, Within MMZ
on lines.	P/T field defined by intersection of reaction lines	<pre>< 5.5kb 450-675°C </pre> <pre>< 5.5kb 500-650°C</pre>	4.2-5.5kb 550-625°C	4.4-7.3kb 620-690°C 5.4kb 650±20°C 5.5-7kb 675-750°C 5-6.6kb 615-700°C	
secton of the individual reaction	ilm/rut (Bohlen et al. 1983)	1 1	1 1	- 600°C : 45kb 600°C : 5-6kb 600°C : 6kb	I
by the inters	A1 ₂ Si0 ₅	and and	sill sill	sill sill sill sill	sill
its on the P-T conditions are given	pg/ab (Holland & Powell 1985)	- 4kb:615±5°C 6kb:674±6°C	4kb:571±3°C 6kb:617±3°C 4kb:607±6°C 6kb:657±7°C	4kb:598±3°C 6kb:647±3°C 4kb:610±8°C 6kb:660±9°C 4kb:648±10°C 6kb:648±10°C 4kb:617±8 4kb:668±9	4kb:590±7 6kb:638±7
estimates. Constrain	gt/bi (Ferry & Spear 1978) T(±50°C)	6kb:507-514°C 8kb:514-521°C 6kb:517-575°C 8kb:525-583°C	6kb:409-569°C 8kb:415-576°C -	6kb:504-540°C 8kb:511-547°C 6kb:494°C 8kb:494°C 8kb:492-607°C 8kb:492-607°C -	I
	SAMPLE	195 554	136 169	302 31 408 516	70.2

Table 8. Summary table listing all analysed schist samples in the Strieden Unit, and the various reactions and methods used to calculate P-T

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5.1.2. K/Ar age Distribution in Altkristallin Rocks of the Study Area and Adjacent Areas

In the light of the considerations discussed above, the Altkristallin in the study area and adjacent regions can be divided into six age provinces, trending W–E, which each show a particular pattern of ages (summarised in Figs. 39, 41). They demonstrate consistant age profiles (Fig. 42). Below the Permo-Mesozoic unconformity, in the Drauzug area, ages are Palaeozoic, becoming progressively younger, towards the N, with concordant K/Ar mica ages of ca. 80 Ma in the Polinik Unit. The latter are interpreted as cooling ages from a Cretaceous burial metamorphism (section 3.3.2) in

Province 1

(see age section on province 2).

In the structurally highest parts of the Altkristallin, immediately below the Permo-Mesozoic unconformity surface in the Goldeck area, BREWER (1969, 1970) has measured uniform K/Ar whole rock and mica ages of 300-320 Ma (Fig. 39). The muscovite, biotite and whole rock ages are all similar and are considered to be true cooling ages from a Variscan regional metamorphism, and have not been disturbed subsequently.

Further west of this area, in the southernmost part of the Kreuzeck area, age province 1 is not apparent.



Here, the Drauzug sedimentary sequences rest unconformably on phyllites (possibly Palaeozoic in age) which in turn rest on higher grade schists of age province 2. The precise age of these phyllites and their structural relationship with the underlying Altkristallin schist is not clear. A structural break along the Drau valley is possible.

Province 2

Immediately north of age province 1, in the northern Goldeck and southern Kreuzeck areas, BREWER (op. cit.) has determined ages which are consistantly ca. 300 Ma for muscovite, and between 100 and 300 Ma for biotite, becoming younger towards the north (Figs. 39, 41). Many of these mineral ages were calculated from whole rock ages, and a knowledge of the volume proportions of biotite and muscovite in the rock. This assumes that muscovite and biotite are the only minerals which contribute to the whole rock age. BREWER (1970) constrained these calculated mineral ages with measured mineral ages from a few samples. He interpreted the ages of province 2 to be a consequence of a brief reheating of rocks at ca. 105 Ma (see below), which last cooled at ca. 300 Ma (as in province 1), at temperatures between the blocking temperatures of biotite (ca. 300°C) and muscovite (350–380°C). The biotites only partially outgassed, while the muscovites were unaffected and retained their Variscan cooling ages (Fig. 42).



Text-Fig. 40.

Sketch diagram illustrating the likely relationship between temperature and the rate of Ar40 loss from muscovite and biotite.

The blocking temperature is defined here as the temperature at which the rate of argon loss is negligible. Above ca. 400° C, the K/Ar mica systems are considered to be fully open to argon diffusion.

An estimate of the age of the peak temperature reached during the Cretaceous metamorphism is given by one sample (G18), from which BREWER (1970) measured an unaffected Palaeozoic age in white mica of 322 Ma, but found that the biotite age has been reset during the Cretaceous, giving an age of 103 ± 2 Ma. This sample must have experienced a critical temperature between the blocking temperatures of muscovite and that of biotite which could not have been maintained over a long period of time and therefore probably records the age of the peak temperature experienced by this sample during the Cretaceous metamorphism.

BREWER (1969) defined surfaces on which the rocks show identical biotite ages and also muscovite ages. He called these surfaces synchrons, which dip at about 10 degrees to the south and are assumed to have been parallel to the earth's surface ca. 100 Ma ago. The dip of these surfaces indicates that towards the north progressively deeper structural levels of the Altkristallin, with respect to these synchrons, are exposed.

Province 3A

Immediately north of province 2, within a ca. 1.8 km thick zone (measured perpendicular to the synchron surfaces), BREWER found rocks with biotite ages of ca. 100 Ma, and muscovite ages between ca. 300 Ma and 100 Ma, becoming progessively younger at deeper structural levels (Figs. 39,41,42). This zone is bound by subvertical fault zones (Fig. 41), which converge towards the east. Synchrons dip more steeply (20 degrees and steeper) near these fault zones. The age pattern in province 3A can be interpreted as a consequence of a brief reheating between the blocking temperature of muscovite, which approximately coincides with the temperature at which biotite became fully outgassed, and the temperature at which muscovite became fully outgassed (Fig. 42). If one assumes that this temperature difference is roughly the same as the temperature difference between the blocking temperatures of biotite and muscovite ($50-80^{\circ}C$, Fig. 40) and that the synchrons reflect the instantaneous temperature distribution, one can calculate a geothermal gradient between 28 and 44°C/km ($50-80^{\circ}C/1.8$ km) across this zone.

Province 3B

North of province (3A), separated by a fault, ages measured by WATERS (1976) define a zone where both muscovite and biotite have ages between 150 and 250 Ma, intermediate between the 300 Ma and 100 Ma cooling ages (Fig. 41). This zone also includes the Strieden Unit. These mixed ages can be interpreted as a consequence of reheating, as in province 3A, somewhere between the blocking temperature of muscovite and the temperature at which muscovite became completely open to argon diffusion. However, the similarity of the biotite and muscovite ages needs to be explained. Rocks in province 3B are distinctly coarser grained (medium- to coarse-grained schists and gneisses with large muscovite and biotite crystals) than the finegrained schists in provinces 3A and 2. BREWER (1969) also remarks on the grain-size difference, which he suggests might have affected the age pattern. It is suggested in this study, that because of the larger grain-size in province 3B, both muscovite and biotite were only partially outgassed, yielding intermediate cooling ages (Fig. 43).

Province 4

This province lies to the north of province 3B, separated from it by the Ragga-Teuchl and Strugenkopf fault zones (Fig. 41). Province 4 is characterised by muscovite and biotite ages of 80–90 Ma, which are interpreted as true cooling ages, and are discussed in more detail in the section dealing with the age of deformation and metamorphism in the Polinik Unit (section 5.2.1).

Province 5

Altkristallin rocks to the NE of the Drau valley, in the Millstätter lake area, yield anomalous K/Ar mica ages (BREWER, 1969, 1970) (Fig. 39). They are anomalous because the biotite ages are much older (by up to a factor of five times) than the co-existing muscovite ages, which is not a normal blocking temperature relation. BREWER (1969) discusses this problem and suggests that the age pattern is a result of excess argon (Fig. 42). The high argon pressure was due to argon outgassing at lower structural levels, as a consequence of reheating, with Ar migrating upwards and being trapped in the micas as the rocks cooled through the mica blocking temperatures.

Province 6

At the margins to the Tauern Window, within the Altkristallin near the basal contact, K/Ar ages (whole rock and mica ages) of less than 70 Ma years are found (OXBURGH et al., 1966; LAMBERT, 1970; WATERS, 1976; HAWKESWORTH, 1976), (Fig. 39). These were then believed to be a consequence of a Tertiary reheating (Tauern crystallisation), which has partly reset the Cretaceous cooling ages. This heating resulted in metamorphism within the Tauern Window, where rocks give K/Ar mica cooling ages ranging between 18 and 22.5 Ma (OXBURGH et al., 1966) (Figs. 39, 41). However, recent Rb/Sr dating along the base of the Altkristallin, of rocks characterised by an extensional deformation fabric which post-dates the Cretaceous metamorphism, has yielded 40-50 Ma ages (WALLIS, 1988). WALLIS (1988) interpretes these ages to date the final stages of an extensional deformation which resulted in the contact relationships of the Altkristallin and the

Matrei Zone. He also argues that these ages cannot be due to the Tauern metamorphism, which is too low in this area to reset the Rb/Sr systematics in white micas.

The sharp break in the pattern of ages across the margin of the Tauern Window which is seen north of the study area (Fig. 41) is interpreted as a consequence of movements along the Mölltal fault zone in the Tertiary.



Text-Fig. 41.

They define various age provinces, discussed in the text. A summary N-S cross-section is illustrated in Fig. 42. Major structural features are: M.T.Z. = Mölltal fault zone; R.T.F.Z. = Ragga Teuchl fault zone; G.F.Z. = Grießenegscharte fault zone; N.F.Z. = Nigglaibach fault zone; MMZ = Main Mylonite Zone.

K/Ar ages in the study area and in adjacent areas, labelled 1, 2, 3a, 3b, 4.



Text-Fig. 42.

Schematic section through the Altkristallin (see Fig. 41), illustrating the vertical distribution of age provinces and their suggested relation to maximum post-Variscan temperatures and maximum post-Variscan depth of burial.

5.1.3. Summary of the Regional Age Pattern

The age pattern depends on the precise temperature that any particular part of the rock pile reached during the Cretaceous metamorphism, how long it remained at that temperature, the grain size, and the argon pressure.

The age pattern can be used to define surfaces of identical ages (synchrons), which are assumed to be parallel to the earth's surface during the Cretaceous. These synchrons suggest that the Altkristallin, away from its northern margins (e.g. outside the Polinik Unit) has not been tilted regionally more than 20 degrees S since the Cretaceous, and that the presently exposed structural thickness (excluding the Polinik Unit), with respect to the synchrons, is probably not more than 6 km. The progressive change in ages with increasing structural depth, with respect to the synchrons, is as-

sumed to reflect the thermal structure of the crust during cooling from the Cretaceous thermal event, and suggests a geothermal gradient during Cretaceous cooling between 28 and 44°C/km.

5.2. Age of Metamorphism and Deformation in the Polinik Unit (Table 9)

5.2.1. Age of Amphibolite Facies Metamorphism in the Polinik Unit

In the Polinik Unit, K/Ar ages of muscovites and biotites are uniform and range between 76-83 Ma (Ox-BURGH et al, 1966), (Fig. 41). The dated muscovites and biotites form part of the equilibrium amphibolite facies mineral assemblage (st - gtll - bi - musll - ky - plag - qz). However, an older generation of muscovite (mus I) can be distinguished using chemical and textural criteria, and biotites can show considerable variation in composition (ionic exchange with other Fe-Mg phases). Never-the-less, K/Ar mica ages are uniform, suggesting that all micas have been completely open to argon diffusion during peak metamorphism at ca. 105 Ma (see section 5.1.2, age province (2), and cooled through their blocking temperatures at ca. 80 Ma.

5.2.2. Age of D₃

The D_2 mica fabric and the inter D_2 - D_3 amphibolite facies mineral assemblages are deformed by D_3 structures. It is suggested, that D_3 occurred shortly after the Cretaceous thermal peak and continued until after the rocks had cooled below their mica blocking temperatures:

- D₃ mica crenulation hinges are often broken and fractured, suggesting low temperatures for deformation.
- 2) Quartz grains in F_3 fold hinges are elongate and aligned parallel to the F_3 fold axial surfaces, showing undulose extinction and no signs of annealing. VOLL (1960, 1976) estimates that temperatures of at least 275°C are needed before quartz recrystallises. Therefore D₃ deformation must have still been active at temperatures well below the blocking temperature of micas.
- 3) The similarity between co-existing muscovite and biotite ages suggests that cooling rates were fast, and hence one would not expect large age differences from samples at different structural depths. The expected uplift rates are estimated at 2-10 mm/a for geothermal gradients varying bet-

Text-Fig. 43. Graph illustra

Graph illustrating the variation of muscovite and biotite K/Ar ages with depth in the Altkristallin in the SE part of the Tauern area.

The pattern of ages in province (3A) and (3B) seem to be dependent on grain-size.



P-T-t PATH POLINIK UNIT



P-T-time trajectory for the Polinik Unit, using available P-T estimates from the Polinik Unit schists and garnet amphibolites.

Ages are constrained by available geochronological data. However, a Cretaceous age the high pressure assemblages, found in the Polinik Unit, is speculative (see text for discussion).

ween 25–45°C/km, and assuming that ΔT of the mica blocking temperatures is between 50–80°C. The observed maximum structural thickness presently exposed in the Polinik Unit is less than 2 km, regardless of whether this is calculated for the Polinik Unit before or after D₃.

5.2.3. Age of Pegmatites

Pegmatites in the Polinik Unit are folded by D_3 and contain an internal D_2 fabric. However, no evidence for D_1 is seen in the pegmatites, and it is therefore concluded that they pre-date D_2 and post-date D_1 . Rb/Sr two-mineral-isochron ages from the coarse-grained centre of the deformed pegmatite are ca. 254 Ma (CLIFF, pers. com.). It is assumed that this age is close to that of pegmatite emplacement, and has not been significantly reset during the Cretaceous metamorphism. Therefore D_2 is younger than 254 Må.

5.3. Ages of Metamorphism and Deformation in the Strieden Unit

5.3.1. Age Data

Geochronological data for the Strieden Unit consist of K/Ar whole rock and mineral ages for schists and mylonites (WATERS, 1976), and Rb/Sr and K/Ar ages for pegmatites (CLIFF, pers. com.).

The Strieden Unit (excluding the MMZ) lies within age province (3B), (Figs. 41-43). It is characterised by similar K/Ar muscovite and biotite ages which are intermediate between the Variscan and Cretaceous cooling ages (ca. 200 Ma). One sample is anomalous, with a biotite age which is older than the co-existing muscovite age (mus197/bi226, possibly due to excess argon). In general, muscovite and biotite ages become progressively older towards structurally higher levels in the south (muscovite ages increasing from 183 to 229 Ma, biotite ages increasing from 175 to 210 Ma).

At the base of the Strieden Unit, within the MMZ, mica ages also show a pattern of increasing ages towards the south, with muscovite ages between 94 and 103 Ma, and biotite ages between 89 and 127 Ma. One sample shows evidence of excess argon, with the biotite age exceeding the muscovite age (mus98/bi127).

The northern margin of the MMZ is marked by subvertical brittle fault zones, which separates the MMZ from the Polinik Unit. However, though there is a change in the age pattern across the Polinik Unit-Strieden Unit boundary, it is not dramatic.

5.3.2. Age of Metamorphism in the Strieden Unit

The high temperature mineral assemblages in the Strieden Unit (formation of andalusite, staurolite, sillimanite) are interpreted as having a Variscan age. The pattern of K/Ar mica ages suggests that rocks in the Strieden Unit have only been partially outgassed during a Cretaceous thermal event (see description of age province 3 in section 5.1.2). It seems likely then that Cretaceous temperatures never exceeded that at which muscovite is completely outgassed (i.e <400°C) (Fig. 42). P/T estimates for the peak metamorphic mineral assemblages, suggest that temperatures exceeded 650°C (section 4). This metamorphism must therefore be older than the Cretaceous and is probably related to the mica cooling ages of province 1. In this case, the oldest muscovite age, which is 320 Ma, yields a minimum age for the Variscan peak metamorphism.

Pegmatites, which occur abundantly within sillimanite-bearing schists, post-date the formation of sillimanite and andalusite and pre-date the D_n deformation. A Rb/Sr whole rock age of 266 ± 6 Ma has been obtained by CLIFF (pers. com., sample PTA2 from the Trögersee area) from a pegmatite showing these structural relations. The Cretaceous metamorphism did not reach high enough temperatures to reset significantly the Rb/ Sr system in a coarse-grained pegmatite, and therefore this age must be close to, or at least is a minimum age for the emplacement of the pegmatite. Therefore sillimanite and andalusite growth occurred prior to 266 Ma, most likely during the Variscan metamorphism.

5.3.3. Age of Mylonites (D_p)

 D_p deformation affected the high temperature mineral assemblages in the Strieden Unit, and pegmatites with a minimum age of emplacement of 266 Ma, therefore putting an upper age constraint of 266 Ma on the age of D_p deformation. D_p structures are cut by undeformed dacitic dykes which yield ages between 30 and 40 Ma (DEUTSCH, 1980), placing a minimum age constraint on D_n .

The following evidence suggests that the D_p deformation occurred at some time between 90 and 60 Ma:

 Quartz microfabrics in the mylonites formed at low temperatures. Quartz forms ribbons with aligned subgrains which have a pronounced crystallographic preferred orientation and show undulose extinction. These textures suggest that quartz has not annealed into polygonal grains, but continued to de-

POLINIK UNIT	STRIEDEN UNIT
D ₁ >250 Ma : stretching lineation > 250 Ma : emplacement of leucocratic rocks (they cut across &1)	D _m , D _n >> 320 Ma : pre-date Variscan metamorphism M _v > 320 Ma : peak of Variscan metamorphism: andalusite overprinted by sillimanite; partial melting
\sim 250 Ma : emplacement of pegmatites	D_{0} between 320 and 266 Ma: restricted to lower sillimanite zone \sim 266 Ma: emplacement of pegmatites
D ₂ <250 Ma, possibly between 135 - 105 Ma : folding; foliation in pegmatites; possibly associated with underplating of the Polinik Unit resulting in high pressure mineral	
assemblages of the eclogite amphibolites. $M_{C} \sim 105 Ma$: peak metamorphic temperatures during the Cretaceous metamorphic event, recorded in metapelites (600°C \pm 25°C, 6 - 7 kb) and in the hydrous amphibolite facies alterations	M $^{\circ}$ 105 Ma : lower greenschist facies metamorphic conditions
of the eclogite amphibolites. D ₃ 80 - 90 Ma : cooling of Polinik Unithhrough the K/Ar mica blocking temperatures;	D _p 90 - 60 Ma : cooling below the mica K/Ar blocking temperatures; formation of the main mylonite zone (MMZ)
<pre>~3 'O'U'U' ~ 65 Ma : final stage of emplacement of Polinik Unit onto the Lower East Alpine and the Peripheral Schieferhülle Units.</pre>	Dq between 60 and 40 Ma : folding of the MMZ
M _T 22 - 35 Ma Tauerncrystallisation: reheating of the base of the Polinik Unit in sub-area 5 D. < 29+6 Ma arching (folding) of the Polinik Unit and the base of the	30 - 40 Ma : emplacement of dykes
Altkristallin (contact surface with the underlying units); flexural slip faults	< 30 Ma : faulting; movement along the Ragge Teuchl fault zone
≤ 22Ma faulting; movement along the Mölltal fault zone and other subvertical faults	

form, by slip along preferred crystallographic planes, well under its recrystallisation temperature (ca. 275°C, VoLL 1960, 1976). It therefore seems that D_p continued after the Strieden Unit had cooled through the mica blocking temperatures (ca. 350°C), i.e. is younger than ca. 90 Ma.

2) WATERS (1976) obtained a whole rock K/Ar age of 61 Ma from a fine-grained sericite- and chlorite-rich mylonite. This can be interpreted as dating the end of D_p deformation, assuming that the system has not cooled significantly below the temperature of beginning of retention of argon and that the mechanical grain-size reduction that occurred during D_p was sufficient to outgass the micas and reset the argon system. The rock therefore became closed when intense deformation ceased.

5.3.4. Age of D_q and Later Deformation

 D_q structures deform the D_p mylonites, but are cut by dykes. It therefore must have occurred before ca. 40 Ma. An older age limit is more difficult to determine and depends on the interpretation of the D_p deformation age and correlation with deformation in the Polinik Unit. Movement on the Teuchl fault zone post-date D_q and also the emplacement of dykes, and therefore are younger than 40 Ma.

5.4. Geological History

The ages of deformation and metamorphism (Table 9), in the Polinik and Strieden Units, place constraints on their burial and temperature histories (Fig. 45). Estimated depths of burial are average depths, and these

are large compared to thicknesses of the individual units. They can be derived from average metamorphic pressures, assuming that tectonic stresses are not significant (<0.5 kbars) and that 1 kbar is equivalent to ca. 3.5 km of overburden. Also, simple models of cooling (outlined in Fig. 40), and assumptions about the geothermal gradient can be used in conjunction with K/ Ar mineral ages to give estimates of depth at particular times. Finally, the sedimentary history of the cover rock sequences, which overlie the Altkristallin, provide valuable information about uplift or subsidence.

The depth of burial history, when combined with the structural and metamorphic history of the study area, forms a powerful method for understanding the tectonic evolution of this part of the Eastern Alps (Fig. 45).

5.4.1. Pre-Variscan and Variscan Events

Evidence for the earliest structural and metamorphic events is well preserved only in the Strieden Unit, where metamorphic conditions were close to the Al_2SiO_5 triple point (5.5±1.5 kbars, 650±100°C, Section 4.3.4). This metamorphism occurred around 320 Ma (oldest muscovite K/Ar Variscan cooling age) and is Variscan in age (M_v). It post-dates at least two deformation events, D_n and D_m. At the time of peak metamorphic conditions, the Strieden Unit was at a depth of ca. 19 km. Temperatures appear to have been locally sufficient for partial melting, though there is no evidence for an unusually high geothermal gradient (e.g. greater than 25–30°C/km). Little can be said about the earlier deformation except that it involved crustal shortening, and may be Variscan or much older. D₁ in the Polinik Unit is also probably of Variscan age



Text-Fig. 45.

b) Temperature-Time graphs for the Strieden and Polinik Units and part of the Pennine Unit in the SE Tauern area, using all the available data (this page). Horizontal and vertical bars represent age and depth uncertainties respectively for a particular geological event. Where depth estimates are based on assumed geothermal gradients, derived from other evidence, the uncertainties are large.

a) Depth-Time (opposite page) and



because it pre-dates pegmatite emplacement, which has a minimum age of 254 Ma.

The regional K/Ar mica age pattern suggests that the Strieden Unit cooled through the biotite blocking temperature of ca. 300°C at ca. 300 Ma. The mixed ages in age province (3B) can be interpreted as a consequence of localised partial resetting of Variscan cooling ages, either during a Cretaceous reheating event, or by pegmatite intrusion at ca. 266 Ma. Assuming that the geothermal gradient during cooling after M, was at least as high as the average gradient during the peak metamorphic conditions during M_v (25-35°C/km), the Strieden Unit was probably not deeper than 12 km at ca. 300 Ma. Furthermore, the primary mineral assemblage in the pegmatites suggests maximum pressures of 3 kb (maximum depth of ca. 10 km) during their emplacement at 266 Ma. Thus, between the time of peak Variscan metamorphism (ca. 320 Ma) and pegmatite emplacement (ca. 266 Ma), the depth of the Strieden Unit progressively decreased by ca. 9 km. This suggests a history of tectonic or erosional unroofing in the period following peak Variscan metamorphism.

5.4.2. Permian to Jurassic History

From the Permian to the Late Jurassic, the Alpine region underwent crustal extension with the development of the Tethyan ocean between Adria and Europe.

In the Goldeck and southern Kreuzeck areas, the upper parts of the Altkristallin consist of low grade schists and phyllites. These are unconformably overlain by sediments (Drauzug sequences), with conglomerates (Gröden beds) at the base. The mid-Permian Gröden beds contain clasts of garnetiferous mica schists, tourmaline-rich schists, pegmatites, and phyllites. The unconformity surface dips at 30–40° SSW, and Variscan isograds in the Strieden Unit, and also the Cretaceous synchrons (discussed in section 5.1.2), generally dip at less than 20°S-SE. This implies that there has only been moderate tilting of the Strieden Unit and upper parts of the Altkristallin since the late Cretaceous. The structural thickness between the base of the Strieden Unit and the unconformity surface, normal to these tilted surfaces, is between 2.5 and 4.5 km. This estimate is based on the interpretation that the Strieden Unit (age province 3B) has had a similar burial history to that of age province 3A, and was overlain by rocks of age province 2 and 1 and the Mesozoic cover rock sequences. Assuming no subsequent thickening or thinning of this sequence, (see discussion in section 5.4.3), the Strieden Unit was at a depth of 2.5-4.5 km immediately prior to the deposition of the mid-Permian Gröden beds. Thus, there may have been several kilometres of further erosion since pegmatite emplacement at ca. 266 Ma, but prior to the mid Permian.

in the Drauzug sequences, the basal Gröden beds pass up into Triassic shallow marine carbonates, reaching a total thickness of ca. 5 km (TOLLMANN, 1977) suggesting slow subsidence. In the mid Jurassic, pelagic sediments, including radiolarites, Aptychus and red nodular limestones dominate (TOLLMANN, 1977). This change in the sedimentation pattern may be due to an increase in the rate at which the Tethyan ocean was opening, producing an increased water depth throughout the Tethyan region, estimated to be 2-3 km (JENKYNS, 1970). Such a water depth suggests that the Jurassic crust was thin and probably less than 12 km thick (Fig. 48). If so, the Strieden Unit must have been at a depth less than this, but deeper than the depth of the Mesozoic cover sequences, i.e. at a depth between 5-12 km.

5.4.3. Cretaceous Events

From the Early Cretaceous onwards, the tectonic development of the Eastern Alps can be described in terms of convergence between the two continental



Text-Fig. 46.

Map showing the distribution of metamorphic grades during the Cretaceous metamorphism in the Eastern Alps.

Modified after FRANK (1987). A = Altkristallin, D = Drauzug, G = Graz Palaeozoic, Gs = Gurktal sheet, F = Flysch Zone, NCA = Northern Calcareous Alps, P = Pennine Units, SA = Southern Alps.

landmasses, Adria and Europe, with a subduction zone dipping underneath Adria (HAWKESWORTH et al., 1978; PLATT, 1986; FRISCH, 1976) (Fig. 48). Evidence of crustal shortening can be found at all crustal levels:

- Deformation in upper crustal levels, during the Neocomian to Cenomanian, pre-date the deposition of the Gosau beds. (SEDGWICK 1830). These were deposited on top of deformed, folded and thrusted Mesozoic cover rock sequences. Sedimentation in the Gosau beds starts with Cenomanian shallow water clastic sediments.
- Early Cretaceous crustal shortening can also be demonstrated at the base of the Gurktal Alps, where Palaeozoic quartz phyllites are thrust onto Mesozoic cover sequences (von GOSEN, 1982). Here, mineral growth, dated at ca. 100 Ma (HAWKESWORTH, 1976) and post-dating the formation of the thrust contact, suggests that movement was older than 100 Ma.

The Strieden and Polinik Units had a complicated geological history during the Cretaceous, and at some point they were brought together. There is no evidence for deformation in the Strieden Unit between ca. 266 Ma and the Late Cretaceous, in contrast to the Polinik Unit, which experienced D_2 deformation during this period. Therefore, the Polinik Unit may have been widely separated from the Strieden Unit, both vertically and horizontally, before they were juxtaposed.

Strieden Unit

K/Ar mica ages suggest that most of the Altkristallin, in the southern Kreuzeck area just below the Drauzug unconformity, was heated above the biotite blocking temperature in the Cretaceous and cooled again through the biotite blocking temperature at c. 100 Ma (Fig. 45B). Temperatures in the Strieden Unit (age province (3B)) did not exceed 400°C during the Cretaceous reheating event. Peak temperatures are assumed to have been reached just prior to the oldest Cretaceous biotite cooling age (103 ± 2 Ma), which was measured in biotite co- existing with an undisturbed Variscan white mica (BREWER, 1970).

The amphibolite facies mineral assemblages in the Polinik Unit are considered to have crystallised during the peak of the Cretaceous thermal event (ca. 105 Ma). An average geothermal gradient of 22 to 34°C/km, at the Cretaceous thermal climax, can be calculated from the P/T conditions (Section 3.3.2). If this geothermal gradient is applicable to the Strieden Unit, then the Strieden Unit was at a temperature of 400°C and at a depth of 12 to 18 km at ca. 105 Ma (Fig. 47). It is interesting to note that the previous depth history of the Strieden Unit, outlined in section 5.4.2, places the Strieden Unit at the end of the Jurassic at a depth of 5-12 km, suggesting that an extra 0 to 13 km of overburden is required to account for the Cretaceous metamorphism. Thus, structural thinning in the Altkristallin, above the Strieden Unit, may have occurred since the Cretaceous. The Strieden Unit cooled through the biotite blocking temperature (ca. 300°C) between 90 and 100 Ma. Using the Cretaceous cooling geothermal gradient, calculated for age province (3A), this suggests that the Strieden Unit was at a depth between 6.5 and 12 km at 90-100 Ma. Typical P-T-time trajectories for a rock-pile undergoing uplift, suggest that the cooling geothermal gradient is usually higher than the peak metamorphic gradient, and this is compatible with the estimated gradients used here (22–34°C/km for peak metamorphism, and 25–45°C/km for cooling).

Polinik Unit

Amphibolite facies metamorphic pressures, deduced from metapelites, suggest that, the Polinik Unit during the temperature peak of the Cretaceous metamorphism (ca. 105 Ma), was at a depth of 20 to 26 km. If one considers that the crust during the Late Jurassic was probably only 12 km thick (see discussion, section 5.4.2), then a doubling of the crust (shortening by 50%) is necessary to account for the metamorphism in the Polinik Unit at ca. 105 Ma. D₂ deformation, which clearly pre-dates the Cretaceous peak metamorphic mineral assemblages in schists, might be an expression of the internal ductile shortening experienced by the deeper parts of the Altkristallin during crustal thickening. D₂ deformation is restricted to the Polinik Unit and no deformation of similar age has been observed in the Strieden Unit.

Eclogite amphibolite pods in the Polinik Unit record a history of progressive hydration and decompression from a high pressure metamorphism (in excess of 10 kbars) prior to the peak temperatures reached during the Cretaceous metamorphism. The hydration reactions suggest a history of warming, expulsion of water from the surrounding schists, and decompression, until P/T conditions were similar to those recorded by the surrounding schists. The ages of the primary eclogite assemblage and its alterations are not known, and it is possible that these are Cretaceous, suggesting that the Polinik Unit may have been at depths greater than 35 km in the Cretaceous, but prior to 105 Ma (Fig. 44). Burial of the Polinik Unit to these depths might have occurred by underthrusting in or near a subduction zone (Fig. 48). High pressure metamorphism in the Pennine units (estimated to have occurred at ca. 100 Ma in the Western Alps (HUNZIKER, 1974) is also attributed to burial in a subduction zone, beneath the structurally higher Pennine unites and the thickened Altkristallin with its stacked up cover rock sequences (Fig. 48).

The onset of subduction in the Eastern Alps is estimated at ca. 110-120 Ma. This is based on both the occurrence of detrital high- pressure minerals in Cenomanian-Turonian flysch (OBERHAUSER, 1968); and PLATT's (1987) estimate of the time required for high pressure rocks to reach the surface from a depth of 30 km (ca. 20 Ma).

Late Cretaceous Juxtaposition of the Polinik and Strieden Units

The history outlined for the Polinik Unit during the Creataceous is markedly different from that for the Strieden Unit, and both the vertical and horizontal separation between the two units must have been large. The vertical separation between the Polinik Unit and the Strieden Unit during the thermal peak of Cretaceous metamorphism (ca. 105 Ma) is difficult to assess. If one assumes that the geothermal gradient, estimated from the P/T conditions of the amphibolite facies assemblages in Polinik Unit metapelites, applies also for the Strieden Unit, then the base of the Strieden Unit lay 7–8 km above the top of the Polinik Unit (Fig. 47) at ca. 105 Ma.



Text-Fig. 47. Pressure-temperature graph, showing the possible vertical separation between the Polinik and Strieden Units at the climax of the Cretaceous metamorphism (ca. 105 Ma).

An average geothermal gradient between 22 and 34°C/km has been calculated, derived from Cretaceous peak temperaure mineral assemblages in metapelites in the Polinik Unit, at this time. Assuming that this geothermal gradient is also applicable to the Strieden Unit, then the Strieden Unit was at a temperature of 400°C and at a depth of 12 to 18 km at ca. 105 Ma.

The history following peak metamorphic conditions is constrained by the K/Ar mica ages. The continuity of the K/Ar ages, right across the Polinik Unit/Strieden Unit boundary, suggests that their juxtaposition must have ocurred during cooling, when both units passed through the 300–380°C temperature interval, recording similar mica K/Ar cooling ages. Since 105 Ma, there has been a change from a possible large vertical structural separation (up to 10 km) between the Polinik and Strieden Units, to an essentially small one.

Significance of D_p deformation

The MMZ is an important discrete high strain zone of deformation, which forms the base of the Strieden Unit and appears to have been originally flat lying. It was probably active between 60 to 90 Ma, shortly after the Cretaceous peak metamorphism. D_p kinematic indicators show that the upper plate (Strieden Unit) moved

towards the NW relative to the lower plate (Polinik Unit). If the MMZ acted as a normal fault, along which the Strieden and Polinik Units were brought vertically closer together, it would account for the marked difference in peak Cretaceous temperatures and pressures (equivalent to 2-8 km structural thickness) between the two units. It would also account for the uplift and cooling of the Polinik Unit at ca. 80 Ma.

Other shear zones in the Altkristallin occur above the MMZ (e.g. Striedenkopf mylonite, Figs. 4, 19) and below, at the base of the Altkristallin, where it passes into the Matrei Zone and upper Pennine unit, (WALLIS, 1988)). Kinematic indicators in these shear zones are similar to D_p fabrics and also indicate NW movement of the upper plate relative to the lower plate.

Given a normal fault interpretation for the MMZ, it is suggested that D_3 in the Polinik Unit, which occurred

at high crustal levels during cooling of the Polinik Unit represents either,

- a widely distributed deformation in the footwall of the MMZ related to movement in the MMZ ($D_p = D_3$) or
- D_3 post-dates D_p and movement in the MMZ and is possibly related to D_q .

Late vertical movements

The Strieden and Polinik Units are presently juxtaposed along steep brittle fault zones, such as the Ragga – Teuchl and the Strugenkopf fault zones (Fig. 4), which truncate dykes dated at 30-40 Ma (DEUTSCH, 1984), (Fig. 20). The absence of the dykes in the Polinik Unit suggests a strike-slip component greater than 5 km (width of the study area) on the late brittle fault zones. This is because, if there were only vertical movements, one might expect to find deeper levels of the dykes exposed in the Polinik Unit.

The minimum estimate (2 km) for the vertical separation of the two units during Cretaceous peak metamorphism, could be explained by vertical movements on these brittle fault zones. However, this interpretation is not satisfactory, because it provides no role for the MMZ, which is a large zone of ductile shear. However, it is certain that the Polinik and Strieden Units were ultimately juxtaposed by movements on the Ragga-Teuchl and Strugenkopf fault zones, albeit relatively small ones (<2 km).

Regional metamorphic, structural and stratigraphic evidence or post-metamorphic crustal thinning

Metamorphic evidence

The high pressure metamorphism experienced regionally in the Pennine Units, attributed to burial in a subduction zone (SELVERSTONE, 1985; HOLLAND & RAY, 1986), is estimated to have occurred at 78±12 Ma (RAITH et al., 1978). Pressures in excess of 8 kbars have been measured in the Upper Pennine Unit (HOL-LAND & RAY, 1986) and pressures of at least 10 kbars in the lower Pennine Unit (SELVERSTONE, 1985). This suggests a depth of burial of 28 km and 35 km respectively. The total preserved overburden above the lower and upper Pennine Units, consisting of ca. 5 km of upper Pennine Unit (CLIFF et al., 1971) and a 10-15 km thick Austroalpine nappe pile (Altkristallin and the Mesozoic cover rocks), is far too thin in both cases to account for these pressures (15-20 km missing above the lower Pennine Unit, and 13- 18 km missing above the upper Pennine Unit).

In the Altkristallin, amphibolite facies Cretaceous metamorphism in the Polinik Unit requires an overburden of 20–26 km (ca. 6.5 kbars). However, the exposed overburden is at least 10 km too thin to account for the metamorphic pressures.

Cretaceous amphibolite facies metamorphism has been found elsewhere in the Altkristallin (Fig. 46), for instance in the Ötztal Alps, in the Schneeberg area (THÖNI, 1983; HOINKES, 1981; HOINKES et al., 1982). Here too, metamorphic conditions (6–7 kbars) require an overburden (ca. 23 km) which is greater than the presently observed overburden (ca. 13 km).

2 Structural evidence

At the base of the Altkristallin, in the SE Tauern area, WALLIS (1988) has recognised a km-wide zone of distributed shear, which can be traced into the underlying

Matrei Zone and the upper Pennine Unit. It post-dates the Cretaceous metamorphism. The extensional fabric in this shear zone formed under greenschist facies conditions, suggesting the same sense of shear as that in the MMZ (top to the NW sense of shear). Rb/Sr ages suggest that deformation was still active at 40-50 Ma. Kinematic and metamorphic studies suggest that at least 10 km of crustal thinning has been accomodated by this deformation zone. The extensional deformation described by WALLIS (1988), together with the D_p deformation in the MMZ at the base of the Strieden Unit, across which 2-8 km of structural thinning may have occurred, could account for 12-18 km of post-metamorphic crustal thinning. This is in good agreement with the estimated missing overburden required to explain the Cretaceous metamorphic conditions in the Pennine Units.

However, this does not explain the 'missing overburden problem' for the Tauern metamorphism at ca. 25 Ma, discussed by DROOP (1985). Pressures derived from mineral assemblages, dated at ca. 25 Ma in the Peripheral Schieferhülle, suggest an overburden between 22-30 km (DROOP, 1985), which appears to be incompatible with the thickness of the presently observed thrust pile. DROOP concludes that this missing overburden formed part of a thick thrust sheet (7±4 km thick), which overlay the Altkristallin (for instance, the Gurktal thrust sheet) and has been eroded away. DROOP suggests that the 'missing overburden' must be above the Altkistallin, as pressures estimated from the 30 Ma Rieserferner intrusion, which lies within the Altkristallin, suggest that it was emplaced at depths between 15 and 17 km (DROOP & CLIFF, unpubl. data).

However, the structural position of the Rieserferner intrusion within the Altkristallin is not clear, and it is possible that it is in fact right at the base of the Altkristallin. Present estimates of the thickness of the Altkristallin (12-15 km) and the overlying Drauzug (minimum thickness of 3 km) at 30 Ma may therefore be compatible with the depths of intrusion of the Rieserferner, without invoking missing overburden. But what about the peak metamorphic pressures in the Peripheral Schieferhülle during the Tauern Metamorphism at ca. 25 Ma? The K/Ar ages at the base of the Altkristallin have been only partially reset in the Tertiary, and hence the base of the Altkristallin was unlikely to have been much above 400°C at 25 Ma. Assuming that the geothermal gradient derived from DROOP's P/T estimates is applicable to the Altkristallin (20-27°C/km), then this suggests that the base of the Altkristallin was not deeper than 20 km at this time, and may have been shallower (say comparable to that of the Rieserferner, at 17 km). In this case, using DROOP's depth estimates for his assemblages, they are between 2 and 13 km below the base of the Altkristallin. This suggests that there may be missing crust at the base of the Altkristallin, though the lower depth estimates are compatible with the likely thickness of the Matrei Zone and Peripheral Schieferhülle (up to 8 km). An interesting possibility exists however, that crustal thinning occurred after the Tauern metamorphism (25 Ma), perhaps accommodated by the Mölltal fault zone, which acted as a normal fault zone. This deletion may be another manifestation of late-stage extension in an orogenic belt revealing the deeply buried metamorphic sequences (DEWEY, 1988).



Cartoon diagram illustrating the possible evolution of the Altkristallin as part of the Austroalpine sheet, since the Upper Jurassic. Diagrams represent NW-SE cross-sections. In the Jurassic, the total crustal thickness probably did not exceed 12 km (see text for discussion). Convergence resulted in crustal shortening, expressed as D2 deformation in the Polinik Unit. Subduction of the basal part of the Austroalpine sheet is based on the assumption, that the high pressure metamorphism, observed in amphibolites in the Polinik Unit, is Cretaceous. Crustal thickening ultimately lead to the mid-Cretaceous burial metamorphism with peak temperatures in the Polinik Unit of ca. 650°C. At that time, the Polinik Unit was at a depth of ca. 22 km (assumed to be the total thickness of the Austroalpine sheet) and was buried ca. 9 km below the Strieden Unit. Post-metamorphic extension in the Austroalpine sheet seems to coincide with the high pressure metamorphism (ca. 8 kb) observed in the Pennine Unit (HoL-LAND & RAY, 1985). During this time, the Austroalpine experienced uplift and extension.

Text-Fig. 48.

Post metamorphic extensional deformation is also a possible interpretation for the Grazer Palaeozoic area, in the eastern part of the Austrian Alps. Here the Grazer Palaeozoic, which has experienced a weak Variscan metamorphism, shows mixed K/Ar mica ages, which fall between the Variscan and Cretaceous mica cooling ages. These are interpreted as partially reset Variscan ages, during the Cretacous metamorphism. However, there is a sudden increase in metamorphic grade in the basal part of the Grazer Palaeozoic and also in the underlying Plattengneiss (a high strain, flat lying shear zone, which forms the top of the Altkristallin). This metamorphism yields Cretaceous K/Ar cooling ages of 80-90 Ma. This configuration suggests that the Plattengneiss shear zone acted as a low angle normal fault, which has substantially thinned the Cretaceous metamorphic sequence. Along this shear zone, hot (Cretaceous metamorphic event) Altkistallin basement rocks (yielding Cretaceous cooling ages) were juxtaposed with the 'cold' rocks (Grazer Palaeozoic sheet) on the down-thrown side of the low-angle normal fault, resulting in the resetting of Variscan mica ages.

3 Sedimentary evidence

At the highest structural level, within the Austroalpine domain, sedimentation of the mid Cretaceous Gosau beds was taking place at the same time as the postulated crustal thinning observed in the Altkristallin and the Pennine Unit. The Gosau beds were deposited in fault bounded basins above the deformed and thickened Austroalpine nappes (Fig. 48). They also unconformably lie on rocks of the Grazer Palaeozoic, where they form 1-2 km thick sequences. Sedimentation started in the Cenomanian with shallow water brackish marine deposits. However, during the Campanian to early Maastrichtian, green claystones were deposited below the carbonate compensation depth (FAUPL & SAUER, 1978), suggesting a water depth of several kilometres. The thickness of sediments (several and the water depth (also kilometres) several kilometres) suggest a major phase of regional crustal extension affecting the Austroalpine. The timing of the syn-tectonic deposition of the Gosau beds in fault bound basins coincides with both, major crustal thinning at deeper structural levels, within the Altkristallin and in part of the Pennine Units (WALLIS, 1988) and with the Cretaceous mica cooling ages found in the Altkristallin (e.g. Polinik Unit).

Summary of the Cretaceous events (Fig. 48)

The convergent tectonic history of the Eastern Alps is associated with crustal thickening, affecting both the Austroalpine and the Pennine units, and resulting in the Cretaceous burial metamorphism.

 D_2 deformation might be the expression of ductile shortening experienced by the Polinik Unit during this stage of crustal thickening. It pre-dates the Cretaceous metamorphism and is restricted to the Polinik Unit. The Cretaceous metamorphism reached amphibolite facies metamorphic conditions at ca. 105 Ma in the Polinik Unit (620±60°C and 6.25±1.25 kbars), whereas in the Strieden Unit temperatures did not exceed the K/Ar blocking temperature of micas (300-380°C).

A major phase of crustal extension took place after crustal thickening and the Cretaceous metamorphism. This affected the Austroalpine and the Pennine Units and caused substantial thinning (12–18 km) of the Cretaceous metamorphic sequence.

It is suggested that post-metamorphic crustal thinning occurred during D_p deformation. During D_p deformation, the Strieden Unit moved over the Polinik Unit in a NW direction, cutting out 2–8 km of the Cretaceous metamorphic sequence.

PLATT (1986) and SELVERSTONE (1985, 1988) have both recognized the metamorphic evidence for postmetamorphic thinning in the Eastern Alps. Contemporaneous regional extension in an overall convergent tectonic setting is explained by PLATT (1986) as a consequence of gravitational collapse at high levels in a weak orogenic wedge, which is being continuously thickened at depth by underplating. This extension also provides a powerful mechanism, along with erosion, for bringing up deeply burried rocks (PLATT 1986; BEHRMANN et al., 1986).

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Folding in the Polinik Unit.

- Fig. 1: View of the southern slopes of Mt. Polinik (2784 m), showing hinge area (arrowed) of F_3 fold.
- Fig. 2: Detail of fold shown in Plate 1 Fig. 1.
- Arrow points to the fold hinge of F_3 fold. Fold axis trends approximately E and plunges gently towards the W (lefthand side of the picture). Other F_3 hinges and limbs can be seen above the arrow. Rock surfaces which dip subparallel to the mountain slope form the longer limbs of asymmetric F_3 folds. Viewed towards the W (down plunge) the asymmetry of F_3 folds is anticlockwise ("S" folds) with fold axial planes dipping towards the north at 30-45° (Plate 1 Figs. 3, 4).
- Fig. 3: Close-up of an F₂ fold closure (also shown in Plate 1 Figs. 1, 2, 4), showing the effect of minor F₃ folds (arrowed with subvertical axial planes) on an F₂ structure (arrowed with subhorizontal axial plane). Hammer (30 cm long) for scale.
- Fig. 4: Approximate profile section of F₃ folds (N-S section, viewed towards the E) from an outcrop above the one illustrated in Plate 1 Figs. 1-3.
 Arrow points to fold closure of F₂ fold. The interference pattern produced by superposition of minor F₂ and F₃ folds is a combination of "S" folds on "Z" folds when viewed down plunge of the F₃ folds (towards the west). Hammer in centre of picture (30 cm long) for scale.
- Fig. 5: Handspecimen of quartzo-feldspathic Polinik schist taken from the hinge area of a larger scale F₃ fold showing the preservation of small scale F₂ folds.
 Locality is marked by an arrow on Plate 1 Fig. 4. Quartzo-feldspathic layers are thickened in the hinge areas and attenuated on the limbs.
- Fig. 6: Looking towards the east along a F₃ fold nose (down plunge) developed in a foliated pegmatite surrounded by metapelites of the Polinik Unit.
 F₃ folds form cascading (subhorizontal) asymmetric folds striking approximately E with fold axes gently plunging towards the E (up to 6°). F₃ fold asymmetry is clockwise ("S" folds) when viewed down plunge (towards the E). Locality: 300 m NE of the lake Bodensee.
 Specimen 439; Harker collection number 147080.



Mineral assemblages and textures in aluminous metapelites of the Polinik Unit.

- Fig. 1: Photomicrograph (pp) of zoned garnet typically found in metapelites of the Polinik Unit. Garnet 1 (gt 1) forms the inclusion-free core-region and garnet 2 (gt 2) forms the rim zone of the whole garnet. The solid line marks the electron microprobe traverse. The mineral zoning profile of this garnet is shown in Fig. 10. Scale bar: 1 mm
 - Specimen 81.4; Harker collection number 147017.
- Fig. 2: Photomicrograph (cp) of twinned staurolite (st) and kyanite (ky) surrounded by muscovite, biotite and quartz. Scale bar: 1 mm

Specimen 81.4; Harker collection number 147017.

- Fig. 3: Photomicrograph (pp) of mineral assemblage typically found in metapelites of the Polinik Unit. It shows a lensoid domain within a mica-schist which contains the mineral assemblage ky - gt - st - qz - mi - bi - rut ilm - plag. Scale bar: 1 mm Specimen 81.2; Harker collection number 147016.
- Fig. 4: Photomicrographs (cp and pp) of S_2 mica cleavage folded by F_3 . Fig. 5: Kyanites (ky) are aligned within S_2 and have been deformed by F_3 . Kyanites are slightly altered along their margins to a fine aggregate rich in sericite and margarite. Scale bar: 1 mm
 - Specimen 85.1; Harker collection number 147021.
- Fig. 6: Photomicrograph (cp) of deformed kyanite (ky). Kyanite is kinked by F₃. Scale bar: 1 mm Specimen 77.16; Harker collection number 147019.





Sillimanite-bearing quartzitic biotite-rich schists of the Strieden Unit.

Fig. 1: Photomicrograph (cp) of sillimanite fibrolite developed in biotite pseudomorphs after garnet. Scale bar: 1 mm Specimen 408.6; Harker collection number 147204.

Staurolite-garnet-andalusite-bearing aluminous metapelites of the Strieden Unit.

- Fig. 2: Photomicrograph (pp, cp) of garnet- and staurolite rich graphitic mica-schist.
- Fig. 4: Staurolite prisms grow across foliation (internal foliation marked by graphite trails and aligned ilmenite). Scale bar: 1 mm
 - Specimen 157; Harker collection number 147247.
- Fig. 3: Photomicrograph (pp) of garnet in graphite- and ilmenite-rich muscovite schists. Garnet core shows internal foliation, the rim is free of inclusions. Locally the garnet rims are replaced by biotite. Also present are staurolite prisms which overgrow the foliation. Scale bar: 1 mm Specimen WÖ2; Harker collection number 147242.
- Fig. 5: Photomicrograph (pp) of euhedral staurolite grain overgrowing a graphite- and imenite-rich crenulated muscovite foliation. Other mineral phases present are biotite, quartz and garnet. Scale bar: 1 mm
 - Specimen 520.1; Harker collection number 147248.
- Fig. 6: Photomicrograph (cp) of euhedral staurolite grains which have overgrown a crenulated graphite- and ilmenite-rich muscovite foliation. Internal foliation of staurolites (graphite trails, ilmenite) are not as intensely folded compared with the external foliation. Scale bar: 1 mm Specimen 520.2; Harker collection number 147248.
- Fig. 7: Photomicrograph (pp) of staurolite- and garnet-bearing mica schist. Garnets are partly altered to biotite. There are two generations of staurolite: Staurolite 1 is rich in quartz inclusions and in this case aligned within the foliation and staurolite 2 (higher Zn-content) is free of quartz inclusions and cross-cuts both the foliation and staurolite 1. Scale bar: 1 mm Specimen WÖ1; Harker collection number 147230.
- Fig. 8: Photomicrograph (cp) of graphitic staurolite, garnet and andalusite-bearing schist.
 - Andalusite and staurolite porphyroblasts show an internal foliation marked by graphite and aligned ilmenite which can have an orientation different to the external foliation. Scale bar: 1 mm
 - Specimen 520: Harker collection number 147248.

Textural evidence for partial melting.

- Fig. 9: Small leucocratic pod (qz ksp plag bi sill) within aluminous metapelites (bi sill plag ksp qz). Pod is mantled by biotite and sillimanite-rich layer, and may represent an early partial melt (leucosome) surrounded by a restite (melanosome). Pod is a few cm across.
 - Locality: 150 m SSE of Latischalm hut.
- Fig. 10: Leucocratic layers (plag ksp qz sill bi) interfolded with biotite- (and sillimanite)-rich layers. This is interpreted as a melanosome and leucosome respectively which have undergone substantial deformation. The foliation is defined by the alignement of biotite laths, and is labelled S_{01} . This foliation is folded by F_{02} folds. Same locality as Plate 3 - Fig. 9.



Quartz mylonites of the Strieden Unit.

- Fig. 1: Polished section of a quartz mylonite which is orientated perpendicular to the mylonitic foliation and quartz elongation orientation (stretching lineation).
 Tight to isoclinal mm-scale folding occurs in upper part of the hand specimen at the contact with fine-grained mica-rich layers. Fold axes are sub-parallel to the mylonitic foliation. The rock is a quartzitic biotite-rich fine-grained gneiss (in the lower part of the photograph).
 Location: 550 NNE Strieden, ridge.
 Specimen: 382.
- Fig. 2: The mylonitc lineation (I_p) and foliation (S_p) are deformed by a tight fold (F_q) which has an axial plane oblique to the mylonitic foliation.

Locality: 700 m W of Latischalm hut.

- Fig. 3: Photomicrographs (pp, cp) of part of the quartz mylonite hand specimen illustrated in Plate 4 Fig. 1.
- Fig. 4: Sections are perpendicular to the mylonitic lineation, which is defined by the shape fabric and preferred crystallographic Fig. 5: orientation of quartz grains. Pure quartz bands are intensely interfolded with mica-rich layers. The fold axes are sub-parallel to the mylonitic lineation.

Scale bar: 1 mm. Locality: 550 NNE Strieden, ridge; same as Plate 4 - Fig. 1.

Specimen: 382.

Fig. 6: Photomicrograph (cp) showing shape and preferred crystallographic orientation in elongate quartz grains (quartz ribbons).

Quartz ribbons show undulose extinction and can by dynamically recristallised to fine-grained elongate quartz grains. Scale bar: 1 mm. Locality: 400 m S of Strugenkopf.

- Specimen: 327.
- Fig. 7: Photomicrographs (cp) of deformed asymmetric kink folds in quartz ribbons.
- Fig. 8: Section cut perpendicular to the foliation and mylonitc lineation. Mica-rich layers mark the fold axial planes of these kink folds. Scale bar: 1 mm.

Locality: 500 m SSE of Strugenkopf. Specimen: 329.





Pegmatite mylonites of the Strieden Unit.

- Fig. 1: Hand specimen of a pegmatite mylonite, cut perpendicular to the foliation.
- Fig. 2: H(A) shows a polished section of a hand specimen cut parallel and H(B) perpendicular to the lineation (marked as L). Cataclastically deformed tourmalines define rods oriented parallel to the lineation (stretching lineation). Location: 500 m E of Möllkopf.
 - Specimen: 331.
- Fig. 3: Photomicrographs (cp) of pegmatite mylonite with plagioclase and tourmaline porphyroclasts surrounded by quartz-sericite-rich matrix.

Thin section is cut parallel to the lineation (extensional lineation). There is a prominent quartz ribbon texture. Plagioclase clasts show narrowly spaced lamellar twinning (sometimes bent). Scale: Shorter side of the photograph is 4 mm. Locality: 450 m E of Möllkopf. Specimen: 251.B.

 Fig. 4: Photomicrograph (cp) of tourmaline-rich pegmatite mylonite. Thin section cut parallel to the extension lineation. Tourmalines are aligned within the foliation and are deformed by parallel micro-faults which are at a high angle to the foliation. Movement on these micro-faults is dextral resulting in extension of the porphyroclast within the foliation. Scale: Shorter side of the micrograph is 4 mm. Specimen: 331.

Microstructures in mylonites. Extensional crenulation cleavages.

Fig. 5: Photomicrograph (cp) of mm-scale banding defined by a compositional variation.

Layers rich in quartz, feldspar and white mica alternate with each other. Section cut parallel to the lineation. Phorphyroclasts include plagioclase and mica-fish. The fine-grained micaceous layers are deformed by a single parallel set of dextral extensional crenulation cleavages. Plagioclase porphyroclasts are cut by a single parallel set of sinistral microfaults. With progressive deformation the sinistral microfaults and the blocks they bound have undergone a clockwise rotation causing extension of the porphyroclast in the plane of the mylonitc foliation. The orientation of the shear bands together with the crystallographic preferred orientation of quartz has been used to deduce an overall dextral shear sense in a simple shear regime (chapter 7).

Scale bar: 1 mm. Locality: 200 m E of Möllkopf.

Specimen: 256.A; Harker collection number 147178.



- Fig. 6: Photomicrograph (cp) of an extensional crenulation cleavage developed in graphite- and mica-rich mylonite. The extensional crenulation cleavage postdates the mylonitic foliation. The mylonitic foliation between the extensional micro-faults is also folded and chemical alteration (chlorite-rich zones) occurs along micro-fractures oriented parallel to the extensional crenulation. Scale bar: 1 mm. Specimen: 171.1.
- Fig. 7: Photomicrograph (cp) of plagioclase porphyroclasts in an augengneiss mylonite. The large plagioclase porphyroclast is cut by parallel sinistral micro-faults. As movement occurred along these microfaults, the faults and the blocks they bound rotated clockwise and the porphyroclast was progressively stretched within the mylonitic foliation. The porphyroclasts are surrounded by a matrix which has undergone extreme grainsize reduction with quartz grains showing a preferred crystallographic orientation. A predominant single set of dextral shear bands (ex-tensional crenulation bands) is developed throughout the specimen. These shear bands have been used to deduce the overall shear sense in this rock (top moves to the right). Scale bar: 1 mm.
 - Specimen: 502.2.



Appendix A

pheng

phengite

		phlog	phlogopite
ab	albite	plag	plagioclase
alm	almandine component	prp	pyrope component
amph	amphibolite (non specific)	py	pyrite
an	anorthite component	qz	quartz
and	andalusite	rut	rutile
ар	apatite	sill	sillimanite
bi	biotite	spess	spessartine component
cc	calcite	sph	sphene
chl	chlorite	st	staurolite
срх	clinopyroxene (non specific)	sym	symplectite
ctd	chloritoid	tour	tourmaline
cum	cummingtonite	tr	tremolite
czo	clinozoisite component	tsch	tschermakite
di	diopside	zo	zoisite
dol	dolomite	zr	zircon
ер	epidote		
fsp	feldspar (non specific)		
g.hb	green hornblende		
gph	graphite		
gt	garnet		A 2 Other Abbreviations
gros	grossular component		A.Z. Other Abbreviations
ĥb	hornblende		
hem	hematite	P	Pressure
ilm	ilmenite	т	Temperature
jd	jadeite	t	Time
ksp	K-feldspar	R	Gas constant
ky	kyanite		(8.314 J K ⁻¹ ·mol ⁻¹ ; 1.9872 cal deg ⁻¹ ·mole ⁻¹
marg	margarite	т (К)	= T (degree centigrade) + 273.15
mc	mica (non specific)	к	Equilibrium constant
mi	white mica	X_i	Mole fraction of component i
mus	muscovite	μ	Chemical potential
mt	magnetite	Υi	Activity coefficient of component i
olig	oligoclase	L	Liquid
omph	omphacite	ai	Activity of component i
ore	unidentified opaque	pp	Parallel polars
or	ortoclase	ср	Crossed polars
par	pargasite	CPO	Crystallographic preferred orientation
pg	paragonite	MMZ	Main mylonite zone

Appendix B

B.1. Electron Microprobe Specification

Hardware

Probe designed and built in Cambridge. It uses a Si (Li) detector and pulse processor system interfaced to a Data General Nova 2 computer.

Software

Spectra processed by iterative peak stripping (STATHAM, 1976). Correction methods after SWEATMAN & LONG (1969). **Detection limits:**

Na		•	•	•	•				•	•	•	•			0.25	%
Mg.															0.15	%
AI,Si															0.10	%
K–Zn															0.05	%

Relative accuracy: 2 % for major elements (with at least 5 % of element present.

B.2. Methods Used in Calculating Fe³⁺ in Minerals from Probe Analyses

The following procedures were adopted for recalculation of Fe₂O₃ from mineral analyses using the NOVA program written by Leslie KANAT (1985) available in the Department of Earth Sciences, in Cambridge.

Garnets X3 Y2 T3 O12

Formula proportions were calculated on the basis of 12 oxygens and the following site assignments were made:

1) Al3+ was allocated to fill the tetrahedal sites and the remainder allocated to octahedral sites.

2) Fe²⁺ was converted to Fe³⁺ until the sum of Al³⁺ + Fe³ + Cr³⁺ + Ti⁴⁺ = 2.0.

A new anion total was then calculated and the formula renormalised to 12 oxygens. The procedure is iterative, until the difference between the initial anion sum and final anion sum is less than 0.00001. The garnet composition can then be expressed in terms of the propotions of the six end members.

Other phases

All iron was assumed to be ferrous in carbonate phases, chlorite, chloritoid, feldspar, sphene, tourmaline, mica, etc.
Appendix C

List of Hand Specimens and Thin Sections Used in this Text

The first number gives the Cambridge University (Depart-ment of Earth Sciences) Harker collection number and the second refers to numbers used in the text.

POLINIK UNIT

	POLINIK UNIT		Kyanite-staurolite-tourmaline-rich lensoid do-
Polinik Schis Garnetiferous	ts s feldspathic paragneisses and schists	147017/ 81.4	mains. Staurolite-kyanite twins. Coord. 46°53'59"N – 13° 9'52"E. Mörnigköpfe E.
Fine-grained (a	average grain size <1 mm)		Medium-grained metapelite.
146977/ 8	Coord. $46^{\circ}53'45''N - 13^{\circ}8'2''E$. Raggaalm E. Biotite-rich gneiss, deformed by kink-band. Gt - mi - bi - chl - qz - plag - tour - zr - or (hem) - ap. In the kinked zone micas are bent, and quartz shows undulose extinction.	147019/ 77.16	Bi – mi – gt – st – ky – chl – qz – plag – rut – sph – tour – or. Kyanite-staurolite-garnet-rich domains. Staurolite-kyanite-twins. Coord. 46°54'16"N – 13°10'20"E. Mörnigköpfe E. Fine-grained crenulated metapelite.
Medium-graine	d (average grain size -2 mm)		$B_i - m_i - g_i - s_i - ky - cn_i - q_i - p_i ag - tour - or.$
146988/ 77.5	Coord. 46°54'16"N - 13°10'20"E. Mörnigköpfe E. Tourmaline-rich, crenulated schist. Gt - mi - bi - qz - plag - or - tour - zr - ap	147020/441.7	Kinked kyanite is aligned within the crenulated mica foliation. Conjugate kink-bands and associated box-fold. Coord. $46^{\circ}54'16''N - 13^{\circ}9'25''E$.
146989/ 77.6	Coord. 46°54′16″N – 13°10′20″E. Mörnigköpfe E. Tourmaline-rich gneiss. Bi – mi – qz – plag – gt – chl – tour – or.		Medium-grained crenulated metapelite. Mi - bi - gt - st - ky (kinked) - qz - plag - tour - rut - ilm - chl - ap - zr. Kyanite deformed by l_2 -crenulation.
146990/ 77.7	Coord. $46^{\circ}54'16''N - 13^{\circ}10'20''E$. Mörnigköpfe E. Crenulated feldspathic plagioclase-rich gneiss. Bi - mi - qz - plag - gt - chl - or. Lensoid domains of plagioclase and quartz within the foliation. Parallel sets of shear- bands are at a shallow angle to foliaton.	147021/ 85.1	Coord. $46^{\circ}53'44''N - 13^{\circ},9'42''E$. Polinik. Crenulated metapelite. Mi - bi- gt - st - ky - marg - chl - qz - plag - tour - rut - sph - or (ilm). Kyanites are altered and surrounmded by fine-grained aggregates of sericite and mar- garite.
146991/ 77.9	Coord. $46^{\circ}54'16''N - 13^{\circ}10'20''E$. Mörnigköpfe E. Blastomylonite. Bi - mi - qz - plag - gt - chlor - tour - or - rut - ap.	147022/ 83.2	Coord. $46^{\circ}53'44''N - 13^{\circ}9'30''E$. Polinik. Coarse-grained metapelite. Mi - bi - gt - st - ky - marg - qz - plag - chl - tour - rut - sph - or (ilm). Fine-grained aggregates of margarite and
146993/ 77.10	Coord. 46°54'16"N - 13°10'20"E. Mörnigköpfe E. F_2 -fold core. Bi - mi - qz - plag ('gefüllte' texture) - gt - chl - ap - zr - or. Micas in fold-hinges recrystallised. Folded quartz band: quartz grains show un- dulous extinction, serrated grain boundaries are partly recrystallised (mortar texture). Coord. 46°54'16"N - 13°10'20"E	147023/ 67	sericite surround and partly replace kyanite. Coord. $46^{\circ}53'42''N - 13^{\circ}9'44''E$. Polinik E. Metapelite with domains rich in deformed pla- gioclase porphyroblasts (blastomylonite). Bi - mi - qz - plag - gt - st - chl - rut - zr - or (ilm) - ap. Quartz shows undulose extinction and has re- crystallised grain boundaries (mortar tex-
140993/ 77.11	Mörnigköpfe E. Crenulated schist, from F_3 -fold core. Bi – mi – qz – plag – gt – chl – tour – rut – sph – zr. Micas are bent; quartz has recrystallised along grain boundaries (mortar texture).	147024/ 68	tures); plagioclase porphyroclasts (augen tex- ture). Coord. 46°53'43"N - 13° 9'44"E. Polinik E. Tourmaline-rich metapelite. Mi - bi - gt - st - ky - qz - plag - chl - rut - or (im)
Coarse-grained 146999/ 81.5	d (average grain size >3 mm) Coord. 46°53′59″N - 13° 9′52″E. Mörnigköpfe E. Garnet-rich gneiss. Gt - bi - mi - qz - plag - chl - rut - ap - zr - or - (hem).	147025/ 77.3	Coord. 46°54'16"N - 13°10'20"E. Mörnigköpfe E. Medium-grained metapelite. Bi - mi - gt - st - qz - plag - chl - zr - rut - ap - or (ilm).
Fresh alumin	ous metapelites	Potrograsses	l metanolitos — various starss
147016/ 81.2	Coord. 46°53′59″N - 13° 9′52″E. Mörnigköpfe E. Medium-grained metapelite. Bi - mi - gt - ky - st - qz - plag -tour - zr - ·rut - or.	147031/ 4	Coord. 46°53'43"N - 13° 7'55"E. Raggaalm E. Coarse-grained retrogressed metapelite con- taining feldspar eyes.

Gt - st (1.5 % ZnO) - mi - bi - chl - qz -plag (2 % An) - tour - rut -sph - zr - or (ilm). Garnet forms elongate grains aligned within the foliation. 147034/ 83.4 Coord. 46°53'44"N - 13° 9'30"E. Polinik. Medium-grained retrogressed metapelite. Bi - mi - qz - plag - gt - chl - st - tour - zr - rut - or. Coord. 46°53'44"N - 13° 9'42"E. 147035/ 85.4 Polinik E. Crenulated retrogressed metapelite (no hand specimen). Bi - mi - qz - plag - gt - chl - tour - rut - zr – ilm Well-crystallised mica-fabric, two generations fine-grained mica of garnets, white pseudomorphs after staurolite and/or kyanite. Plagioclase porphyroblasts overgrowing crenulated mica fabric. 147032/ 7 Coord. 46°53'44"N - 13° 8'0"E. Raggaalm E. Blastomylonitic retrogressed metapelite: Bi - mi - qz - plag - gt - chl - st - tour - zr - ap - sph - or.

Quartz shows strain textures and serrated quartz grain boundaries which can be recrystallised (mortar texture).

Pegmatites

147080/439 Coord. 46°54′28″N - 13° 9′32″E. Bodensee S. Foliated pegmatite. Contact pegmatite/ metapelite: Qz - plag - mus - chl - bi - gt - ap - or. Sample taken from hinge-region of F₂ fold. Additional sample of separated kinked muscovite books.

STRIEDEN UNIT

Metapelites

Staurolite-rich

- **medium- to coarse-grained metapelites** 147238/195.1 Coord. 46°52'15"N - 13° 7'35"E.
 - 900 m S of Striedensee. Kink-bands in staurolite-rich quartzitic mica schist. St – bi – mus – or – chl – qz – plag. Euhedral staurolites (0.3 cm) overgrow mica foliation. Ilmenite exsolution in biotite. Micas are kinked. Staurolite is altered along fractures to a fine-grained aggregate of sericite and chlorite.
- 147241/554 Coord. 46°52′50″N 13° 5′54″E.
 350 m ENE of Steinwandalm hut. Andalusite-bearing, garnet-staurolite-mica schist. And - st - gt - qz - plag - mus - bi - ilm - zr - chl - ap - ?sill. Andalusite and staurolite porphyroblasts over-grow a crenulated (wavy) mica foliation. Alter-

ations of garnet and staurolite to biotite and chlorite (randomly orientated). Kink-bands developed in all micas. Andalusite can contain fibrolitic mineral with a high relief (posiibly sillimanite). Quartz bands with polygonal texture.

Graphite-rich metapelites

147242/WÖ2 Coord. 46°53'46"N – 13° 4' 3"E. 400 m S of Wöllatratten. Graphitic garnet-staurolite-bearing muscovite schist. Gt - st - mus - bi - ilm - chl - gph - plag - ep - zr.

Garnets (0.5–1 cm) have graphite-rich cores and clear rims. They have pressure shadows filled with quartz, biotite and chlorite which are aligned within the foliation and parallel to the lineation. Staurolites cut across the graphite-rich muscovite foliation. Coord. $46^{\circ}52'21''N - 13^{\circ}6'53''E$.

147247/157 450 m NNE of Striedenkopf. Graphitic garnet-staurolite-rich mica schist: Gt - st - mus - plag - bi - chl - qz - ilm - zr - tour - gph. Graphitic, plagioclase-muscovite-rich foliation overgrown by kinked biotite and euhedral staurolite. Garnet altered to biotite and chlorite. Biotite altered to chlorite. Coord. 46°52'39"N - 13° 5' 3"E. 147248/520 800 m W of Steinwandalm hut. Graphitic staurolite-garnet-rich schist. Gt - st - mus - bi - plag - qz - ilm - gph zr - or (hem). Graphitic muscovite-plagioclase-rich foliation cross-cut by biotite blades and euhedral staurolite grains. Garnets are altered to biotite and chlorite. Coord. 46°52'39"N - 13° 5' 3"E. 147249/520 800 m W of Steinwandalm hut. Graphitic garnet-staurolite-rich schist. Gt - st - bi - mus - qz - plag - ilm - gph -zr - tour. Graphite-rich plagioclase porphyroblasts are aligned within graphite-rich muscovite foliation. Complex folding defined by graphite trails. Biotite-rich zones surrounding garnets. Staurolite (cm-size) cross-cus folded graphitic foliation.

Prismatic and alusite and alterations in guartz-segregation-veins

147194/399	Coord. 46°52′51″N - 13° 6′18″E. 200 m SW of Saukopf. Prismatic andalusite crystals with muscovite inclusions. Andalusite is altered along frac- tures to sillimanite, phengite, chlorite and rare margarite. Sillimanite fibrolite with up to 3 mm long fibres form fan-shaped patterns with the apex of the fans aligned along fractures. And - sill - pheng - chl - qz - sph - marg - ap.
147195/And	Coord. $46^{\circ}52'42''N - 13^{\circ}8'08''E$. 200 m SW of -Schneestallkopf. Prismatic andalusite crystals locally replaced by sillimanite, sericite, chlorite and chloritoid. Chloritoid seems to have grown prograd within the sericite aggregate. And - sill - mus - ctd - qz - chl - ab. Fan-shaped sillimanite fibrolite along frac- tures replaces andalusite.
Sillimanite-b	earing quartzitic biotite schists
147202/516.1	Coord. 46°53'14"N – 13° 6'58"E. 150 m S of Latischalm hut.

147202/516.1 Coord. 46°53'14"N - 13° 6'58"E.
150 m S of Latischalm hut. Medium-grained sillimanite schist. Sill - bi - mus - qz - plag - zr - ap - ilm - tour - rut. Biotite intergrown with sillimanite fibrolite. Plagioclase contains ca. 25 % anorthite. Ilmenite exsolution in biotite. Sillimanite fibrolite concentrated together with biotite in lensoid cm-sized domains which are aligned within the foliation.
147203/516.2 Coord. 46°53'14"N - 13° 6'58°E. 150 m S of Latischalm hut.

Medium-grained mica schist. Sill – bi – mus – qz – gt – plag – ilm – rut. Banded rock with alternating layers rich in

muscovite, feldspar and biotite. Biotite intergrown with sillimanite fibrolite. Rare are coarse-grained prismatic sillimanite grains. Plagioclase contains up to 38 % anorthite. Coord, 46°53'16"N - 13° 7' 0"E. 147204/408.6 100 m S of Latischalm hut. 147217/169 Garnet relics in sillimanite-bearing quartzitic schist. sill - bi - mus - plag - qz - gt - tour - ilm rut Biotite with exsolved ilmenite and intergrown with sillimanite fibrolite. Garnet relicts surrounded by sillimanite fibrolite and biotite. Representative garnet composition: pyrope 147219/ 91 1.2 %, spessartine 1.09 %, andratite 7.44 %, almandine 18.07 %, grossular 72.21 %. Sillimanite fibrolite partly altered to sericite. Crenulated quartz-sillimanite-rich layers. 147205/336 Coord. 46°53' 4"N - 13° 7' 0"E. 500 m SES of Latischalm hut, ridge. Medium-grained quartzitic biotite-sillimanite schist containing shear band. sill - bi - mus - plag - qz - chl - ap - or tour. **Pegmatites** Sillimanite partly replaced by sericite which mimics fibrolite growth. Sillimanite fibrolite in-147142/251.1 tergrown with biotite. Biotite intergrown with sillimanite shows a higher TiO₂ contnet (up to 2.7 wt-% TiO₂) compared with sillimanite free biotite (up to 2.3 wt-% TiO₂). 147207/ 31.3 Coord. 46°53' 6"N - 13° 7' 2"E. 400 m S of Latischalm hut. Medium-grained feldspathic biotite- rich sillimanite-bearing schist located within a metre of a pegmatite. sill - bi - mus - ksp - plag - qz - gt - tour ox – zr rut. Plagioclase porphyroblasts locally show 'gefüllte' textures. Ilmenite exsolution in biot-147151/251.A ite; sillimanite fibrolite can cross quartz and feldspar grain boundaries; inclusion of garnet in plagioclase (ca. 30 % an). Fluid inclusions in plagioclase and quartz. Coord. 46°52'47"N - 13° 7'37"E. 147211/302 50 m E of Striedensee. Medium-grained sillimanite-bearing biotite schist. sill - gt - bi - mus - qz - plag - ilm - ap tour - gph - zr. Ilmenite exsolution in biotite. Sillimanite fibrolite intergrown with biotite. Rare occurrence of prismatic sillimanite. Plagioclase contains approximately 16 % anorthite. Coord. 46°52'37"N - 13° 7'20"E. 147214/136 147157/331.A 400 m SW of Striedensee. Sill - st - gt - qz - plag - bi - mus - chl - zr - tour - ox (ilm) - gph. Web-textured garnets; graphite inclusions in plagioclase- and muscovite-rich layers; plagioclase contains 20 % anorthite. 147215/167 Coord. 46°52'59"N - 13° 7'59"E. 650 m NE of Striedensee. Medium-grained sillimanite-bearing garnetstaurolite schist. Sill – bi – mus – qz - plag - gt - st - ap - tour - zr - gph - ilm.Web-textured garnets in quart-rich zones. Garnet can be partly replaced by biotite 147168/331.B which can be intergrown with sillimanite fibrolite. Coord. 46°52'55"N - 13° 6'17"E. 147216/397 200 m W of Saukopf. Medium-grained sillimanite-rich kyanite-bearing metapelite. Sill - ky - plag - qz - ilm - tour - ap - mi bi.

Lensoid domains rich in biotite intergrown with sillimanite fibrolite. Kyanite (0.5 cm) in muscovite-rich domains. Specimen deformed by anastomosing shear foliation which is rich in opaques.

Coord, 46°53' 3"N - 13° 8'13"E. 400 m SW of Poliniksee, ridge. Medium-grained sillimanite-bearing guartzitic schist. Sill - bi - mus - qz - plag - chl - or (ilm) tour. Sillimanite fibrolite locally replaced by sericite. Plagioclase contains 11 % anorthite. Coord. 46°52'54"N - 13° 9' 7"E. 700 m SW of Kehlluckerlkopf. Medium-grained sillimanite-bearing biotiterich feldspathic schist. Sill - mus - bi - qz - plag - tour - hem - gt - chl - rut - ilm. Sillimanite fibrolite is altered to a fine-grained aggregate of sericite. Rare are coarse-grained prismatic sillimanite grains. Plagioclase con-tains 21-28 % anorthite. Plagioclase por-phyroblasts show 'gefüllte' textures.

Coord. 46°53'26"N - 13° 6'35"E. 450 m E of Möllkopf. Blastomylonite. Plag - ksp - qz - mus - or. Porphyroblasts (eyes) consist of fractured Kfeldspar, plagioclase and muscovite (cm-size). They are aligned within a foliation and surrounded by fine-grained quartz (quartz-ribbon texture) and sericite. Pull-apart fractures in feldspars are filled with quartz. A set of parallel micro-faults defines an extensional fracture cleavage.

Pegmatite mylonites

Coord. 46°53'26"N - 13° 6'35"E. 450 m E of Möllkopf. Blastomylonite with plagioclase and tourmaline phenocrysts. Thin section cut parallel to the lineation and perpendicular to the foliation. Plag – tour – gt – mus – ksp – or –ap. The mylonitic foliation consists of alternating layers rich in quartz (ribbon texture) and fractured plagioclase, fine-grained K-feldspar and tourmaline respectively. Fractured phenocrysts (cm-size) are all aligned with their long axes parallel to the lineation. Conjugate fracture-sets are developed in both plagioclase and tourmaline. Coord. 46°53'26"N - 13° 6'38"E. 500 m E of Möllkopf.

Blastomvlonite.

Thin section cut parallel to the lineation and perpendicular to the foliation.

Plag - qz - ksp - tour - gt - or - mi. Tourmalines are aligned within the mylonitic foliation and show pull-apart fractures. Plagioclase porphyroblasts are stretched within the foliation. They are deformed by a single set of microfaults oriented at shallow angle to the foliation.

Coord. 46°53'26"N - 13° 6'38"E. 500 m E of Möllkopf. Blastomylonite. Thin section cut perpendicular to both foliation and lineation. Plag - qz - ksp - tour - gt - or - mi. In this section no clear foliation is developed. Rounded porphyroclasts of plagioclase, rare K-feldspar and tourmaline are surrounded by a fine-grained aggregate of quartz-sericitefeldspar.

Augengneisses

Biotite-rich augengneisses

- 147174/502.1 Coord. 46°53′16″N 13° 8′14″E. 200 m W of Poliniksee. Biotite-augengneiss mylonite. Ksp - plag - bi - sph - or - qz - mus - chl ap - tour - zo - ep - rut. Eye-shaped composite porphyroclasts of granular plagioclase, K-feldspar and quartz. Groundmass is rich in sphen and apatite. Quartz shows crystallographic preferred orientation (ribbon-txture). Alternating zones rich in quartz and sphene-biotite-plagioclase. Biotite can be retrogressed to chlorite.
- 147179/256.A, 256.B Coord. 46°53'24"N 13° 6'22"E. 200 m E of Möllkopf. Mylonite with feldspar and muscovite porphyroclasts. Thin section cut parallel to the lineation and perpendicular to the foliation. Plag - ksp (microperthite) - mus -qz - or tour - chl - gt - ep. Alternating layers rich in sericite and quartz contain aligned porphyroclasts of muscovite (bent and kinked) and fractured plagioclase ('gefüllte' texture). Layers rich in granulated feldspar. In quartz layers grains are externely fine-grained and show crystallographic preferred orientation (c-axes oriented at a shallow angle to mylonitic foliation). Granulated finegrained garnets.

Quartzites

Quartz mylonites

147188/382.1 Coord. 46°52′52″N – 13° 6′54″E. 550 m NNE of Mt. Strieden, along ridge. Preferred crystallographic orientation in pure quartz layer.

Section cut perpendicular to the mylonitic foliation and lineation. Pure quartz bands (1-1.5 cm wide) are interlayered with tourmaline- and mica-rich layers.

Qz - tour - ap - mus - or. Tourmaline is fractured; mica is mainly sericite; rare are mica porphyroclasts. Large quartz grains show undulous extinction. Micaand ore-rich layers define mylonitic foliation.

All grain sizes less than 0,5 mm.

147186/382.2 Coord. 46°52'52"N - 13° 6'54"E. 550 m NNE of Strieden, ridge.

Pure quartz mylonite.

Section A cut perpendicular to the mylonitic foliation and lineation. Microtexture: quartz ribbons consist of single quartz grains with 0.05 mm-sized subgrains which show a common crystallographic preferred orientation and are orientated differently from the large ribbon-shaped grains (host grain). The contact of the quartz layer with the fine-grained schist is folded on a mm-scale with the fold axes oriented parallel to the mylonitic lineation.

147187/327 Coord. 46°53'32"N - 13° 6'29"E. 400 m S of Strugenkopf. Pseudotachylite-beraring quartz-mylonite. No hand specimen.

Qz - mus - or (hem) - chl - pseudotycylite. Quartz ribbons show undulose extinction. Length of quartz ribbons vary from 6 mm to 0.03 mm.

147190/329 Coord. 46°53'30"N - 13° 6'33"E. 500 m SSE of Strugenkopf.

Interfolded pure quartz layers with mica-rich layers.

Quartz ribbons have been deformed into asymmetric kink bands with mica-rich layers aligned axial planar to these folds. This deformation texture is locally reworked into a microbreccia along single sets of parallel extensional faults.

Qz - gt - tour - mus - or (hem).

BRITTLE FAULT ZONE

147278/171.1 Coord. 46°53'01"N - 13° 8'14"E.

500 m NNE of Schneestallkopf, ridge.

Conjugate set of extensional faults deforming mica-quartz-graphite-rich ultramylonite. Movement on these faults caused offsets and folding of the mylonitic foliation. Quartz crystallographic preferred orientation is deformed (folded) by these faults.

Qz - plag - mi - chl - rut - or - grph - tour - py.

Appendix	۵	
	Appendix	

Selected Microprobe Analyses

TINU XINI	85.4-2 37.370 21.370 22.331 32.331 4.671 1.712 91919	2,995 0,005 2,167 2,167 2,167 2,167 2,167 2,1767 0,130 0,130 0,130	76.520 4.897 18.583 20.474	81-8 37.555 31.555 21.635 35.185 35.185 2.895 2.264 2.264 2.264 2.264 2.264 100.355	3.006 2.042 2.042 2.356 2.356 0.3545 0.3545 0.154 0.154	85.784 2.408 11.808 12:787
POL 1	85.4-1 21.676 21.298 21.298 32.267 32.267 5.316 1.834 0.752 9.143	3.008 	76.579 2.162 21.259 22.695	81-5 81-5 38.740 21.976 21.958 31.958 6.226 0.226 0.226	3.022 2.035 2.035 2.035 0.741 0.741 7.967	71.820 2:839 25:341 26:219
	4-25 -18 37.741 21.634 21.634 34.600 34.600 0.371 0.571 100.140	2,992 0,008 2,013 2,013 2,294 2,294 0,013 0,013 7,997	77.709 1.625 20.665 21.188	81-1 F1m 37.680 21.204 21.204 21.204 2.443 2.443 2.443 2.443 2.443 2.443 2.443	3.026 2.007 2.393 2.393 0.322 0.166 0.166 7.971	87.122 1:933 10.945 11:844
	4-5 71m 38.210 21.521 21.521 33.253 33.253 33.253 33.253 33.253 33.253 33.253 33.253 33.253 33.253 33.253 33.210	3.052 - 2.026 - 2.221 0.137 0.137 0.137 0.137 0.137	82.508 3,492 14,000 15,262	81.1-17 81.1-17 37.542 21.097 21.097 33.853 3.590 3.591 3.501 0.557 0.557	3.026 2.005 2.262 2.262 0.371 0.371 0.371 7.971	85.740 1.636 12.624 13.990
	4-2 71m 38.345 21.337 0.269 33.641 4.954 0.315 100.587	3.020 9.016 2.216 2.216 0.0581 0.021 7.980	75.543 4:819 19.638 20.787	81.1-16 37.728 21.335 21.335 21.335 21.335 21.335 21.335 21.335 21.469 2.465 0.465 0.465	3.009 2.006 2.322 2.322 0.412 0.410 0.200 0.200	84.798 1:336 13:866 15.080
	8-11 37.872 21.425 29.452 29.452 29.452 2.348 6.250 6.250 100.824	3.014 2.010 1.960 2.21 0.278 0.278 0.291	80.552 10:028 9.419 12.439	81.1-15 37.831 21.539 33.498 1.358 1.00:223	3.008 2.019 2.227 2.227 0.516 0.162 0.162 7.983	80.823 1.709 17.468 18.820
	68-3 38.663 38.663 32.568 32.360 5.114 5.114 1.367 1.681 1.681 100.873	3.026 2.001 2.118 2.118 0.597 0.141 0.141	74.969 4.785 20.246 21.974	81.1-13 38.053 38.053 38.053 38.053 38.053 1.25 1.25 1.25 1.001152	3.016 2.010 2.137 2.137 0.563 0.196 0.196 0.196	79.006 1.942 19.053 20.838
	68-7 38-433 21-597 32-556 32-556 4.755 1.542 1.542 1.765	2 oxygens 3.022 2.022 2.141 2.141 0.560 0.1149 0.1149 7.977	76.006 5.038 18.956 20.721	81.1-12 .000 6 37.772 21:389 22:188 32.188 4.762 3.2708 0.708 0.708	2 oxygens 3.004 2.006 2.141 2.141 0.564 0.217 0.217 7.993	79.054 2.023 18.923 20.863
	68-1 riac 38.056 38.056 0.123 0.123 32.444 1.820 1.820 1.820 2.042 2.042 2.042	ne basts of 1 3.030 1.993 0.007 2.160 2.160 0.484 0.123 0.123	77.619 5.922 16.459 18.307	85.4-7 Fim 21.340 21.340 32.668 32.668 3.313 3.527 3.527 3.527 3.527 3.527	ne basis of 1 3.058 2.014 2.188 2.188 0.276 0.276 0.159 0.159	84,798 5,558 9,644 11,204
	81.3-38 >r1m 37.556 21:158 21:158 21:158 2.455 1.357 1.357 1.357 1.357 1.357	cations on ti 3.029 2.013 2.351 2.351 2.354 0.354 0.118 0.118 0.118	83.837 41.044 12.119 13.094	85.4-3 ->rim 37.594 21.55 2.596 32.596 3.532 2.030 99.241	cations on ti 3.022 2.05 2.192 2.192 2.192 0.423 0.423 0.423 0.423 0.423	79.719 5.930 14.351 16.184
Garnet	ox/ID S102 S102 Fe203 Fe203 Fe0 H100 M100 M100 Ca0 Ca0	formula S1 Aliv Aliv Fe3 Fe2 HM Mn2 Ca Ca	elmSps graAnd prp Mg no.	Garnet ox/ID 5102 8102 81203 76203 76203 7600 Mfn0 CaO CaO	formula Si Aliv Fe3 Fe2 Mn2 Mn2 Ca Sum5	almSps graAnd prp . Mg no.
NIK UNIT	77-9 rum 37.370 21.482 35.792 35.792 2.682 1.344 1.969 1.969	2.991 2.019 2.336 2.336 0.091 0.169	83.574 5.675 10.751 11.780	81.3-37 38.075 21.450 34.53 4.139 1.1088 1.1088 1.1088 1.1088	3.021 2.007 2.292 2.292 0.489 0.073 7.976	80.214 3.184 16.602 17.599
POLI	77-6 77-6 37-917 21-703 21-703 35-643 35-643 1.240 1.240 1.240 1.272 101-324	3.005 - - - - - - - - - - - - - - - - - -	82.915 5.677 11.408 12.468	81.3-36 38.027 21.1430 33.573 4.612 1.206 99.978 99.978	3.019 2.006 2.229 2.229 0.013 0.106 0.106	77.929 3.599 18.472 19.666
	77-3 20:503 20:853 0.201 34.999 0.124 0.124 0.124 3.092 0.596 2:141 2:141 99:509	3.021 9.012 0.012 2.358 0.018 0.031 0.041 0.185 0.185	81.181 6.255 12.564 13.602	81.3-35 38.118 21.590 33.030 4.625 1.336 1.334 100.053	3.019 2.016 2.188 2.188 2.188 0.016 7.973	77.555 3.853 18.582 19.969
	77-2 38.030 21.570 35.200 35.200 3.675 0.384 1.988 100.847	3.011 2.013 2.331 2.331 2.334 0.026 0.169 7.983	79.644 5.700 14.656 15.686	81.3-34 38.149 21:273 21:273 32:394 1.269 1:269 1:269 9:933	3.023 	76.383 3:705 19.912 21.507
	77-1 27.338 27.338 2.121 0.121 34.663 34.663 3.571 0.359 0.359 99.482	3.004 1.93 0.007 2.332 2.332 0.428 0.206 7.996	78.813 6.881 14.306 15.511	81.3-33 38.207 21.461 2.0462 32.462 32.462 1.809 1.809 1.257 1.257 1.267	3.020 2.000 2.146 2.146 0.587 0.587 0.121 0.128 7:980	76.579 3.596 19.825 21.1475
	83-26 r1m 37.468 37.468 21.201 21.201 34.212 0.492 3.429 3.429 2.016 1.329 10.147	2.997 0.996 1.996 1.996 2.289 0.030 0.030 0.137 7.974	82.270 3.864 13.866 15.154	81.3-32 0016	3.018 	76.240 3.989 19.772 21.580
	83-19 37.291 37.291 21.403 34.432 34.432 34.432 34.233 1.3092 1.300 99.541	2.994 0.006 2.312 2.312 0.063 0.083 0.112 7.993	80.572 3.762 15.766	81.3-11 37.701 21.360 35.52 35.52 35.52 35.52 35.52 1.403 1.696 1.696 1.696 1.696 1.696 1.003 349	3.019 2.016 2.016 2.379 2.379 0.318 0.0145 0.1145 7.973	84.217 4:953 10.830 11.796
	83-18 37.487 21.599 33.468 4.51 1.595 1.639 100.139	2 oxygens 2.984 0.016 2.011 2.011 2.228 2.228 0.1140 0.1140 0.1140 8.003	78.073 4.673 17.254 18.809	81.3-3 37.982 21.464 33.649 33.649 1.4018 0.439 2.227 2.227 99.779	2 oxy8ena 3.022 2.014 2.239 2.239 0.476 0.030 0.190 0.190	77.298 6.469 16.233 17:545
_	83-17 37.409 21:572 33.033 4.300 1.881 1.464 1:464 99.749	re basts of 1 2.987 2.013 2.017 2.017 2.206 2.206 0.127 0.127 0.127 0.127 7.998	78.272 4.202 17.526 19.148	777-11 37.998 21.646 3557 35.157 2.4477 1.14511 1.145111 1.145111 1.145111 1.145111 1.1451111111111	ne basis or - 3.029 2.034 2.384 2.384 2.384 0.098 0.098 0.115 0.115	85.834 3.988 10.177 10.988
(METAPELITES)	83-16 core 37.371 21.116 0:009 33.930 33.930 33.930 1.676 1.676 1.344 1.344 99:523	cations on ti 3.002 1.999 0:001 2.279 0.114 0.114 0.114 7.998	79.856 3.860 16.284 17.636	77-10 37.398 21.369 35.811 35.81 3.199 3.199 0.575 1.615 1.1815	cattona con ci 2.997 0.003 2.016 2.016 2.400 2.400 0.382 0.382 0.156 7.993	81.931 5.235 12.834 13.733
CARNETS (ox\ID \$102 Fe203 Fe203 Fe0 H102 H102 H102 H102 H102 H102 H102 H10	formula Si Aliv Alvi Fe3 Fe2 Mm2 Ca Sum5	almSps graAnd prp Mg no.	carnet ox/ID \$1203 \$100\$	formula Aliv Aliv Aliv Fe2 Fe2 Fe2 Mn2 Ca Sum5	almSps grsAnd prp Mg no.

NIK UNIT	4–9 35,965 18,421 19,421 19,421 2,025 9,312 9,312 0,110 8,526 94,031	5,515 3,539 2,539 0,234 2,135 2,135 2,135 0,018 1,668				
FOLI	8-12 36.842 18.524 18.524 2.146 9.962 0.136 0.136 9.62 0.136	5.560 3.3247 2.3247 2.3244 2.3244 2.244 0.2244 0.2440 0.2440 0.2440 0.2440 0.2440 0.2440 0.2440000000000				
	68-9 30.641 19.150 25.464 11.20 11.20 11.20 2.57 2.536 89.895	4,959 4,959 3,4448 3,4448 3,444 6,44 0,055 1,447 0,055 1,478 1,478				
	68-18 27,115 21,115 24,800 0,473 11,2998 11,2998 11,2998 0,230 0,230 0,230 2,128 2,128 2,128 89,611	4,401 3,375 3,375 3,375 3,375 3,375 0,058 0,058 0,058 0,029 0,028 0,029 0,028 0,029 0,029 0,029 0,029 0,029 0,029 0,020 0,00000000	81-23 36.554 19.647	1.579 10.416 8.043 95.289 95.289 3.477	0.178 2:326 0.018 1.537 15.379	
	68-10 36.265 19.326 19.326 1.372 9.757 9.757 8.765 8.765 95.006	5, 487 33, 487 2, 2482 2, 260 2, 156 2, 256 2, 262 1, 62 1, 62 1, 62 1, 62	81-14 36.690 19.607	94.415 5.501 5.501 5.501 5.501 5.501 5.501	2.485 2.485 1.587 1.587	
	81.3-13 35.794 19.549 1.383 9.645 - - - - - - - - - - - - - - - - - - -	5.458 3.458 2.4518 2.192 2.192 2.192 2.192 1.594 1.594 1.541 15.430	81-7 36.394 19.758	10.016 10.016 10.016 8.560 8.560 95.343 5.474 5.474	2:245 2:245 1.542 1.542 1.542 1.542	
	81.3-8 34:533 19:067 21.337 0.888 9.571 9.571 9.571 0.094 0.104 0.104	5.387 3.784 2.784 0.104 2.225 2.225 0.016 1.436 1.436 1.487	81-5 35.441 19.168	8:907 8:907 8:442 8.442 94.093 5.452 5.452	2:042 2:042 1.657 1.657 15.456	
	77-13 35.667 19.176 20.989 1.871 8.071 8.071 6.071 9.048 9.048 94.883	2 oxygens 5.469 3.466 3.466 3.466 1.844 1.844 1.844 1.70 1.770 1.770 15.477	81.1-18 36.219 19.419	1.503 10.072 6.621 93.838 93.838 93.838 5.510 5.510	2:284 2:284 1.673 1.673 1.673	
~	77-5 35.281 20.196 20.178 0.855 8.518 8.518 8.518 8.518 8.518 8.518 8.518	s basts of 26 5.426 2.662 2.662 1.952 1.952 1.952 1.952 1.952 1.952 1.952 1.478	85.4-8 34.911 18.515 21.637	1.353 7.975 0.213 0.063 8.958 93.611 93.465 5.465 5.465	0.159 1.861 1.861 0.028 1.78 1.70 15.570	
(METAPELITES	83-27 35.105 18.105 2.015 2.015 7.1869 0.1105 0.1105 0.1105 0.105 0.080 0.080 93.943	ations on the 5.463 3.7368 3.7368 0.2366 0.236 1.825 1.825 0.014 1.792 1.792 0.021 15.511	4-27 35.741 18.903 22.493	1.654 6.998 6.998 7.77 94.566 94.566 94.566 5.528 3.447	0.192 1.613 1.732 15.422	
BIOTITES	oxVID S102 F 41203 F 4203 F 400 MaC MaC CaO Ma20 K20 CaO CaO CaO CaO CaO CaO CaO CaO CaO CaO	formula o S1 A1 T1 T1 Mm2 Mm2 Ca Na K K Ca Sum5	Biotites ox\ID SiO2 Al203 Fe0	T102 Mg0 Mn0 Ca0 Na20 K20 C1 S1 S1 S1 S1	TI Ma Mhr2 Ca Ca Ca Ca Ca Sums	
NIK UNIT	4-16 46.248 46.248 34.836 0.494 0.1323 0.103 0.113 9.783 9.783 9.783 9.783 9.783 9.783 9.783	6.240 5.542 0.140 0.035 0.035 0.035 0.015 0.015 0.015 1.3.836	41.557 0.946 0.008			
FOLI	68-16 46.638 16.638 0.697 0.208 0.208 0.208 0.237 1.248 8:537 8:537 8:537	6.226 5.2693 0.078 0.078 0.078 0.323 1.454 1.454 1.454	0.818 0.182	81-11 49.364 33.3698 33.598 1.567 0.647 1.294 - - 8.687 8.687 8.687 - - 95.936	6.445 6.445 6.171 0.171 0.252 0.252 0.252 1.447 1.447 1.447 13.768	59.539 0.854 0.146
	68-11 35.140 35.140 0.784 0.466 0.466 0.436 0.436 1.104 8.932 8.932 8.932	6.235 5.235 5.568 0.087 0.087 0.087 0.287 1.230 1.230	49.774 0.842 0.158	81-29 47-934 35.224 35.254 35.	6.317 5.143 5.143 0.143 0.143 0.143 0.143 1.133 1.135 1.135 1.136 1.136 1.136 1.136 1.136 1.136 1.136 1.146	50.660 0.882 0.118
	81.3-24 45.493 37.033 0.186 0.186 - - 1.426 8.279 92.852	6.111 5.005 5.005 0.019 0.019 0.019 0.371 1.371 13.833	0.793 0.207	81.1-10 45.031 36.046 36.046 36.046 36.046 36.046 1.546 1.546 1.546 1.546 1.546 1.473	6.136 5.032 0.040 0.040 0.408 0.408 1.331 13.798	- 0.765 0.235
	81.3-18 53.296 32.831 32.831 0.483 0.219 0.219 0.219 1.332 1.332 7.054 7.054	6.850 4.975 0.052 0.012 0.332 1.157 13.407	44.690 0.777 0.223	85.4-26 45.265 36.2182 36.234 0.333 0.334 0.105 9.368 0.105 9.388 9.368 9.368 9.368 9.368	6.101 5.101 5.105 5.105 0.047 0.047 0.047 0.047 1.13 1.13 13.884	31.400 0.936 0.055 0.008
	77-19 48.349 30.711 0.720 1.572 1.572 1.572 1.572 1.572 0.316 9.316 92.319	6.581 4.892 0.074 0.319 0.319 0.319 0.083 1.591 1.591 13.736	62.081 0.950 -	85.4-18 46.145 05.783 05.783 0.499 0.306 0	6.183 5.050 0.050 0.061 0.061 1.507 13.845	40.654 0.833 0.167
	77-16 345-616 345-616 0.577 0.577 0.577 0.577 0.544 1.123 8.588 8.588 8.588 8.588	6.194 5.587 5.587 0.054 0.094 1.488 13.845	42,254 0,834 0,166	85.4-16 46.084 35.4489 35.4489 0.531 0.533 0.533 0.533 0.533 1.457 1.457 1.457 1.457 339 8.339 93.035	6.198 6.198 0.054 0.053 0.053 0.053 0.053 1.431 1.431 1.431 13.840	34.958 0.790 0.210
	77-15 46.335 1.4750 1.4750 0.748 0.748 0.931 0.931 0.931 8.838 8.838 8.838 8.838 8.838	2 oxygens 6.285 5.365 5.365 0.076 0.076 0.188 0.188 0.188 1.530 13.830	53.021 0.875 0.125	85.4-9 45.828 35.65 0.7615 0.582 0.582 0.582 0.282 7.394 7.394 7.394 7.394 7.394	2 oxygens 6.153 5.692 5.692 5.692 0.055 0.056 0.056 1.266 1.266 13.839	39.772 0.706 0.294
ES)	85-13 45.336 0.5253 0.551 0.551 0.551 0.457 0.551 0.457 0.551 0.5530 0.5530 0.5530 0.5530000000000	e basis of 2 6.166 6.165 0.056 0.093 0.093 0.382 1.386 1.386 13.835	чв.063 0.784 0.216 -	4-29 47.481 30.5774 2.5569 0.765 1.861 1.861 1.861 1.861 2.566 8.580 8.566	(1) (1) (1) (1) (1) (1) (1) (1) (1) (1)	56.349 0:913 0.087 -
IS (METAPELIT	83-4 141.781 355.628 15.162 0.217 0.217 0.213 0.213 0.513 9.526 9.526	ations on ti 6.127 6.127 0.128 0.054 0.054 - - 136 136 13.877	29.501 0.924 0.076 -	4-20 45,503 1,4130 1,413 0,407 0,407 0,407 0,407 0,408 9,4082	cations on ti 5.102 5.172 0.162 0.064 0.081 0.015 - 1.687 1.687 13.822	33.449 0.991 0.009
MUSCOV ITE	ox (ID \$102 \$102 \$102 \$102 \$102 \$102 \$102 \$102	formuta (Si Alvi Fe2 Fe2 Mm2 Mm2 Ca Na Na Ca Sums	Mg no. K Na Ca Muscovite	ox\ID \$102 \$102 \$102 \$760 \$7102 \$7102 \$7102 \$7102 \$720 \$120 \$20 \$20 \$20 \$20 \$20 \$20 \$20 \$20 \$20 \$	formula S1 Fe2 Fe2 Mm2 Ca Mm2 Ca K K Sum5 Sum5	Mg no. Na

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					NIK UNIT	81-20 32.276 44.57 41.372 41.312 8.617 8.617 8.617 7.981 84.83 0.208 0.208 0.208 0.206 0.207 0.208 0.206 0.208 0.208 0.206 0.208 0000000000
					TES) POLI	81.3-42 31.6154 31.6154 31.6154 0.5718 0.5718 0.5718 2.504 2.504 2.504 0.0739 0.0739 0.0739 0.0739 0.0739 0.059 0.059 0.059 0.059 0.059
					AES (METAPELI	81.3-41 36.064 36.170 5:506 5:506 0.17426 0.141 7.425 0.141 2.141 2.141 2.141 2.141 2.141 2.141 2.141 2.322 0.966 0.967 0.967 0.967 0.967 0.967 0.972 0.967 0.972 0.967 0.972 0.967 0.9720 0.9720 0.9720 0.9720 0.9720 0.9720 0.9720 0.9720 0.9720 0.97200 0.97200 0.97200 0.97200000000000000000000000000000000000
					TOURMALIN	ox/ID S102 S102 Fro Fro Fro Fro S1 S1 S1 S1 S1 S1 S1 S1 S1 S1 S1 S1 S1
INIK UNIT	85.4-21 rutule 0.552 0.556 95.133 96.21 11mentte 0.006 0.909 0.909				TINU NIK	81-26 264-184 264-184 0.155 9-175 9-175 9-175 9-175 9-175 9-175 9-172 9-174 9-172 9-172 9-174 9-172 9-174 9-174 9-174 9-172 9-174 9-
POL	85.4-19 11mentte 0.5512 0.5512 0.5512 9.0.423 5.0.423 9.0.124 9.0.124 9.0.125 1.722 1.722 1.722 1.978 1.978 1.978 1.978 1.978 3.968 3.957 3.958 3.9577 3.9577 3.9577 3.9577 3.9577 3.9577 3.9577 3.95777 3.957777 3.95777777777777777777777777777777777777	28.331 - 11.627 0.297	87.196	(8 oxygens 5.167 5.110 5.1110	I'IOd	4-28 26.745 26.745 1.526 10.626 9.223 9.223 2.928 1.079 1.079 1.079 1.079 1.079 0.909 0.909 0.928
(METAPELITES)	68-19 11mentte 0.554 0.553 71.853 71.853 71.853 71.853 71.853 71.853 71.853 71.950 0.125 0.033 0.033 0.125 0.125 3.993 3.932 3.9333 3.93333 3.93333 3.93333 3.93333 3.93333 3.933333 3.933333 3.933333333	21.022 - 12.367 0.124 0.223	87.115	ie basis of 2 5.272 5.569 5.272 5.505 5.272 5.505 5.505 5.272 5.272 5.505 5.272 5.505 5.272 5.505 5.272 5.272 5.505 5.272 5.505 5.272 5.505 5.505 5.505 5.272 5.505 5.272 5.505 5.272 5.505 5.505 5.505 5.272 5.505 5.50	ITES)	8-5 61.716 24.230 5.201 5.301 8.569 9.093 9.093 9.093 1.257 1.257 1.257 1.257 1.257 1.257 1.257 0.095 0.005 0.712 0.712 0.712 0.205
AND RUTILE	77-12 11mentte 0.230 0.230 0.235 9.263 9.45,363 9.463 9.463 9.042 0.014 1.975 1.975 3.975 3.975 3.975 2.195 2.1	0.1687 0.124 0.1687 0.163	85.933	ations on th 5.269 5.062 5.077 5.077 5.077 5.077 0.028 0.028 0.028 19.894	SES (METAPEL	77-18 21.5.349 21.5.349 2.5.349 2.309 1.662 99.967 99.967 1.123 1.123 1.123 1.123 1.123 1.123 1.123 1.123 0.9109 0.9109 0.910 0.993
ILMENITE	ox/ID SiO2 SiO2 FiO2 FiO2 MHO MHO2 MHO2 MHO2 SiA SiA SiA SiA SiO2 SiO2 SiO2 SiO2 SiO2	T102 Mg0 Mn0 K20	SILLS	formula S1 A1 A1 Fe2 M6 Mn2 K K Sums	PLAGIOCLA	ox/ID 5102 5102 71203 760 760 8412 541 76 81 81 81 81 81 81 81 81 81 81 81 81 81
LINU XINIJ	81.1-9 25.807 25.1788 21.1880 21.2005	. •				
PO	81.1-6 31.61 54.144 12.185 54.144 12.185 0.62 91.042 91.042 91.042 0.659 91.042 14.672	ı				
	4-22 28.55142 555.142 555.142 555.142 0.550 0.550 0.550 0.550 0.193 7.194 3.943 8.943 8.943 8.943 8.943 0.0113 0.0113 0.013 0.023	8,265				
	4-18 28 745 54.7478 0.5778 0.576 0.5778 0.576 0.576 0.576 0.576 0.578 0.578 0.578 0.578 0.578 0.178 0.175 0.	8.970			2 oxygens	
	4-17 28.007 54.552 54.552 0.1772 0.1775 0.2756 0.210 0.2756 0.212 96.519 96.519 96.519 96.519 9.519 0.157 0.157 0.157 0.157 0.2125 0.157 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2157 0.2125 0.2122 0.2125 0.2122 0.2125 0.2122 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2125 0.2155 0.2125 0.21555 0.21555 0.21555 0.215555 0.21555555555555555555555555555555555555	10.761		81-22 r1m 37.881 21.585 21.585 35.370 35.370 35.370 1.585 1.595 1.	: basis of 12 3.015 - 2.026 - 2.355	0.389 0.126 0.0126 7.372 84.667 13.278 14.182
	714 32.203 52.203 52.203 0.557 0.557 0.557 0.265 98.023 98.023 98.023 98.023 98.023 98.023 98.023 98.023 0.091 1.215 0.091 0.127 0.127	6.959		81-9 11m 37.922 21.415 32.849 32.988 2.198 2.115 1.639 1.00.238	ttions on the 3.015 2.007 2.184	0,474 0,163 0,1163 7,982 7,982 1,280 1,2717 16,003 11,823
	68-20 27,140 27,140 55,673 1610 11,610 11,610 0,286 0,286 0,286 0,286 0,286 0,286 0,286 0,286 0,286 0,015 0,0000000000	10.212	Garnet	ox\ID 8102 8102 81203 7102 7102 7102 7102 7100 7100 7100 7100	formula ca Si Aliv Alvi Fe3 Fe2	T1 MG MG2 CCa Sums almSpa almSpa RT3And MG no.
	81.3-28 28.203 53.2203 53.2203 53.2203 53.2203 1.404 1.404 1.404 1.404 98.239 98.239 98.239 98.239 98.239 0.205 0.205 0.205 0.205 0.203 0.203 0.203 0.203 0.203 0.203 0.203 0.203 0.200000000	15.244	81-32	27.895 54.030 13.690 0.704 1.273 0.1143 0.1143 0.380 0.380 0.380	oxygena 3.862 8.818 8.818 1.585 0.073 0.263	0.017 14.656 14.215
ES)	83-23 27.359 54.504 0.519 0.519 0.518 0.519 0.519 0.519 0.519 0.519 0.512 0.071 0.071 0.071 0.072 0.0120000000000	8.164	81-28	28,176 54,169 12,819 0.618 0.1483 0.116 0.1483 0.116 0.334 0.334	basis of 23 3.918 8.930 8.930 1.491 0.065 0.100	0.014 - 0.034 6.292 6.292
\$ (METAPELIT)	83-2 28.254 53.317 53.317 53.317 0.578 0.578 0.578 0.254 0.254 0.254 0.254 0.255 0.980 95.554 1.337 0.980 0.980 0.980 0.980 0.914 0.102 0.014 0.012 0.014	10.515	81-6	28.163 54.565 12.527 1.2.527 1.262 1.062 1.062 1.062 0.457 0.457 0.457	ions on the 3.906 8.926 1.453 0.036 0.220	
STAUROLITE	artID art1202 5760 460 460 460 460 460 461 11 11 11 11 11 11 11 11 11 11 11 11 1	fg no. Staurolites	UI / NO	1102 11203 1102 1102 1100 1100 1100 1100	ormula cat 1. e.2 13 13	5 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1 1

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GARNETS	(METAPELITE	2)							STRIEDEN	UNIT k	VIUS COVITES	(METAPELI'	ES)						STRI	EDEN UNIT	0.7.0
0X/ID	136-1	136-7	136-4	302 111 imani ta-1	31 bearing	408.3	408.16	516.2	554.5 and-hear	554.7 ¹	OX/ID	516.1-8 coarse-g. (554-22 soarse-g.	136-12 recryst.	302-11 after sill	And-7 after and	336-1 after and	514 after and	4/8.21 fine-g.	4/8.5 fine-g.	coars
5102 A1203	rim 37.589 20.923	core 38.431 20.889	rim 37.309 20.883	37.414 20.989	37.781 21.837	38.196 21.013	incl.in pla 37.988 20.874	19 37.696 20.892	rim 37.274 21.047	20.929	sio Al 03 Feo	46.001 35.788 1.129	45.468 34.983 1.221 0.366	45.414 36.154 0.731 0.536	45.038 35.282 1.169 0.439	45.973 41.465 0.207 -	44.278 36.534 0.811 0.168	45.704 37.919 1.682 0.221	45.557 39.016 -	46.715 37.376 0.693 0.145	46.17 38.03 0.26 0.26
Fe ²⁰³ T102 Mg02 Mn0	35.124 - 1.540 3.790	31.376 0.122 1.330 1.961	- 35.083 1.825 3.692	34.540 - 3.996	27.532 - 1.532 10.553	30.347 2.529 5.071	32.017 32.017 1.827 7.823	- 29.911 1.976 8.055	- 38.536 2.082 1.279	35.760 1.139 3.121	Mg0 ² Cr Na ₂ 0	0.275	0.467	0.242	0.003		- - 0.729 0.329 9.561	0.348 - 0.823 9.544	- 0.850 4.280 3.119	0.302 - 0.148 0.557 9.690	- 10.00
CaO TOTAL	1.390 100.356	6.462 100.571	1.064 99.857	1.017 99.521	1.310 100.618	3.275 100.431	0.373	1.066 99.595	0.273 100.492	1.913	K2 ^U TOTAL	93.891	92.455	92.865	92.148	96.672	92.409	96.241	92.823	95.625	95,05
Formula c	ations based o	n 12 oxygens									Formula cati	one haced on	22 DYNAPHS								
Si Aliv Fe2+ Fe2+	3.036 - 0.007 2.366	3.058 - 1.960 0.040 2.044	3.026 - 0.003 2.377	3.040 - 2.010 2.347	3.025 - 2.061 - 1.844	3.045 - 0.025 1.998	3.051 - 0.024 2.126	3.051 - 0.007 2.018	3.011 2.005 2.604	3.024 2.003 2.428	Formuna caun Si Alvi Al ² + Fe ²⁺	6.164 6.164 1.836 3.817 0.127	6.189 6.189 1.811 3.803 0.139	6.127 1.873 3.878 0.082	6,159 1,844 3,841 0,132	6.202 1.798 4.404 0.022	6.039 1.961 3.914 0.092	5.996 2.004 3.561 0.185 0.022	5.996 2.004 4.050 -	6.130 1.870 3.912 0.076 0.014	3.97
i Be s	0.185 0.259 0.120	0.007 0.158 0.132 0.551	0.221 0.254 0.093	0.190 0.275 0.089	0.1183 0.716 0.112	0.300 0.343 0.280	0.219 0.532 0.032	0.238 0.552 0.092	0.251 0.088 0.024	- 0.138 0.215 0.166	r Spo	0.055	0.095	0.049	0.013	- - 0.115	0.106	0.068	0.120	0.059 - 0.021 0.142	0.01
TOTAL	7.967	7,954	7.970	7.949	7.940	7.967	7.961	7.952	7.982	7.974	Na K	0.100	0.136	0.198	1.606	0.423	1.664	1.597	0.524	1.622	1.54
Mol% pyr Mol% gros Mol% spes Mol% alm Mol% alm	6.29 4.08 8.81 0.35 0.35	5.36 18.67 4.47 69.50 2.00	7.49 3.15 8.62 80.59 0.15	6.549 3.068 9.479 80.903	6.41 3.92 25.08 64.59	10.14 9.47 11.60 67.54 1.25	7.44 1.09 18.07 72.21 1.20	8.18 3.16 18.96 69.34 0.35	8.46 0.80 2.97 87.77	4.683 5.633 7.267 82.388	TOTAL Mg/Mg+Fe Mol% mus Mol% parag Mol% marg	13.813 30.231 94.26 5.74 -	13.847 40.541 91.65 7.69 0.66	13.822 37.090 88.74 11.26 -	13.873 87.17 12.83	13.963 22.19 71.80 6.01	13.881 - 5.73 4.69	13.848 26.934 93.25 6.75	13.785 	13.846 43.733 90.90 7.93 1.16	13.83 - 86.33 86.33 0.71
OTHER 1 ox/ID	IINERALS SILLIMANIT 169-6 fibrolite	E 169-21 prismatic	ANDALUSIT And	E CTD And	STAUROLI' 136-20 core	TE (sill-in 136-21 rim) (sill-out) 554-18	CHL 554-16	11.M 31.16	AP 336-5	BIOTITES 0x/1D	(METAPELI7 554-15	(ES) 554-21	136-2	136	336-4	514-7	516.2-12	408.5-5		
510, A1,0,	38.014 61.593	37.799 61.810	36.688 63.434	24.127 41.753	27.899 52.767	27.989 53.475	27.500 53.354	41.245 15.964	1.266 0.472	1.509		and bil	zone bill i	ntergrown	ill - in z in cont. wit armet	one incre h	iasing s tr uct	cural deptn -	~		
Fe0 ' Fe203 C123	0.241	- 0.439 -	0.224	23,186 - -	12.973 - -	13.018 - -	13.732 - -	24.809 - -	42.397 - -	- - 0.075	sio ₂ Al ₆ 0,	33.910 19.458	35.130 19.736	37.167 20.900	34.461 21.041	36.169 19.437	37.331 20.412	36.948 19.958	36.955 16.736 15.030		
T102 Mn02				0.317	0.525 0.239	0.494 0.257	0.389 0.108		52.888 4.331		Fe6 3 T102	23.223 1.614 6.487	22.203 1.703 6.950	19.173 1.620 8.917	20.156 1.575 7.234	19./33 2.390 8.017	1.530	2.981 2.981 8.480	1.882		
Cal Cal						-	0,8U3	061.1		54.533	0 Out	5 	1 1			0.119 -	0.130				
					0.405	- 0.480		0.094			Na 20 K 20	- 8.542	- 8.721	- 9.092	8.811	- 9.085	7.680	9,084	160.6		
P ₂ 05 Total	- 99.848	- 100.048	- 100.456	- 92,110	- 95,712	- 96,535	- 95.885	- 89.302	- 101.355	42.668 98.784	TOTAL	93.233	94.443	96.869	93.277	94.949	93.085	97.183	660 .66		
Formula c	ations based o 5 ox	n 5 ex	5 ox	12 ox	23 ox	23 ox	23 ox	28 ox	3 ox	25 ox	Formula cat	tions based c	n 22 oxygen:	5		501	5 500	5 470	5.615		
Si Al ₂₊	1.456	1.395	1.444	1.986	3.946 8.799	3.926 8.715	3.887 8.891	8.119	0.031	0.249	si iv Alvi	5.359 2.641 0.985	5.434 2.566 1.033	5.493 2.507 1.135	5.348 2.652 1.198	2.499 0.986	2.401	2.530	2.385		
Fe ³⁺	0.003	0.007	0.003	1.596 -	1.535	1.527	1.623	4.084	0.872		Fe ²⁺ Ti	3.070	2.872 0.198	2.370 0.180	2.616 0.184 1.673	2.510 0.273 1.817	0.173	0.332	0.215 3.034		
5 E E	• •			0.022	0.056	0.052	0.041		0.978		₽£.	97 5 .1		100.1		0.015	0.017				
G Rg				0.334	0.190	0.172	0.169 -	2.109	• •	9.626	╏뚭 ^ᄍ	1.722	1.721	- 1.714	- 1.744	-	1.470	1.716	1.762		
톱 주 년	• •	• •				1 1 0	• •	0.024			TOTAL	15.497	15.428	15.363	15.415 20.000	15.364	15,185 54 655	15.314 43 369	15,535 61,351		
54					-	-				5,951	Mg/Mg+Fe	33.234	35,809	45.318	900.55	nne. 1 +	600° • 60				
TOTAL	2.848	2.850	2.924	7.989	14.598	14.600	14.625	18.040	1.984	15.846											

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