

**The Kinematic and Structural Development of the Austroalpine - Pennine
Boundary in the S.E. Tauern, Eastern Alps.**
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The Structural and Kinematic development of the Austroalpine - Pennine
Boundary, S.E. Tauern, Eastern Alps.

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ABSTRACT

The Eastern Alps are a belt of major deformation formed by the convergence and collision of Europe with the Adriatic microcontinent. A major tectonic boundary can be defined between the dominantly continental Austroalpine domain, which represents the northerly fringe of Adria; and the Pennine oceanic domain, which once lay between the Austroalpine domain and Europe. This boundary is one of the key areas for unravelling the convergent tectonic history of the Eastern Alps. The main emphasis of this thesis is on the deformational behaviour of the Austroalpine domain and its implications for the regional tectonic processes that were instrumental in forming the structure of the Eastern Alps. The data for this work are derived from structural and metamorphic studies in the S.E. Tauern in profiles straddling the boundary, and a review of the published regional geology.

The first manifestation of convergence is the development of a broad zone of thickened crust, including both Pennine and Austroalpine units, which underwent Cretaceous burial metamorphism. Postdating this thermal event in the Austroalpine domain, there was a further phase of regional deformation, which caused substantial reduction in the structural thickness. To the south of the Tauern Window, a study of the associated kinematic development gives a minimum estimate of 10km for this thinning, which radiometric dating suggests took place while convergence was still active. A comparable 10km post-metamorphic thinning is suggested throughout the Austroalpine domain in the Eastern Alps by the shortfall in the overburden compared to the depth of burial indicated by the Cretaceous metamorphic conditions.

Major extension in a dominantly convergent tectonic setting cannot be accounted for by the paradigm of plate tectonics and implies that body forces acting on the thickened crust of the destructive margin were a major driving force for deformation.



LONG ABSTRACT

The Eastern Alps form part of the collisional^{belt} developed during the convergence between Europe and the Adriatic microcontinent. The northerly fringe of Adria is represented in the Eastern Alps by the Austroalpine domain separated from the European foreland by the Pennine domain, which represents the ocean that once lay between the two land masses. The Pennine ocean basins were formed when Adria was rifted away from Europe in Permian to Early Cretaceous. Convergence probably began around 120Ma and continued until 40Ma. The Austroalpine and Pennine domains are divided by a major tectonic contact, which can be traced throughout the Alpine chain. This is one of the key areas for unravelling the convergent tectonic history of the Eastern Alps.

The Austroalpine - Pennine boundary is well-preserved along the southern margin of the Tauern Window and the suture zone in this region is marked by a narrow belt of intense deformation known as the Mafrei Zone, which has lithological similarities to both the overlying Austroalpine and the underlying Pennine domains. However, the two domains have incompatible early histories. The two main lines of investigation followed in this thesis are:

- (i) the nature of the disruption in the Mafrei Zone and;
- (ii) the deformational behaviour of the overlying Austroalpine domain.

The discontinuous nature of the Mafrei Zone can be largely explained as the result of tectonic processes acting within the zone. There is, however, a significant phase of syn-sedimentary instability, which formed large sedimentary blocks (olistoliths) that also contribute to the disruption. The general age of both the olistoliths and the surrounding matrix suggest that this phase of syn-sedimentary instability is most easily explained as the result of Jurassic block faulting, which developed before convergence began as the Pennine Ocean basins were being formed by the rifting of Adria away from Europe. There is no evidence that the Mafrei Zone formed in a subduction trench during convergence. The intimate association within the Mafrei Zone, of continental fragments and serpentinite lenses with pelagic sediments probably reflects an originally heterogeneous make up of the Pennine oceanic crust, consisting of thinned slivers of the original continental crust and serpentinite. Even before convergence, the Mafrei Zone had a complex association of different lithologies, which was due to the action of extensional tectonics during rifting. These relationships were further complicated by deformation during convergence.

The first manifestation of convergence in both the Pennine and Austroalpine domain was a phase of crustal thickening, which produced a broad zone of thickened crust involving a shortening of about 200km in the Austroalpine domain in a WNW direction. Shortening in the Austroalpine domain is expressed as folding and thrusting within the cover nappes, and a phase of ductile deformation under increasing metamorphic grade at lower structural levels within the basement. Crustal thickening caused Cretaceous metamorphism, characterized by high P/T conditions in the Pennine domain and Barrovian facies in the Austroalpine domain. The contrasting early metamorphic history demonstrates the initially distinct tectonic settings for these two domains. It was only after the Austroalpine basement (Altkristallin) had cooled from this metamorphism at around 80Ma that it was juxtaposed with the low-grade units of the Pennine domain and the Mafrei Zone. Radiometric dating suggests that cooling of the Altkristallin was locally very rapid, which could be related to the underthrusting of relatively cool low-grade rocks beneath the higher-grade Altkristallin.

Along the southern margin of the Tauern Window, the Altkristallin was first emplaced over the lower grade units by an episode of thrusting, which can be subdivided into two distinct phases. The second phase may have been associated with a top to NE sense of shear. This emplacement of the Altkristallin onto the Matrei Zone took place after thickening had already begun in the Pennine and Austroalpine domains and represents a major out-of-sequence thrust. The Austroalpine - Pennine boundary in this area does not, therefore, represent the site of the first-formed subduction zone. The sudden jump in metamorphic grade across this boundary in the study area is probably due to the transitional material having been subducted away.

Immediately postdating imbrication, the whole suture zone underwent penetrative ductile deformation, Ds, which produced a lower greenschist, kilometre-scale zone of distributed shear across the Austroalpine - Pennine boundary. Microstructural evidence suggests that deformation took place on a falling temperature path. At structural levels above the zone of Ds distributed shear the Austroalpine domain is cut by a number of Ds shear zones, which overprint the Cretaceous metamorphic fabric. Kinematic indicators show that Ds is associated with a consistent top to the NW sense of shear and radiometric dating indicates that this deformation was continuous from at least 80Ma to 40Ma, i.e. while convergence was still active. Along the southern margin of the Tauern Window, a study of the kinematics of Ds deformation suggests it caused a minimum of 10km thinning of the Austroalpine sheets. Later deformations within the area postdate collision and are all low-strain.

The evidence for substantial tectonic thinning within the study area can be related to a regional phase of crustal extension. Throughout the Eastern Alps there is insufficient overburden to account for the Cretaceous metamorphic pressures in either the Pennine or the Austroalpine domains. The shortfall in the overburden indicates a minimum regional thinning of 10km in the Austroalpine domain. This is in good agreement with the estimated thinning during Ds in the study area. Ds can be correlated with deformation in many parts of the Eastern Alps and locally these deformation phases can be demonstrated to be associated with extensional tectonics. Estimates of crustal thickness before and after Ds suggest that the Austroalpine domain was thinned by 10km from an average thickness of 25km to the present-day value of 15km, which represents an extension of 66%. To develop the present lateral dimensions of the Austroalpine domain, this implies an extension of the original thickened Austroalpine domain during Ds of 80km in a NW direction.

Most workers in the Eastern Alps have assumed that during its emplacement the Austroalpine domain behaved as a largely rigid plate and caused strong deformation in the underlying Pennine units. However, metamorphic and structural evidence throughout the Eastern Alps clearly demonstrates that both major shortening and extension have taken place during convergence in the Austroalpine domain. The leitmotiv is one of deformation and not stability, and there is no evidence for the proposed great rheological contrast between the Austroalpine and Pennine domains.

In general, destructive margins develop regions of thickened crust, which have a wedge-shaped cross-section. These regions may become gravitationally unstable if their surface slope exceeds a critical value, which is dependent on the bulk rheology of the rock. Gravity spreading may become an important deformation process if the

accretionary wedge is continually thickened at depth by underplating causing a corresponding increase in the topographic slope above.

The evidence for regional syn-convergence crustal thinning in the Eastern Alps strongly suggests that gravity-driven extension was an important tectonic process in the area. The Eastern Alpine orogenic wedge began to develop at the onset of convergence around 120-130Ma, and probably became overthickened and gravitationally unstable at around 80Ma, although localized extension may have occurred at an earlier date. Extension in the upper Austroalpine domain was maintained by concomitant subduction and underplating of Pennine material beneath. Subduction to depth caused the high P/T metamorphism seen within the Pennine units and probably continued until collision at around 40Ma. The evidence for extension in the Austroalpine domain after collision can be interpreted as continued gravity-driven deformation acting to reduce the taper of the orogenic wedge. This implies that during convergence the taper was at least partially maintained by frictional resistance to sliding along the base, suggesting that the wedge as a whole had a very low long-term yield strength.

The primary driving force for mountain building is the convergent plate motion of the colliding landmasses. However, the paradigm of plate tectonics cannot explain the phenomenon of regional extension in a dominantly convergent tectonic setting. Gravitational forces acting on a region of overthickened crust have played an important role in the tectonic development of the Eastern Alps and the kinematics of deformation in the Austroalpine domain represent the interplay between plate motion and gravity-driven extensional flow. Although largely independent of one another, kinematic indicators suggest that deformation in the Austroalpine domain associated with both crustal thickening and thinning produced a consistent top to NW sense of shear. The compressional forces causing the initial crustal thickening are likely to be associated with a movement direction directly related to plate motion. However, the relative movement direction during gravity spreading will be controlled by the line of maximum slope on the orogenic wedge. Although not directly controlled by plate movement, the associated spreading direction is unlikely to occur at an angle greater than 90° to the convergence direction. The consistency of the kinematic indicators in the Austroalpine domain, therefore, suggests that the movement vector of the Adriatic microcontinent with respect to Europe was to NW throughout convergence. The most recent work on the motion of Africa with respect to Europe shows that the convergent path of these two major continental masses was to the N or NNE throughout the Cretaceous. This is incompatible with the suggested motion of Adria with respect to Europe and suggests that Adria moved as an independent microplate from the onset of subduction.

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CHAPTER 1 INTRODUCTION

1.1 REGIONAL SETTING OF THE EASTERN ALPS

Throughout the Alpine chain the Austroalpine - Pennine boundary represents a major tectonic contact which separates a dominantly continental Austroalpine domain from the dominantly oceanic Pennine domain. These two domains were created by the rifting away of Africa from Europe in Permian to Cretaceous times, which was probably related to the opening of the Northern Atlantic to the west (Dewey et al 1988, Bernoulli & Jenkyns 1974). The Tethyan Ocean formed in the rifted and thinned area between the two major continental masses, and is represented in the Alps by three distinct palaeogeographic areas (i) the northern continental margin (European foreland) (ii) oceanic basins (Pennine domain) (iii) the southern continental margin (Austroalpine and Southern Alpine domains).

The Austroalpine domain represents the most northerly fringe of a major landmass, Adria, which was rifted away from Europe along with the main African continent. The extent to which Adria constitutes an independent microplate or an African promontory is still disputed (Dewey et al 1973, Lowrie 1986).

The Austroalpine domain consists of a cover sequence of dominantly Mesozoic carbonates overlying thinned metamorphic continental basement. Variations in the sedimentary facies and their depositional thicknesses imply the existence of palaeo-relief during deposition. This is also reflected in differences in the pre-rifting, i.e. pre-Permian erosion surface. In some places (e.g. Oetzal Alps) this cuts deep enough into the underlying basement to reveal relatively high-grade pre-Alpine metamorphics (Hoinkes 1973), while in others Permian sediments

unconformably overlies only weakly metamorphosed Palaeozoic phyllites (Oxburgh 1968).

From Permian until Mid Jurassic times extension was concentrated in the Austroalpine and Southern Alpine domains and throughout this time sedimentation kept pace with subsidence, maintaining a relatively shallow environment. In the Early Jurassic, however, a sudden increase in the rate of extension caused fracturing and rapid subsidence of the carbonate shelf. This phase of tectonic activity is reflected in the formation of fault bounded carbonate platforms, separated by basins that received clastic sediments derived from the nearby basin highs, including turbidites and breccias (Bernoulli & Jenkyns 1974). Continued extension and subsidence resulted in widespread pelagic sedimentation (Winterer & Bosellini 1981). These pelagic deposits include aptychus limestone, radiolarian cherts and red nodular limestone indicating water depths of several kilometres (Bernoulli & Jenkyns 1974), suggesting that the Austroalpine crust had been strongly thinned by this time.

Immediately to the north of the Austroalpine units, the Pennine domain represents the Tethyan oceanic basins and their sedimentary fill. The common and sometimes extensive occurrence of serpentinite associated with pelagic sediments, suggests the basins were mainly floored by oceanic crust, although some large and many small fragments of rifted continental crust are also found in this domain, e.g. the Briançonnais microcontinent and the possibly correlative Tauern Gneisses in the Eastern Alps (Frisch 1979). Oceanic crust probably did not begin to form until Mid Jurassic and the overlying Pennine sediments are, therefore, mainly late Jurassic and younger, although palaeontological evidence to demonstrate this is very sparse (Tollmann 1977a). The Penninic lithologies are very variable but are dominated by calcareous graphitic schists and greenschists both of which are commonly associated

with serpentinite. Chemical analyses of the greenschists suggests they are derived from tholeiitic ocean basalts (Bickle & Nisbet 1972).

The Tethyan Ocean reached its maximum N-S extent sometime in the Early-Mid Cretaceous when convergence began (Frisch 1979, Tollmann 1986, Trümpy 1980). No discrete spreading centre has been identified and the lack of a sheeted dyke complex in the ophiolitic fragments suggests that extension of the area may have been accommodated by displacement along a low-angle normal fault (Lemoine et al 1987). The full width of the oceanic basins is not well constrained and certainly varied along their length. At the time of maximum extension, the Pennine and Austroalpine domains together probably represented a depositional area with a N-S dimension of several hundred kilometres (Trümpy 1980) and may have been as much as the 1000km suggested by Tollmann (1977a). A proposed palaeogeographic reconstruction for the Late Jurassic is given in fig.1.1.

Before the onset of convergence, the original transition between the Pennine and Austroalpine domain was a gradational change in sedimentary facies, which reflected a corresponding change in sedimentary environment. The striking difference in the characteristic sedimentary facies of these two domains means that a particular unit in the present-day Alps can usually be assigned to one or the other palaeogeographic domain simply on the basis of lithological associations. However, the gradational nature of the boundary defined on the basis of rock type, inevitably leads to ambiguities; and the sometimes heated disputes about the affiliation of some transitional zones (e.g. Klammkalk Zone) are largely an artefact of this nomenclature, as recognized by Clar (1965).

Extension of the Tethyan region from Permian to Early-Mid Cretaceous produced a broad zone of oceanic and thinned continental crust. During the subsequent convergence between Adria and Europe, this

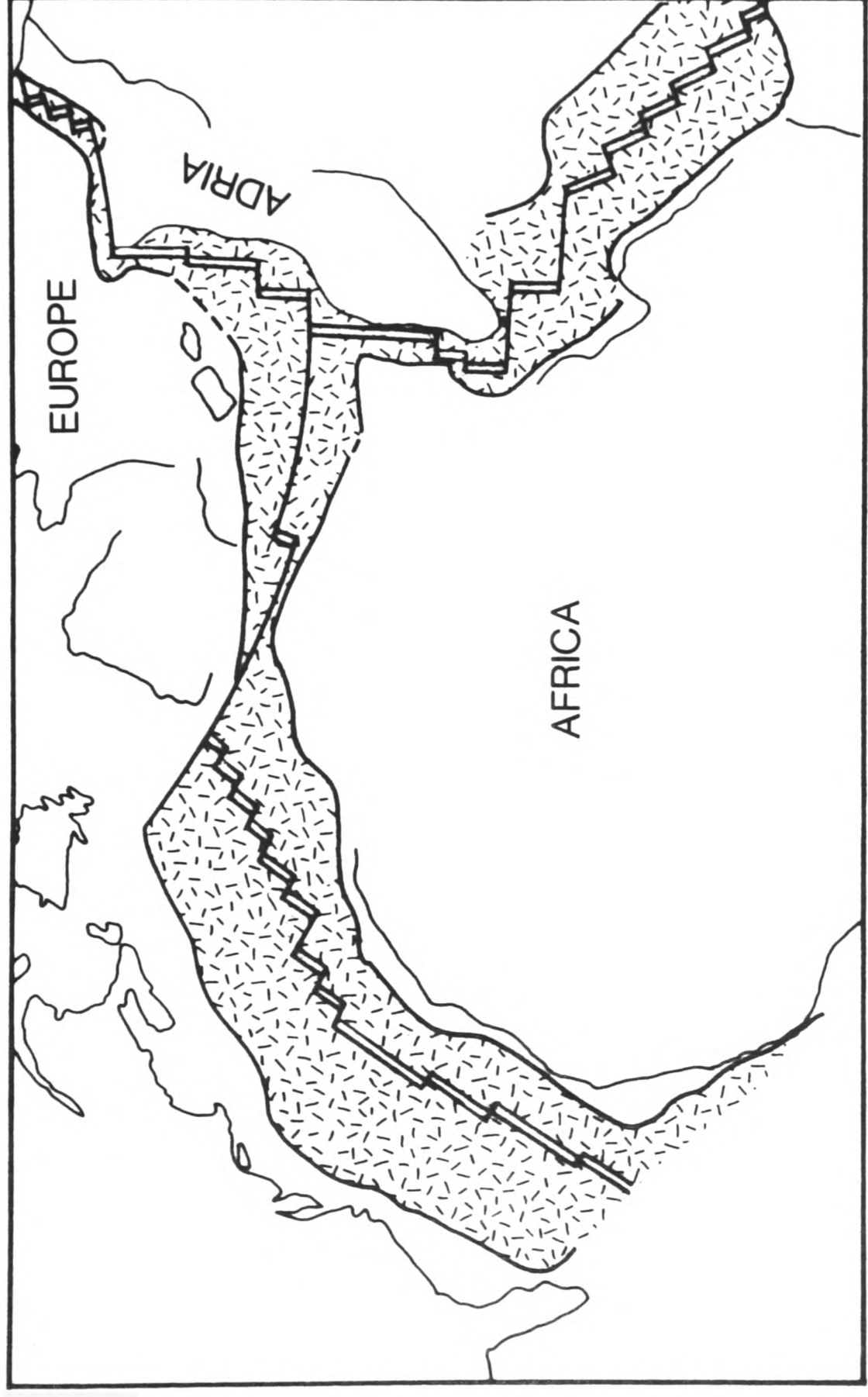


Fig.1.1 General palaeogeography for the Tethyan and North Atlantic realms in the Late Jurassic. Ornamented area = oceanic crust. (Modified after Dewey et al 1973, Knott & Turco in prep.)

basement and its sedimentary cover was drastically shortened, causing the emplacement of the Austroalpine nappes onto the Pennine domain.

The present-day Austroalpine - Pennine boundary in the Eastern Alps clearly represents a major tectonic break, which separates a dominantly continental domain from a dominantly oceanic domain (fig.1.2). In most plate tectonic reconstructions of the area this suture zone has been interpreted as marking the initial site of subduction, which formed in the Early-Mid Cretaceous (Frisch 1984, Dietrich & Franz 1976). The identification of the Austroalpine - Pennine boundary as a plate suture allows the area to be divided up into a upper and lower plate separated by a fossil subduction zone.

Along the suture there is commonly a narrow complexly deformed zone sandwiched between units of clear Austroalpine affinities and those of clear Pennine affinities, which shows lithological similarities with both. Along the southern margin of the Tauern Window this tectonic unit is known locally as the Matrei Zone (fig.1.3).

Many workers are in favour of this zone representing a semi-continuous nappe complex, which is the strongly-deformed palaeogeographic transition zone between the Austroalpine continental margin and the Pennine Ocean basins. This would account for the lithological affinities with both domains. In this interpretation the transition zone does not represent truly oceanic conditions and it is, therefore, described as Lower Austroalpine (Clar 1965, Tollmann 1977a).

Bickle & Hawkesworth (1978) consider the chaotic nature of this zone is dominant and describe it as a tectonic *mélange*; they assign it to the Pennine domain. Frisch et al (1987) have a similar interpretation and suggest that most of the disruption is related to olistostrome formation in an active subduction zone trench.

Evidence for the previous existence of a destructive plate margin comes from the regional high P/T metamorphism found mainly (but not

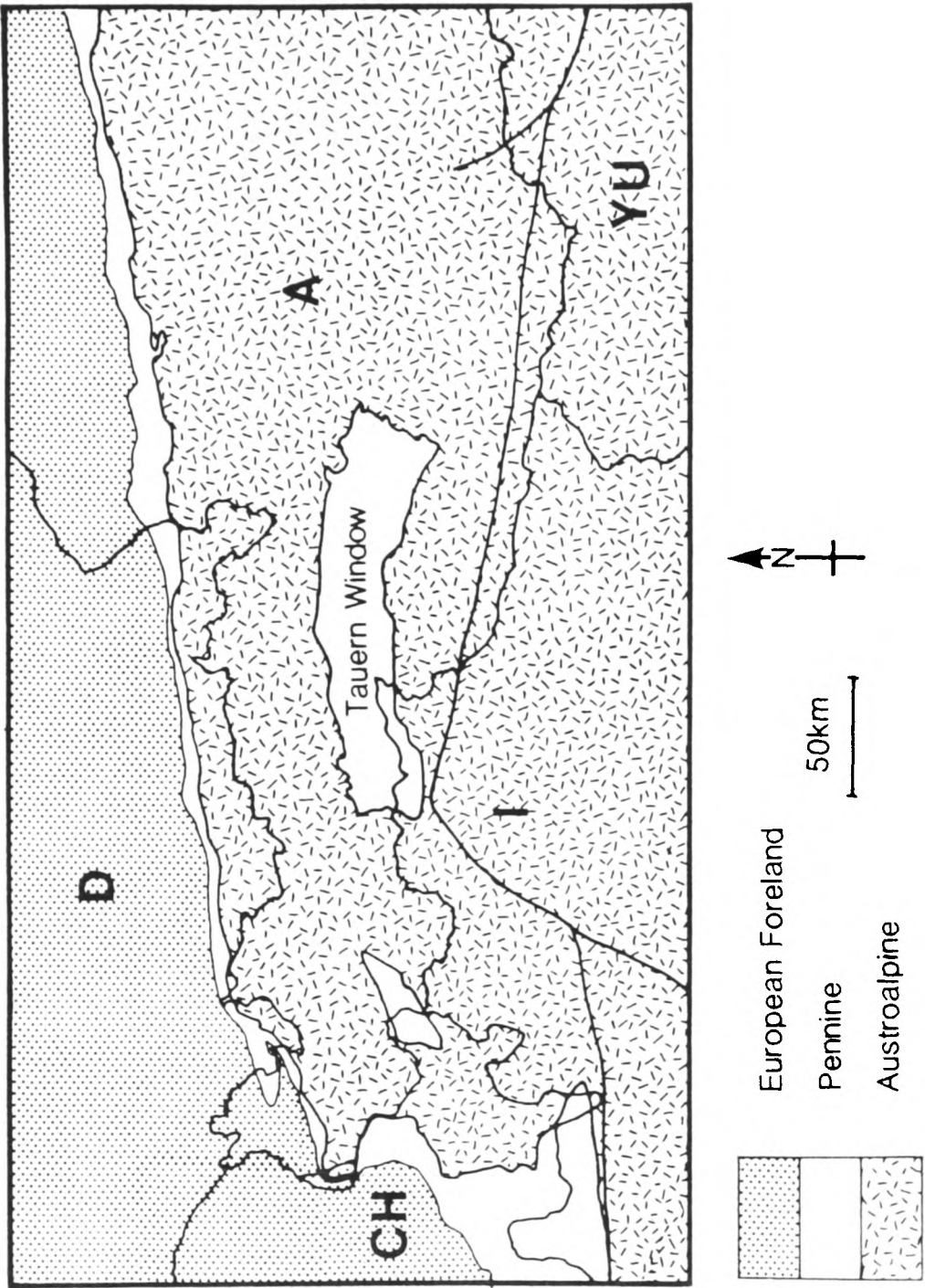
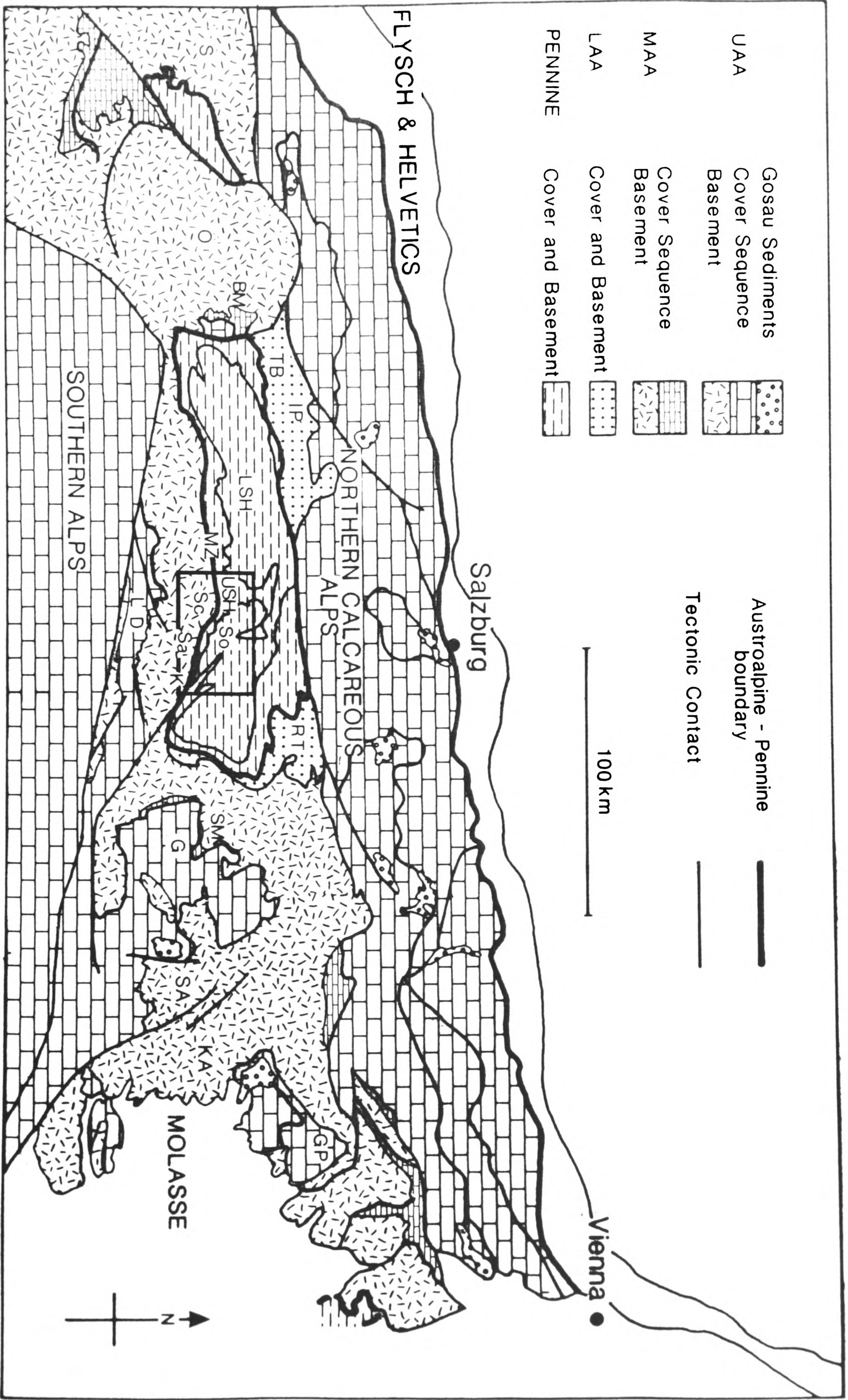


Fig.1.2 The three palaeogeographic domains of the Eastern Alps: an upper Austroalpine plate, the European foreland and the oceanic Pennine units between.

Fig.1.3 The principal tectonic divisions of the Eastern Alps (after Tollmann 1977a).

UAA = Upper Austroalpine, MAA = Middle Austroalpine, LAA = Lower Austroalpine. O = Oetztal Alps, S = Silvretta, BM = Brenner Mesozoic, LSH = Lower Schieferhülle, USH = Upper Schieferhülle, Sc = Schober Group, Sa = Sadnig Group, K = Kreuzeck Group, LD = Lienzer Dolomites, RT = Radstädter Tauern, MZ = Matrei Zone, TB = Tarentaler Berge, SM = Stangalm Mesozoics, G = Gurktaler Alps, KA = Koralpe, SA = Sau Alps, GP = Graz Palaeozoics, IP = Innsbruck Quartz Phyllite.

The boxed region indicates the region covered by fig.2.1.



exclusively) within the Pennine units of both the Western and Eastern Alps (Ernst 1973). These metamorphic conditions are produced by rapid burial of relatively cool rocks in a subduction zone, producing a crustal-scale thermal perturbation with abnormally low geothermal gradients (Oxburgh & Turcotte 1970). Radiometric dating in the Western Alps indicates that metamorphism was probably a continuous process lasting from Mid Cretaceous until Early Tertiary (Platt 1986).

The greatest age of the high P/T metamorphism in the Pennine units of the Western Alps is $\approx 100\text{Ma}$ (Hunziker 1974, Frey et al 1974, Deutsch 1983) and this gives a minimum age for the onset of subduction. Indirect evidence in the Eastern Alps, from dating sediments containing detrital high P metamorphic minerals and the eroded remains of oceanic crust (Oberhauser 1968, Winkler & Bernoulli 1986) also suggests that subduction had begun by Cenomanian times ($\approx 100\text{Ma}$). Relative and absolute dating of tectonic, metamorphic and sedimentary events shows a general younging towards the European foreland away from the Austroalpine boundary (Milnes 1978, Frisch 1979, Platt 1986). This implies the suture zone progressively prograded towards the European margin and suggests a subduction zone dipping beneath Adria.

There is, therefore, good evidence for interpreting the tectonic history of the Alps in terms of the relative motion between two large continental masses, Adria and Europe, which were separated by the ocean basins of the Pennine domain. Convergence began in the Early to Mid Cretaceous, during which the Pennine domain was subducted beneath Adria.

The high P/T metamorphism found within the Pennine domain can be explained as a result of subducting material beneath the thickened Austroalpine domain. In contrast to the dominantly high P/T metamorphism seen in the Pennine units, the Austroalpine nappes underwent a Cretaceous Barrovian type metamorphism with widespread $\approx 80\text{Ma}$

cooling ages (Brewer 1970, Hawkesworth 1976). The very different contemporaneous metamorphic conditions in the Pennine and Austroalpine domains emphasizes their initially distinct tectonic environments. The Cretaceous metamorphism in the Eastern Alps has been compared to the paired metamorphic belts of the circum-Pacific region (Ernst 1973, Hawkesworth et al 1975, Roeder 1977). In this analogy the metamorphism in the Austroalpine domain represents the high T/P side of the paired belts, which in the circum Pacific region is associated with an igneous arc. Frank (1983) suggested that the metamorphism may have developed before convergence began and be related to heating caused by rifting. In contrast to both these suggestions Tollmann (1977a) relates the metamorphism to a phase of crustal thickening.

The lack of any associated igneous activity and the evidence for an originally thin crust, implies a phase of crustal thickening is required to produce the Cretaceous metamorphism in the Austroalpine domain. The high P/T metamorphism in the Pennine domain can, therefore, be explained as the result of subduction beneath a thickened Austroalpine domain. However, a quantitative study shows that in the Eastern Alps there is no simple relationship between the depth of burial indicated by the metamorphic conditions and the thickness of the available overburden.

A regional P_{max} of >10kbar has been determined by Selverstone (1985) in the lower Pennine units of the Tauern Window. Assuming a $\rho = 2.9 \text{ g cm}^{-3}$, this indicates depths of burial in excess of 35km. Around the Tauern Window the overburden for these metamorphic rocks, consisting of the structurally higher Pennine material and the complete stack of Austroalpine nappes, has a maximum of about 20km (Cliff et al 1971, chapter 5). This is 15km thinner than the 35km of overburden needed to account for the metamorphic pressures within the high P Pennine units.

This deficiency in the overburden has been explained by Platt (1986) as the result of gravity spreading during convergence. However,

the Austroalpine domain has generally been regarded as a rigid sheet, which was largely unaffected by Alpine deformation (Bickle & Hawkesworth 1978). There is clear evidence for the involvement of Mesozoic cover rocks in strong Alpine deformation (Tollmann 1977a) but this deformation has been interpreted largely as the result of thinskin tectonics (Clar 1965). A few related zones of penetrative deformation have also been recognized in the basement (Tollmann 1977a). Oxburgh (1972) suggested that a rigid flake of upper crust had been detached during convergence and this is now represented ^{by} _^ the Austroalpine domain. In almost all tectonic reconstructions of the area, a fundamental rheological contrast is assumed to have existed between the Austroalpine and Pennine domains. The model proposed by Platt (1986) requires the Austroalpine units to have been weak enough to undergo substantial thinning by deforming under their own weight. One of the main aims of this thesis is to investigate these contrasting hypotheses for deformation within the continental Austroalpine domain.

The end of convergence represents continental collision between Adria and Europe. Around the S.E. margin of the Tauern Window, Oxburgh et al (1966) and Cliff et al (1971) report ≈ 65 Ma K-Ar radiometric ages at the base of Austroalpine sheet, which they relate to the emplacement of the Austroalpine nappes over the Pennine units marking the collision between the Austroalpine and European plates. The consequent metamorphism, the Tauern Metamorphism, has a peak T around 30 Ma and the thermal modelling of Oxburgh & Turcotte (1974) suggest that these effects can be explained as the result of static thermal reequilibration after a single major thrusting event at around 65 Ma. However, subduction of Penninic material began in Early-Mid Cretaceous and the emplacement of the Austroalpine nappes was a progressive process, not a single datable event. Continental collision is more satisfactorily defined by the onset of deformation in sediments which were deposited on

the stable European foreland. In the Central Alps this can be dated $\approx 40\text{Ma}$ (Milnes & Pfiffner 1977). Collision was probably diachronous along the Alpine chain.

Instead of a steady progressive emplacement, Tollmann (1977a, 1986) suggests the Austroalpine units were thrust N towards the European foreland in a series of orogenic paroxysms, separated by periods of tectonic quiescence. He suggests the periods of tectonic activity can be related to regionally correlatable deformations. The same general N to NE movement of the Austroalpine nappes, has been assumed by most workers in Eastern Alps (Cliff et al 1971, Tollmann 1962, Selverstone 1985). This is based on the general E-W trend of the Eastern Alps; fold vergence; and the assumption that fold hinges and stretching lineations form, and remain, perpendicular to the movement direction throughout deformation. However, fold axes may progressively rotate into the stretching direction with increasing strain (Bell 1978, Williams 1978), and fold orientation and vergence are very unreliable kinematic indicators. On the basis of sedimentary evidence, other workers have suggested that convergence may have been oblique to the trend of the orogen (Frisch 1979, Bechstädt 1978). Very little reliable kinematic data has yet been published from the Eastern Alps.

The relative motion between the African and European plates can be constrained by matching magnetic anomalies and tracing fracture patterns in the North Atlantic (Dewey et al 1988). These movement paths have often been compared to the kinematics of deformation seen throughout the Alpine arc (e.g. Dewey et al 1973, Baird & Dewey 1986). Regionally consistent kinematic indicators are commonly assumed to mirror plate movement, and recently Ratschbacher (1987) has proposed a three phase translation of the Austroalpine domain on the basis of stretching lineation data.

Underpinning these ideas is the basic assumption that large-scale

tectonics of mountain belts can be described in terms of displacing rigid sheets on the Earth's surface. However, the evidence that regions of thickened crust may become gravitationally unstable and deform under their own weight (England 1987, Platt 1986), shows this paradigm of plate tectonics needs to be reconsidered.

Mechanical analyses of wedge-shaped continua deforming due to the action of body forces, show that the direction of motion is determined by the line of maximum surface slope (Chapple 1978, Platt 1986). The geometry of an orogenic system will be controlled to some extent by the large-scale plate movements. Although the movement direction associated with gravity spreading may not be directly related to plate movement, it is unlikely to be at an angle greater than 90° to the convergence direction. Throughout the convergent history of Austroalpine domain, kinematic indicators documented in this thesis suggest a top to the NW sense of shear. The associated deformation can be divided into an early phase of shortening and thickening followed by regional extension. I interpret this as a phase of crustal shortening forming an area of thickened crust, which then deformed by gravity spreading. The consistent tectonic transport direction suggests that the large-scale convergence between Adria and Europe was to the NW.

1.2 AIMS OF THE RESEARCH

The Austroalpine - Pennine boundary is one of the key areas for unravelling the convergent tectonic history in the Eastern Alps. The main emphasis of this thesis is on the role that the Mafrei Zone played in the evolution of the Eastern Alps and the deformational behaviour of the overlying Austroalpine nappes.

In the Mafrei Zone, I set out to document the nature of the disruption and to distinguish between tectonic and synsedimentary

causes. This information has an important bearing on the palaeogeographic interpretation and tectonic setting that can be proposed for this zone.

Structural profiles were traced from the Matrei Zone into the overlying Austroalpine units with the aim of establishing a structural and metamorphic framework for the Alpine development of the Austroalpine - Pennine boundary in this area. On a regional scale, deformation and metamorphism within the study area are correlated with events in other parts of the Austroalpine domain. These data are used to test the suggested deformational behaviour of the continental Austroalpine domain during convergence, and to constrain the convergence vector between Adria and Europe.

1.3 CHOICE OF STUDY AREA

An area was chosen on the southern margin of the Tauern Window, where several continuous profiles straddling the Austroalpine - Pennine boundary are exposed. This particular location avoids areas where the boundary is overprinted by late-stage movements such as probably occurred along the base of the Northern Calcareous Alps (Clar 1965) and the Silvretta Nappe (Masch pers. com.). Similar continuous profiles are exposed to the S.W. of the Tauern Window, however, in this area the original microstructural relationships along the boundary have been overprinted by the post-collisional Tauern Metamorphism (Stöckhert 1984, Hoffmann et al 1983). Choosing a study area in the S.E. Tauern has the further advantage that several detailed geochronological studies have been carried out in this region (Waters 1976, Brewer 1970, Hawkesworth 1974, Oxburgh et al 1966, Lambert 1970). This earlier work could, therefore, be incorporated in this study to help date some of the tectonic events.

CHAPTER 2 THE MATREI ZONE - DISRUPTION AND PALAEOGEOGRAPHY

2.1 INTRODUCTION

The Matrei Zone is a narrow belt of intense deformation involving a large number of different rock types, which lies along the southern margin of the Tauern Window (fig.2.1, 1.3) and is tectonically sandwiched between the underlying Pennine units to the north and the Austroalpine nappes to the south. Rocks similar to those of the Matrei Zone are distributed in the same tectonic position all around the margin of the Tauern Window (fig.1.3). In the NW corner of the Tauern Window the sedimentary sequence of the Radstädter Tauern can be identified as an extended Austroalpine continental margin (Tollmann 1977a). Most authors link these different outcrops, and look upon the Matrei Zone as the tectonically thinned root zone of a semi-continuous Lower Austroalpine sheet (e.g. Tollmann *ibid*). However, Frisch et al (1987) have proposed that the Matrei Zone formed in an active subduction trench beneath the Lower Austroalpine continental margin sequence. They interpret the complex relationships within the zone largely in terms of olistoliths, derived from the overriding Austroalpine units that were resedimented in a Penninic matrix of Lower-Mid Cretaceous age. However, the Matrei Zone has been strongly deformed and tectonic processes will clearly have contributed to the disruption of the original sequence. Bickle (1973) describes it as a tectonic *mélange*, which formed prior to the emplacement of the Austroalpine units. The two main aims of this chapter are to examine the nature and the palaeogeographic significance of the disruption within the Matrei Zone.

The Matrei Zone has many phases of deformation concentrated within a narrow zone; each subsequent phase partially obliterating the

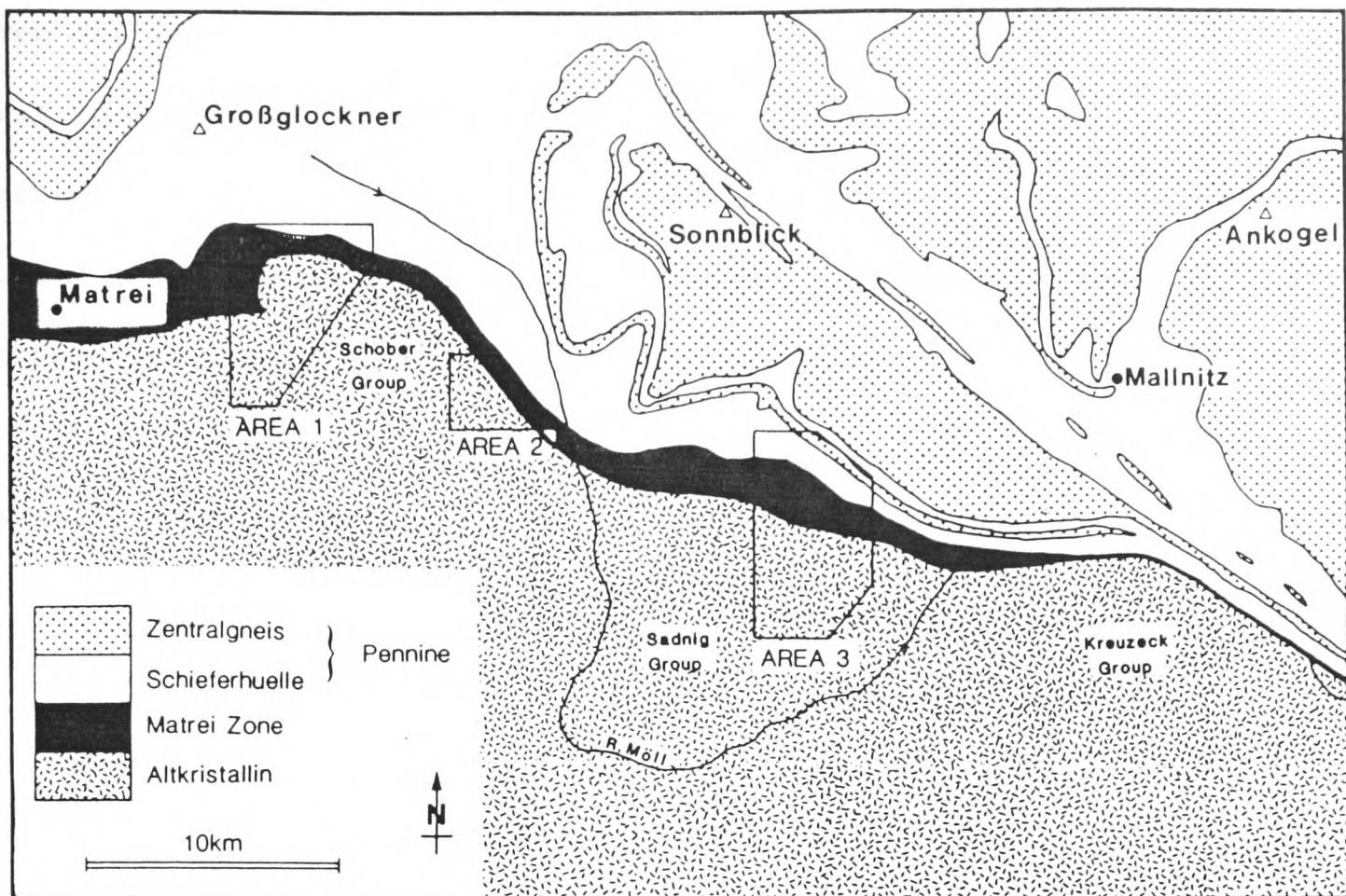


Fig.2.1. Geology of the S.E.Tauern (after Cliff et al 1971). The three boxed areas indicate the approximate extent of regions covered by enclosures 1,2 and 3. The divisions on this map are based largely on lithology and not on tectonic contacts as in fig.1.3.

previous. This geological palimpsest can be read only incompletely but even the earliest marks may be partially deciphered. It will be argued that the disruption of the Matrei Zone is dominantly the result of deformation within the zone. There is, however, a significant olistolithic component, as well as other evidence of syn-sedimentary instability, which is related to Jurassic extensional block faulting. There is no evidence for the Matrei Zone having been formed in a subduction zone trench.

The Matrei Zone can be divided into two main structural units which are separated by a major tectonic contact and derived from quite distinct palaeogeographic settings. The upper unit comprises mainly quartz phyllite associated with smaller amounts of Triassic carbonate and slivers of metamorphic basement. I suggest this represents the extended southern continental margin, and include it as part of the Lower Austroalpine tectonic sheet. The lower structural unit of the Matrei Zone is dominated by Pennine Bündnerschiefer deposits, which are associated with fragments of continental crust and serpentinite lenses. The heterogenous basement material within this zone probably reflects the original make up of the Pennine ocean floor.

2.2 ESTABLISHING A STRATIGRAPHY FOR THE MATREI ZONE

The Matrei Zone has yet to yield a single identifiable fossil and stratigraphy has to rely on lithological comparisons with other, better preserved areas. The Matrei Zone is most closely associated with the Upper Pennine and Lower Austroalpine sheets and also shows some affinity with the Lower Pennine and Middle Austroalpine tectonic sheets (see fig.1.3. for location). A general stratigraphy for each of these four units is considered, in order to construct the most likely stratigraphy for the Matrei Zone (fig.2.2).

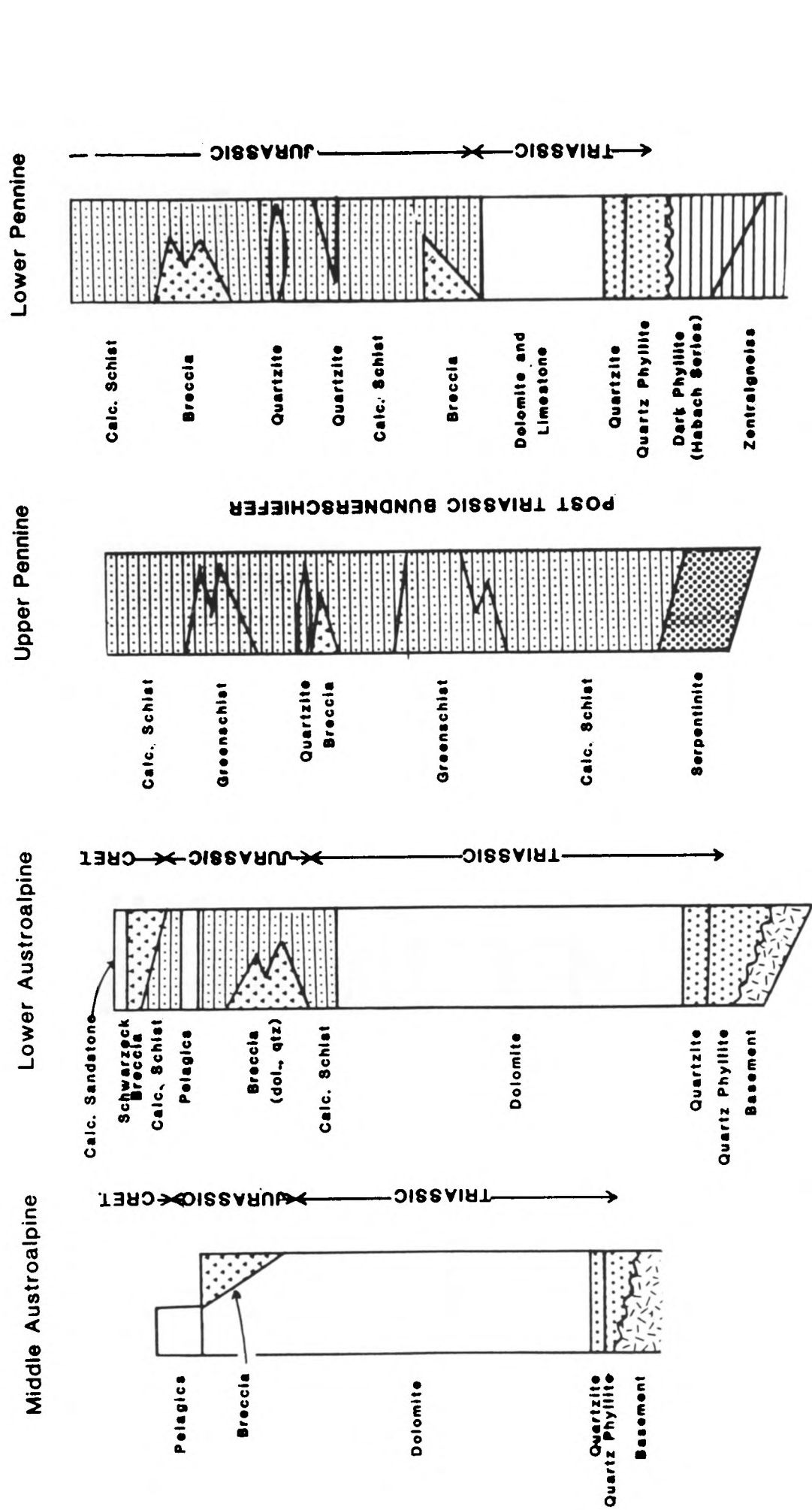


Fig.2.2.a. General stratigraphic columns through the four principal tectonic domains, which are most closely allied to the Matri Zone (after Gwinner 1978, Tollmann 1977, Frasl 1958).

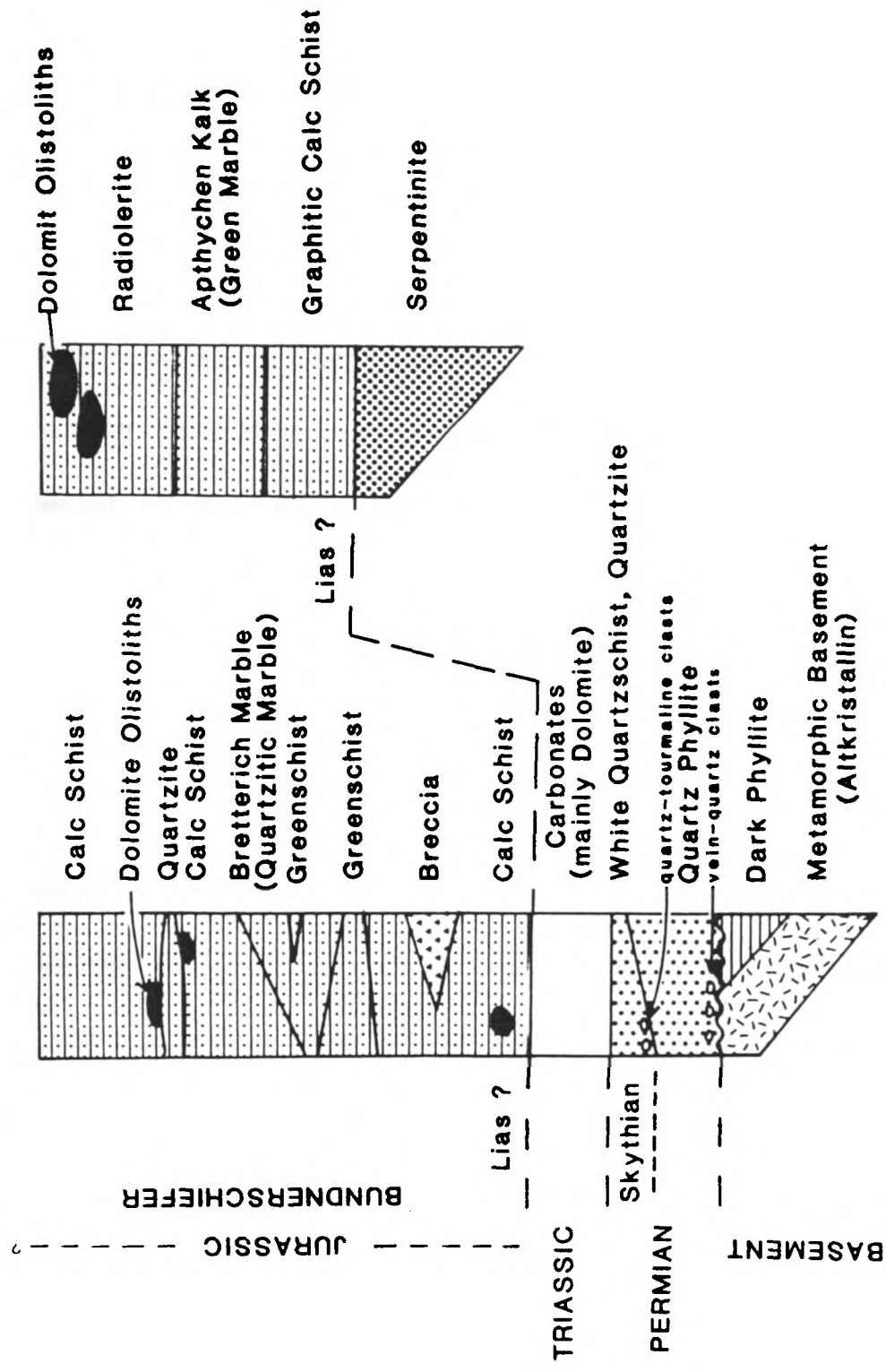


Fig.2.2.b. Lithologies of the Matrei Zone and suggested stratigraphy based on a comparison with fig.2.2.a.

The Permo-Triassic development of all these units is broadly similar, they are most readily distinguished by differences in the younger sediments and the basement on which the relevant sequence was deposited. The geological history of the Pennine and the Lower Austroalpine domains is most easily determined from the north of the Tauern Window, where the late-stage metamorphism (Tauern Metamorphism) was of lower grade and the tectonic disruption less severe than in the south. In this northern area fossils are still preserved, e.g. the famous *Perisphinctes* find in Malm limestone, Lower Schieferhülle (Tollmann 1977a). The Middle Austroalpine Permo-Mesozoic sequence is only poorly preserved and several different sections have been considered to derive a general stratigraphic column, all of which show a similar development, e.g. Brenner Mesozoic, Stangalm Mesozoic (fig.1.3).

The Alpine Permo-Mesozoic sequence rests on basement that is rather different for each of the four tectonic domains. The Middle Austroalpine basement consists of a metamorphic sequence (Altkristallin), which has been affected by several pre-Alpine events and is overlain by patches of Permo-Mesozoic sediments. Parts of the basement have been affected by an Alpine metamorphism dated at $\approx 100-80$ Ma (Mid Cretaceous), which locally reached amphibolite grade (fig.5.3). In the Lower Austroalpine nappes the basement is similar but the Mid Cretaceous metamorphism only reached lower greenschist conditions (Janoschek & Matera 1980). In this area the basement can be subdivided into older pre-Alpine metamorphic slivers (now retrogressed) overlain by weakly metamorphosed Palaeozoic sediments. The Palaeozoic rocks are mainly quartz-phyllites with occasional marker horizons of siliceous shales and marbles (Flügel 1963).

The Pennine nappes lie structurally beneath the Lower Austroalpine units and can be subdivided into an upper and lower tectonic unit

separated by a major tectonic contact (Selverstone 1985, Tollmann 1977a). The Upper Pennine (Upper Pennine = Glockner Decke; Lower Pennine = Venediger Decke) sequence was deposited largely on oceanic crust (Höck & Miller 1980, Bickle & Pearce 1975), which was first formed in the Jurassic (Winterer & Bosellini 1981). The Lower Pennine sheet, however, still contains parts of the original, older continental basement in the form of the gneiss bodies (Zentralgneis) and Palaeozoic phyllites (fig.2.1). The Palaeozoic phyllite is typically represented by a group of calcite-free, dark phyllites with characteristic bands of graphitic quartzite known as the Habach Series (Frasl 1958) found in the north of the Tauern Window.

In the three tectonic domains preserving old continental stratigraphies, the Permoskythian sedimentary facies is a coarse clastic deposit known in many areas as Verrucano. The base of the Verrucano is commonly a conglomerate with occasional volcanic horizons. This grades into quartz-rich deposits, and the Skythian is typically represented by a white quartzite (Gwinner 1978). The Permoskythian clastic sequence is overlain by shallow water reef-carbonates of the Triassic. This change in facies is taken to reflect the extension and gradual subsidence of the developing sedimentary basins (Dewey et al 1973, Frisch 1979) and is seen throughout the Tethyan region.

At some stage in the Early-Mid Jurassic, extension rates increased causing the Triassic carbonate sequence to fracture and form discrete platforms. Eventually the growth of these carbonate platforms was no longer able to compensate for the tendency to subside. The formation and foundering of these carbonate platforms has been documented by Bernoulli & Jenkyns (1974) for parts of the southern margin of the Tethyan Ocean.

As a result of continued extension, oceanic crust began to form, and a number of distinct facies belts were produced. The Upper Pennine

sheet represents a former ocean basin and consist largely of greenschists derived from oceanic tholeiites (Bickle & Nisbet 1972), thick graphitic schists and volumetrically less important pelagic sediments and clastics. During their formation the margins of the ocean basins were occupied by carbonate platforms lying on thinned continental crust, e.g. Radstädter Tauern. Most of the carbonate platforms continued to grow into the early Jurassic and were then overlain by pelagic sediments. At the margins of these carbonate platforms, active extensional faulting produced instability and detritus from the platforms was shed into the neighbouring basins as turbidites and breccias (Bernoulli & Jenkyns 1974, Wiedenmayer 1963). This phase of extension is commonly associated with neptunian fissures (Wendt 1971). The breccias and neptunian dykes are typically Early Jurassic in age, but other similar breccia deposits are also known from the Early Cretaceous. In most cases the younger breccias can be distinguished by the presence of serpentinite clasts, although the Schwarzeck Breccia of the Radstädter Tauern is an exception to this rule (Tollmann 1977a, p.110).

The Alpine sediments of the Middle Austroalpine sheets are thin and not very well preserved. The rock types are very similar to the Lower Austroalpine sequence, but were deposited on substantial continental crust and their primary thickness was probably never very great. The youngest lithologies of the Middle Austroalpine are pelagic cherts and nodular limestones of Lower Cretaceous age.

Using the stratigraphies from these different tectonic sheets, allows a general stratigraphy to be proposed for the Mafrei Zone on the basis of lithological associations (fig.2.2).

2.3 ROCK TYPES PRESENT IN THE MATREI ZONE

2.3.1 Basement

It has long been suggested that the Matrei Zone contains parts of the basement on which the Mesozoic sequence was originally deposited (Cornelius & Clar 1939, Schmidt 1950-52, Exner & Prey 1964). The lack of fossils and the ubiquitous low-grade metamorphism makes it difficult to draw a clear distinction in every case between these rock types and the Alpine, Mesozoic sequence. In general, however, it is possible to distinguish basement rocks which show retrogression from a higher grade of metamorphism (Altkristallin) from those, which on the basis of lithological comparisons, appear to be related to the pre-Alpine Palaeozoic cover. Some attempt has been made to further subdivide the Palaeozoic lithologies, but the main aim of this section is to identify areas where part of the original basement is still preserved.

a. Altkristallin

Throughout the study area the regional metamorphism of the Matrei Zone was lower greenschist facies, and the best method for identifying slivers of retrogressed Altkristallin (known as Diaphthorit in the German literature - Becke 1909) was recognizing relics of higher grade metamorphic minerals. In one case (281, area 3) small heavily-chloritized garnets occur. More commonly, small chloritized crystals of a brown pleochroic minerals, possibly biotite or stilpnomelane, are still preserved. Even where higher-grade minerals are no longer preserved, the presence of large micas (>5mm) can be used as an indication of a former higher grade of metamorphism. Large micas are also known from some Palaeozoic sediments (v.Gosen pers. comm.).

Finally, plagioclase which has either a gefüllte texture (Frasl 1953, Cliff et al 1971), or which has an earlier internal fabric unrelated to Ss (chapter 3) is used to indicate that the host rock has been through a previous metamorphic development. Using these criteria it is possible to distinguish a number of different basement slivers within the Mafrei Zone. In area 3 the positions of these are in close agreement with those suggested by Prey (1961, Exner & Prey 1964) during his investigation of the same area.

b. Dark Phyllite

This is represented by a dark quartz-phyllite containing little or no carbonate and lacking a gradational contact with the Jurassic carbonate sequences. These phyllites have been distinguished by many workers in the Mafrei Zone who assign them a Palaeozoic age (Schmidt 1950-52, Cornelius & Clar 1939, Exner & Prey 1964). This Palaeozoic dark phyllite is distinguished from the superficially rather similar post-Triassic graphitic schist (see later p.24) by the common iron-staining, the presence of feldspar, the general silvery sheen and the lack of any gradation with carbonate bearing lithologies, especially the dolomite breccias.

A particularly characteristic member of the dark phyllite group is the graphitic quartzite which has also been distinguished on the maps. These quartzite bands are in general <1m thick and contain finely disseminated graphite. Similar rock types are known from other areas, e.g. Habach Serie (Tollmann 1977a, p.20), Inner Schieferhülle (Exner & Prey 1964, p.31), where they are always associated with dark carbonate-free phyllite, and are dated as Palaeozoic.

The collective terms Glanz Schiefer, e.g. Angel (1929) and dark phyllite, e.g. (Bickle 1973) have been used to describe both these and the Jurassic graphitic schists. Although the distinction between the

two rock types is not always straightforward, I avoid using either of these terms in their general sense and dark phyllite is restricted to the rocks described above.

c. Quartz chlorite schist

A third rock type which is included in this group but is restricted in its occurrence to area 1, is a light coloured quartz-chlorite schist with abundant rutile needles within the chlorite-rich layers. The rutile needles may indicate exsolution from higher T minerals such as biotite. There is also a variable amount of secondary carbonate. This rock type is associated with dark phyllite and may correspond to the Chloritfleckschiefer described by Cornelius in the Palaeozoic Habach Series (Cornelius & Clar 1939, p.200).

These greenschists are distinguished from the greenschists of the Jurassic by the lack of epidote and the presence of the rutile needles.

d. Quartzite

In area 1, the three rock types above are commonly associated with a light-coloured quartz-rich material which in places can be identified as highly deformed white augengneiss (Bickle 1973, p.30). Elsewhere this rock type is without a significant amount of feldspar and is best described as a meta-quartzite (e.g. B125). These quartzites could be part of the Permo Skythian sequence or, more likely on the basis of their lithological associations, they are meta-quartzites of pre-Alpine age.

2.3.2 Quartz phyllites (Permoskythian) (Helle Matreier Phyllite - Angel)

Locally possible original sedimentary contacts between basement and chloritic quartz phyllite can be identified. In places lenses of vein-quartz are present that may be strongly deformed clasts, and the

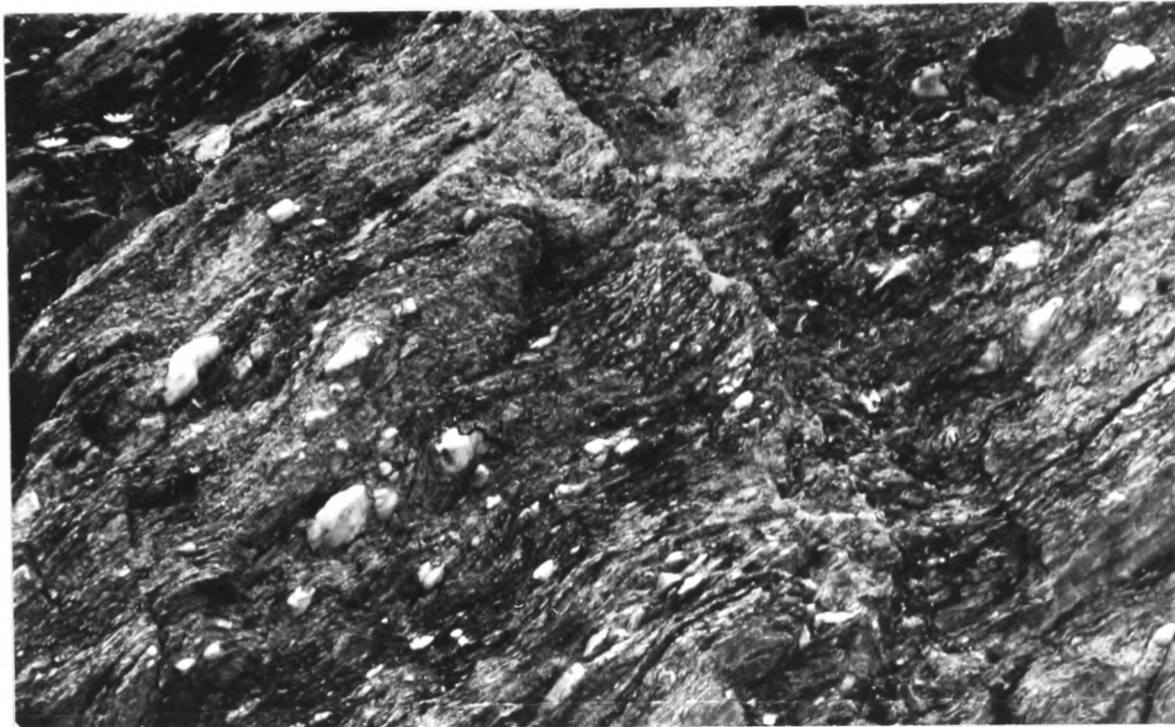
deposit can be compared to the Verrucano. It was not always easy to distinguish the quartz phyllite from retrogressed basement. One of the characteristic features of these quartz-phyllite is the presence of ankeritic veins. It is unclear whether these are a primary or secondary feature of the rock. Going up section the quartz-phyllite grades into more quartz-rich sediments usually assigned to the Skythian.

2.3.3 White Quartz Schists and Quartzite (Buchstein - Angel)

The quartzites are rather irregular, both in thickness and lateral extent, and vein quartz material is relatively common (e.g. 515 area 3). It is unclear to what extent the more massive appearance of the quartzite, compared to the quartz schist, is due to primary differences in sediment type and to what extent this is due to tectonic processes.

Locally the white quartz schist and quartzite contain characteristic quartz-tourmaline clasts up to 15cm in length (fig.2.3). Many of these clasts contain radiating aggregates of green-blue pleochroic tourmaline reminiscent of Luxullianite (tourmalinized granite). At a few localities rose quartz clasts also occur, which are typical for the Permoskythian (Clar pers. comm.). There is no obvious source area for these clasts, either to the palaeogeographic north or south, but pre-Alpine tourmaline is a common mineral in the Altkristallin and tourmaline pegmatites are known in the Kreuzeck Group (Hoke 1987, Exner & Prey 1964). Similar quartz-tourmaline clasts are known from the Lower Austroalpine Semmering quartzite (Cornelius & Clar 1939, p.250), where they are described as lydite, and from the Permoskythian deposits of the Lower Schieferhülle in the Tauern Window (Lammerer pers. comm.).

This work agrees with nearly all the previous workers in this area (Senarclens Grancy 1964, Angel 1928, Cornelius & Clar 1939) by relating the white quartz schists to the Permoskythian of the Matrei Zone.



10cm

Fig.2.3.a Permoskythian quartz schist with dark quartz-tourmaline clasts and light vein-quartz clasts (B61, area 1).



Fig.2.3.b. Quartz schist with strongly deformed clasts of the same composition as above. The large dark clast in the centre of the photo is $\approx 20\text{cm}$ long (loose block east of Ofenspitze area 3).

However, the presence of gefüllte plagioclase led Bickle (1973) to suggest that much of the white quartz schist in area 1 around Peischlach Törl was basement orthogneiss. These schist is, however, associated with lithologies typical for the Mafrei Zone, e.g. greenschist, dolomite, marble, and are also linked to the Mafrei Zone by the presence of the luxillanite clasts, which are found throughout the Mafrei Zone. The gefüllte texture can be best explained as clastic material derived from the erosion of basement which already had feldspar with this type of retrogressive texture (see chapter 3 for further discussion).

2.3.4 Marble, Dolomite and Rauhwake^c (Triassic)

The quartz-rich deposits are commonly overlain by a calcareous rock type, either a marble or a dolomite, with a yellow weathering colour. There is little dispute that these rocks are Triassic, and by analogy with fossiliferous areas, represent a shallow water reef facies. Patches of rauhwake^c are commonly associated with dolomite and marble and are probably of a comparable age.

Rauhwake is a porous calcareous rock, which is commonly associated with evaporites and shallow-water carbonates. Rauhwake associated with gypsum has been reported from various localities in the Mafrei Zone (e.g. Cornelius & Clar 1939, p.255; Exner & Prey 1964), and in area 3 possible rectangular pseudomorphs after gypsum were seen in thin-section.

Two phases of early extension are recorded in the dolomite blocks (i) dolomitized veins cutting many of the outcrops and (ii) neptunian dykes filled with calcite-rich material which does not show any secondary dolomitization (fig.2.4a,b). Bedding can be identified in one locality within the fill of a neptunian dyke, but is not apparent in any of the other Triassic carbonates. In places, however, strong Ds

3cm

Fig.2.4.a. Dolomite olistolith (1) surrounded by chloritic schist of the Bündnerschiefer (2). The dolomite block is cut by an early phase of dolomitized fractures (3) and a later phase of extensional fractures (4) filled with a pure calcite deposit, both of which developed before the block's incorporation into the Bündnerschiefer.

50cm

Fig.2.4.b. Fractured dolomite block with Neptunian fissures filled with fragments of the surrounding block in a pure calcite matrix (MS99 area 3). The larger fragment of the dolomite block contains early dolomitized veins (3).



deformation has produced a foliation in the dolomite and a fine lamination in the marble which can resemble bedding. No fossils were found.

2.3.5 Bündnerschiefer- Post Triassic Sediments

Bündnerschiefer is a useful collective term derived originally from work in the Swiss Alps and adopted to describe similar sequences in the Eastern Alps (Kober 1955, Tollmann 1977a, Frisch 1984). The term covers a wide range of interbedded, post-Triassic rock types, often with gradational contacts. Bündnerschiefer in its various guises is normally thought of as restricted to the Pennine tectonic units, and the presence of some comparable lithologies, has been the main reason for the suggestion that the Mafrei Zone contains Penninic elements (Schmidt 1950-52, Exner & Prey 1964, Gommeringer 1985). A number of basic sediment types belonging to this group can be distinguished in the Mafrei Zone, all of which may be locally modified either by a gradation between each other or the addition of clastic material usually dolomite or quartz.

The most readily identifiable and widespread clastic component is dolomite sand which is interpreted as being derived from the Triassic carbonates. The presence of this material is an important tool in identifying post Triassic rock types and helping identify olistoliths (Frisch et al 1987, Gommeringer 1985). The dolomite sand grains are normally quite well-rounded and commonly have an opaque rim (fig.2.5). The form of these clastic grains distinguishes them from the euhedral rhombs of secondary dolomite found at many localities. The opaque rim is interpreted by Frisch et al (1987) as an Fe-Mn crust produced by subaerial erosion or by prolonged periods of low sedimentation rate above a basin high. Although this may be true in some cases, the

secondary euhedral rhombs of dolomite may also be pseudomorphed by opaques (fig.2.6). Dolomitic sand has been identified elsewhere in the Eastern Alps e.g. Tollmann (1977a, p.109) in the Radstädter Tauern, where it is present in Lower Jurassic sediments.

a. Calc Schists (phyllitic calc phyllite and massive calc phyllite - Bickle)

Quartzitic calc-schist with minor amounts of graphite and platy minerals is the most typical Bündnerschiefer lithology. It is the most widespread rock type in the Upper Pennine units (fig.2.7) and forms an important part of the Mafrei Zone. The calc-schists show a gradation with all the other Bündnerschiefer rock types and can be looked upon as the background sediment for these other types. Commonly the calc-schists grade into horizons of quartzitic marble, the most important of which have been distinguished on the maps.

b. Quartzitic Marble (e.g. Bretterich Marble)

The most important outcrop of this lithology is the Bretterich Marble in the Makerni Spitze area (area 3), where a quartzitic marble makes up the prominent Bretterich Ridge immediately to W of Fragner Hütte. The deposit is lenticular and at its thickest it is some 50m, but pinches out to the east and west. The lenticular shape seems to be a primary feature, but may have been accentuated by later tectonic effects. Similar rock types occur in other parts of the Mafrei Zone, where they are always associated with graphitic calc-schists. In the Makerni Area the Bretterich Marble is overlain by, and in part interbedded with greenschist and underlain by graphitic calc-schist. Prey (Exner & Prey 1964) describes minor dolomite breccia horizons within this unit and on this basis assigns it to the Jurassic. The high proportion of both quartz and dolomite clastic material within this marble suggests that it

1mm

Fig.2.5. Quartzitic marble containing rounded sand grains of micritic dolomite. (q = quartz, d = dolomite).

0.5mm

Fig.2.6 Rhombs of secondary dolomite in a quartz phyllite being replaced by opaques (B3, area 1).

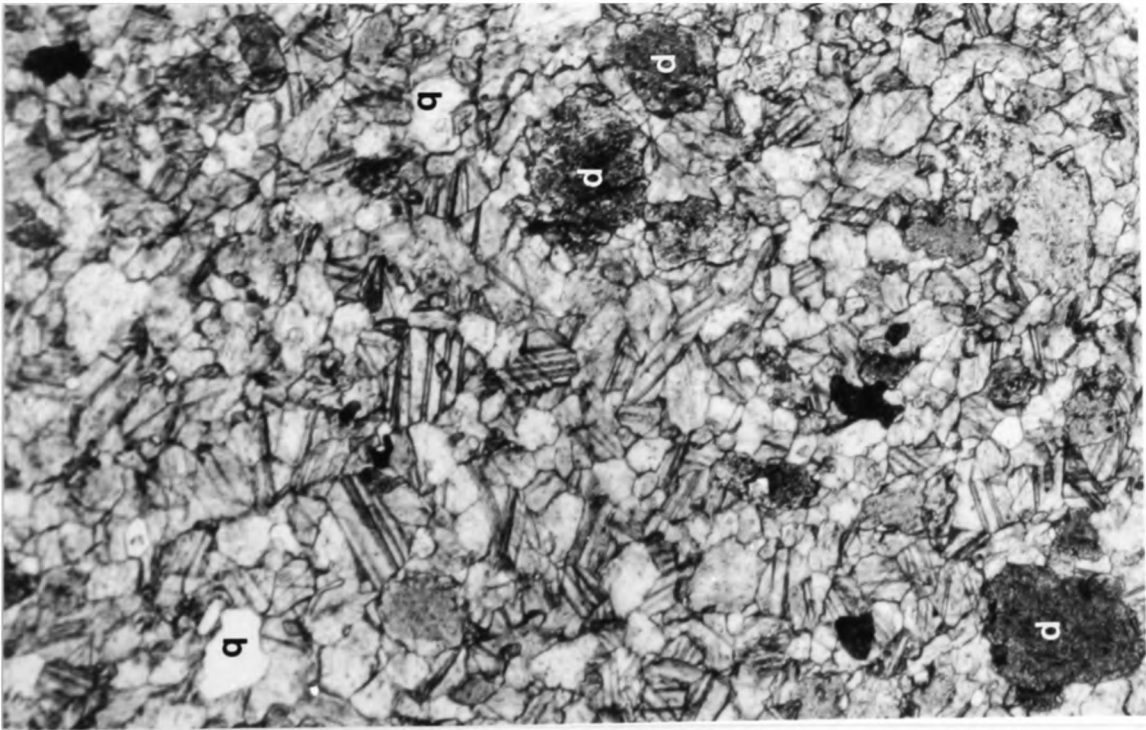
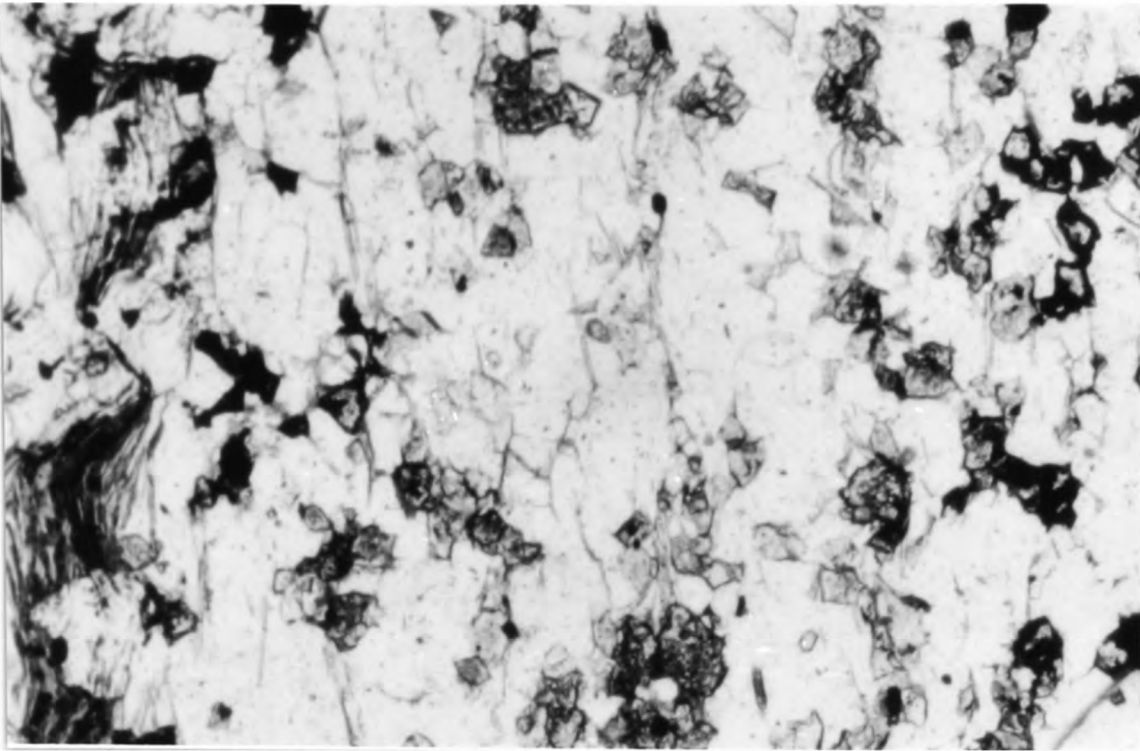




Fig.2.7. The imposing Bratschenhänge of the Pennine Bündnerschiefer composed entirely of calcareous schist, area 1 looking north (Studl Hütte for scale, bottom left).

was not formed in a low-energy carbonate environment and Frisch et al (1987) suggest this represents a mass-flow deposit. A fine lamination is commonly developed within this lithology, which is defined by flattened quartz grains and is not a primary bedding feature as suggested by Gommeringer (1985) and Frisch et al (1987).

c. Greenschist

Two types of greenschist can be identified both of which I interpret as metavolcanics. The more typical metabasic rock type contains amphibole, feldspar, epidote, and chlorite commonly with minor amounts of quartz, Fe^{++} biotite (amphibole bearing greenschist is commonly referred to as Prasinit in the Eastern Alps). Both rock types grade into green calcareous phyllite and are associated with dolomitic sand. Some of these rocks have been chemically analysed, (Höck & Miller 1980, Bickle & Nisbet 1972) and these authors suggest that they are derived from oceanic tholeiite precursors. One locality (Bl27) contains a rhombohedral shaped inclusion of epidote with minor amounts of calcite, which may be a pseudomorph after lawsonite. Similar pseudomorphs have been described from the greenschists of the Pennine area (Höck 1974, Droop 1978).

The second type of greenschist is a chlorite phyllite, which is only well developed in a few places, but the calc schist commonly contains chlorite-rich lenses which may be equivalents. The mono-mineralic nature of the rocks suggests there may have been some metasomatic influence (e.g. MS88, see appendix 2 for chemical analysis).

d. Quartzites

Quartz sand is present throughout most of the Bündnerschiefer material, and the typical calc schists often contain interbedded quartz-

rich horizons. In contrast to the older Permoskythian quartz phyllites, the younger sediments nearly always have a high calcite content and contain quantities of dolomitic sand. However, in a few cases the quartz and feldspar component dominates over the calcareous material, and the sediments are very similar to the Permoskythian deposits. Work in the Penninic Bündnerschiefer has already revealed that there are feldspathic quartzites of post Triassic age (Frasl 1958 p.359-361, Braumüller 1939). In the Matrei Zone, as with the other localities, the lack of fossils means the younger quartzites have to be identified by interbedding and gradation with other Bündnerschiefer rock types. Outcrops of such quartzites can be seen north of Saukopf, area 1 (interbedded with greenschist); SW ridge of the Makerni Spitze area 3 (interbedded with quartzitic marble and greenschist). The largest deposit of this kind is in area 3, where a 1km long lenticular body of quartzite with a thickness of up to 20m, is enclosed within a matrix of graphitic and chloritic schists. The upper and lower margins of the quartzite are decorated with blocks of dolomite and the surrounding matrix contains large quantities of both dolomitic and to a lesser extent quartzofeldspathic clastic material. Along strike the quartzite layer passes into isolated dolomite olistoliths. The association with Bündnerschiefer lithologies containing a high proportion of clastic material and olistoliths demonstrate the post-Triassic age of this quartzite.

The quartz and feldspar detritus is derived from basement or older quartz phyllite, and although very difficult to recognize, it is sometimes possible to find larger coherent blocks of the retrogressed basement, which have become incorporated within the quartz phyllites (e.g. Area 3, 281; Cornelius & Clar 1939, p.248).

c. Graphite Schists (Lower Jurassic?), (dark phyllite - Bickle; Glanz Schiefer - Angel)

Dark graphite schists with only minor amounts of calcite are rare in the studied areas, and are best developed to the north of the Medel Spitze. These rocks show a millimetre layering of fine-grained graphite with varying quantities of dolomite and quartz sand. Some sedimentary structures are still preserved within the graphite schists, and north of the Medel Spitze (B101) centimetre-scale graded bedding and cross-lamination could be discerned (fig.2.8). The sedimentary structures together with the high proportion of clastic material suggests these rocks were deposited by turbidity currents.

f. Breccias

Interbedded with the graphite schists there are breccia horizons with dominantly dolomite clasts up to 30cm, although locally quartz-phyllite, quartzite and retrogressed basement clasts also occur (e.g. Prey 1964). In some places rip-up clasts of the surrounding graphite schists are also present, which can be used as a way-up indicator (fig.2.9). Similar breccias are characteristic of the Lower Jurassic in the other fossiliferous areas e.g Radstädter Tauern (Tollmann 1977a), Tarntaler Berge (Clar 1940) and were probably formed at the unstable margin of carbonate platforms (Bernouilli & Jenkyns 1974).

2.3.6 Olistoliths

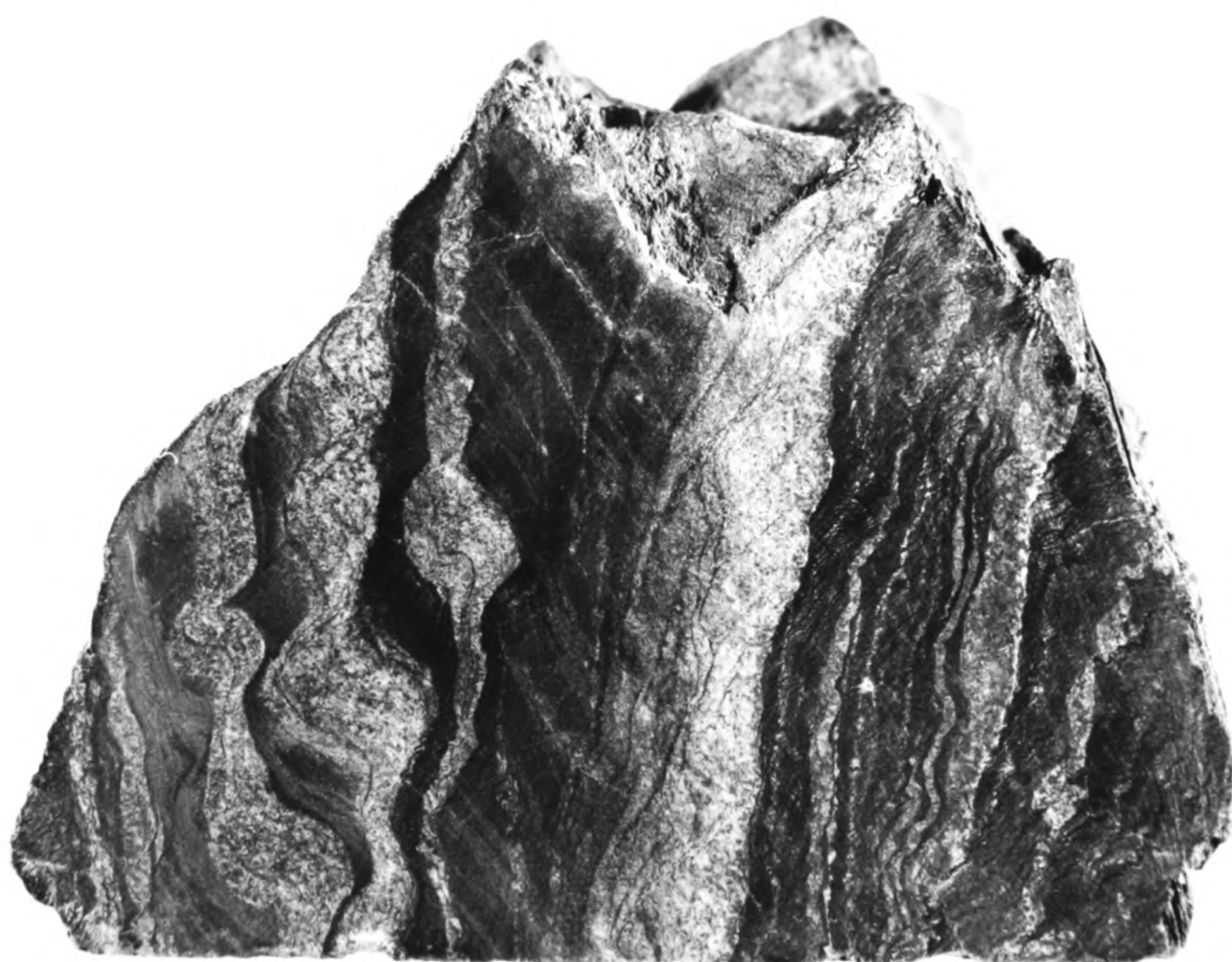
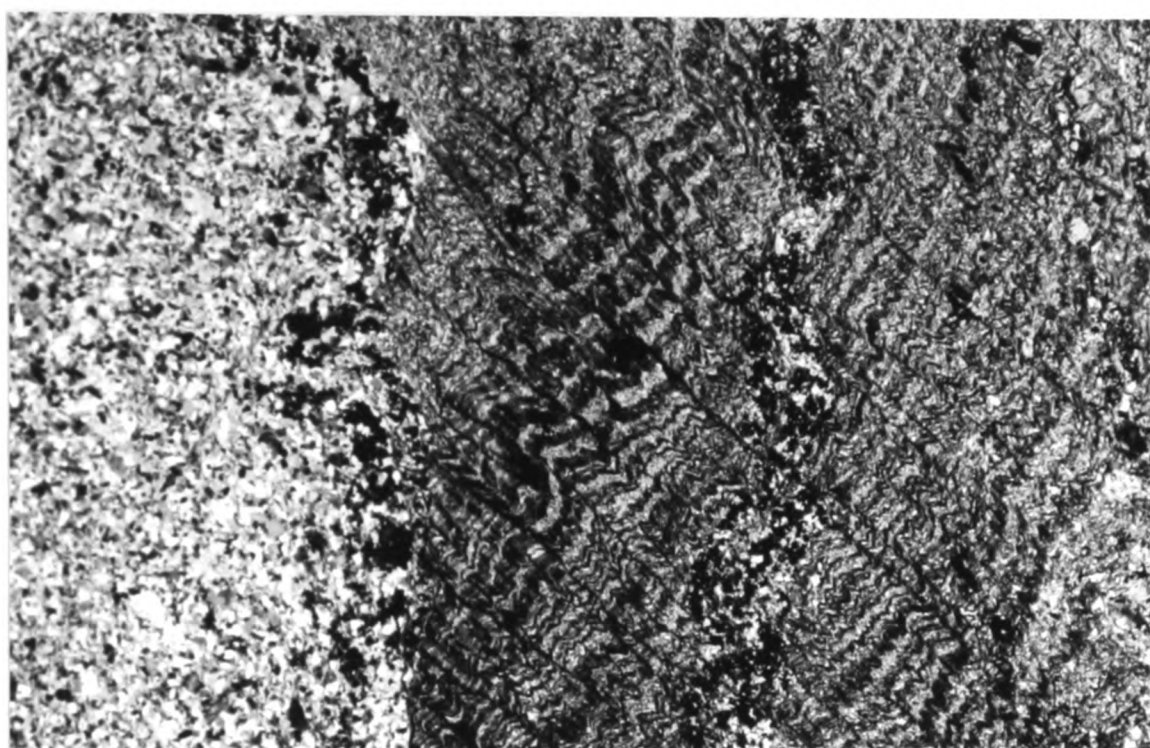
As well as the turbidites and breccias it has long been suspected that large sedimentary blocks, or olistoliths, are present within the post-Triassic sediments of the Mafrei Zone (Exner & Prey 1964, Frisch et al 1987). To test this idea, criteria need to be established to allow such sedimentary blocks to be identified and to be distinguished from tectonically emplaced slivers.

2cm

Fig.2.8.a. Rhythmically bedded, graded units within graphitic shists. The coarser layers are composed of dolomite and quartz sand (B101, area 1).

2mm

Fig.2.8.b. Thin section across the erosional base to the main coarse layer in hand specimen (B101), which overlies a cross-laminated graphitic layer.



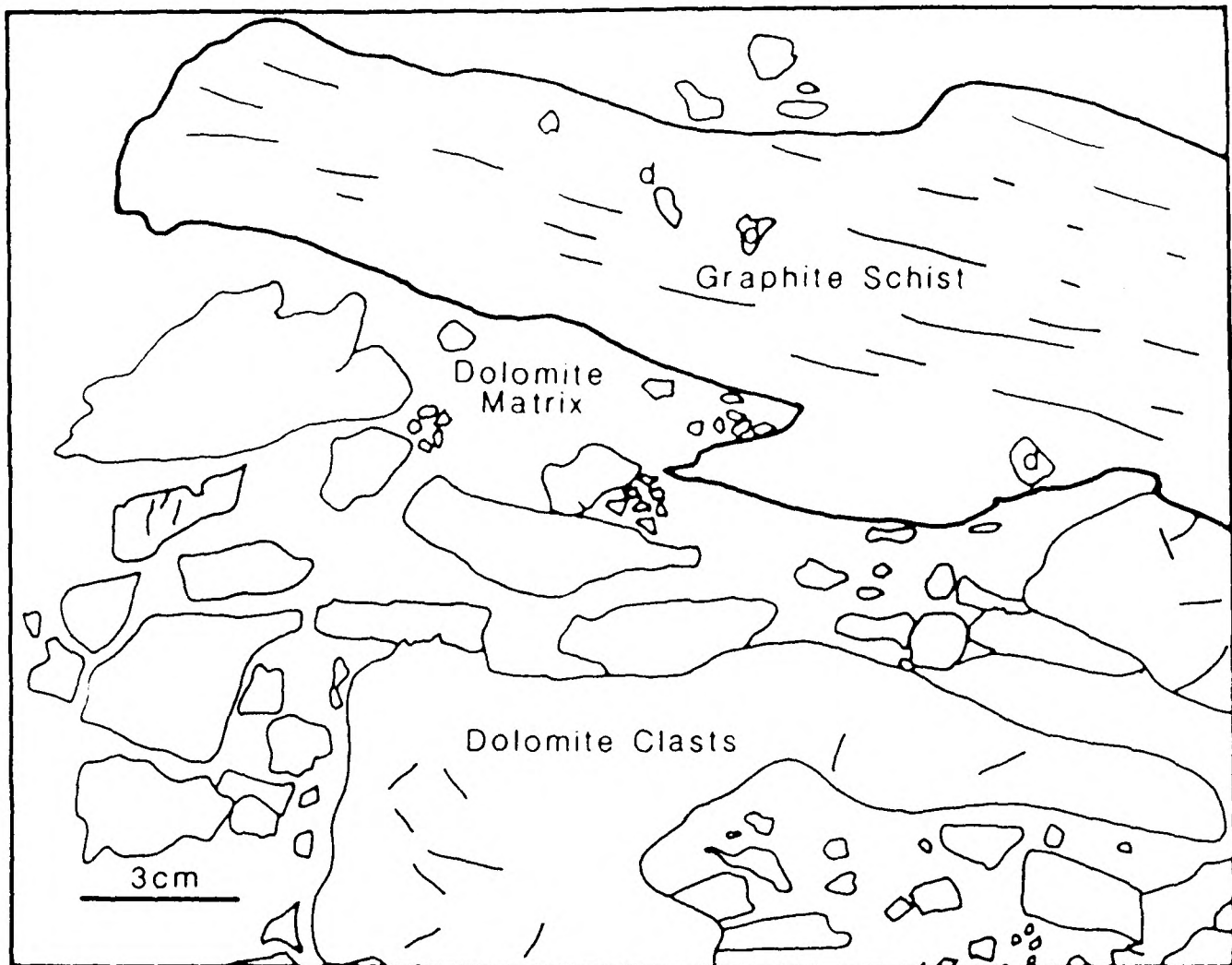
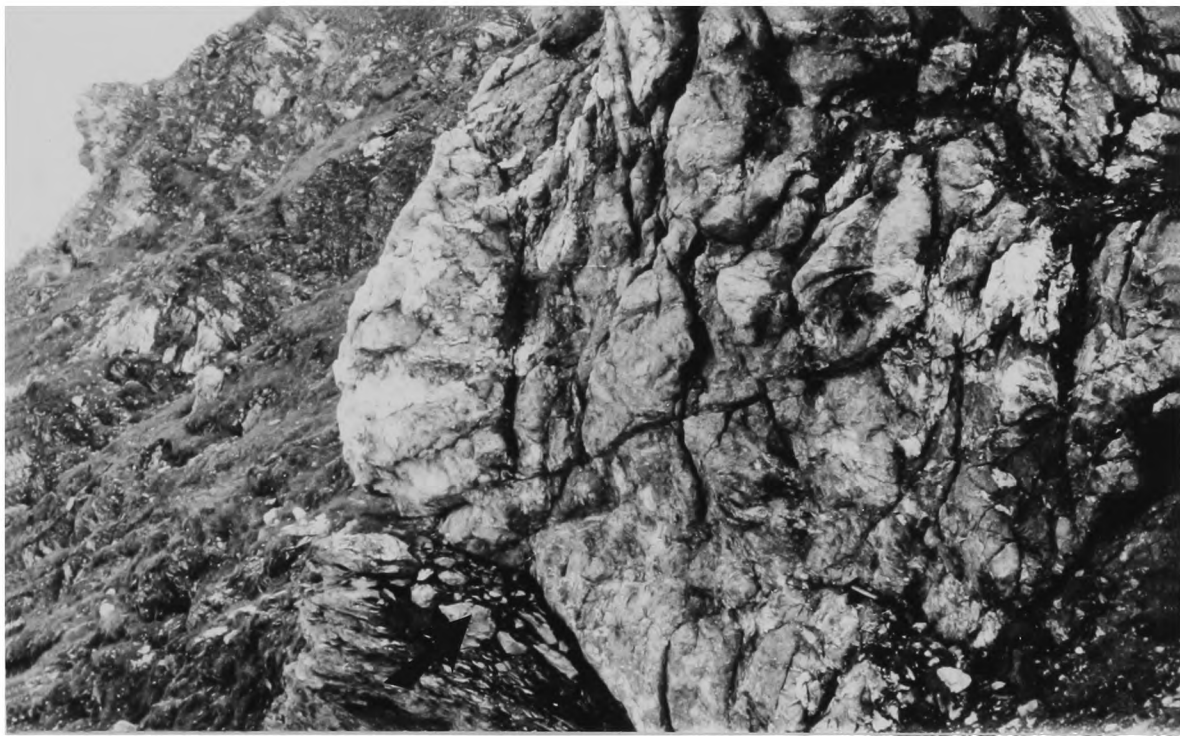


Fig.2.9. Dolomitic breccia containing a rip-up clast of graphitic schist (drawn from field photograph).



1m

Fig.2.10. Dolomite olistolith embedded in calcareous schist of Bündnerschiefer. Immediately adjacent to the main block, the matrix contains smaller fragments of clastic dolomite (see arrow).

An olistolith has to be deposited in a background sediment that is younger than the block itself. Within the Matrei Zone there are many prominent Triassic dolomite blocks, which occur within Skythian white quartz schists. Since the enveloping material is older, the blocks must have a tectonic rather than a sedimentary origin. In many other cases, however, these blocks are surrounded by younger greenschist or calcareous schist and, therefore, represent possible olistoliths.

The sediment immediately adjacent to many of these lenses contains abundant dolomite sand fragments. Frisch et al (1987) interpret these fragments as a shower of fine-grained clastic material dislodged at approximately the same time as the larger dolomite olistolith. The increase in dolomite sand towards an isolated dolomite block is strong evidence in favour of a sedimentary origin (fig.2.10). The same idea can be applied to putative Permoskythian and basement olistoliths. In this case the associated finer-grained clastic component would be dominantly quartz and such olistoliths would be surrounded by a matrix of quartz phyllite. The lithological difference between the olistolith and its reconstituted matrix is, therefore, likely to be very minor and difficult to recognize. A possible example of one such olistolith is 281 (area 3), where a basement block lies within a quartz-rich sediment which grades into greenschist and Bretterich Marmor (see also Cornelius & Clar 1939, p.248).

As well as studying the background sediment, the blocks themselves can preserve evidence about their tectonic or sedimentary origin. To produce the discrete olistoliths, the margin of a raised platform has to be fractured until it becomes unstable and the dislodged material can move down the escarpment. Before the final stages of dismembering, fractures have the opportunity of filling with any overlying sediment. A pure calcite deposit fills fractures within many of the isolated dolomite blocks. The vein material is restricted to the blocks and is

quite distinct from the enveloping phyllite and schist. These veins are probably an original sedimentary fill within an extensional fracture (fig.2.4). Similar neptunian fissures have been described from elsewhere in the Alps (Bernoulli & Jenkyns 1974, Wendt 1971) where the fill is usually a Lower Jurassic red biomicrite. In the Matrei Zone the fill shows no signs of a red colouring or associated iron oxides, but is composed entirely of calcite with brecciated dolomite derived from the fractured block.

A study using cathodo-luminescence reveals a primary lamination wrapping around the fractured margin of some veins (fig.2.11), which may be of algal origin or due to original crystal zonation. Both possible explanations suggest the infilling of an open fracture. In one case possible calcareous neptunian dykes within a quartz phyllite block were seen.

At locality MS99 a large dolomite olistolith contains a ten metre scale neptunian fissure, which still preserves bedding defined by layers of angular dolomite fragments probably derived from the host block (fig.2.12). With the help of other criteria, the presence of neptunian fissures is a relatively good way of identifying olistoliths. These fissures may, however, be intruded into platform carbonate material without it becoming dismembered and forming discrete sedimentary blocks.

Using the three criteria of background sediment, surrounding fine clastic material, and neptunian veins, a number of olistoliths were identified. These are mainly dolomite, but a few quartzite and basement blocks were also identified. The results suggest that the Matrei Zone contains an important but not dominant olistolithic component.

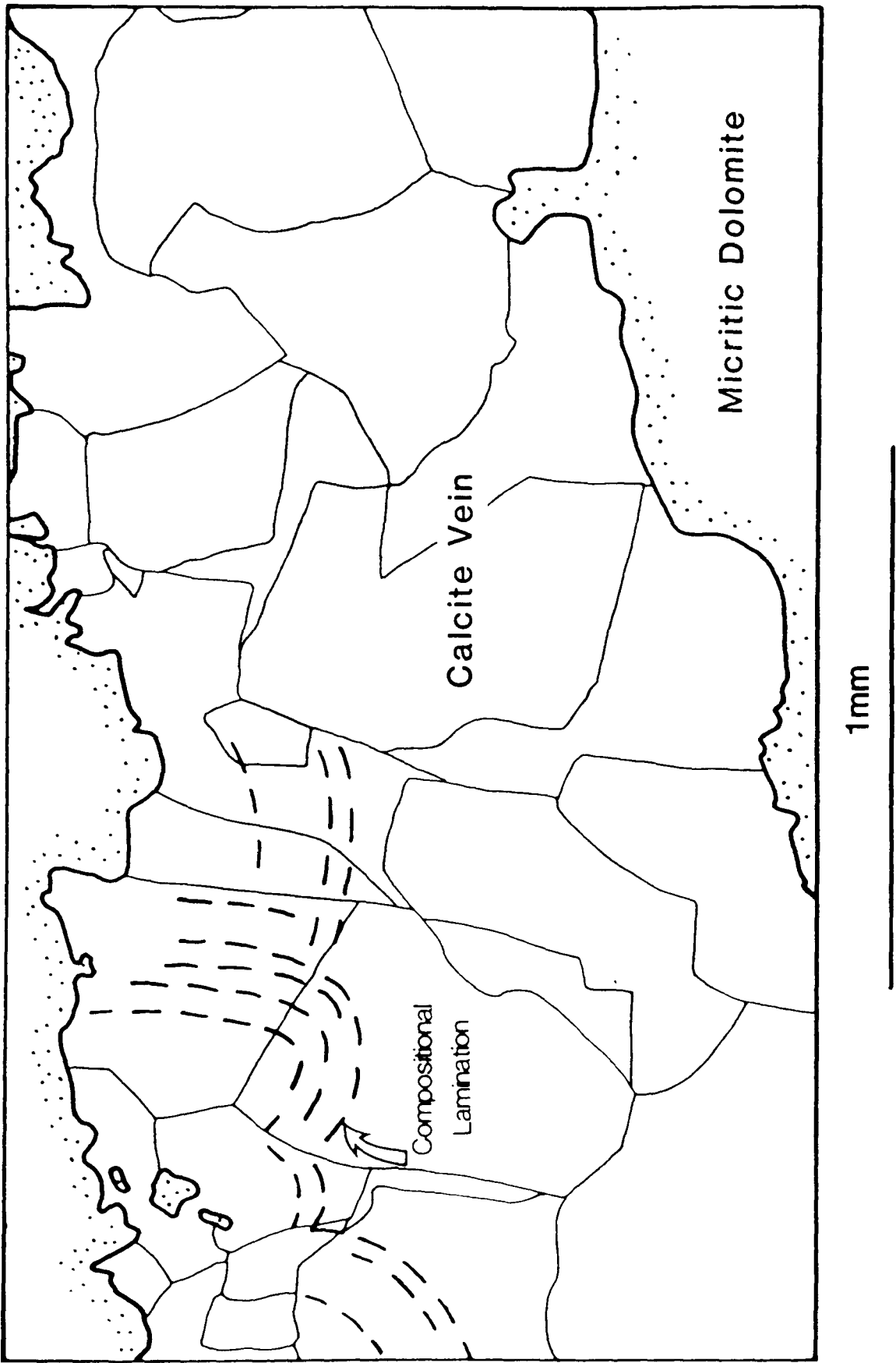


Fig.2.11 An early fracture in a dolomite block with a possible neptunean calcite fill. The compositional lamination cuts across grain boundaries in the calcite and is probably a primary feature.



Fig.2.12. Neptunian fissure within a dolomite olistolith. Bedding is defined within the fill by the alignment of dolomite fragments, which are probably derived from the surrounding block. Hammer for scale, MS99 area 3.

2.3.7 Serpentinite

Serpentinite is present only in minor quantities in the areas which were investigated, although according to the mapping done by Schmidt (1950-52) serpentinite occurs more commonly in the west of the zone. In area 3 the outcrops were entirely restricted to thin, much-altered slivers now present as talc schist, which occur along the base of retrogressed Altkristallin layers and probably mark major tectonic contacts (figs. 3.40, 3.41). In areas 1, 2 there are a few larger masses of serpentinite, which may represent fragments of former oceanic crust. No relic igneous minerals or textures were found.

2.3.8 Pelagic Sediments

Two volumetrically minor rock types overlie the serpentinite body of the central Glorier Hütte area (area 1). Immediately above the serpentinite is a thin layer of graphitic schist. Above this is a green fissile marble, which has been interpreted by Cornelius & Clar (1939, p.266) as Aptychen Kalk (a pelagic limestone found in the Eastern Alpine region - Bernouilli & Jenkyns 1974). Directly above the calcareous sediments lies a quartz phyllite with bands up to 10cm thick of Mn-bearing quartzite. The Mn content of the quartzite was high enough to have made it commercially viable to exploit a continuation of this layer (Clar pers. comm.). Cornelius & Clar (1939, p.625) describe these quartzites as

" the only definite outcrop of radiolarite in the map area (das einzige ... sichere Vorkommen von Radiolarit innerhalb der ...Karte)".

The quartz phyllite contains layers of dolomitic sand and to the south contains 10m sized dolomite olistoliths, implying that the area of deposition of the radiolarite was subject to frequent incursions of clastic material from a nearby basin high. The radiolarite also contains rhombs of dolomite scattered throughout the matrix; these,

however, probably grew secondarily and do not represent a clastic dolomite input.

Serpentinite bodies directly overlain by pelagic sediments are relatively common in some parts of the Tethyan realm, and Lemoine et al (1987) interpret these as remnants of ocean crust.

2.4 PRESERVED SEQUENCES AND STRATIGRAPHY

With the exception of the serpentinite and associated pelagic sediments in the Glorer Hütte area, the Mafrei Zone can be divided into four stratigraphic units: basement, quartz phyllite, Triassic carbonates, and the Bündnerschiefer. By comparison with other areas this fourfold division clearly implies a general stratigraphy for the Mafrei Zone (fig.2.3).

Transition analysis (Naylor & Woodcock 1977) was applied to the area as a purely statistical test of whether this crude stratigraphy could be used to help understand the complex structure of the Mafrei Zone or whether the original stratigraphy had been completely obliterated. The results show that proposed sequence is significant in both areas 1 and 3 at the 0.05% level (appendix 1) and suggests that although strongly disrupted the Mafrei Zone still contains fragments of the original stratigraphic sequence.

The serpentinite body in area 1 does not fit into this stratigraphic scheme and is probably a fragment of oceanic crust with overlying pelagic sediments.

Two other specific localities warrant particular attention, as conflicting ages have been assigned to these rocks by other workers. To the north of the Bretterrich Kamm in area 3 there is a poorly exposed chloritic quartz-phyllite associated with feldspathic quartzites, some Triassic dolomite and marble with lesser amounts of calc-free dark

phyllites, greenschist and calc phyllites. The feldspathic quartzites show no gradation with the calcareous schists and in agreement with Prey (Exner & Prey 1964) and Schmidt (1950-52) I have interpreted them as Permoskythian. The main body of chloritic quartz phyllite has been strongly deformed and has a prominent differentiated layering. No large micas or other signs of a basement precursor were discovered. In contrast to this work Frisch et al (1987) suggest these phyllites may be of Jurassic age, which implies the associated dolomites and quartzites are surrounded by younger sediments and may be olistoliths. The lack of any corroborative evidence that these quartzite and dolomite layers are olistoliths, and the lack of calcareous gradations within the quartz phyllite lend support to the suggested Permoskythian age for these units.

North of the Medel Spitze there is a band of fissile chloritic quartz phyllites surrounded by calcareous schists, in which Cornelius & Clar (1939 p.248) found a garnet-bearing lens of amphibolite and they suggested the whole layer could be retrogressed basement. In support of this suggestion, in this study a number of lenses of carbonate free dark phyllite were also discovered in the same body of quartz phyllite. However, the quartz phyllite clearly contains interbedded layers of sedimentary dolomite breccias and in places it grades into a calcareous Bündnerschiefer rock type. The most likely explanation for these observations is that the layer consists dominantly of Bündnerschiefer phyllites containing lenses of older material either as tectonically emplaced slices or olistoliths.

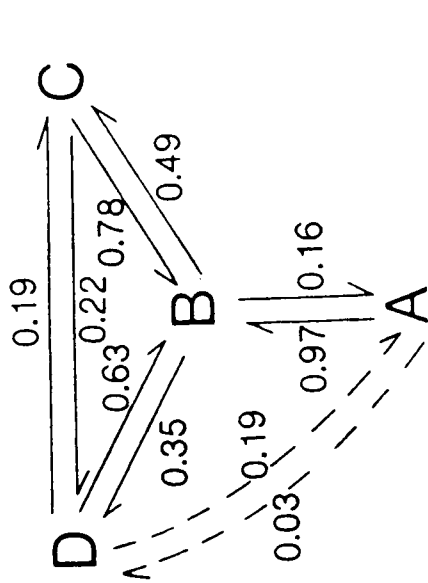
2.5 TECTONIC SETTING FOR THE MATREI ZONE

2.5.1 Mechanisms producing disruption

The discontinuous nature of many units within the Matrei Zone can be explained as either the result of tectonic processes or the sedimentation of large blocks of material as olistoliths, and the tectonic interpretation of this zone is largely dependent on the suggested nature of this disruption. The role of these two processes has been discussed by a number of workers (Exner & Prey 1964, Bickle & Hawkesworth 1978, Schmidt 1950-52) and recently Frisch et al (1987) have emphasized the importance of olistoliths surrounded by a Bündnerschiefer matrix both in the Matrei Zone and other parts of the Eastern Alps.

If the Matrei Zone consisted entirely of olistoliths set in a Bündnerschiefer matrix, a statistical analysis of the number of transitions should not be significant for the proposed general stratigraphy and transitions to Bündnerschiefer should be likely from all of the different lithologies. The probability matrix, P (fig.2.13) represents the likelihood of a particular transition occurring in areas 1 and 3. In area 3, the two most likely sequences are basement- quartz phyllite- Triassic carbonate - Bündnerschiefer; and Basement - Quartz phyllite - Bündnerschiefer. These represent partially preserved sedimentary sequences within the Matrei Zone. The greater disruption of area 1 is shown by there being no clearly defined preferred sequence. The transition matrix is, however, significantly non-random against the proposed stratigraphic sequence (appendix 1), which suggests that although strongly deformed this area still contains parts of the original stratigraphy. Clearly neither area can be interpreted as a series of discrete blocks entirely enclosed within a Bündnerschiefer matrix.

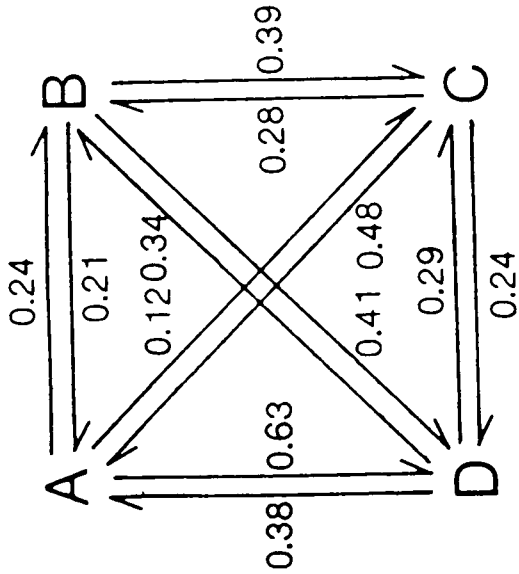
AREA 3



P_{ij}

	A	B	C	D
A	1	0.97	0.0	0.03
B	0.16	1	0.49	0.35
C	0.0	0.78	1	0.22
D	0.19	0.63	0.19	1

AREA 1



P_{ij}

	A	B	C	D
A	1	0.24	0.12	0.63
B	0.21	1	0.39	0.41
C	0.48	0.28	1	0.24
D	0.38	0.34	0.29	1

Fig.2.13. Probability matrices for transitions within the Matrei Zone in areas 1 and 3. A = basement , B = quartz phyllite, C = Triassic carbonate, D = Bündnerschiefer (see appendix 1 and text for further discussion).

An alternative hypothesis, favoured by Frisch et al (1987) is that the Bündnerschiefer contains large sedimentary blocks, which preserve a partial stratigraphy within them. This situation cannot be distinguished from tectonic effects on the basis of transition analysis alone.

The Permo-Skythian quartz-phyllite sheets immediately adjacent to the Altkristallin in areas 3 and 1, lack any suitable background sediment to be olistoliths (see cross-sections, figs.3.40, 3.42), and Frisch et al (ibid) assign these to the Austroalpine tectonic units. They define the Mafrei Zone as lying to the N of these sheets. In the study area, they identify two particular units in the Mafrei Zone as large olistoliths; the Medl Spitze dolomites (area 1) and the Makerni Spitze dolomite-quartzite complex (area 3). Both these units comprise imbricated quartz phyllite - dolomite sequences, which are enveloped by Bündnerschiefer material. The contact between these units and the surrounding matrix is the critical area for deciding if these represent olistoliths or tectonically emplaced slices.

Where they can be positively identified, olistoliths are characterized by fractured margins and the presence of neptunian fissures. The surrounding material is commonly full of finer-grained clastic material. None of these criteria is fulfilled for the Medl Spitze or Makerni Spitze units. I therefore interpret both these units as tectonically emplaced slices whose complex geometries are the result of tight folding of already imbricated sequences.

Most of the disruption in the Mafrei Zone is best explained as the result of tectonic processes. However, the complex relationships cannot be explained as a result of thrusting and folding of a simple sedimentary sequence. In area 3 the transition from metamorphic basement to quartz phyllites and Triassic carbonates is commonly preserved, but it is less common to find the boundary between Triassic

carbonates and the overlying Bündnerschiefer (fig.2.13). This particular stratigraphic boundary is the junction between rock types with very different rheologies, and its poor preservation potential can be explained if this contact was used as a detachment horizon during thrusting. Another possible contributing factor is syn-sedimentary extensional faulting, which could remove parts of the sedimentary sequence (fig.2.14). This effect could also account for the relatively common occurrence of Bündnerschiefer lithologies overlying Permoskythian deposits with no intervening Triassic, in area 1. Later tectonic processes such as boudinage and footwall plucking during thrust propagation (Platt & Leggett 1986) can also lead to similar effects (fig.2.14).

Another potential contributing factor to the disruption is shale diapirism (Barber et al 1987), which is known to operate in recent accretionary complexes. Only very minor evidence for the mobility of Bündnerschiefer is present in the form of veining, and this process is not considered important in this area. However, the block of glaucophane- and aegerine-bearing high P metabasic found by Cornelius and Clar (1939, p.246) in the westward continuation of the Matrei Zone, where there is little other evidence of high P metamorphism, suggests diapirism from depth may be a more important mechanism in other parts of the zone.

2.5.2 Tectonic setting for olistolith formation

The major part of the disruption in the Matrei Zone is due to tectonic processes. There is, however, clear evidence for the existence of syn-sedimentary instability represented by a clastic input ranging in scale from 100m olistoliths to sand particles. The components of the clastic particles suggests a source area consisting mainly of dolomite with lesser amounts of quartzo-feldspathic material. The clastic

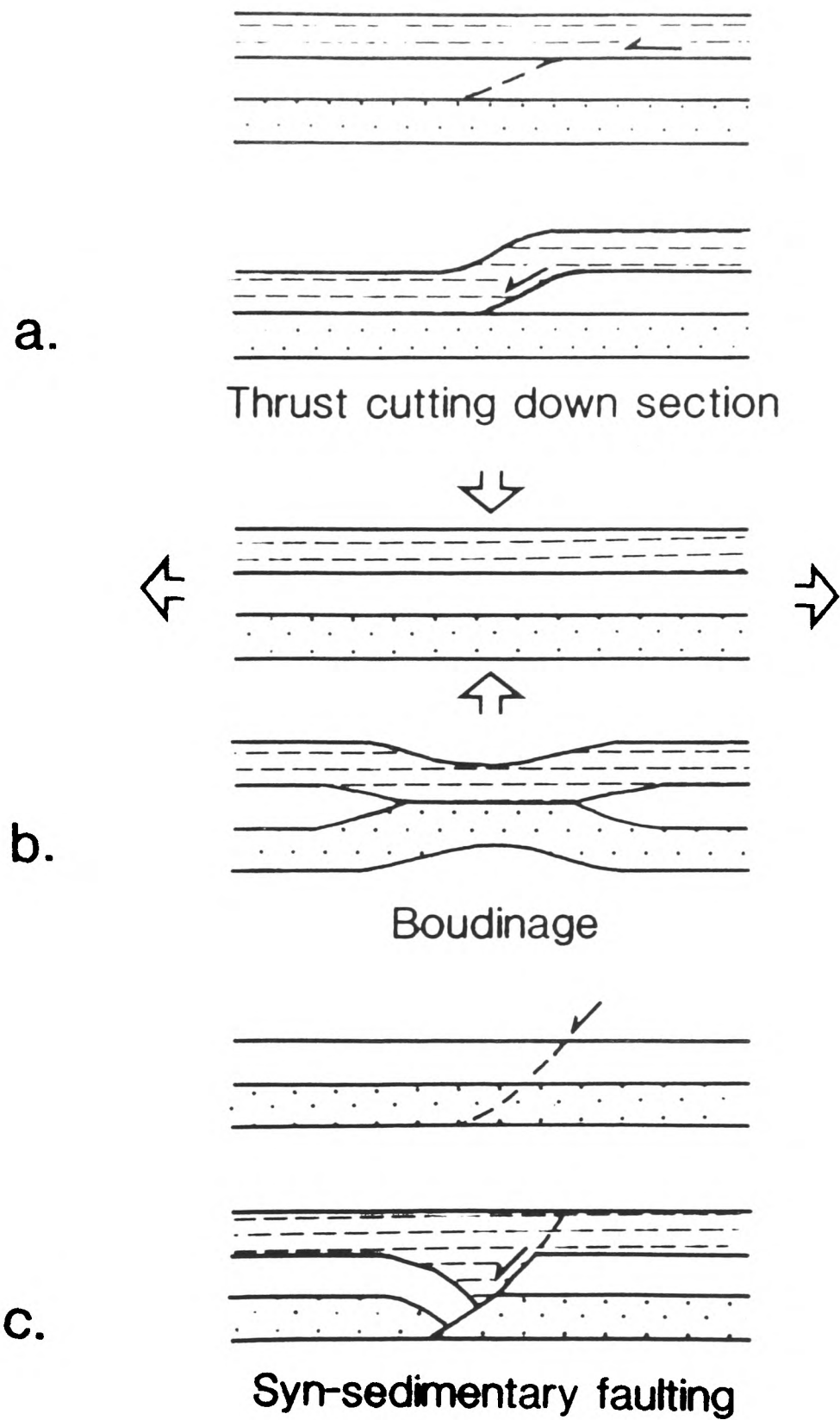


Fig.2.14. Possible tectonic mechanisms for locally removing material from within the stratigraphic sequence.

sediments are all associated with similar graphitic calc-schists, but the possibility that more than one episode of clastic input is present cannot be excluded.

Traditionally, the tectonic instability represented by similar clastic sediments found in many parts of the Alps, is explained as the result of Jurassic block faulting (Bernouilli & Jenkyns 1974). This faulting is related to the rapid increase in spreading rate of the Tethyan region, which took place at around 180Ma (Winterer & Bosellini 1981), with the consequent foundering of the existing carbonate platforms (fig.2.15).

However, Frisch et al (1987) consider the clastic material found within the Mafrei Zone to be due to olistoliths and finer-grained material falling down from the overriding Austroalpine tectonic units into a Mid Cretaceous subduction zone trench with a background sediment of Bündnerschiefer (fig.2.16).

The critical difference between the model proposed by Frisch et al and the sedimentary instability caused by extensional tectonics, is the predicted age of olistolith formation and sedimentation.

Subduction probably began in the Early Cretaceous (see chapter 5), whilst instability due to extension took place before convergence began and is of Mid to Late Jurassic age. Without the help of fossils it is impossible to come to any definite conclusions about the age of the sediment surrounding the olistoliths. However, a number of educated guesses can be made. By comparison with other fossiliferous areas the dominantly dolomitic breccia of area 1 is likely to be of Jurassic age. The radiolarite (area 1), which contains a number of dolomite olistoliths, is also likely to be of Jurassic age (Malm?). The greenschists and graphitic schists which envelop the majority of the olistoliths are not stratigraphically restricted in occurrence and nothing definite can be said about their age. A further observation

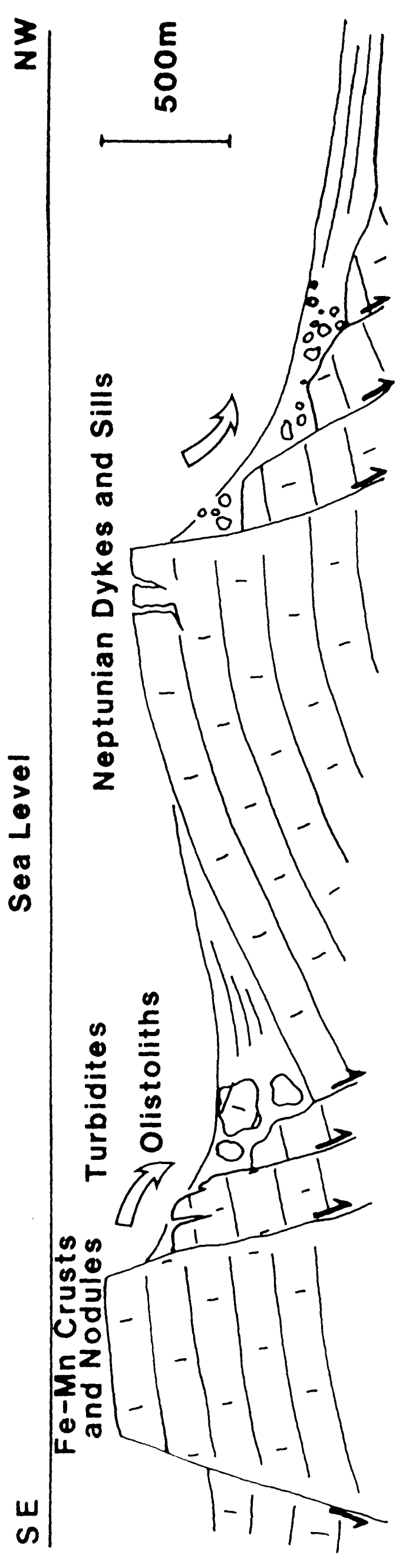


Fig.2.15. Tectonic setting for synsedimentary instability during rifting (after Bernoulli & Jenkyns 1974).

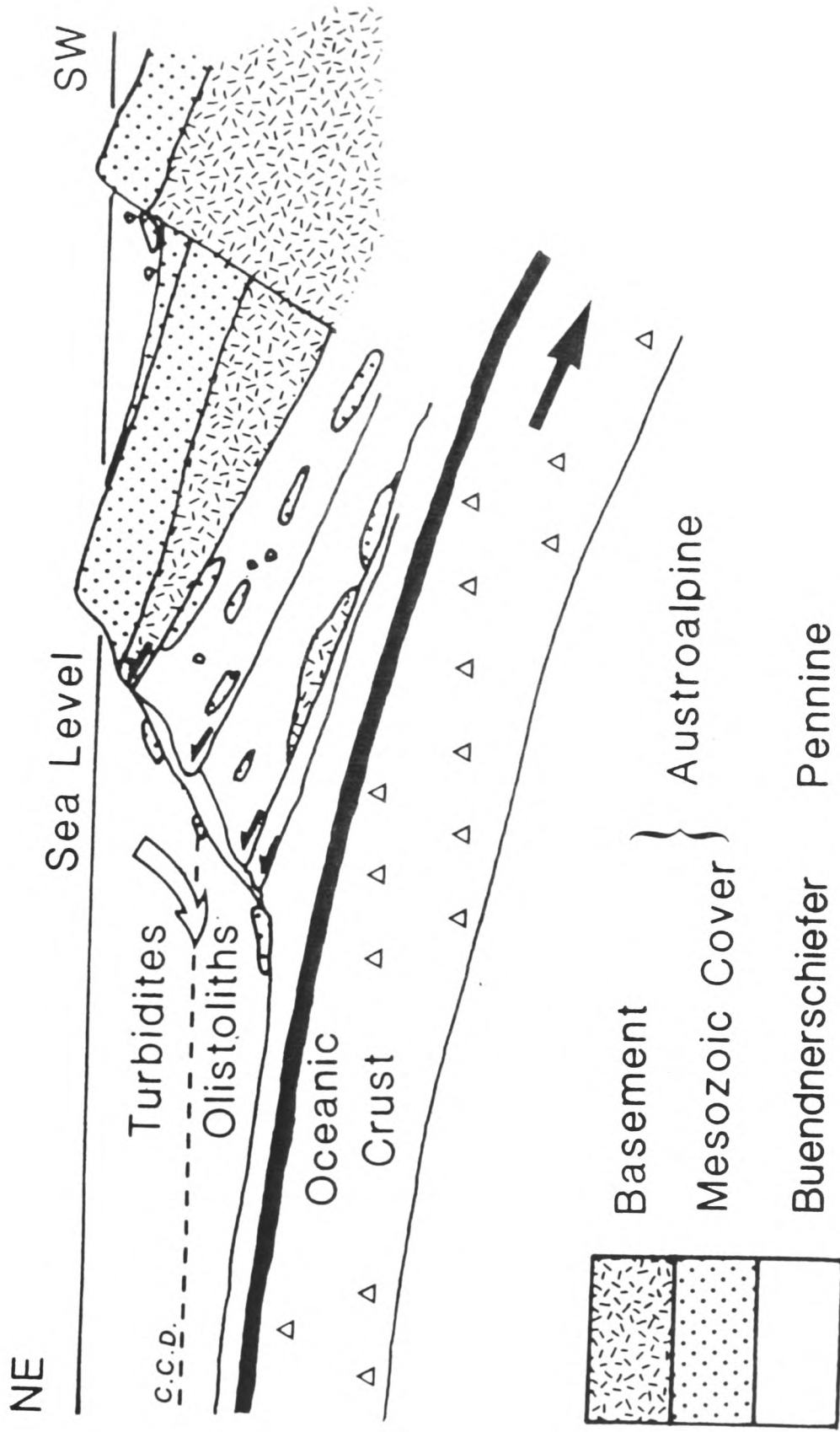


Fig.2.16. Syn-sedimentary instability in an active subduction trench. Olistoliths and breccias derived from the Austroalpine nappes are enveloped in a Bündnerschiefer matrix N.B. minor pre-existing extensional fault due to rifting. *After Frisch 1984.*

germane to this question, is that all the identified olistoliths were of Triassic age or older. No blocks of lithologies normally associated with the Jurassic were found. In conjunction with the lithological comparisons, this is taken as evidence in favour of the general Jurassic age of the sedimentary instability.

In conclusion, Jurassic syn-sedimentary instability is already well-documented from other parts of the Alps and there seems little problem in interpreting the sedimentary disruption of the Mafrei Zone in terms of Jurassic extensional tectonics. The possibility that some of the blocks did indeed slide into the subduction trench at a later stage cannot be entirely excluded, but it is at best a minor effect.

2.6 PALAEOGEOGRAPHY

The area has suffered too much deformation to allow sensible reconstruction with methods such as balanced cross sections, but parts of the original stratigraphy are still preserved and these can be combined to suggest the palaeogeography of the area prior to subduction.

Cross-sections (fig.3.40, 3.42) show that the Mafrei Zone can be divided into two distinct structural units. The upper quartz phyllite sheet is associated with thin slivers of metamorphic basement in area 3, suggesting that it was deposited on thinned continental crust. Bündnerschiefer lithologies only occur locally. This sheet is found in all three map areas and I interpret it as part of the thinned Austroalpine margin, which is therefore included in the Austroalpine tectonic units. The base of the quartzphyllite sheet is a major tectonic discontinuity which in area 3 is marked by serpentinite lenses.

The structurally lower part of the Mafrei Zone is dominated by Bündnerschiefer rock types and there is a strong similarity between

these lithologies and the Bündnerschiefer of the Pennine (Schmidt 1950-2, Exner & Prey 1964, Gommeringer 1985). The main difference is that the Matrei Zone sediments have a larger clastic component. This suggests the two depositional areas were linked, but the Matrei Zone was closer to a clastic source area. In area 1 this lower structural unit is associated with serpentinite overlain by pelagic sediments which can be interpreted as part of the Pennine ocean floor and its sedimentary cover. At deeper structural levels in the same area there is a tectonically emplaced sliver of thinned continental material overlain by Bündnerschiefer (Medel Spitze unit). The lithological associations and its structural position suggest that this fragment also formed part of the Pennine ocean floor. Similar thinned continental sequences overlain by Bündnerschiefer are present in area 3, e.g. Rote Wand Gneiss Lammela (Exner 1962). This suggests the Pennine ocean floor in this area was a heterogeneous mixture of thinned continental fragments and serpentinite. A modern analogue for this could be the Tyrrhenian Sea, which is floored by a heterogeneous mixture of thinned continental fragments and serpentinite (Leg 107 Shipboard scientific party 1986).

The strong clastic component in most of the Bündnerschiefer sediments of the Matrei Zone, shows that topographic highs supplying detritus into the area were not too far away. In the Medel Spitze area the dolomite breccias are in close proximity to massive Triassic dolomite and may be locally derived. Elsewhere there is no obvious source area for the clastic material and the presence of large olistoliths, up to 100m, suggests there may have been a substantial topographic high, which supplied detritus into the area and has subsequently been subducted or eroded away.

CHAPTER 3 : STRUCTURAL AND METAMORPHIC EVOLUTION OF THE
SOUTH EASTERN MARGIN OF THE TAUERN WINDOW.

3.1 INTRODUCTION

Along the southern margin of the Tauern Window, the Pennine units and the Mafrei Zone are tectonically overlain by Austroalpine basement, known as the Altkristallin. This tectonic boundary divides the area into two distinct domains which were both strongly deformed before their juxtaposition.

This chapter sets out to establish as rigorously as possible the structural and metamorphic evolution of the Mafrei Zone and its neighbouring Altkristallin. Radiometric dating is used to obtain an estimate of the age for some of these events.

In the Mafrei Zone the early deformation was dominantly brittle and locally thrusts that may be related to this deformation can still be identified. In contrast the Altkristallin underwent an early phase of ductile deformation, D₁. The initially distinct structural development of these two domains is emphasized by the contrasting metamorphic history. Deformation in the Altkristallin took place under increasing metamorphic conditions, which locally reached a maximum of amphibolite facies. The Mafrei Zone rocks have not undergone a regional metamorphism above lower greenschist facies and the early deformation may have been associated with high P/T conditions.

The Mafrei Zone and the Altkristallin were juxtaposed around 70-80Ma by a phase of thrusting which produced an imbrication of the two units and formed the present boundary. The whole structural pile was subsequently deformed by a phase of ductile deformation, D₂, which affects the basal few kilometres of the Altkristallin, the whole of the Mafrei Zone and the upper parts of the Pennine domain. In the S.E.

Tauern the main deformational fabric across the Austroalpine - Pennine boundary is produced by this deformation. Later events are all relatively low-strain.

One other detailed study of this contact has been carried out in the same region by Bickle & Hawkesworth (1978). There are several points of disagreement between their work and the conclusions set out in this chapter. Bickle & Hawkesworth suggest that the Altkristallin was thrust onto the Mafrei Zone as a rigid amphibolite grade sheet. They did not recognize the evidence for major penetrative deformation, Ds, under lower greenschist conditions, which affects the basal few kilometres of the Altkristallin. The same deformational fabric can be traced into the Mafrei Zone and Pennine units, which argues against the commonly held belief that there was a great rheological contrast between the Altkristallin and the rocks of the Mafrei Zone (Bickle & Hawkesworth *ibid*, Tollmann 1977a, Kober 1955, Clar 1965).

3.2 METHODOLOGY

Within the Mafrei Zone the dominant fabric is labelled Ss, and the corresponding deformation, Ds. Ds is characterized by high-strain crystal plastic deformation under lower greenschist conditions, which is associated with a top to the NW sense of shear. The mesoscopic and microscopic features can be traced continuously from the Mafrei Zone into the overlying Altkristallin, where earlier higher-grade metamorphic fabrics are also present. Within the study area these earlier fabrics all developed under higher grade conditions than Ss. In the Mafrei Zone Ds microstructures are partially annealed by the late-stage static Tauern Metamorphism, which distinguishes Ds from later events.

The features of Ds, in particular the associated schistosity, Ss, are used as a reference to which other structures can be related.

Ductile deformation preceding Ds in the Altkristallin is termed Dr. The letter suffixes are used in preference to numbers in order to avoid some of the difficulties of nomenclature encountered when correlating different tectonic units (Platt et al 1983). In this study Dq is the oldest ductile deformation and Dv the youngest.

Three areas were mapped, which straddle the Austroalpine - Pennine boundary and the structural and metamorphic development studied in each. For the structural analysis both the Altkristallin (areas 1,2,3) and the Matrei Zone (areas 1,3) have been divided into a series of subareas. The orientation data are plotted for each of these subareas to show the changes with position in the structural pile (fig.3.1). The upper boundary of the Matrei Zone immediately underlies the Altkristallin. There is, however, no clearly defined lower boundary and this may be arbitrarily placed along a late-stage shear zone (Bickle 1973) or lithologically defined (Exner & Prey 1964).

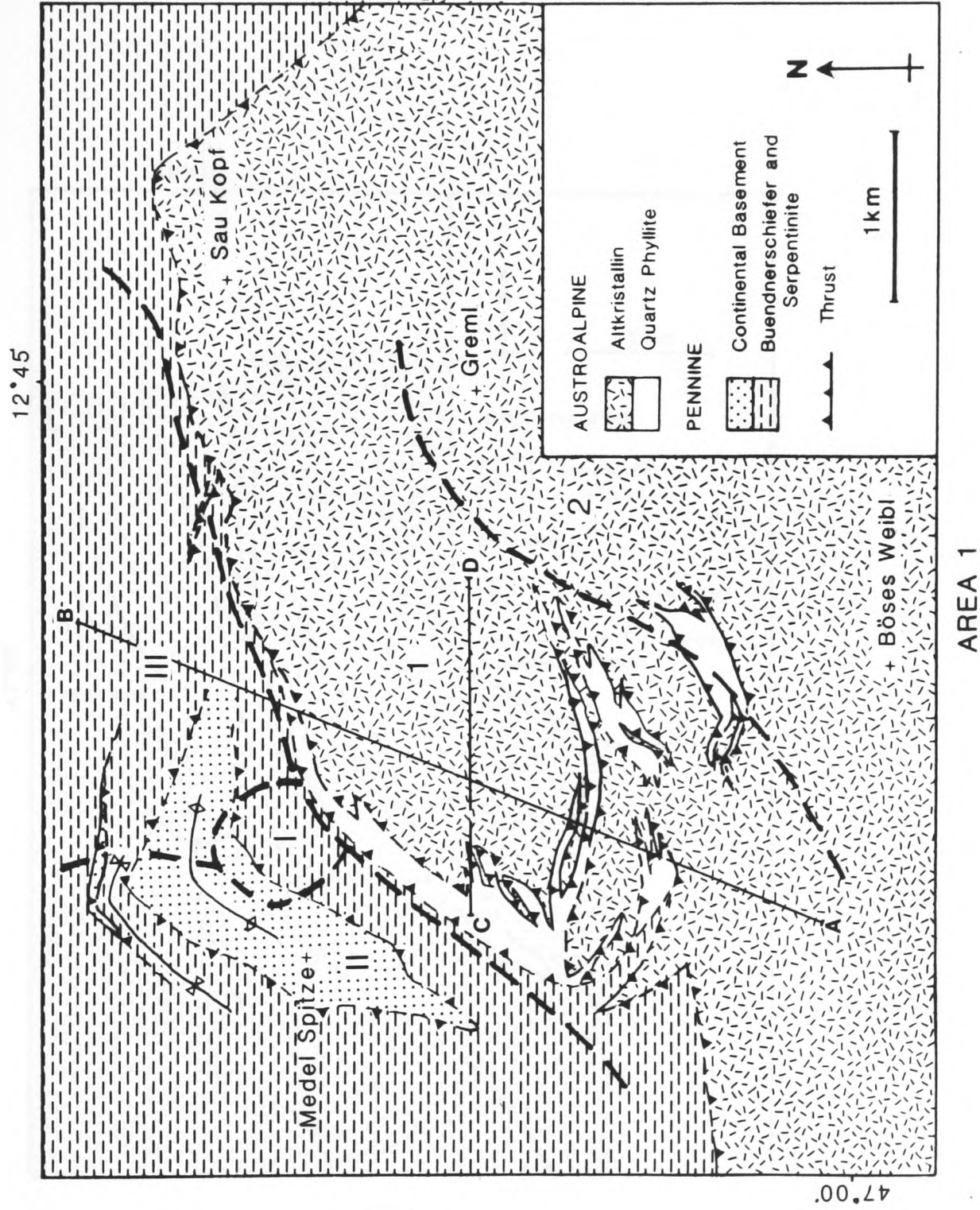
A Clar-type compass (Clar 1954) was used to measure planar and linear features of the area with an estimated accuracy of $\pm 2^\circ$ for dip and $\pm 1^\circ$ for dip-direction.

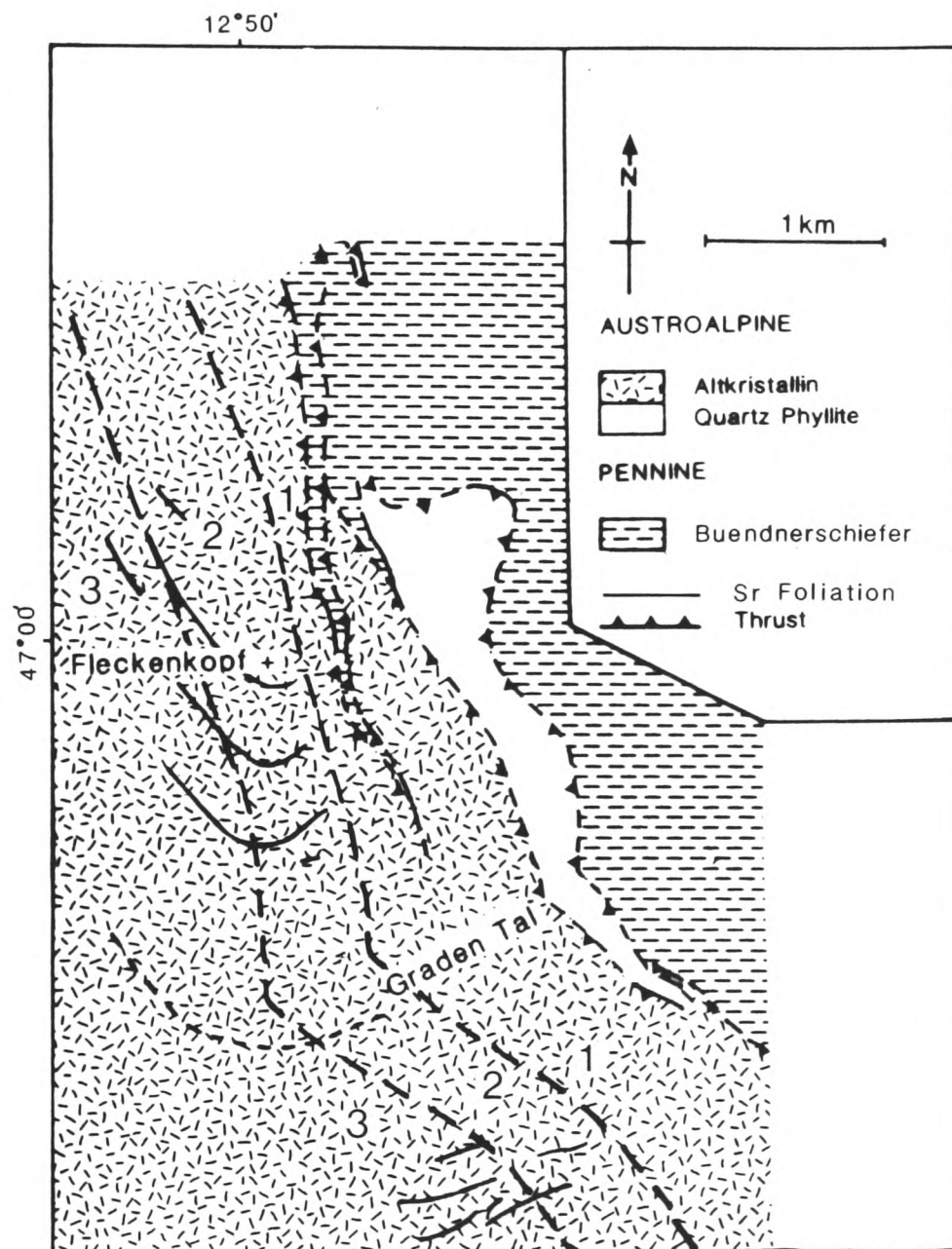
METAMORPHIC AND STRUCTURAL EVOLUTION OF THE ALTKRISTALLIN.

3.3. ROCK TYPES OF THE ALTKRISTALLIN

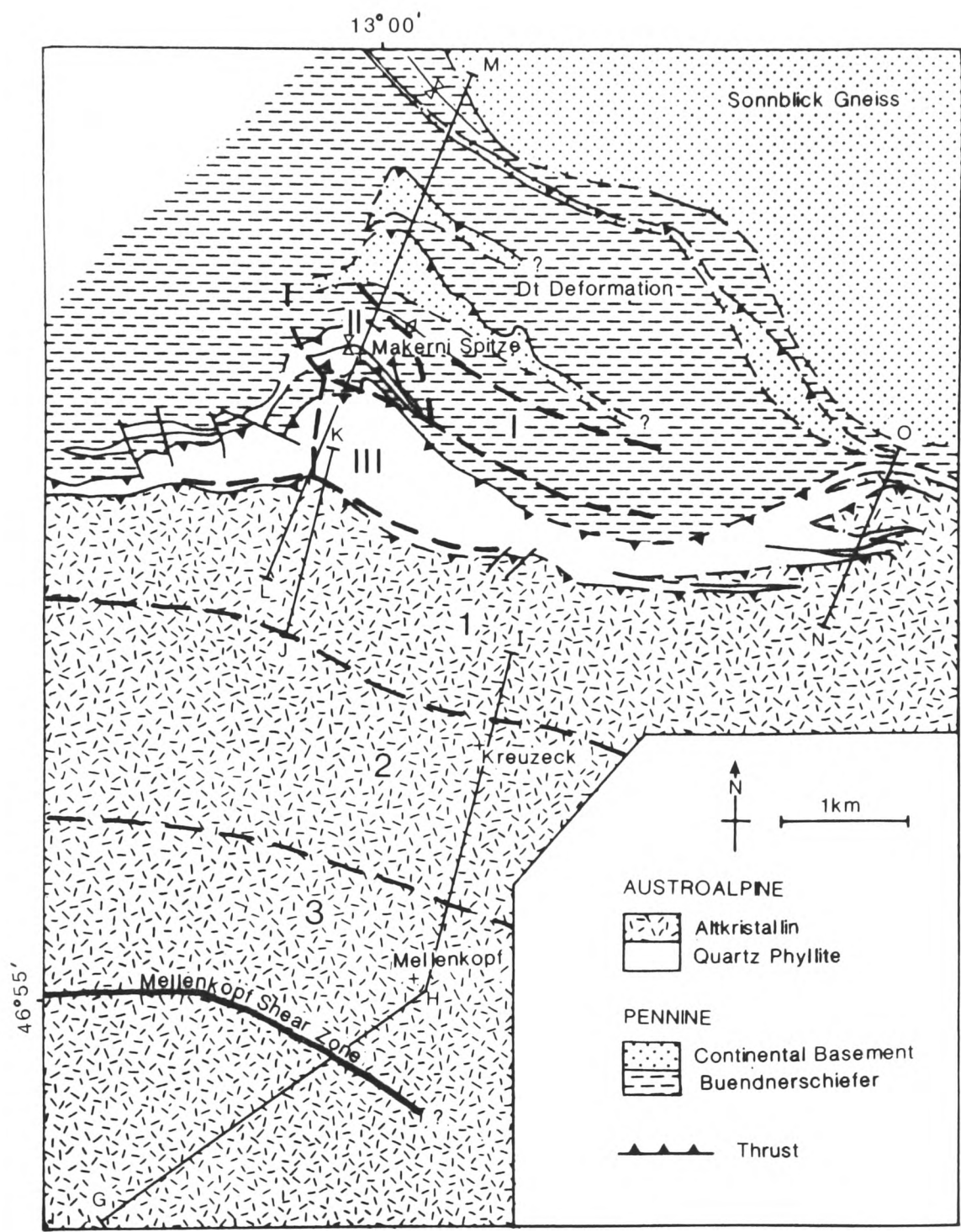
The major part of the Altkristallin in the mapped areas consists of grey-green mica schists with porphyroblasts of albite and chloritized garnet. In places the mica schist grades into more quartz-rich horizons, which have been distinguished on the maps (see enclosures) where this aided mapping of structures within the basement. Some horizons within the mica schists contain a high proportion of feldspar porphyroblasts and were mapped as augengneiss (e.g. Zeneberg in area 3).

Fig.3.1a, b, c. Simplified geological maps of map areas 1,2 and 3, more detailed maps are included as enclosures. For the location of these map areas see fig.2.1, fig.3.16. In all three areas the Altkristallin has been divided into three subareas 1,2 and 3. The Matri Zone is also divided into subareas I,II and III, for maps 1 and 3. The section lines refer to figs.3.40, fig.3.42 and fig.4.29.





AREA 2



AREA 3

The augengneiss has a greenish colour with a moderately well-developed foliation, and is distinct from the white orthogneiss which is also preserved in the area. This orthogneiss consists almost entirely of quartz and feldspar surrounded by films of white mica. The orthogneiss is commonly associated with amphibolite, and some occurrences are garnet-bearing (SW11, area 2). Calcareous rocks are very rare in the Altkristallin and only two outcrops were found. A small 50cm pod of marble lies along the ridge in area 1 just to the WSW of Hahnl Berg (see enclosure 1). In area 2 a marble layer is present which has a complicated structure. Due to the inaccessibility of the outcrop, this structure could not be investigated properly.

To the south of area 3 there is a noticeable increase in the grain-size from the zone of Ds deformation to where Mr amphibolite grade mineral assemblages are preserved (see later). This change probably corresponds to the boundary between mica schist to the north and the paragneiss to the south as recorded on the 1:1 000 000 geological map of Austria, issued by the Geologische Bundes-Anstalt (1964).

3.4. MAIN PHASE DEFORMATION (Ds)

3.4.1 Mesoscopic features

Ds forms the dominant fabric of the study area and is associated with a series of L-S tectonites. The schistosity, Ss, can be traced continuously from the Altkristallin into the underlying Mesozoic sediments of the Mafrei Zone, which clearly demonstrates its Alpine age. Ds defines a broad zone of distributed deformation within the base of the Altkristallin, which is ≈ 3 km wide in map areas 1 and 3 and 500m in area 2. In all three areas there is a similar decrease in the intensity of this fabric to the south, indicating a corresponding decrease in Ds strain. This strain gradient is also reflected in the increasing

opening angles of folds (see next section) and the decrease in intensity of quartz preferred orientation patterns (see chapter 4), both of which are qualitative strain gauges.

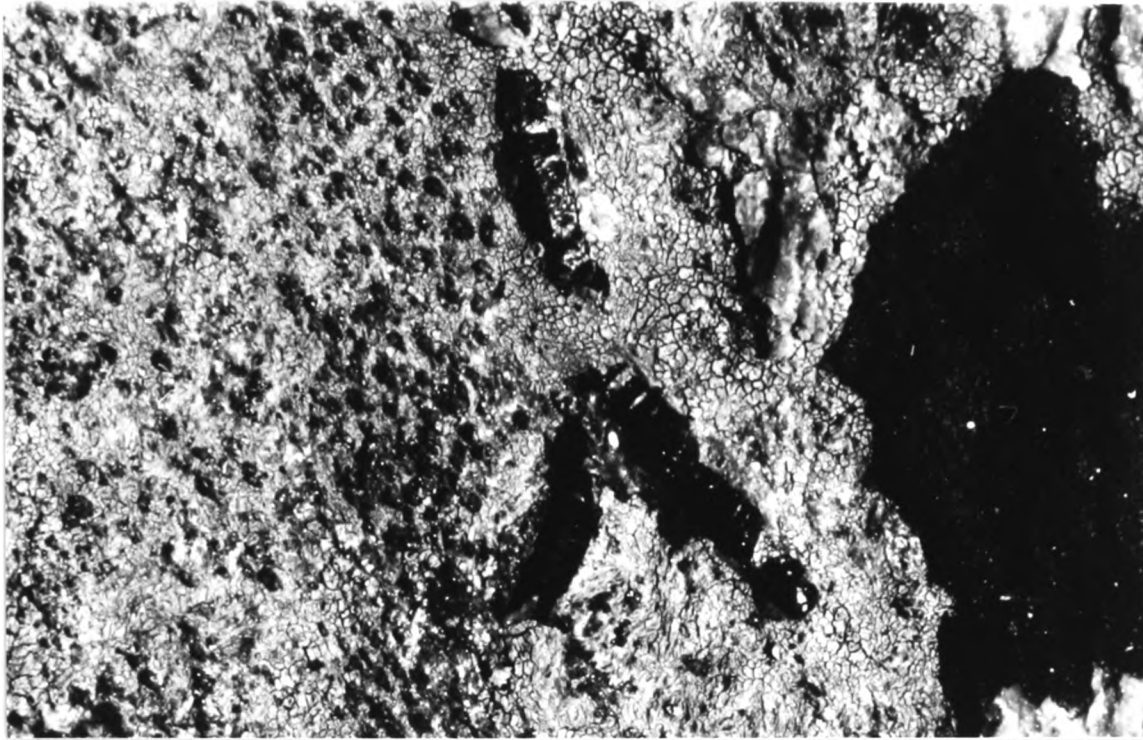
In the following, Ds is first characterized in the high strain zone, near the base of the Altkristallin and then the progressive changes with decreasing strain to the ^{South} are analysed.

a) High-Strain Ds deformation

Within the basal 100m or so of the Altkristallin, Ds tectonites are characterized by a strong platy or schistose foliation Ss. The associated lineation, Ls, is predominantly defined by the preferred alignment of quartz grains in deformed vein quartz material (recorded as a quartz rodding by many earlier workers e.g. Bickle 1973, Hawkesworth 1974). This lineation is commonly sub-parallel to elongate pressure shadows around garnet, mica mineral lineations and locally an extension direction defined by the pull-apart fracturing of minerals such as tourmaline (fig.3.2.). All these linear features were mapped as a stretching lineation, Ls, which reflects the orientation of the maximum finite extension direction, X, of the Ds strain ellipsoid. This has a consistent NW-SE orientation which generally plunges gently to the SE (fig.3.3.).

The L-S fabric is most strongly developed in area 1 and less so in the other two areas. The dominance of S-L tectonites in area 1 is a qualitative indication of plane strain conditions and the dominance of S>L tectonites in area 3 might imply a flattening strain.

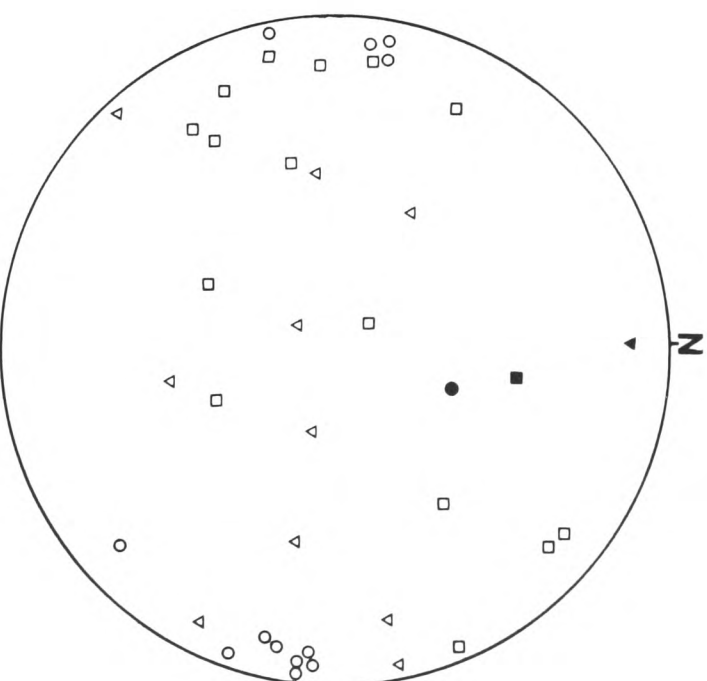
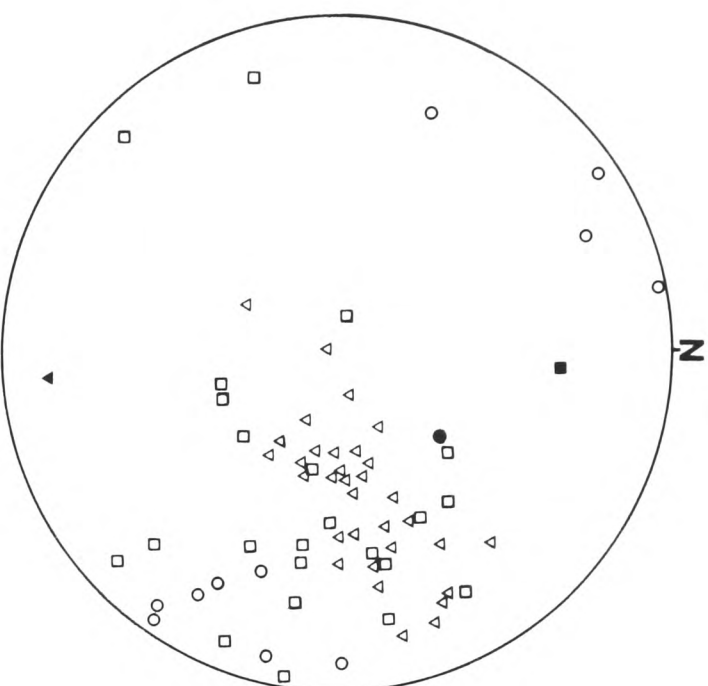
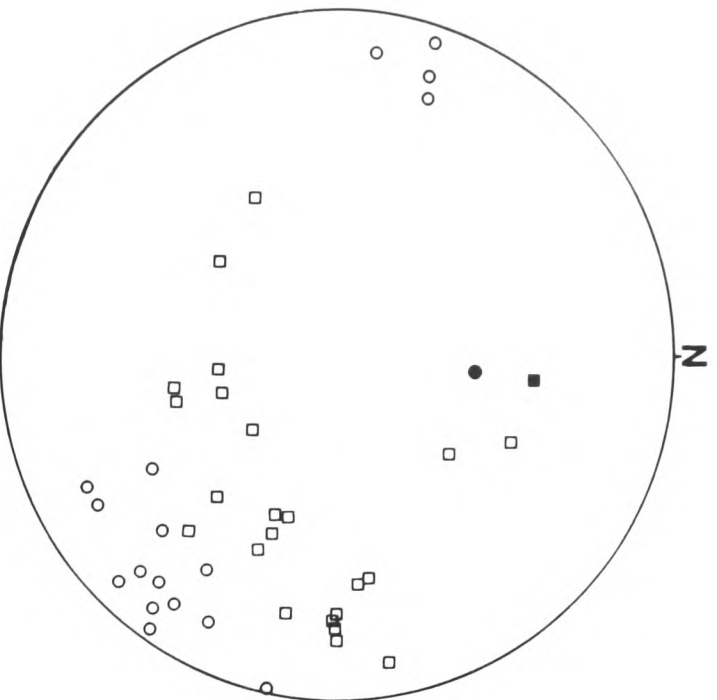
Within the high-strain zone Ss is commonly axial planar to rootless isoclinal Ds folds. These folds are ill-defined and have no identifiable vergence. The intense penetrative nature of the Ds fabric and its association with isoclinal folds, imply a high finite Ds strain. Where they can be measured, the Ds fold axes are consistently sub-parallel to the mesoscopic stretching lineation, Ls (fig.3.3.),



3cm

Fig.3.2. Ds stretching defined by the pull-apart fracturing of tourmaline crystals and the elongation of partially chloritized garnets. Kasteneck, area 1.

Fig.3.3. Mesoscopic Ds structural data from the Altkristallin, plotted for subareas 1,2 and 3. The open symbols represent fold axes and the solid symbols poles to averaged Ss in the different subareas. Circles from subarea 1, squares from subarea 2 and triangles from subarea 3 corresponding to decreasing Ds strain. The number of readings for determining the average schistosity is:
map 1: subarea 1 N = 100, subarea 2 N = 35
map 2: subarea 1 N = 70, subarea 3 N = 20, subarea 3 N = 20
map 3: subarea 1 N = 60, aubarea 3 N = 30, subarea 3 N = 16
The numbers plotted on the second set of diagrams represent the orientation of Ds stretching lineations measured in the respective subareas.

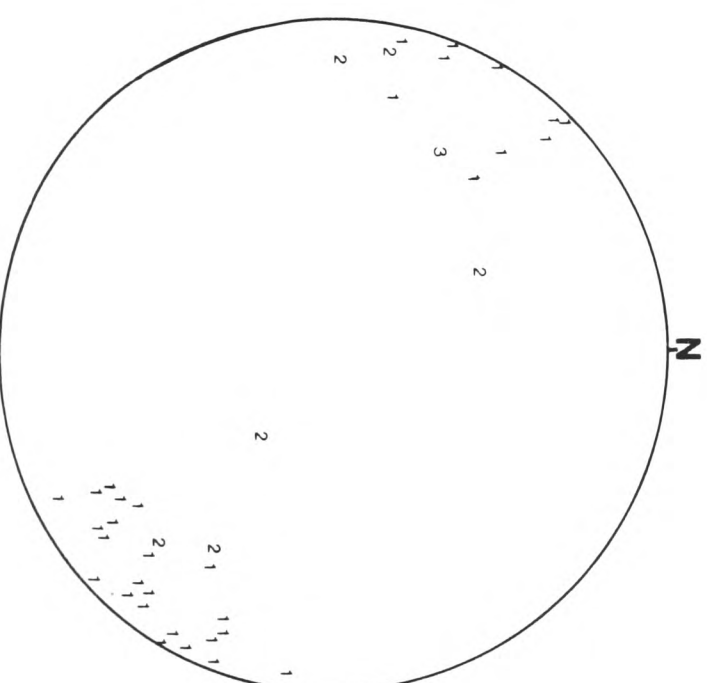
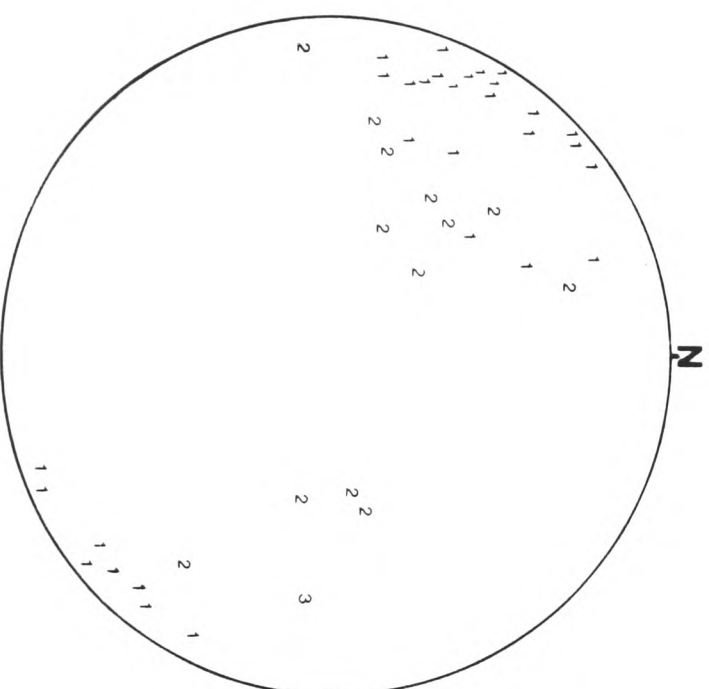
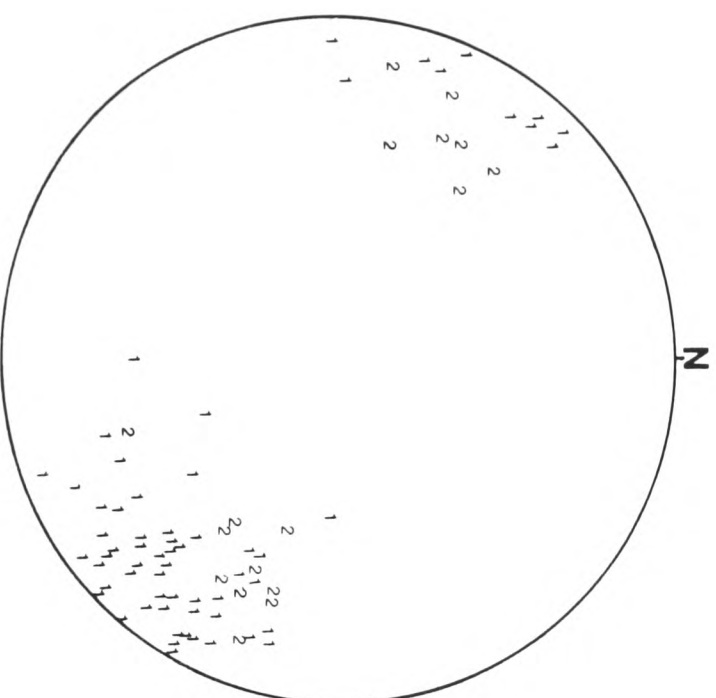


**Ds Fold Axes &
Pole to Averaged Ss**

MAP 1

MAP 2

MAP 3



**Ds Stretching
Lineation**

which may reflect their rotation into the stretching direction during deformation (see next section).

Locally Ss is cut by a series of spaced minor shear bands or extensional crenulation cleavages (eccs) (fig.3.4. - Platt & Vissers 1980). The offset associated with each individual shear is small and displacement dies away along the length of an individual zone. These shears are restricted to the high-strain micaceous lithologies and are interpreted as Ds microshears offsetting the Ss foliation, which is assumed to approximate to the orientation of the X-Y plane of the Ds finite strain ellipsoid.

b) Low-Strain Ds Features and Progressive Changes

The Ss fabric can be traced several kilometres to the south of the high strain zone, into the higher structural levels of the Altkristallin. The changes in the mesoscopic features have been documented by dividing the mapped areas into a series of E-W trending subareas and considering the geometries developed in each (fig.3.1.)

The most obvious qualitative change is a general decrease in the intensity of the L-S fabric, although this is locally strongly dependent on lithology. The orthogneiss and quartzite are more competent than the mica schist and at some localities (e.g. B41) Ss is a weak shear band cleavage in the orthogneiss that passes into an intense fissile schistosity in the immediately adjacent but less competent layers. As well as the decrease in intensity, Ss gradually steepens from a southerly dip of about 45° near the base of the Altkristallin to almost vertical at higher structural levels (fig.3.3.). Although the dip of Ss increases to the south, its strike and the pitch of Ls on this surface remain approximately constant, i.e. where Ss dips steeply Ls also plunges more steeply than at the base (fig.3.5.). This change in dip may be a primary Ds feature related to the increase of Ds strain towards

SE

NW

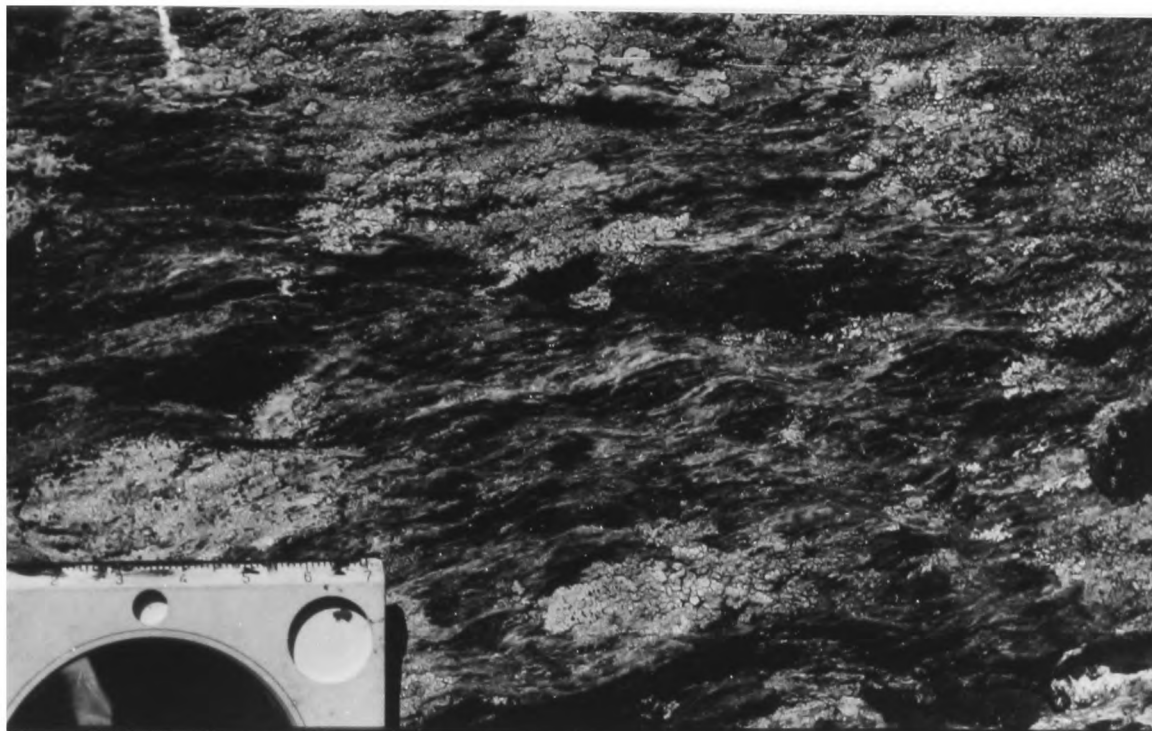


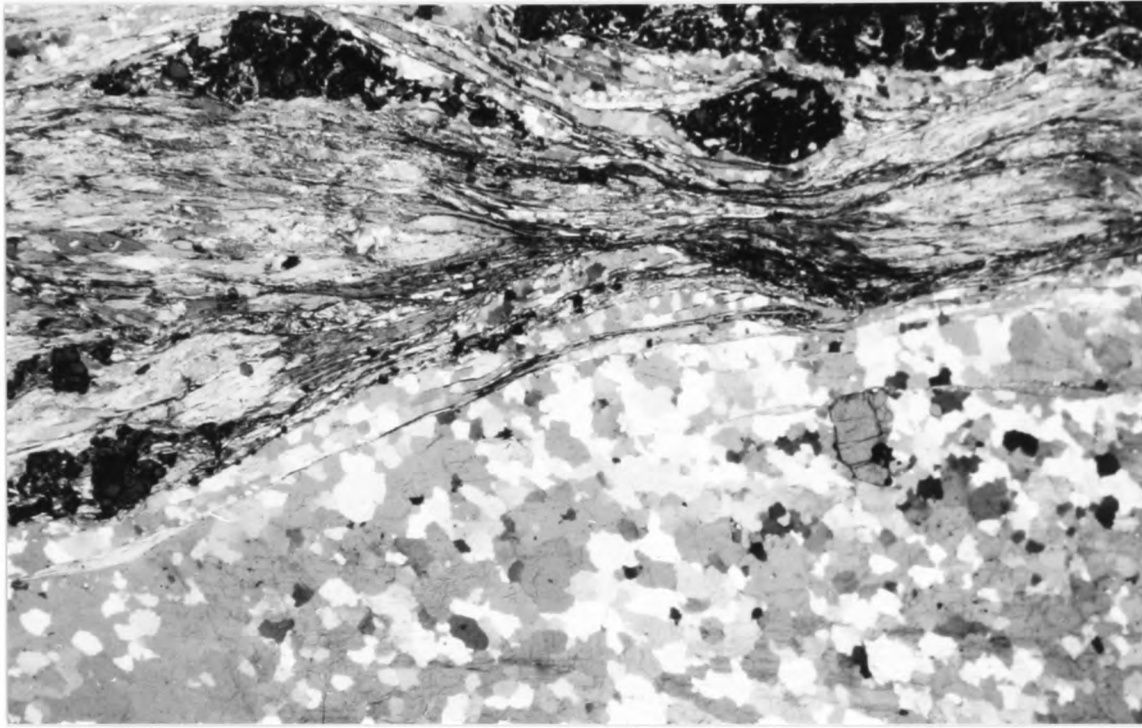
Fig.3.4a. Field shot of extensional crenulation cleavage offsetting the Ss foliation. The dominant set of crenulations indicates a top to NW sense of shear. Kasteneck, area 1, looking S.W.

NW

SE



Fig.3.4b. Sectioned Ds tectonite showing undulating aspect of Ss due to movement along eccs. The displacements are shown clearly by the offset of the layer of vein quartz. Bl29, area 1.



3mm

Fig.3.4c. Under the microscope the same vein quartz layer as in fig.3.4.b. shows no signs of internal strain associated with the formation of the eccs, suggesting that the quartz microstructure and the eccs developed during the same phase of deformation. The irregular dark patches are fractured and chloritized garnet. Cross polars with a sensitive tint plate. B129, area 1.

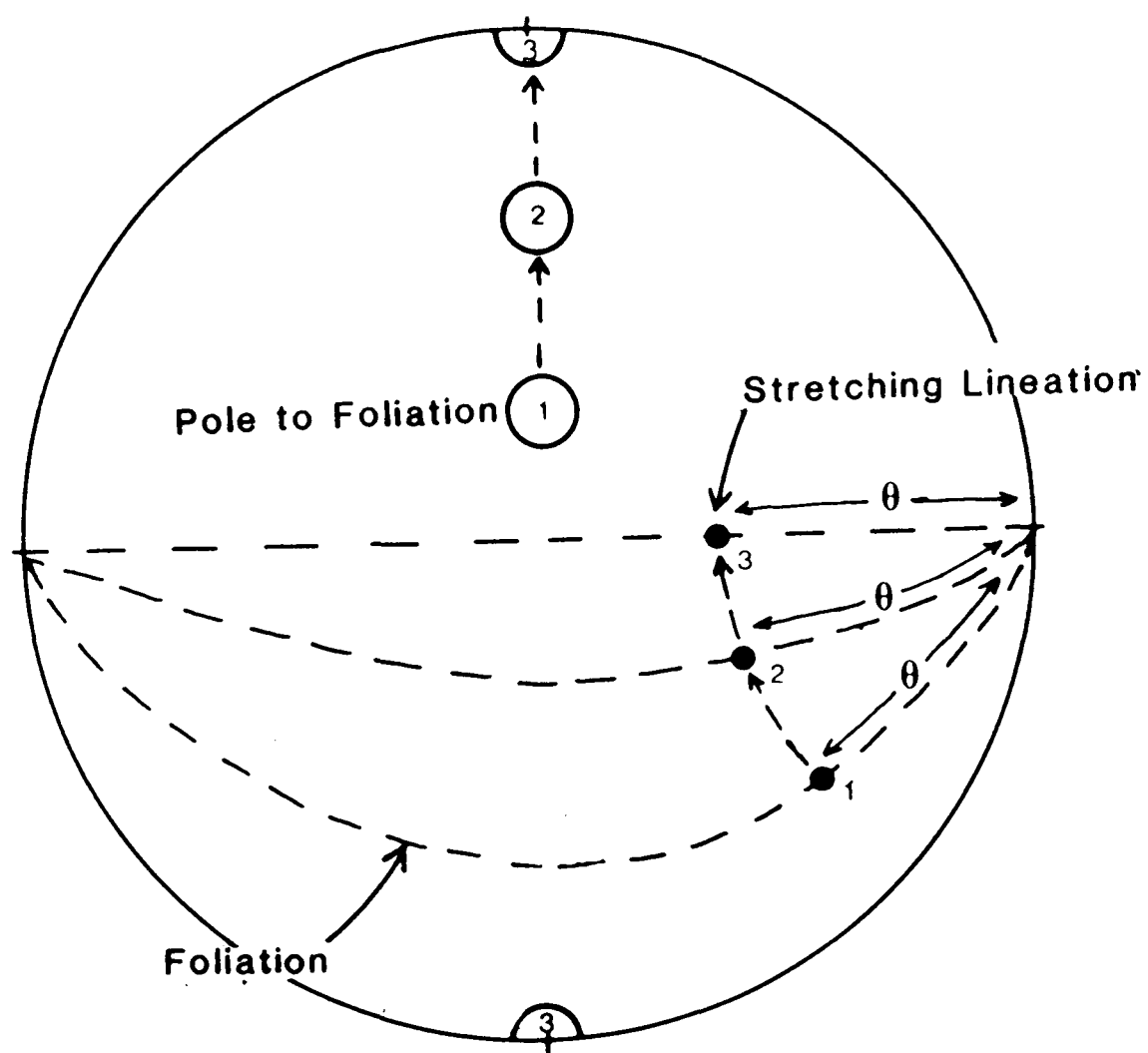


Fig.3.5. Illustration of the suggested relationship between the Ls stretching lineation and Ss foliation, plotted on a Schmidt net. The positions 1 to 3 correspond to decreasing strain in the Altkristallin. θ is the pitch of the lineation on Ss.

the base but could also be explained as the effect of the late-stage updoming of the Tauern Window (fig.3.6.).

Ds overprints a pre-existing metamorphic fabric, Sr, which is defined by the preferred orientation of minerals characteristic of upper greenschist to amphibolite conditions. Sr is completely transposed near the base of the Altkristallin due to high Ds strain. However, in the lower strain area, to the south, the earlier fabric is clearly preserved. Overprinting is either as a shear band cleavage or folding, depending on the original orientation of Sr with respect to the principal axes of the Ds strain.

In general, Ds folds have a progressively larger opening angle away from the base of the Altkristallin (fig.3.7.), although local fluctuations in strain and lithology complicate this relationship. The vergence of the Ds folds is almost exclusively to the N. The paucity of south vergent folds implies a lack of larger-scale folds within this part of the Altkristallin.

Ds fold axes have a wide range of possible orientations and a progressive change is discernible from low structural levels to high. With decreasing strain, the orientation of the Ds fold axes shows a general increase in both variability and angular separation from the stretching direction, Ls (fig.3.3). The progressive reorientation of the fold axes towards the stretching direction with increasing Ds strain strongly suggests that the Ds fold axes have been rotated away from their nucleation orientation towards the direction of maximum finite extension. The folds nucleated in an orientation at a high angle to the stretching direction. The present-day orientation of these fold axes in the low strain areas depends on both the original orientation of the pre-existing planar anisotropy, Sr, and the effects of post-Ds deformation.

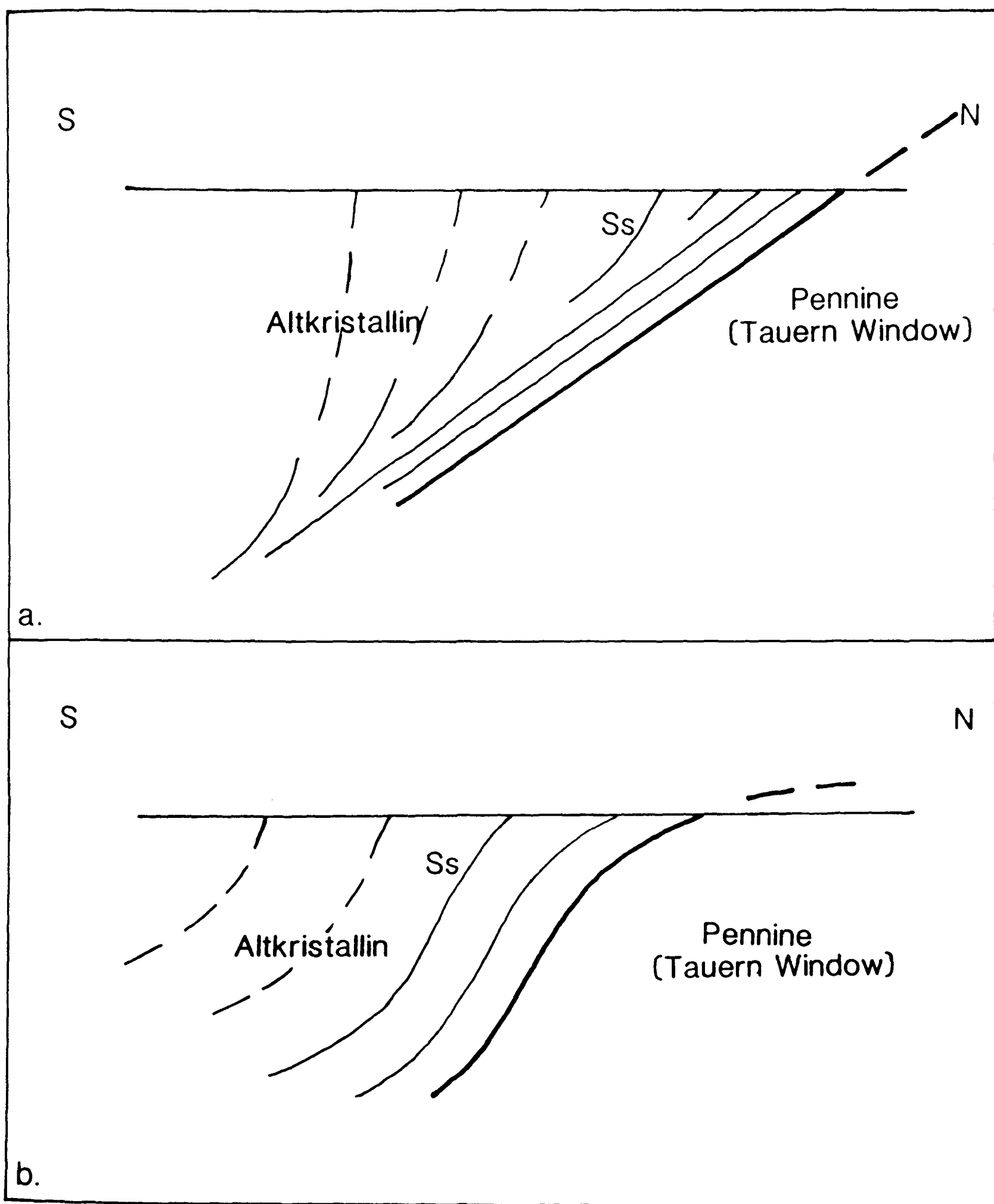
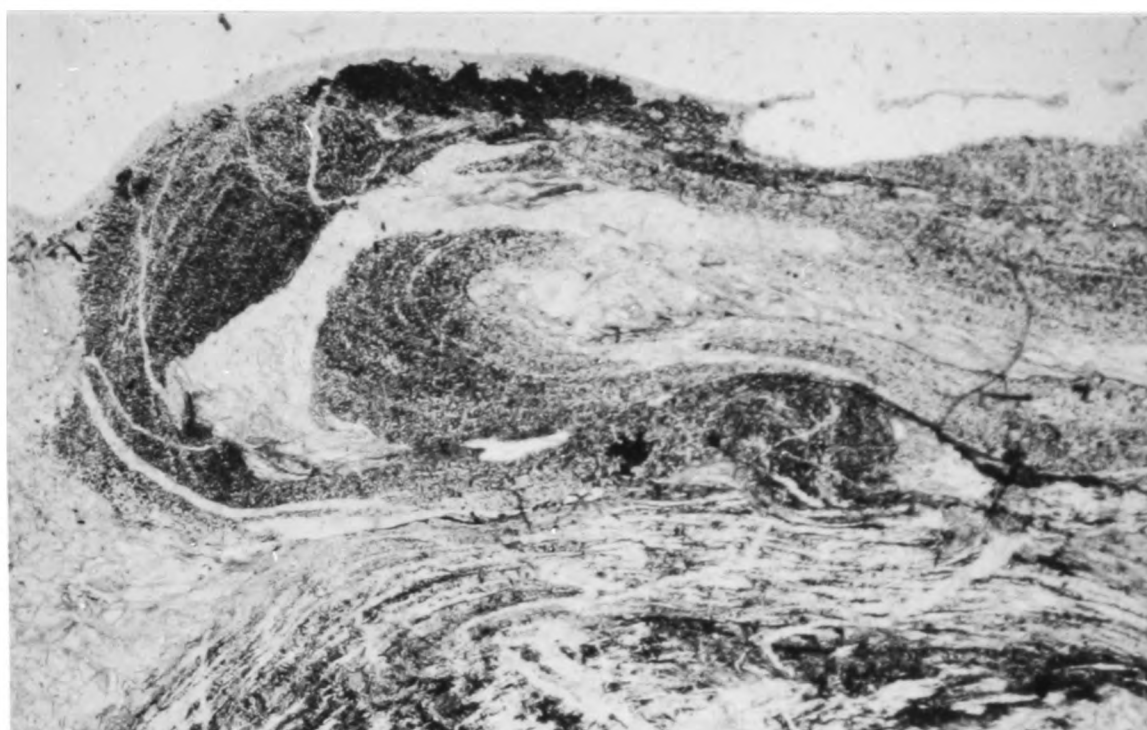


Fig.3.6. The increase in dip of Ss with increasing distance from the base of the Altkristallin can be explained as a) a primary feature related to decreasing Ds strain, or b) a late-stage feature related to the updoming of the Tauern Window.



Fig.3.7a. Tight Ds folds 3km to the south of the base of the Altkristallin. The preexisting Sr layering is defined by amphibolite and white orthogneiss layers. Looking east on Wienerhöhen Weg, leading to Böses Weibl, area 1.



4mm

Fig.3.7b. Isoclinally folded quartz-tourmaline clast from the white quartz-schists of the Matri Zone. B117, 500m to the north of the base of the Altkristallin.

Due to the lack of suitable strain markers it was only possible in one case to show that at low Ds strains the stretching direction was at a high angle to the fold axis (MS81, fig.3.8.). Direct evidence for the rotation of fold axes comes from the presence of a curvilinear Ds fold in area 2 (fig.3.9.). No sheath folds were found and the consistent N vergence of the Ds folds throughout the study area suggests in general the Ds folds rotated with the same clockwise sense.

Folds with steeply plunging fold axes are referred to as Schlingen Tektonik by many Austrian workers (e.g. Tollmann 1977a). Some consider this style of deformation characteristic of Variscan-aged deformation. However this work shows that in the three mapped areas, this style of tectonics in fact formed during Alpine deformation (see also Schmid & Haas 1987).

Summary

The basal few kilometres of the Altkristallin have been deformed by a penetrative, ductile Alpine deformation, Ds. Ds strain intensity increases towards the base of the Altkristallin and is reflected in changes of the mesoscopic structures. In particular, increasing Ds strain progressively reoriented the Ds fold axes into the stretching direction and is responsible for the stronger development of the L-S fabric near the base. The gradual increase in the dip of Ss with decreasing strain may also be a primary Ds feature related to the strain gradient.

3.4.2 Ds microstructures

Microstructurally the mesoscopic Ss foliation is defined by the preferred orientation of platy minerals (mainly chlorite and white mica), which is commonly accentuated by a grain shape fabric in the quartz (fig.3.10, fig.3.11.). Where Ss is offset by ecc formation, exactly the same mineralogy is found within the localized shear zone as

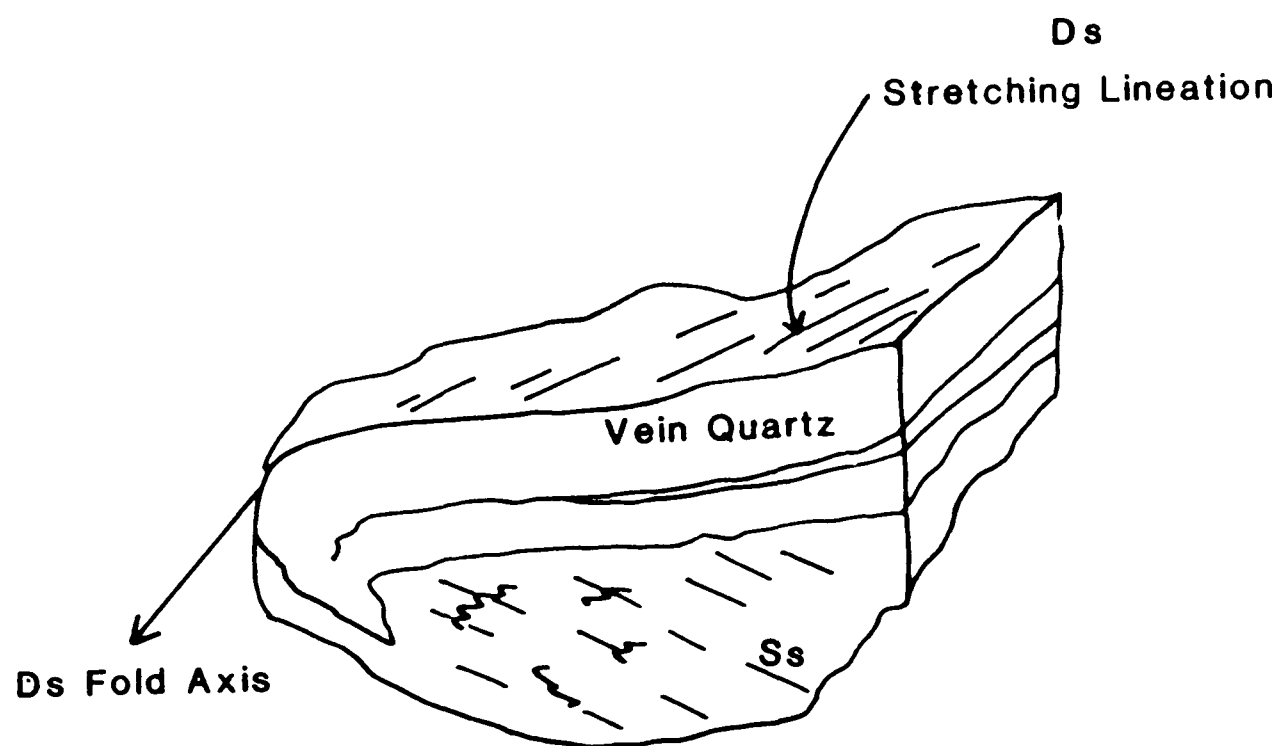
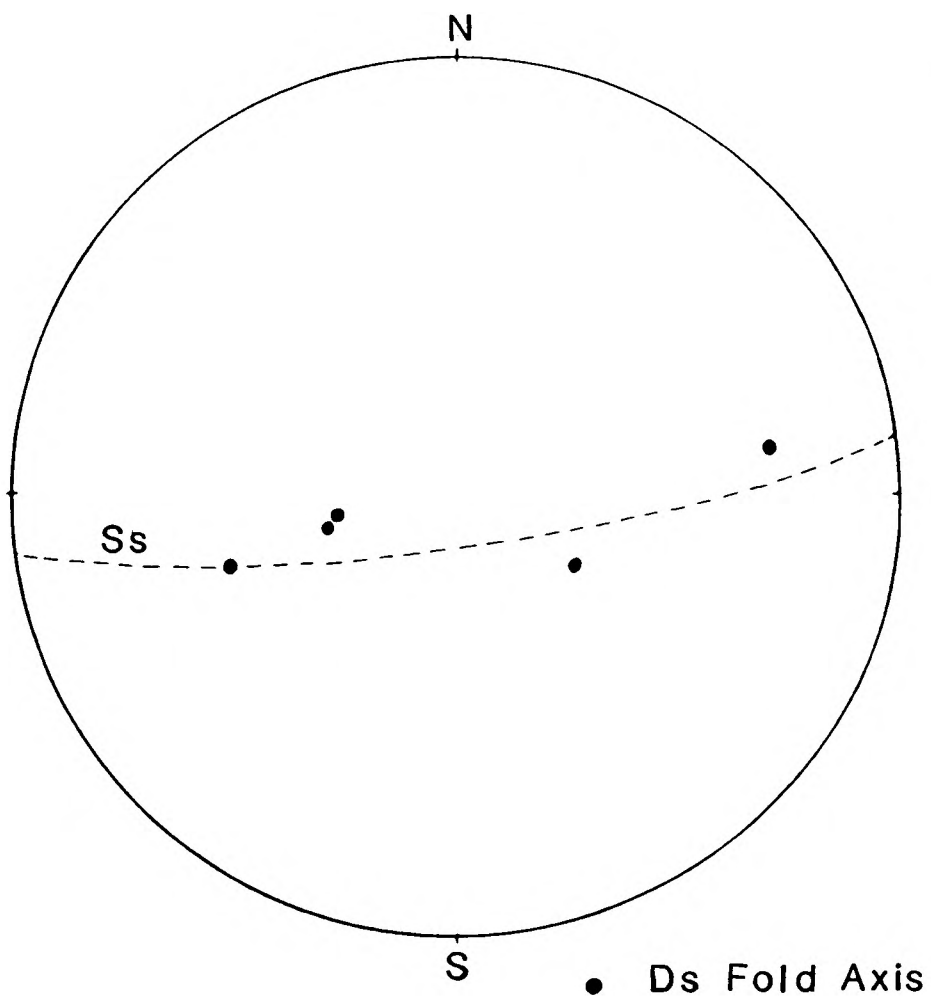


Fig.3.8. Ds fold axis and the associated stretching lineation at an angle of $\approx 45^\circ$ to one another. MS81, 2km to the south of the base to the Altkristallin, area 3.



Fig.3.9a. Curvilinear Ds fold 200m W of Fleckenkopf, 1km from the base of the Altkristallin, area 2.

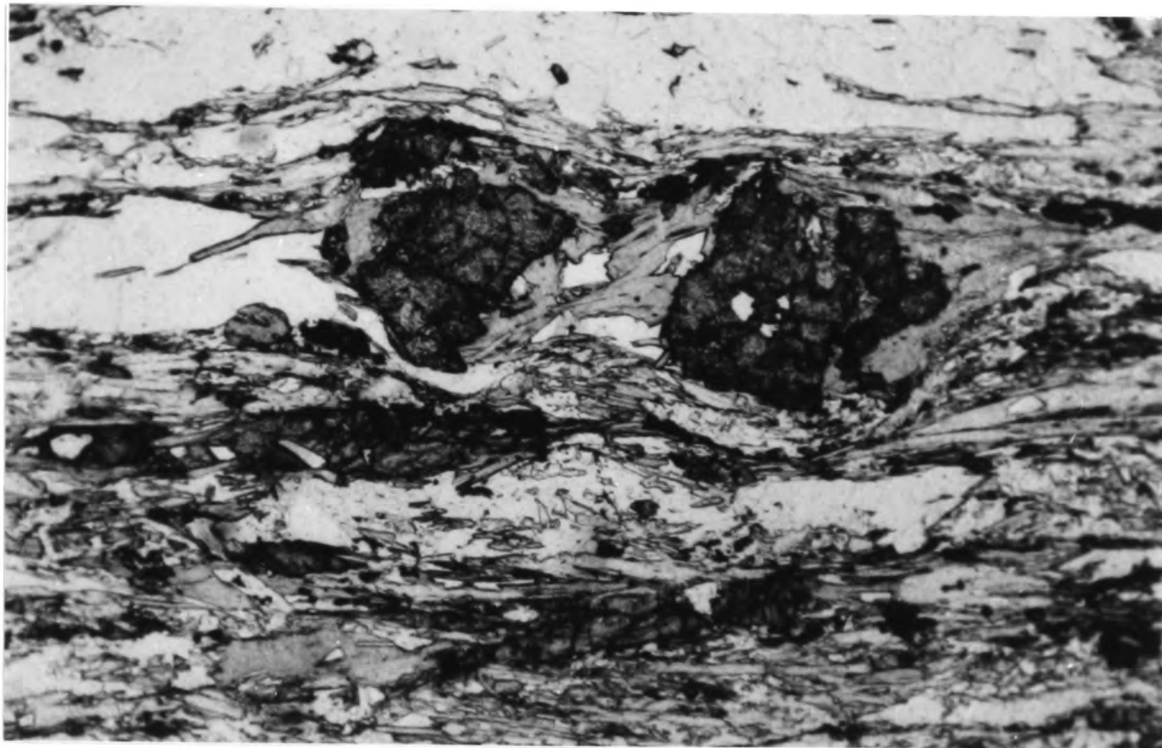
Fig.3.9b. Stereographic plot on a Schmidt net of fold axes from the neighbourhood of the photo above.



associated with the main fabric. In some cases the ecc's formed immediately adjacent to quartz-rich layers deformed by Ds. No late-stage disturbance was recorded in the quartz and there was no discernible disruption of the Ds crystallographic preferred orientation (fig.3.4.). These microstructural observations therefore support the suggestion that the ecc's are related to Ds and not due to later deformation. In thin section it is also possible to confirm that Ls is sub-parallel to the long axis of deformed quartz grains and the pressure shadows around rigid inclusions such as garnet and feldspar.

In the structurally lower, higher-strain, parts of the Altkristallin the most common stable mineral assemblage is chlorite + quartz + white mica + albite, which is characteristic of the lower greenschist facies. Relics of minerals characteristic of higher grade metamorphic conditions are also present, which represent the earlier metamorphic fabric deformed by Ds (Sr). Where Ds strain is relatively low, Sr can be directly observed, and is defined by the preferred alignment of large white micas, biotite and amphibole. Garnet and albite are also commonly associated with this fabric. There have been two reports of staurolite from within the zone affected by Ds, which may also be related to Sr (Angel 1928, Exner & Prey 1964). However, despite searching the reported localities these finds could not be confirmed.

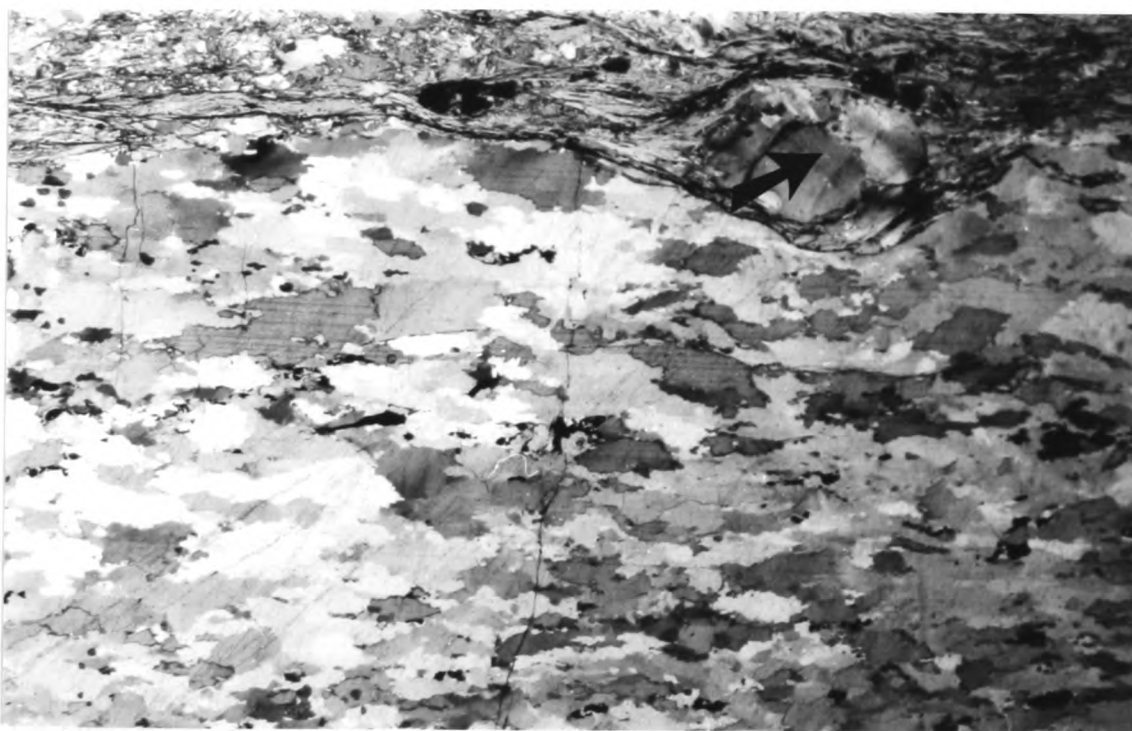
With increasing Ds strain towards the base of the Altkristallin, the metamorphic minerals associated with Sr are progressively retrogressed. The amphibole, garnet, biotite are all replaced by chlorite (fig.3.10.). Chlorite after biotite is commonly strongly pleochroic and stained with iron oxides. An approximate boundary can be defined, which separates areas to the south, where remnants of incompletely altered biotite are still present, and those to the north where biotite is completely retrogressed. The metamorphic minerals associated with Sr are clearly deformed and fractured during Ds and most



1mm

Fig.3.10. Ss is defined by the preferred orientation of platy minerals. Garnet and amphibole developed during Mr are fractured and retrogressed to chlorite by Ds. MS78, area 3.

SE



NW

↑ Up

3mm

Fig.3.11. Ss is defined by the preferred orientation of platy minerals and the grain shape fabric in the quartz. The striations within the quartz are scratches on the thin section surface. A large Mr mica (see arrow) is recrystallized to finer-grained Ds micas. The asymmetry indicates a top to NW sense of shear. B109, area 1.

of their retrogression is due to this deformation. This is contrary to the suggestion of Bickle & Hawkesworth (1978), that the retrogression was due to the static Tauern Metamorphism (see 3.8.2).

The large Mr white micas are deformed and recrystallized to give the finer-grained Ds fabric but can commonly still be recognized in strongly deformed Ds tectonites (fig.3.11.). Garnets are also relatively resilient against the effects of Ds and commonly occur as small altered fragments throughout much of the high-strain lower part of the Altkristallin. The presence of garnet and large micas, with the general absence of carbonate, is the best way of distinguishing retrogressed Altkristallin lithologies from the quartz phyllite of the Matrei Zone and Pennine units.

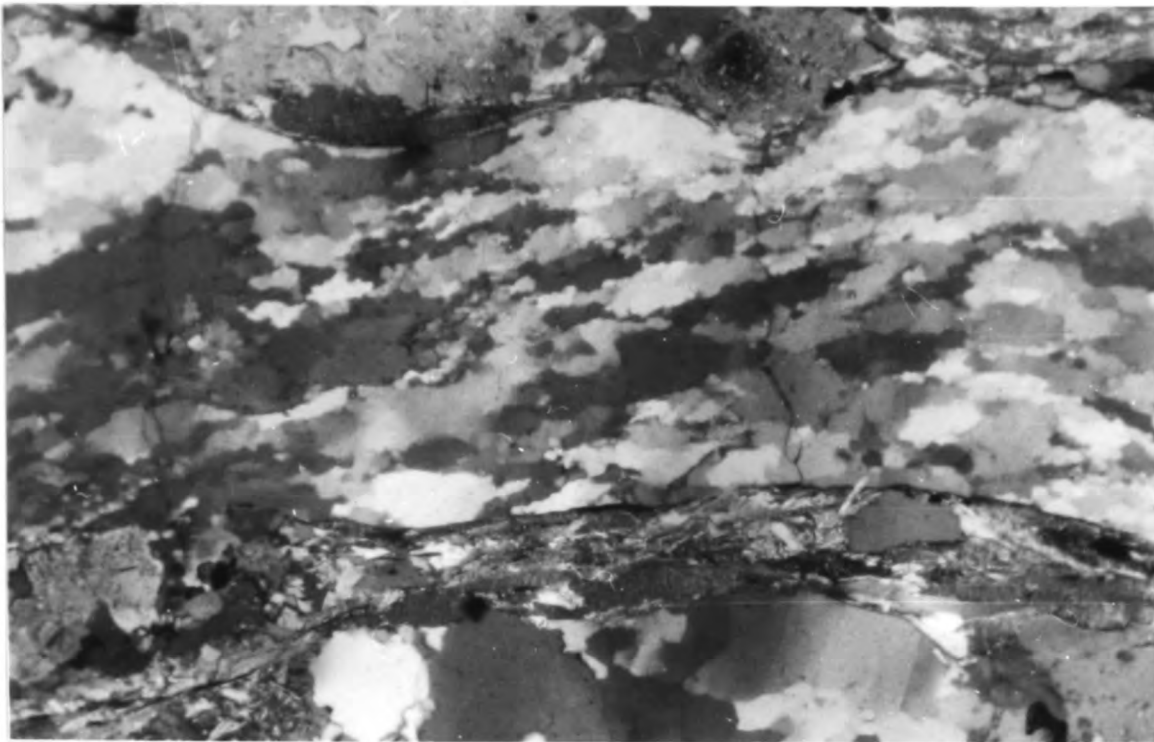
Quartz-rich Ds tectonites have strongly deformed quartz grains with sutured grain boundaries, subgrains and undulose extinction, all of which are characteristic of crystal plastic deformation (fig.3.12.). The high strain has produced a strong crystallographic preferred orientation in many of these tectonites (chapter 4). Near the base of the Altkristallin the microstructures have been partially annealed by the late-stage Tauern Metamorphism but this was not sufficient to destroy the preferred orientation in the quartz (see chapter 4).

In the more micaceous lithologies slip along the foliation is relatively easy and deformation by grain boundary sliding was probably also an important deformation mechanism. In many Ds tectonites the harder minerals such as amphibole, garnet and feldspar often show brittle deformation and retrogression to platy minerals.

The two main Ds strain-softening mechanisms that were operative were: (i) retrogression from hard minerals to weaker platy minerals (e.g. amphibole → chlorite); and (ii) an increase in crystallographic preferred orientation. There is no evidence that the higher strain

SE

NW



↑
Up

1mm

Fig.3.12. Deformed quartz showing undulose extinction and sutured grain boundaries indicative of crystal plastic processes. The mesoscopic Ss foliation is defined by the preferred orientation of platy minerals. The obliquity of the quartz grain-shape fabric suggests a top to NW sense of shear. K5, area 1, crossed nicols with a sensitive tint plate.



1mm

Fig.3.13. Pull-apart fracturing of Mr garnet with fresh interstitial biotite growing approximately parallel to the axial planes of Ds folds in the white mica layers. This suggests that the biotite grew during Ds. MS58, area 3.

parts of the sequence are associated with higher temperatures, which suggests that strain heating (Poirier 1980) was of little importance.

3.4.3. Thermal history during Ds

The progressive retrogression of biotite, amphibole and garnet from S to N in the Altkristallin can be explained as a consequence of increasing Ds strain under lower greenschist conditions. However, there is also evidence that the Ds microstructures at higher structural levels were formed at slightly higher T than those near the base.

MS58 is located several kilometres to the south of the boundary with the Pennine units, where Sr is the dominant foliation. In this sample Ds has crenulated the Sr fabric and caused extensional fracturing of the Mr garnet. Biotite has grown within the fractures, which indicates that Ds took place under conditions where biotite was stable but not garnet (fig.3.13.). Several other less clear examples also show Mr garnets associated with biotite parallel to Ss, although in these cases the biotite has been partially retrogressed to chlorite.

There is further evidence of a temperature gradient from the microstructures of albite porphyroblasts. At high structural levels albite locally overgrows the cores of Ds folds indicating that their growth is post- to syn-Ds (fig.3.14.). Near the base of the Altkristallin the albite porphyroblasts commonly have an internal fabric defined by the grain shape of quartz and mica. In many cases this fabric is clearly oblique to the external fabric, Ss, which suggests that the growth of albite in this part of the Altkristallin was pre-kinematic with respect to Ds. However, in one sample, K5, near the base of the Altkristallin, there is evidence that at least some of these albites also grew syn-kinematically with Ss. A 1cm porphyroblast contains a crenulated mica fabric defined by large white micas, which is cross-cut by smaller grains approximately parallel to Ss. At the

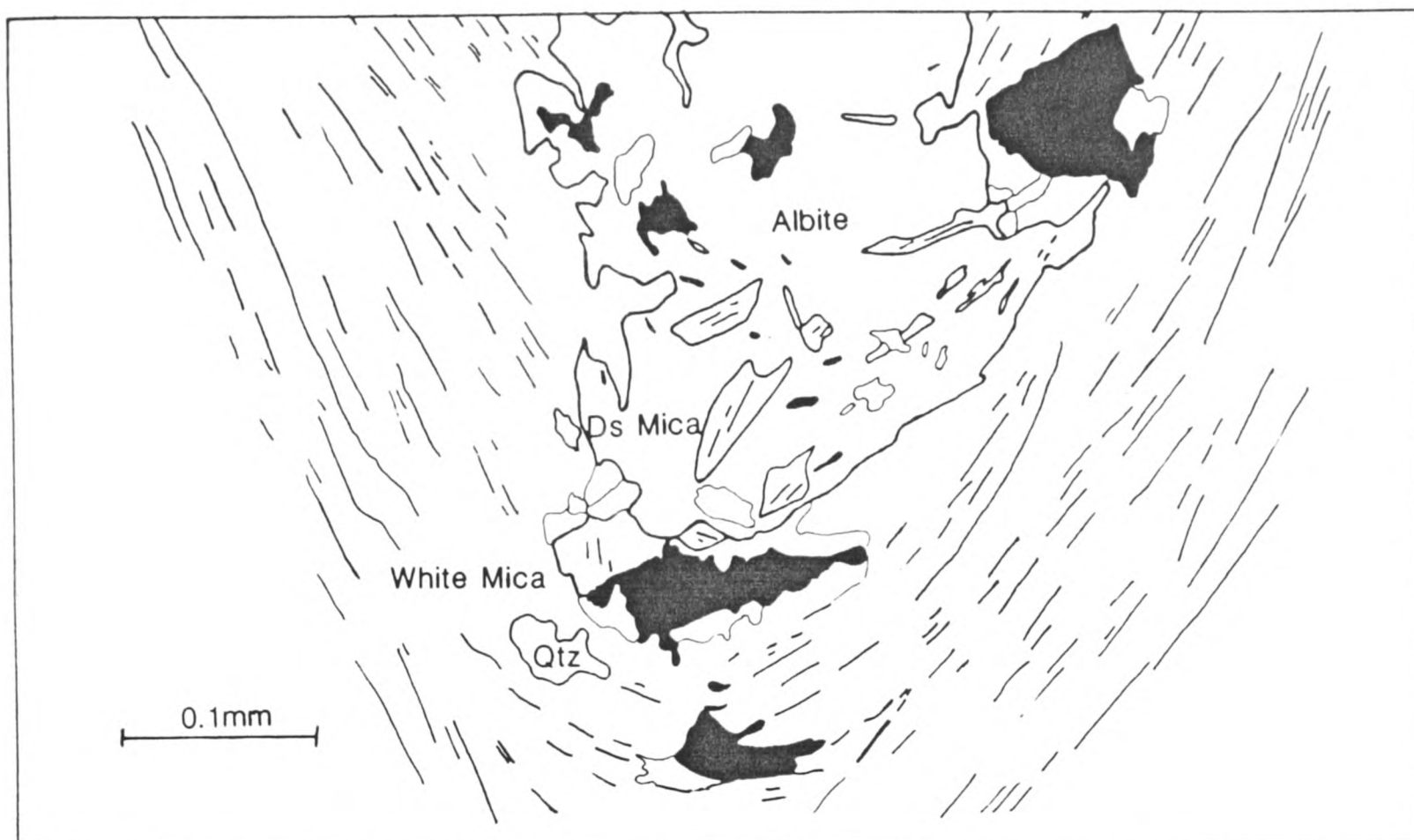


Fig.3.14. Postkinematic albite overgrowing the core of a Ds fold at high structural levels in the Altkristallin. MS57, area 3, taken from a colour photomicrograph.

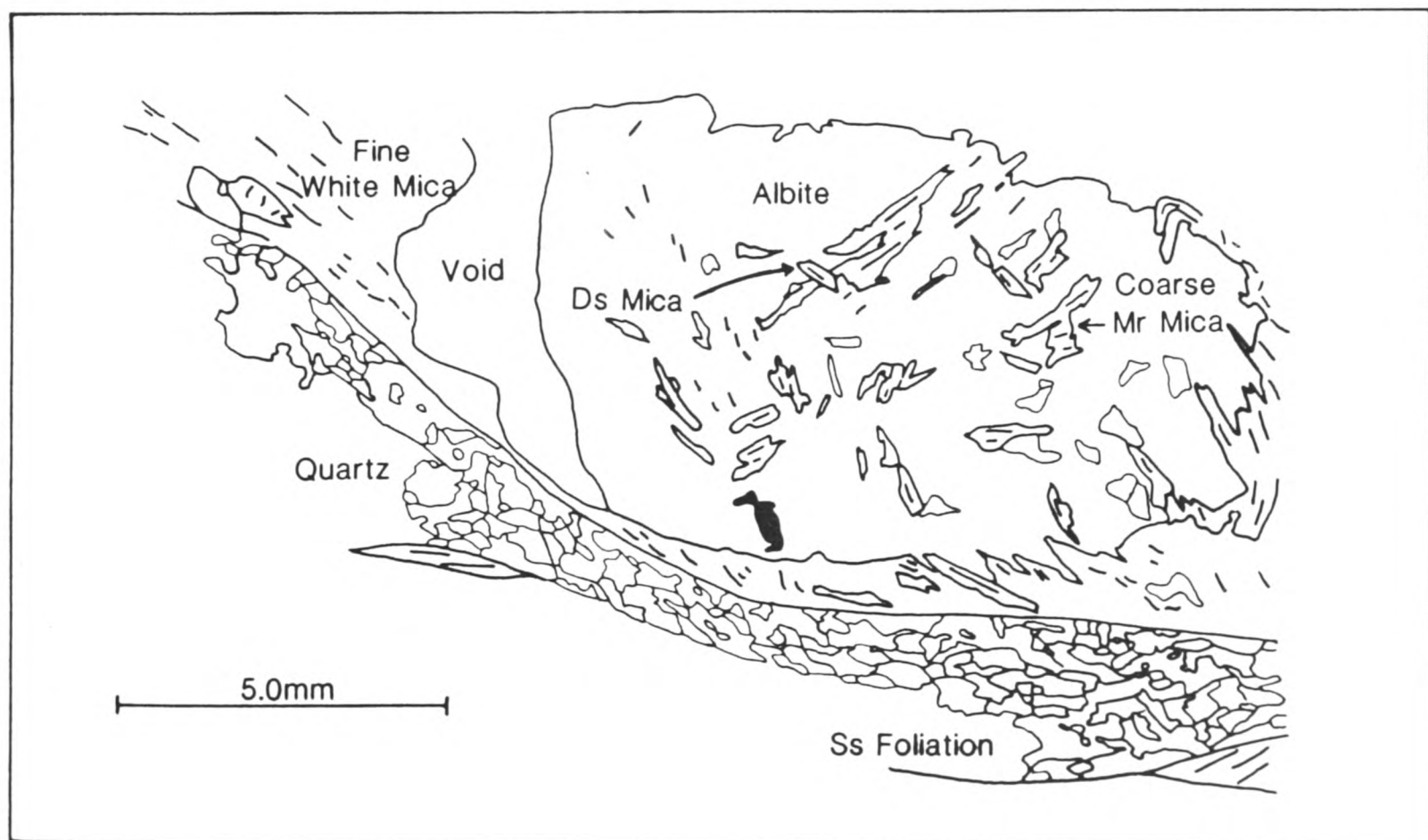


Fig.3.15. Large albite prophyroblast containing an Mr mica fabric which defines an open Ds crenulation. Outside the porphyroblast the Mr micas have been recrystallized to a fine-grained matrix. This suggest the albite grew at an early stage during Ds. K5, area 1, drawn from a photomicrograph.

margins of the porphyroblast, individual large micas can be traced into the finer-grained surrounding Ss fabric (fig.3.15.). These features suggest that the albite completely overgrew a Ds crenulation as it was developing, and that at this structural level Ds deformation continued after the albite had ceased to grow.

On the basis of this rather sparse microstructural evidence, Ds can be tentatively divided into an early slightly higher T stage when biotite and albite were stable and growing; and a later lower T stage when deformation was concentrated in the lower parts of the Altkristallin. The early microstructures are preserved at relatively high structural levels and are progressively overprinted under lower greenschist conditions at deeper levels, i.e. Ds took place under falling T conditions. Extensional crenulations and, in area 3, quartz fabrics also suggest a concentration of Ds in the lower structural levels at a late stage in the deformation.

An important consequence of this division of Ds into late and early microstructures, is that the kinematics of Ds deformation can be studied at different times. The orientation of the Ds stretching lineation remains roughly constant throughout the Altkristallin, which implies the extension direction did not significantly change from one stage of Ds to the next.

Summary

The Ds microstructures show that Ds overprints a higher-grade metamorphic fabric, Sr, and the metamorphic minerals associated with Sr are progressively fractured and retrogressed during Ds. Large Mr white micas and fragments of garnet are the most persistent features. The greater retrogression of these minerals near the base of the Altkristallin is partly due to a corresponding increase in Ds strain but may also be related to initially higher T of deformation. The

higher-temperature Ds microstructures are preserved at higher structural levels.

3.4.4 Ds shear zones

Within area 3 a shear zone, called the Mellenkopf Shear Zone (MSZ), separates the main volume of Ds deformation from the largely unretrogressed amphibolite grade metamorphic rocks to the south (fig.3.16.). The MSZ is not well exposed but is marked by patches of highly sheared and retrogressed rock. By following a prominent topographic depression and finding small patches of sheared rock on the northern side of the zone, the MSZ can be traced to the NW out of the map area 3 and into the main zone of Ds deformation. The MSZ is steeply dipping with an associated NW-SE stretching lineation. The tectonites within this shear zone show deformation under lower greenschist facies conditions, with biotite and garnet being retrogressed to chlorite. The microstructure and orientation of the mesoscopic structures suggests that this shear zone belongs to the Ds phase of deformation.

To the south of the MSZ there are a number of millimetre-scale shear zones and an occurrence of pseudotachylite which may also be related to movements along this zone. The displacement associated with the pseudotachylite formation is top to the NW.

To the east of area 3, Waters (1976) and Hoke (1987) have mapped several similar shear zones within the Kreuzeck Group (fig.3.16.). Locally the boundaries of these shear zones are strongly overprinted by cataclastic deformation. In some areas, however, the contacts are still welded to the surrounding rock, demonstrating the continuity of deformation.

Within the amphibolite grade metamorphics of the Kreuzeck Group Hoke (ibid) maps a zone where two of these discrete shear zones merge with a zone of distributed D₃ deformation, which is also associated with

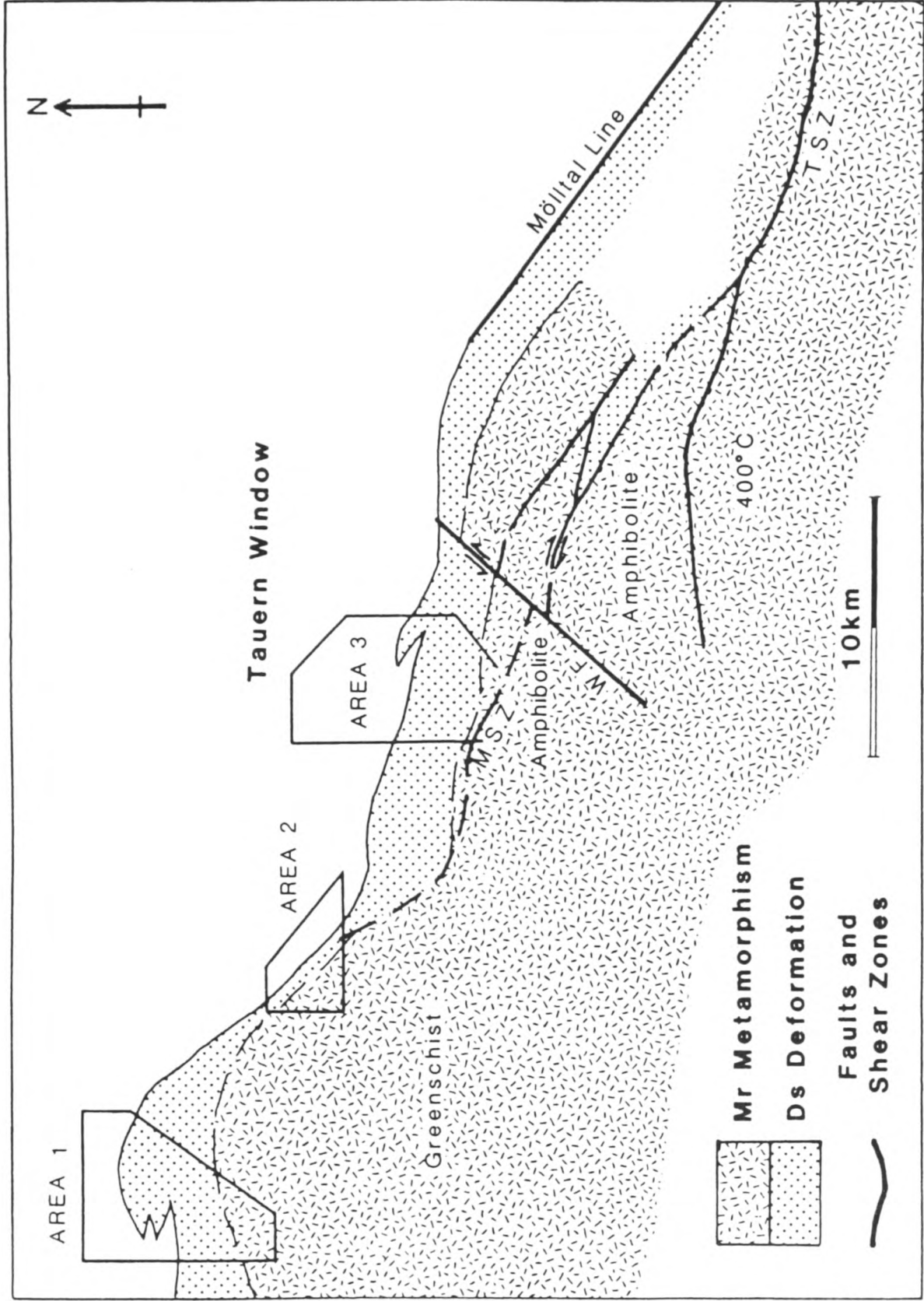


Fig.3.16. Cretaceous metamorphism in the Altkristallin to the S.E of the Tauern Window, and the position of major Ds shear zones. The mapping to the east of the Wöllratten Fault is after Waters (1976) and Hoke (1987). The outlined areas represent the approximate positions of map areas 1,2, and 3. WF = Wöllratten Fault, TSZ = Teuchl Shear Zone. MSZ = Mellenkopf Shear Zone. The metamorphic grade refers to Mr.

strong NW-SE trending stretching lineations and suggests that the formation of the shear zones took place during D_3 . To the north of this area a zone of distributed D_3 shear is also developed along the base of the Altkristallin in the Kreuzeck Group, where it is associated with a zone of lower greenschist retrogression (fig.3.16.). This forms the eastward continuation of the zone of D_s deformation described in this thesis and implies that D_3 , D_s and the formation of the discrete shear zones are all related to the same phase of deformation. This suggestion is supported by evidence from radiometric dating (see 3.9).

3.5 THE PRE- D_s ALPINE FABRIC

3.5.1 Introduction

To the south of the area affected by penetrative D_s deformation, a series of metamorphic rocks are preserved which have a weak foliation. To the east of map area 3 very similar features have been described in the Polinik area of the Kreuzeck Group by Waters (1976) and Hoke (1987), who show that the growth of these minerals took place during the Cretaceous with associated cooling ages $\approx 80\text{Ma}$. Similar cooling ages were measured in areas 1 and 2, which suggests that M_r mineral growth throughout this region is related to the same Cretaceous event. The comparison between the areas studied in this thesis and those to the east plays an important part in the conclusions about the timing of the events in this area and is discussed in more detail later (see 3.8).

It is not immediately obvious that the undisturbed metamorphic fabric is the same as the S_r fabric, which has been deformed by D_s to the N. The undisturbed metamorphic fabric is, therefore, given the provisional label S_r' and the associated phase of mineral growth M_r' .

3.5.2 Mr' Microstructures

In the area to the south of the zone influenced by Ds deformation, the rocks show an upper greenschist to amphibolite grade metamorphism, which overprints a deformational fabric. In area 3 the metamorphic mineral assemblages are typically kyanite + staurolite + garnet + biotite + quartz + muscovite; hornblende + biotite + garnet, both of which are characteristic of the amphibolite facies (fig.3.17a.). In the other two areas no fresh amphibolite grade assemblages were recorded and in these areas, Mr' is probably a maximum of upper greenschist facies. In all three areas a foliation, Sr', is defined by the general preferred orientation of platy minerals (fig.3.17b.). Cross-cutting relationships show that at least some of the mineral growth postdated the formation of Sr'. In some outcrops the preservation of a biotite mineral lineation indicates that some mineral growth may also have been synkinematic. Further evidence for synkinematic mineral growth comes from studies to the south of area 1. In this area Behrmann (1987b) describes a deformation, Dj, which is directly comparable with Dr' of this thesis, and is locally associated with the synkinematic growth of garnets. This microstructural evidence suggests that Dr' and Mr' mineral growth were in part synchronous but that the mineral growth outlived deformation.

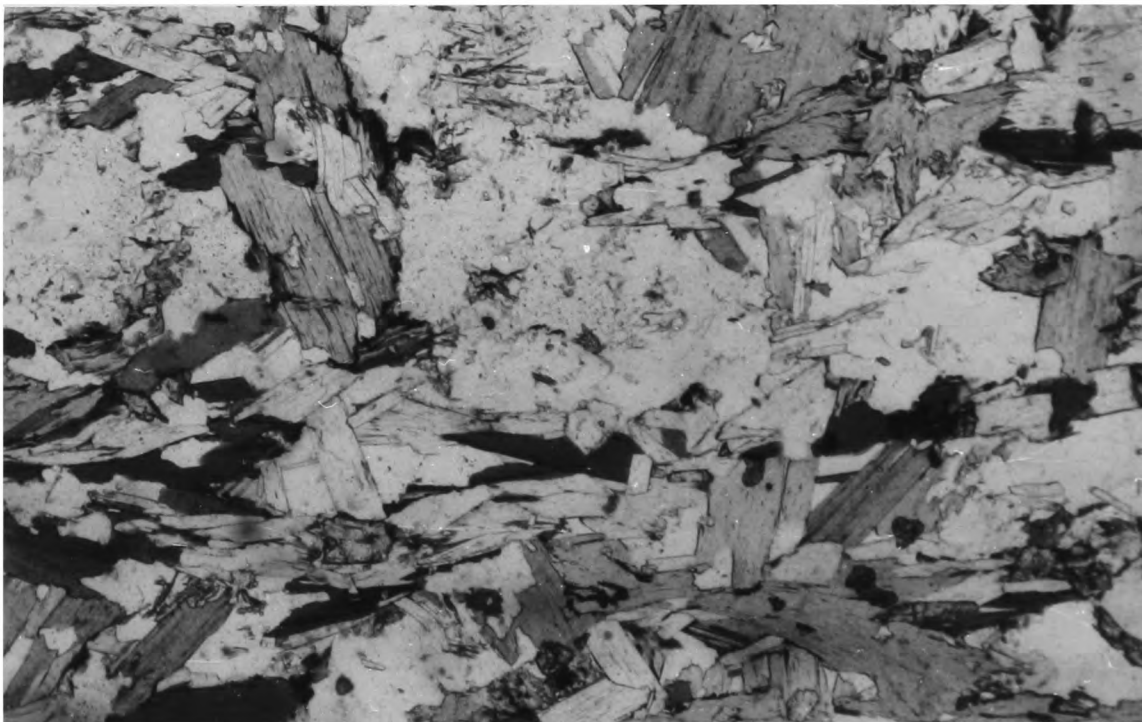
Besides the fresh Mr' garnets there are also some garnets which are relics of an earlier metamorphism, Mq (fig.3.18.). These can be identified by their inclusion trails at a high angle to Sr' and common alteration to biotite. Mr' garnets do not show alteration to biotite and they only rarely contain abundant inclusions.

In area 3, another common feature of the rocks affected by Mr' is the late-stage overgrowth of garnet and in some cases staurolite and biotite, by albite porphyroblasts (fig.3.19.). Several other late-stage effects can also be distinguished. Some of the Mr' porphyroblasts become slightly chloritized; this retrogression is particularly



3mm

Fig.3.17a. Amphibolite grade metapelite, Mr schist MS101, from south of the Mellenkopf Shear Zone. St = staurolite, Ky = kyanite, Gt = garnet, Bi = biotite.



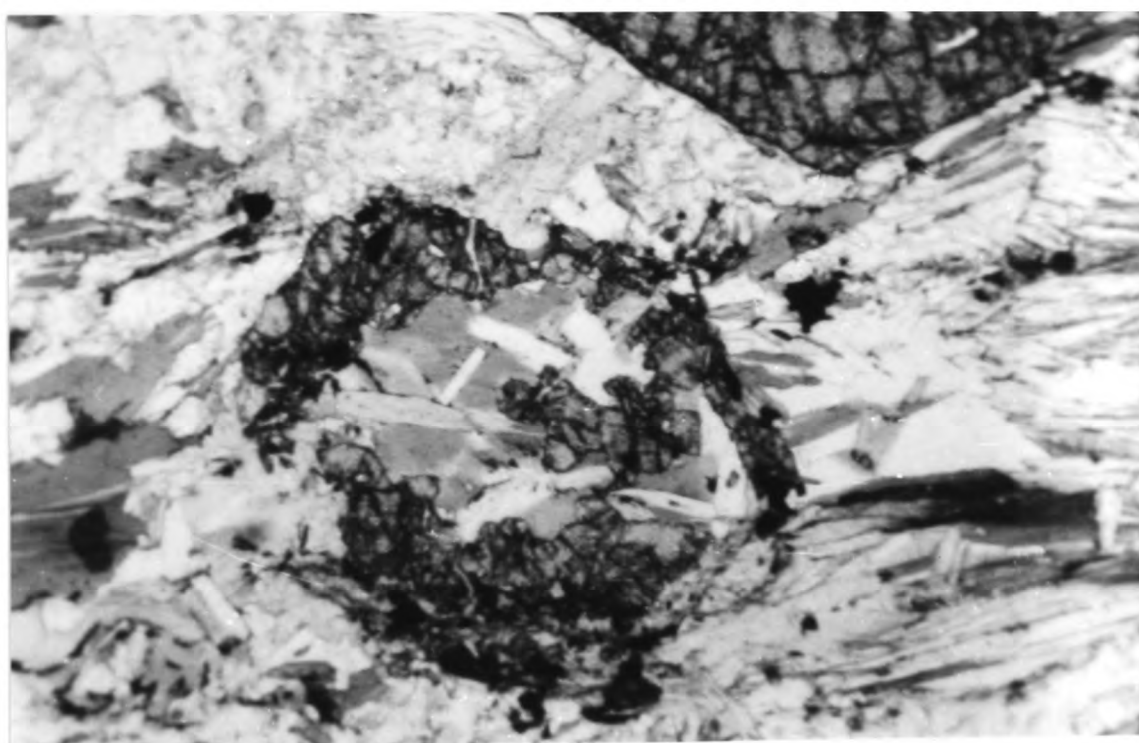
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Fig.3.17b. Mr biotite cross-cutting Sr foliation, SW16, area 2.



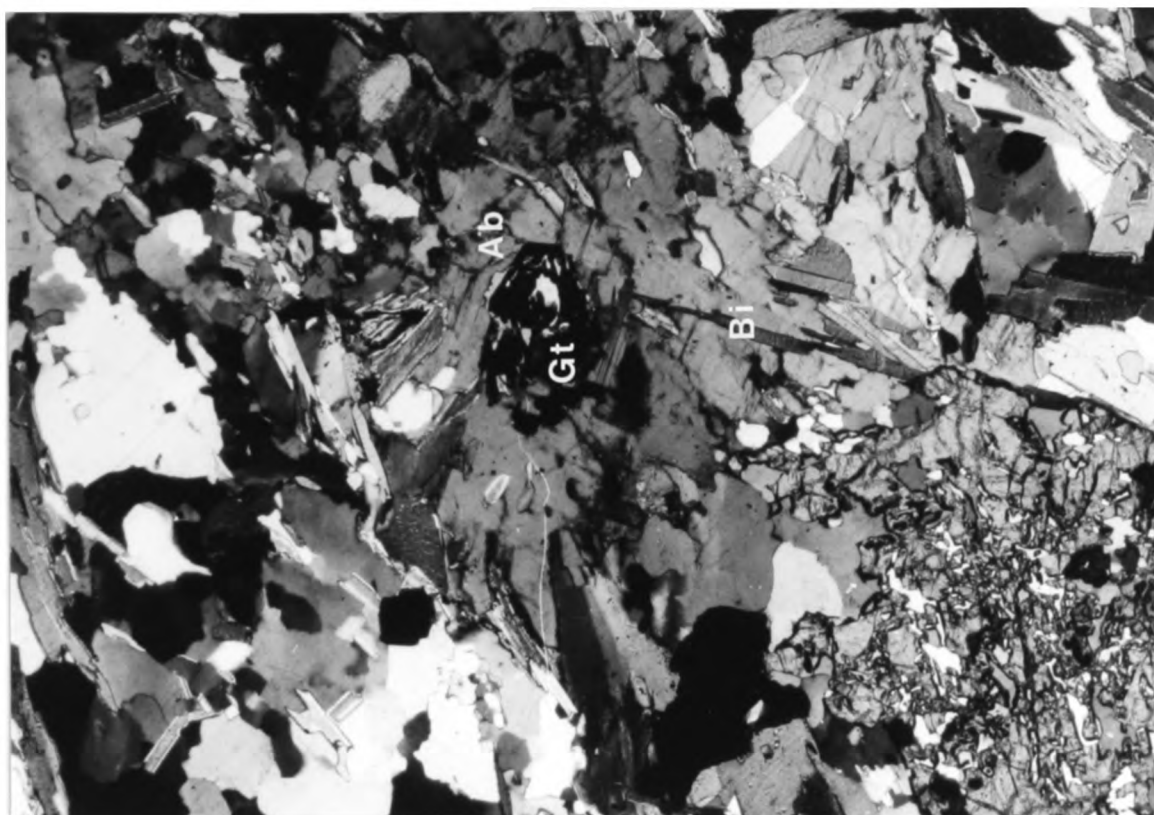
3mm

Fig.3.18a. Mq garnet with inclusion trails at a high angle to Sr (parallel to the long side of the photograph). A small Mr garnet has grown immediately above the Mq garnet and contains an inclusion trail parallel to Sr. MS101, area 3.



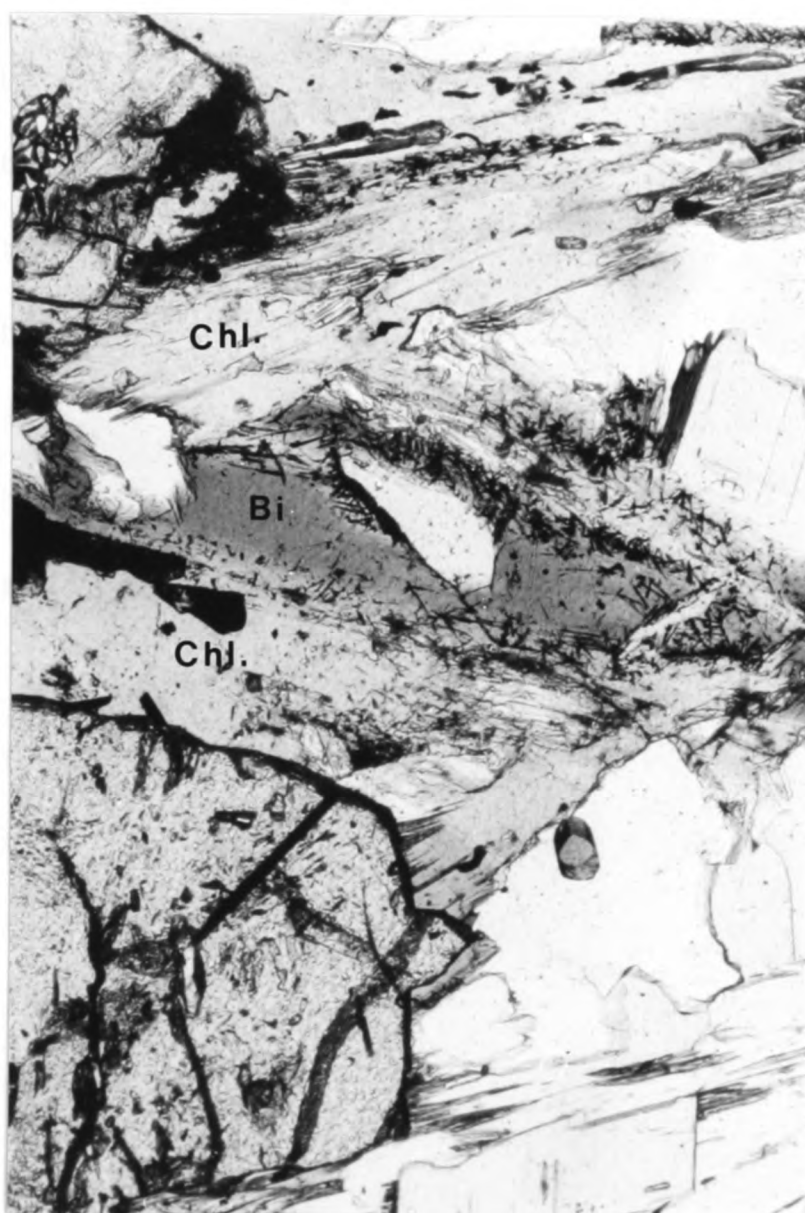
3mm

Fig.3.18b. Mq garnet partially replaced by biotite, MS94, area 3.



1mm

Fig.3.19. Mr garnet and biotite overgrown by albite. MS101, area 3.



1mm

Fig.3.20. Chlorite porphyroblast grown after Mr, the associated retrogression of Mr biotite causes the local exsolution of rutile needles along the chloritized margins, MS101, area 3.

noticeable in the biotites, which exsolve rutile needles near their chloritized margins (fig.3.20.); and a number of new mineral phases crystallize, e.g. fresh chlorite and zoisite porphyroblasts.

3.5.3 Relationship between Sr', Mr' and Sr, Mr

The cooling ages for Mr' clearly represent a minimum age for any deformation which overprints this fabric. It is, therefore, important to establish that Sr' to the south of the zone of distributed Ds shear and the similar grade metamorphic fabric deformed by Ds is one and the same. Three lines of argument are followed:

- (i) tracing the mesoscopic foliation;
- (ii) comparing the microstructure; and
- (iii) radiometric dating.

(i) In area 3 the zone of undisturbed amphibolite Mr rocks is neatly separated from the zone of Ds deformation by the Mellenkopf Shear Zone. Sr' cannot, therefore, be traced into the zone of Ds deformation. In areas 1 and 2, however, there are no such discontinuities and it is very likely that it is the same metamorphic fabric which is deformed by Ds as is present to the south.

(ii) For the most part, the Sr microstructures within the zone of Ds deformation are too altered for a close study, but in MS95 the relationships between the pre-Ds fabric and the growth of the pre-Ds metamorphic minerals are well preserved. Garnets contain good spiral inclusion trails, which are continuous with Sr (fig.3.21.). The garnets therefore grew synkinematically with Sr (Schonefeld 1977). The inclusion trails within the garnets can be traced continuously through to the centre of the crystal and do not show any obvious variation in intensity with position in the crystal. If the garnet had begun to grow at the very onset of Dr deformation then the Sr fabric could not exist in the core, as it would not have yet been formed. I therefore suggest that Mr

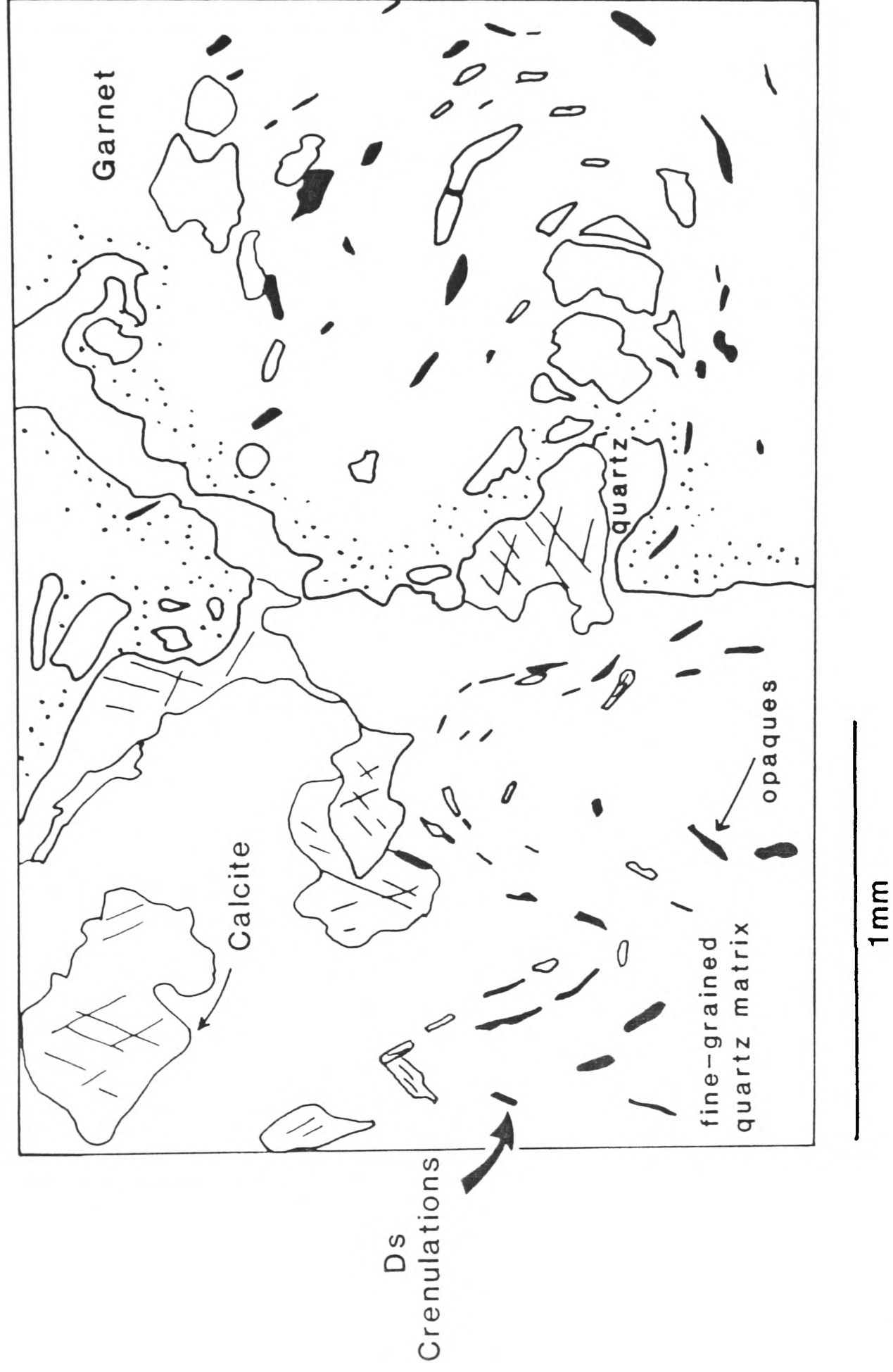


Fig.3.21. Synkinematic Mr garnet found within the zone of Ds deformation. The internal fabric of the garnet is continuous with Sr, which outside the prophyroblast has been tightly crenulated by Ds. The presence of calcite in the rock may explain the lack of retrogression despite the relatively high Ds strain (see text for further discussion).

and Dr were in part contemporaneous, but that Sr had begun to develop before growth of garnet and the peak metamorphic conditions were reached.

The same sample contains relatively large amphibole grains which also have an internal fabric (Sr). The variable angle between the long axis of the amphiboles and the internal fabric can be explained by their rotation during Ds and suggests that these minerals cross-cut the pre-Ds fabric and grew post-kinematically (fig.3.22.).

The microstructural development preserved within MS95, therefore, suggests that the pre-Ds fabric Sr began to develop before major growth of metamorphic minerals and was overprinted by peak metamorphic conditions. This is the same as deduced in the zone affected by Dr' and Mr'(see 3.5.2).

A characteristic feature developed in rocks affected by Mr' is the late-stage overgrowth of higher-grade minerals by albite porphyroblasts. In the zone of distributed Ds shear, similar relationships are found between albite and small euhedral garnets, which can be compared to Mr' garnets. There are also larger garnets present within the zone of Ds deformation which commonly contain abundant inclusions. These can be compared to the relic Mq garnets found associated with Sr'.

There is, therefore, very good correspondence between the microstructures found associated with Mr' and Dr' and those which are related to Sr, Mr deformed by Ds, implying that they are one and the same.

(iii) The third and final method is by radiometric dating. Radiometric dating in the area of undisturbed amphibolite grade metamorphism in the Kreuzeck Group indicates that the mineral growth took place during the Cretaceous, possibly reaching a peak T around 100Ma (see section on dating). The continuation of this zone is represented by the amphibolite grade part of area 3 associated with Mr' and Sr'.



3mm

Fig.3.22. Mr amphibole crystals with Sr inclusion trails at various angles to Ss, suggesting a post kinematic growth of the amphibole with respect to the Sr foliation and subsequent rotation during Ds. MS95, area 3.

In area 1, three samples were selected from the zone of high-strain Ds deformation for Rb-Sr dating. Two of the samples, K1 and K2, were characterized by fine-grained micas parallel to Ss. The third sample, K3, contained markedly coarser micas which were commonly overgrown by albite porphyroblasts. The microstructures suggest that these micas of K3 may be related to Mr. The two samples with Ds microstructures gave ages of 50 ± 1 Ma and 40 ± 1 Ma but the third sample has an age of 109 ± 1 Ma, which corresponds well with the proposed peak T suggested for Mr. It cannot be proved that this Rb/Sr age dates the formation of the mica fabric, it could also be a mixed age, but this is another piece of circumstantial evidence that Sr is continuous throughout the basal part of the Altkristallin and that $Sr' \equiv Sr$; $Mr' \equiv Mr$.

Summary

The evidence for the phase of mineral growth, Mr', and the schistosity, Sr', defined beyond the influence of Ds, being the same as the metamorphic fabric deformed by Ds, is compelling rather than irrefragable. To prove this rigorously the age of the growth of minerals deformed by Ds and those associated with Mr' would have to be shown to be the same. There is, however, strong evidence that these two fabrics are indeed identical, and this will be assumed for the rest of this thesis.

3.5.4 Mesoscopic structures

It is not always easy to distinguish Ds effects from Dr in the field, particularly in the deformed orthogneiss, which only poorly preserve fabrics. Where Ds overprints Sr in the form of a shear band cleavage and not folding, the new-formed fabric may also not be obviously distinct from the earlier Sr. Dr is characterized by its association with higher grade minerals than were stable during Ds. Where undeformed and unretrogressed biotite and garnet were seen on a

foliation surface, the foliation was recorded as Sr. In the areas where both Ds and Dr are present Ds is relatively low-strain and the characteristic NW-SE stretching lineation is absent. In these areas Ss, therefore, has to be identified by the presence of deformed, greenschist facies mineral assemblages.

Sr has a variable orientation, which is due mainly to subsequent folding. Where late-stage folds are lacking, Sr has a moderate southerly dip, similar to Ss in areas 1 and 3. However, in the southern part of area 2 there is a marked discordance in the orientation of the two foliations, and Sr dips steeply to the NW (fig.3.23, fig.3.1c.).

Dr folds can only rarely be identified (fig.3.24.) and their axes have no obvious preferred orientation. No systematic changes in vergence were observed in this study. However, at least some of the repetitions of a characteristic white augengneiss along the ridge to the east of Böses Weibl (area 1) are due to isoclinal Dr folding.

In a few localities a mesoscopic stretching lineation, Lr, can be identified, which in areas 2 and 3 is oriented roughly E-W and defined either by a biotite mineral lineation or the preferred alignment of quartz grains. In area 1, however, where Ds strain is still relatively high, four localities have a pre-Ds stretching lineation oriented NE-SW. Their unusual orientation is interpreted as the result of Ds folding and passive rotation during deformation (chapter 4 - part B). The lack of south vergent Ds folds within the Altkristallin suggests that all the Ds folds have rotated with the same clockwise sense. Assuming the Ds folds originally nucleated at a high angle to the Ds stretching direction, suggests these Dr lineations had an original W-E to NW-SE orientation.

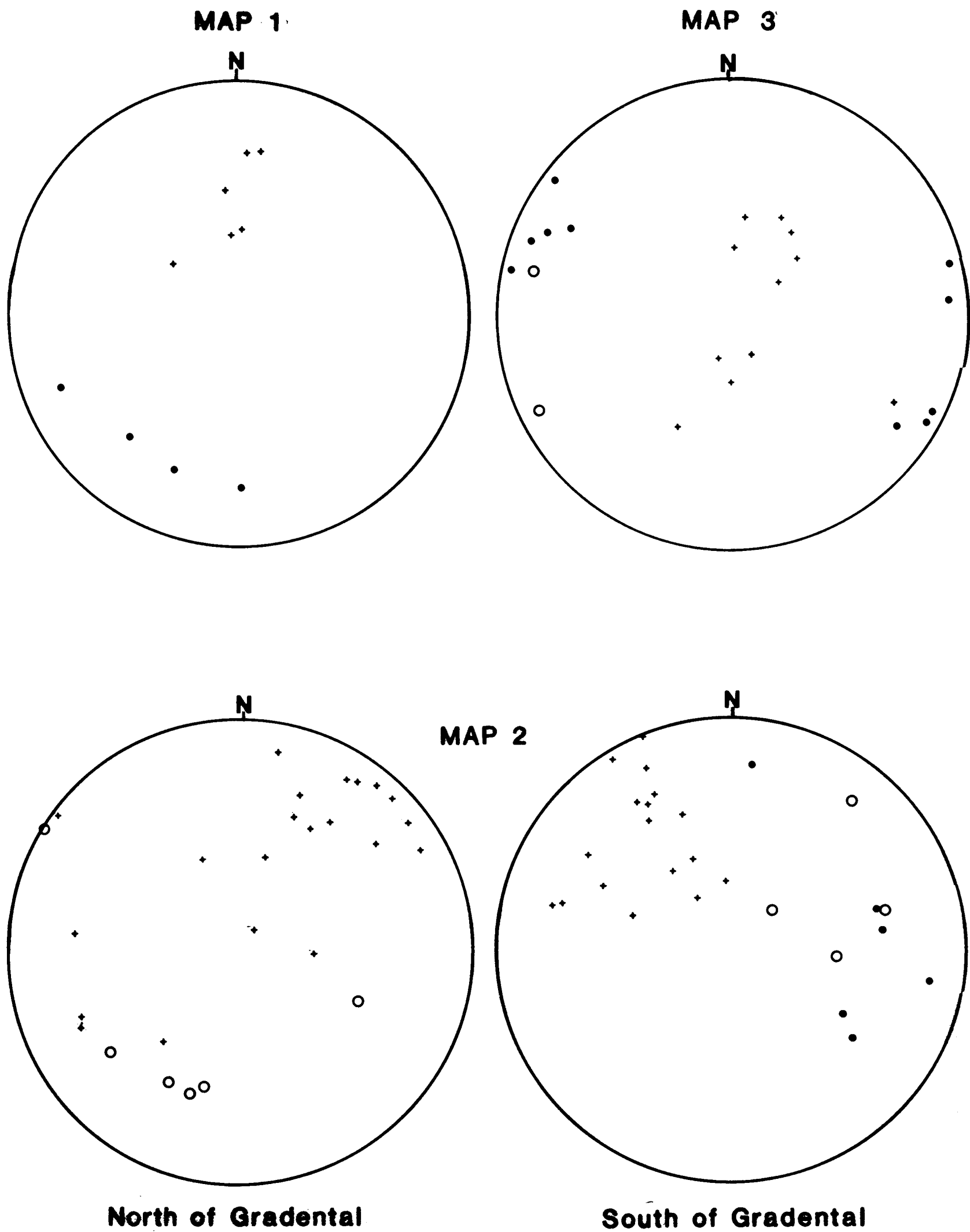


Fig.3.23. Mesoscopic Dr structures. The change in orientation of the Sr north and south of Gridental in area 2 suggests the presence of a large D_S fold (see fig.3.1b.).

• = stretching lineation, + pole to foliation, o = fold axis.



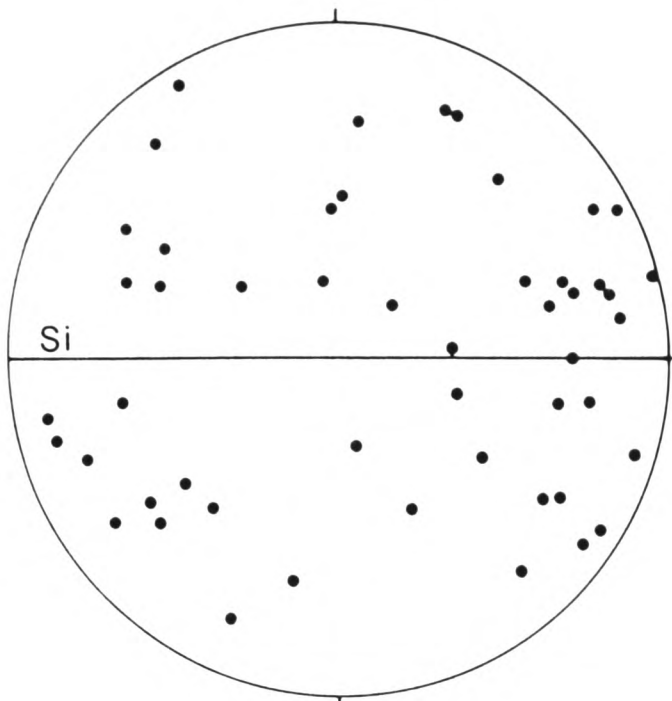
Fig.3.24. Dr isocline refolded by a Ds fold. MS57, area 3.

3.6. PRE-ALPINE Dq DEFORMATION; Mq METAMORPHISM

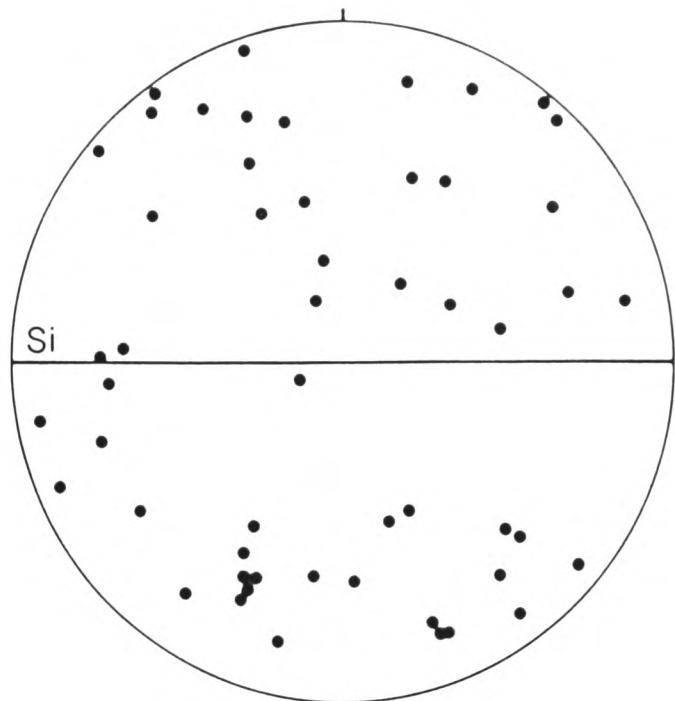
Work by Behrmann (1987b, unpublished) to the south of area 1 shows that the effects of Mr and Dr gradually fade out and reveal an earlier deformation and metamorphism, Mq, Dq. The three profiles studied in this work were not traced back far enough to see where Sq became the dominant fabric but something about its nature can be deduced from microstructures.

Mq garnets can be identified in areas which are dominated by Mr metamorphism, where they are distinguished by their large size, inclusion trails at a high angle to Dr, and their common alteration to biotite during Mr (fig.3.18a, 18b). Their replacement by biotite can be explained by their being formed under conditions specific to Mq and no longer being stable during the Mr re-equilibration. Most of the garnets contain abundant quartz inclusions, often with a sponge texture (skeletal garnets). In the cases where a definite inclusion trail could be identified, it was defined in part by the alignment of grain boundaries rather than a grain shape fabric. The internal fabric in these Mq garnets can also be distinguished from the Sr fabric overgrown by Mr garnets, by lacking a quartz preferred orientation (fig.3.25.). In one sample (SW13, area 2) flattened opaques defining curved inclusion trails were also included in the porphyroblasts, indicating synkinematic growth. Locally the larger Mq garnets contain biotite, suggesting that this was also stable during Mq. In one sample (MS101) Sr is axial planar to tightly folded biotite layers defining Sq, which confirms the association of Sq with biotite. Sq, therefore, formed under at least upper greenschist facies and probably higher in view of the large garnet size (up to 5cm adjacent to the marble band in area 2).

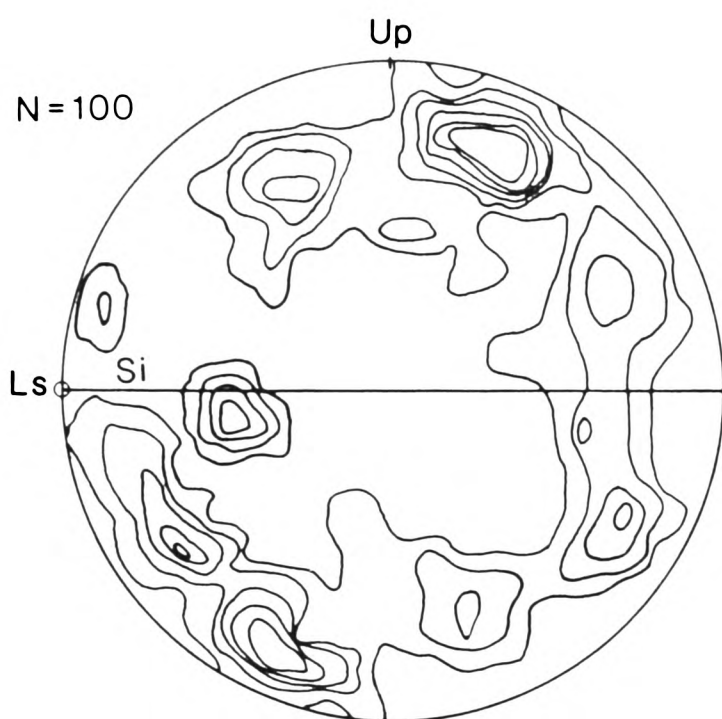
Fig.3.25. Quartz c-axis fabrics measured within Mq garnets, pre-Ds albites and an Mr garnet. The internal fabric of the porphyroblasts was used as a reference frame for plotting the orientation of the quartz c-axes. The fabrics from the Mq garnets in samples K5 and SW13 are taken from only one porphyroblast, all the other fabrics are a combination of measurements from several within the same this section. The c-axes from the inclusions within the Mq garnets show no preferred orientation, whereas the younger porphyroblasts preserve a clear preferred orientation. The c-axis fabrics within the albite porphyroblasts were originally measured in sections parallel to Ls and have been rotated about an axis perpendicular to the reference plane in order to make the fabric symmetrical. The position for Ls is shown after this rotation has been performed. The fabric within the Mr garnet in sample B90 has not been rotated. Contours are 1 to 6% per 1% area, shaded areas within the fabric have a density of <1% per 1% area.



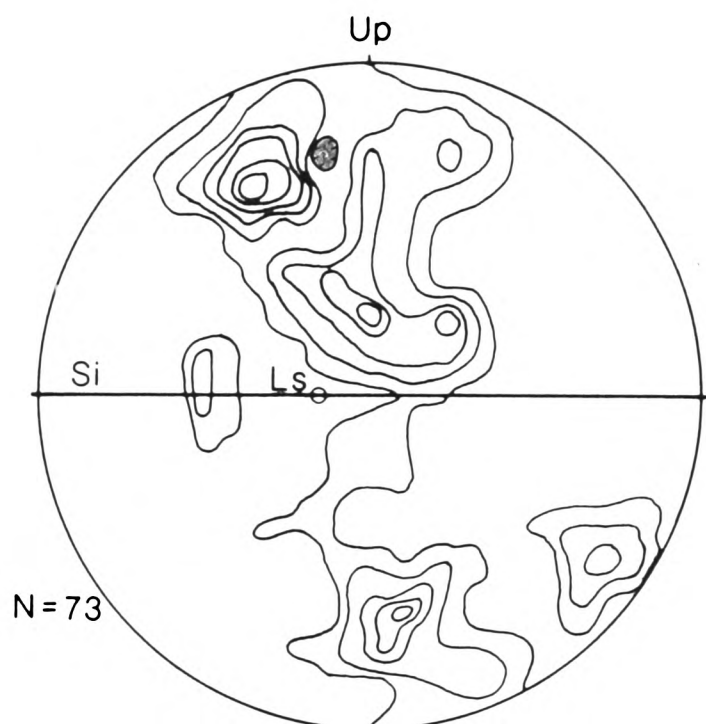
K5
Mq Garnet



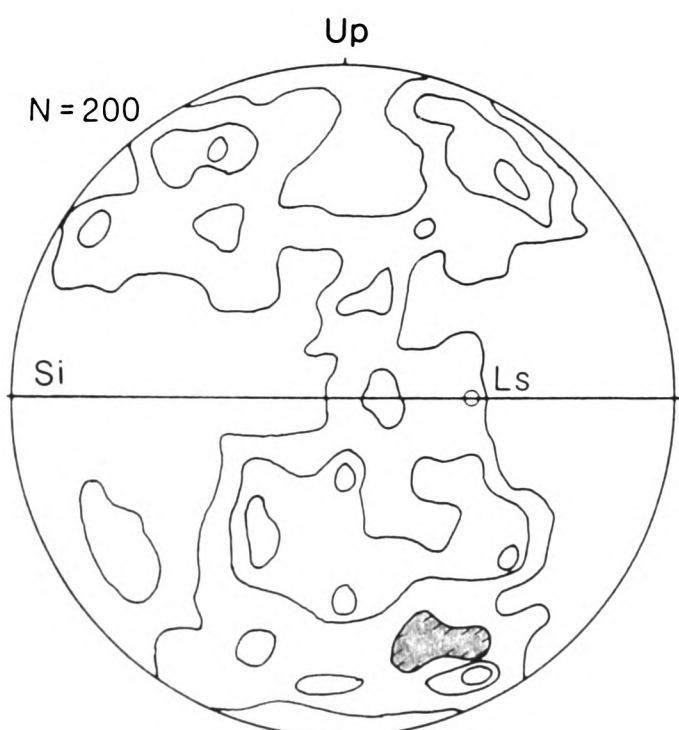
SW13
Mq Garnet



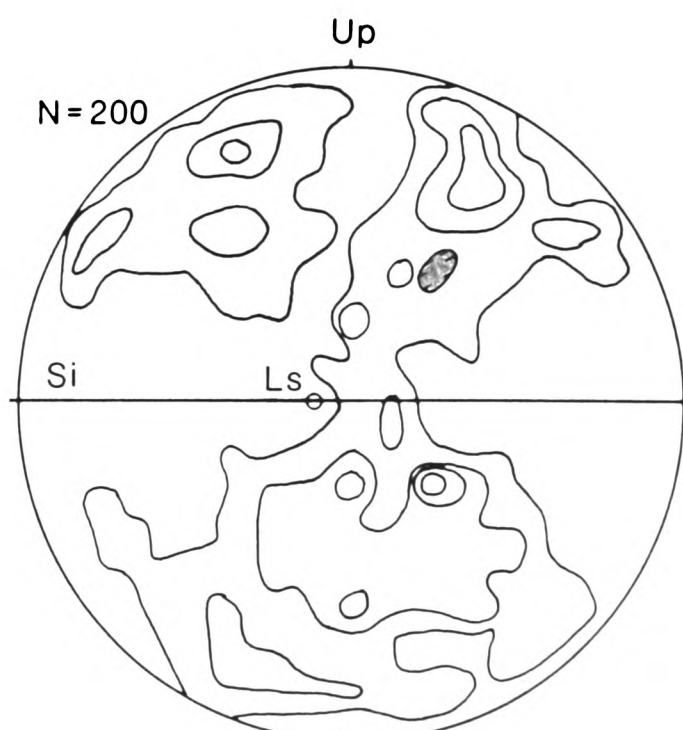
B90
Mr Garnet



K5i
Pre-Ds Albite



MS63
Pre-Ds Albite



K5ii
Pre-Ds Albite

The greatest Alpine ages in this region come from areas strongly affected by Mr, implying that Dq and Mq are pre-Alpine events. Radiometric dating in the area to the south east of the Tauern Window shows that there were at least two pre-Alpine events, which have been dated at around 430Ma (Troll et al 1976) and ≈ 350 Ma (Brewer 1970). Mq compares well with the amphibolite grade upper Palaeozoic metamorphism seen to the south of the Kreuzeck Group (Brewer 1970, Hoke 1987, Waters 1976).

3.7. CAUSES OF RETROGRESSION ALONG THE BASE OF THE ALTKRISTALLIN

Having established the main features of the microstructural development of the Altkristallin, it is now possible to look more closely at the causes of the retrogression seen within the basal few kilometres of the Altkristallin.

Nearly all previous workers have related this retrogression to the Tauern Metamorphism. In contrast, in this thesis I recognize a penetrative deformation Ds, which took place under lower greenschist conditions and was responsible for nearly all this retrogression. In detail, however, there are 3 greenschist events which may, at least in part, be related to the retrogression of the Mr fabric and whose effects need to be distinguished. The three events in increasing order of age are:

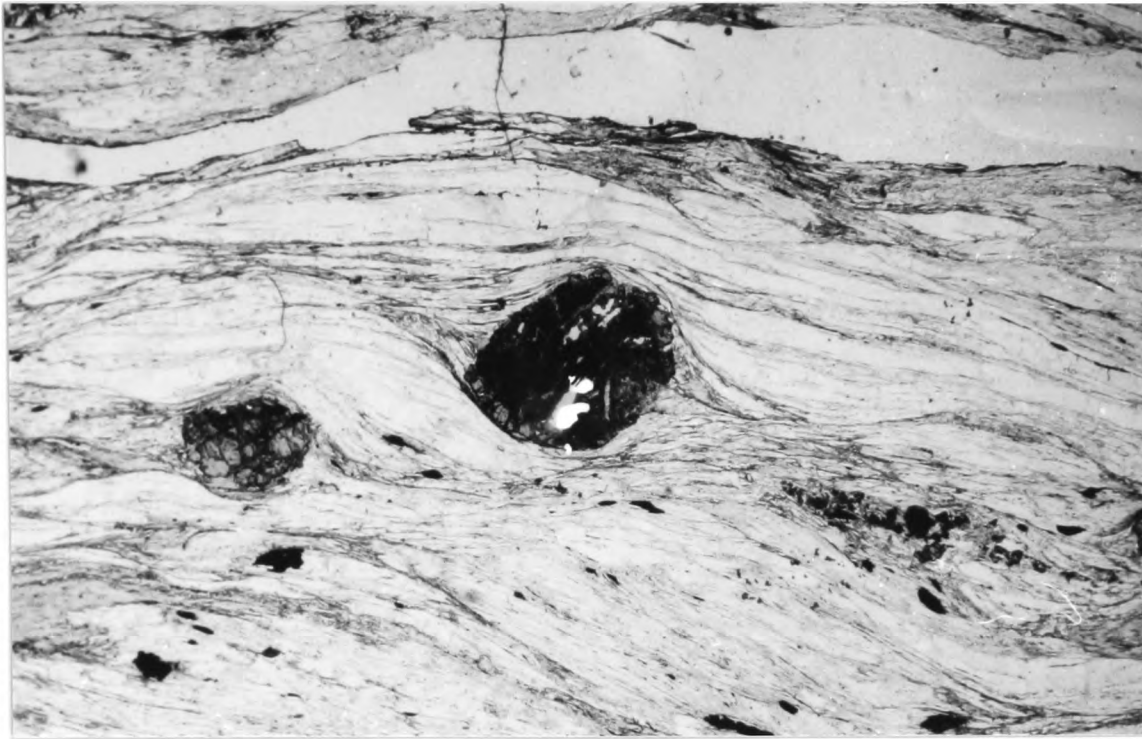
(i) Tauern Metamorphism, (ii) Ds deformation, and (iii) the greenschist minerals developed after the peak of Mr and before Ds.

Within the Matri Zone the effects of the Tauern Metamorphism can be seen where chlorites cross-cut Ss, and quartz develops an equilibrium foam texture, characteristic of static secondary recrystallization. These effects die away up structural section towards the contact with the Altkristallin. Within the basal zone of the Altkristallin the

greenschist minerals lie parallel to the main foliation. Bickle (1973), Hawkesworth (1974), Waters (1976) interpret this as mimetic crystallization of the greenschist facies minerals onto a pre-existing higher-grade fabric. However, if the extensive retrogression of several kilometres thickness of basement, were all due to a static metamorphism, at least some cross-cutting low-grade minerals would be expected. The microstructure of the garnets and biotites shows that the retrogression was, in fact, largely syndeformational. Fractured and chloritized garnets elongated parallel to Ls are common throughout the zone of Ds and porphyroblasts also commonly have chlorite-filled pressure shadows which are parallel to Ls (fig.3.26.).

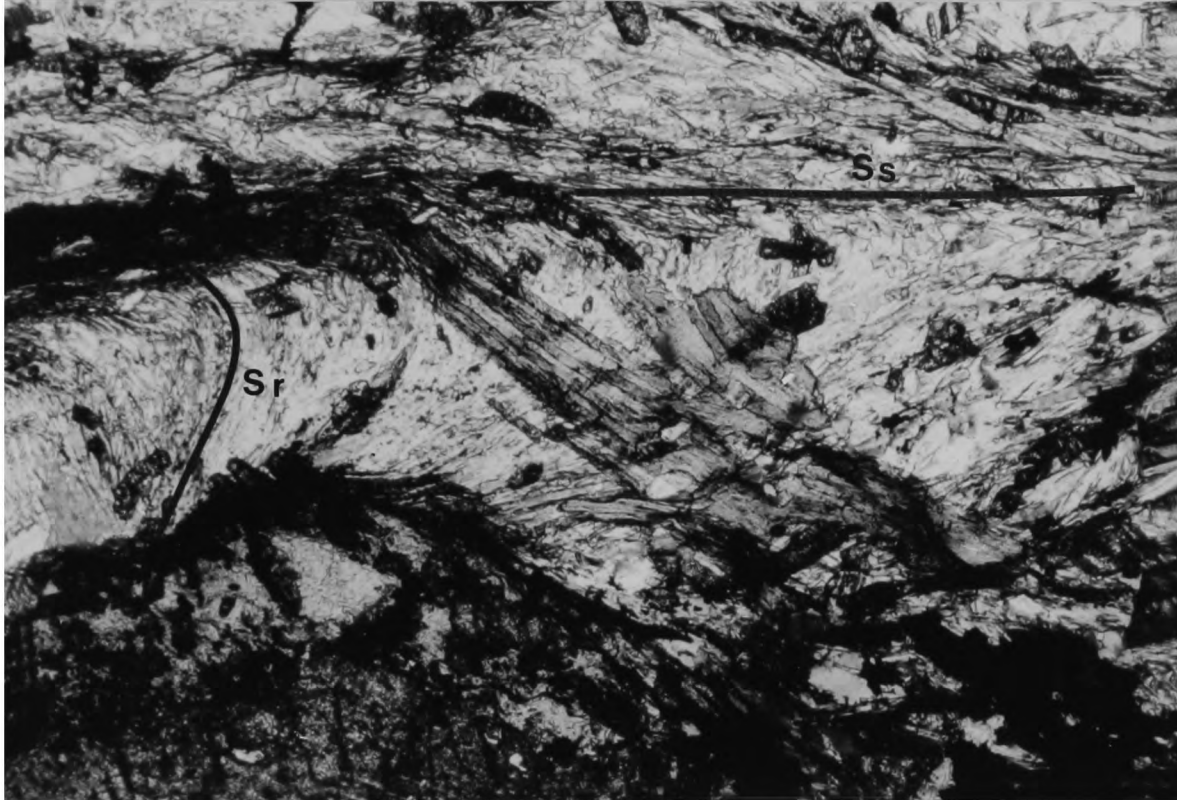
To the south of area 3, there are some distinct retrograde effects which involve the growth of chlorite. The chlorite porphyroblasts are undeformed and these could mistakenly be identified as the effects of the Tauern Metamorphism. These chlorites are, in fact, part of the greenschist event, which took place after the peak of Mr but before Ds and produced a variety of greenschist minerals (see 3.5.2) whose occurrence is entirely restricted to the higher structural levels of the Altkristallin, not near the base where the effects of the Tauern Metamorphism were greatest. In sample MS101 chlorites which cross-cut Sr are deformed by Ds (fig.3.27.).

Within the zone affected by Ds a few samples contain chlorite pseudomorphs after Mr garnets which still preserve the original garnet crystal shape (fig.3.28.). Chlorite is far too incompetent to have been unaffected by the Ds deformation, which implies that the retrogression either took place during the late stages of Ds or was a result of the Tauern Metamorphism.



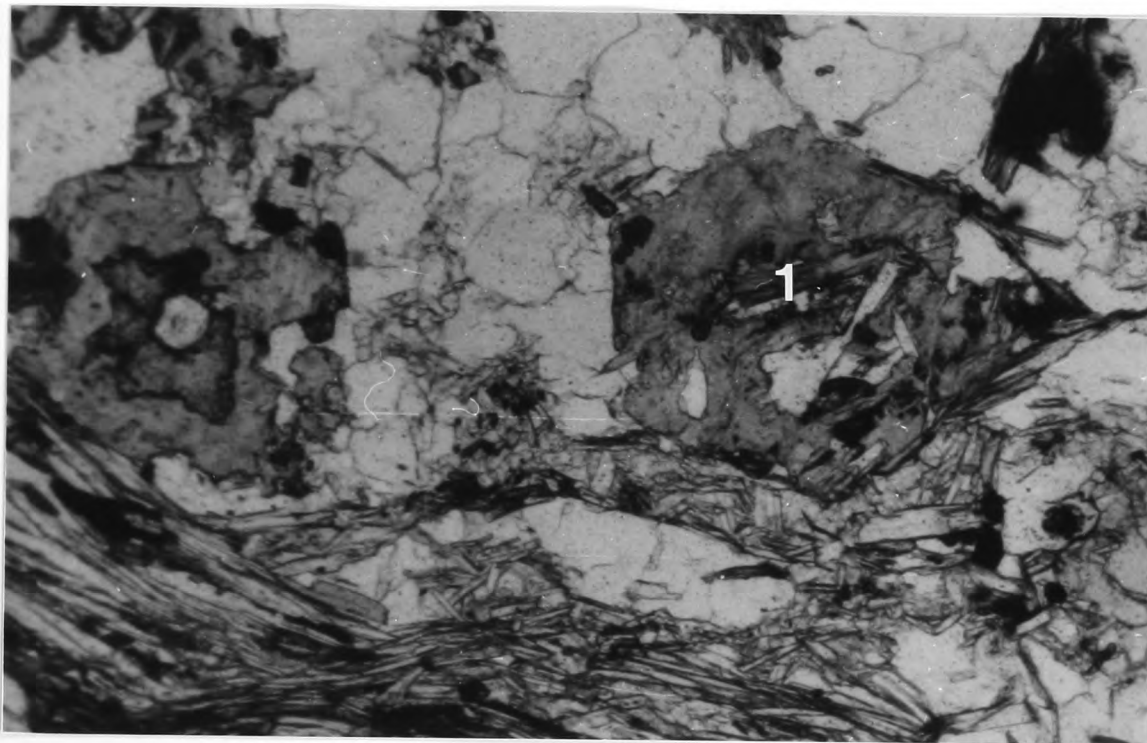
3mm

Fig.3.26. Asymmetric chlorite-filled pressure shadows around a Mr garnet, formed during Ds. The main planar fabric of the sample is Ss, which is slightly offset by a Ds extensional crenulation cleavage. See also fig.3.4.



0.5mm

Fig.3.27. Sr, defined by the preferred orientation of white micas, is cross cut by a chlorite porphyroblast, which has been deformed by Ds. This represents a pre-Ds post-Mr phase of mineral growth under greenschist conditions. MS100, area 3.



1mm

Fig.3.28. Mr garnets one of which overgrows an Mr biotite (1) have been pseudomorphed by chlorite but despite the relatively high accompanying Ds strain these pseudomorphs have not deformed. These could have developed during a late-stage in Ds or be due to the Tauern Metamorphism. Plane polarized light, MS 111, area 3.

3.8. COMPARISONS WITH PREVIOUS WORK IN THE ALTKRISTALLIN

3.8.1 Pre-Ds history

To the east of the map area 3, in the Kreuzeck Group, several detailed geochronological and petrological studies have been carried out (Waters 1976, Wright 1973, Hoke 1987, Brewer 1970). These studies are complimented by the work of Hawkesworth (1974) to the east of the Tauern Window and Bickle (1973) in area 1. Parallels between the geological history of the Kreuzeck Group and area 3 allow the radiometric dating undertaken in the Kreuzeck Group to be used to date events described from the study area.

One of the main points of reference was the Teuchl Shear Zone. In the Kreuzeck Group this is a discrete zone several tens of metres wide, which can be traced for a distance of several kilometres. In the west of the Kreuzeck Group the Teuchl Shear Zone divides into two branches (fig.3.16.); the northern branch cuts through an amphibolite grade metamorphic sequence and the other separates a province dominated by Alpine ages from high-grade Variscan-aged rocks. Deformation within these shear zones took place under lower greenschist conditions and is associated with NW-SE stretching lineations and a top to the NW sense of shear (Hoke 1987). Alpine amphibolite grade rocks are preserved between the two branches of the Teuchl Shear Zone and immediately north of the northern branch. However, a short distance further N, closer to the base of the Altkristallin the amphibolite grade metamorphics are overprinted by a zone of greenschist retrogression associated with penetrative D_3 deformation.

The Mellenkopf Shear Zone in area 3 also separates an area to the south, where fresh amphibolite grade metamorphics are preserved, from a northern area where Sr and Mr are progressively overprinted by a lower greenschist facies deformation. Deformation within the Mellenkopf Shear

Zone is lower greenschist and associated with a NW-SE stretching lineation.

Area 3 and the Kreuzeck Group are separated by a long straight section of the Möll Valley along which the late-stage Wöllratten sinistral fault has been postulated (Hoke 1987, Tollmann 1977b). Projecting the MSZ and the northern branch of the Teuchl Shear Zone towards the Möll Valley, shows they are separated by a horizontal distance of 2km (fig.3.16.). However, the similarities in both microstructure and geological position strongly suggest the two shear zones are part of the same structure. The offset can be explained by the left-lateral movement on the Wöllratten Fault. This direct comparison plays an important part in relating the studies in this thesis to the studies of Waters, Wright and Hoke in the Kreuzeck Group.

South of the Teuchl Shear Zone, Waters (1976) defines a D_2 deformation which is mimetically overprinted by an amphibolite metamorphism. This is exactly the same relationship as seen in area 3 south of the MSZ. D_2 can, therefore, be straightforwardly correlated with D_r . This is supported by directly comparable microstructural relationships recorded in both areas. Waters suggests that M_2 and D_2 may have been in part contemporaneous but has only circumstantial evidence for this. The presence of well-preserved syn-kinematic M_r garnets in map area 3 (MS95, fig.3.21.) confirms his suggestion.

Waters also defines an earlier D_1 and M_1 event which corresponds to the pre-Alpine M_q and D_q found in area 3, and describes very similar microstructural features of the garnets, to those related to M_q and D_q in this work. Hoke (1987) relates some sillimanite-bearing assemblages in the Kreuzeck Group to this metamorphism. This indicates temperatures in excess of 550°C and confirms the suggested amphibolite grade for M_q .

Summary

In both the Kreuzeck Group and area 3 there is a zone of unaltered amphibolite grade metamorphism, which is cut by a lower greenschist facies shear zone associated with NW-SE stretching lineations. To the north of both areas the amphibolite grade metamorphics pass into a zone of lower greenschist retrogression. The geographical proximity and directly comparable geological features of the two areas strongly suggests that they have undergone the same early geological development, i.e. $M_1, D_1 = M_q, D_q$; $M_2, D_2 = M_r, D_r$.

3.8.2 Ds and the retrogression of the Altkristallin

The disagreement between this thesis and the consensus of opinion from earlier work in the Altkristallin of the S.E. Tauern, relates to the structures within the basal few kilometres of the Altkristallin, and in particular the causes of retrogression within this zone. The majority opinion of workers in the Kreuzeck Group was that the foliation within the zone of retrogression was the same S_2 which is present to the south, associated with the amphibolite grade metamorphism (Waters 1976, Cliff et al 1971, Bickle 1973). Retrogression was considered by most workers to be a result of the late-stage static Tauern Metamorphism.

Bickle & Hawkesworth (1978) studied the basal contact of the Altkristallin and their work apparently confirmed this idea. They suggested that penetrative Alpine deformation affected only the lower-most twenty meters of the Altkristallin. The implication made most explicit by Bickle (1973) was that the Altkristallin was emplaced onto the underlying metasediments of the Matri Zone as a rigid amphibolite grade sheet.

Two dissenting opinions were offered by Wright (1973) & Hoke (1987), who both recognized a penetrative deformation which post-dates M^2 and is at least in part responsible for the retrogression of this

fabric. In the Kreuzeck Group the relationship of this deformation to the emplacement of the Altkristallin over the Matri Zone cannot be investigated as the contact is overprinted by a late-stage brittle fault, the Mölltal Line (Exner 1962).

In the Schober and Sadnig Groups mapped in this study, overprinting relationships are present which show that the Mr ($=M_2$) amphibolite - upper greenschist grade metamorphism in the Altkristallin and its associated deformational fabric, Sr, is extensively overprinted by an Alpine, penetrative deformation, Ds. In the Kreuzeck Group Hoke (1987) defines a D_3 deformation, which has the same meso- and microstructural relationships as Ds. D_3 in the Kreuzeck Group is expressed as a zone of penetrative deformation associated with greenschist retrogression along the base of the Altkristallin and as discrete shear zones at higher structural levels. The deformation scheme erected by Hoke can be directly compared to that suggested for the study areas in this thesis and implies $Mq=M_1$, $Dq=D_1$, $Mr=M_2$, $Dr=D_2$, $Ds=D_3$. The retrogression in the base of the Altkristallin was mainly caused by Ds ($=D_3$).

The justification for defining Ds as a separate deformation phase and for relating the retrogression along the base of the Altkristallin largely to this phase has already been discussed (see 3.7.).

Within the zone strongly affected by Ds there are, however, some relationships which at first glance suggest that Ds did indeed take place under amphibolite grade conditions as suggested by Bickle. As this is a major point of disagreement with other workers these relationships are briefly discussed below.

Amphibole, partially altered to biotite, and garnet are found in one particular area within area 3 without any apparent retrogression to greenschist facies despite the accompanying high Ds strain. Garnets contain spiral inclusion trails, which show them to have grown synkinematically. These observations could lead to the erroneous

conclusion that Ds took place under upper greenschist to amphibolite conditions. Closer inspection shows the garnets are in fact synkinematic with the pre-existing deformational fabric, Sr. Sr has subsequently been transposed to give Ss. Where Ds crenulations are present it is obvious that the internal fabric of the garnets is continuous with the Sr fabric not Ss (fig.3.21.). The good preservation of these minerals is explained by the presence of calcite in the rock which would have the effect of lowering the a_{H_2O} during Ds deformation and thus impeding the normal retrograde hydration reactions.

3.9 RADIOMETRIC DATING

3.9.1 Introduction

Throughout the S.E. Tauern area, a large number of mainly K-Ar age determinations have been carried out across the Austroalpine - Pennine boundary. These demonstrate the existence of a consistent age profile. Furthest away from the margin of the window the ages are dominantly Palaeozoic (Brewer 1970, Troll et al 1976). To the north of this zone at structurally deeper levels, the isotopic systems have been influenced by Alpine events, which define a broad zone throughout the base of the Altkristallin, dominated by 100-40Ma ages. The degree to which the Palaeozoic isotopic systems have been reset depends on a variety of factors including temperature, the mineral involved, grain size and the isotopic system under consideration.

In the study area I have shown that there are at least two potentially datable Alpine events within the base of the Altkristallin, i.e. a phase of mineral growth, Mr, and a penetrative ductile deformation, Ds. Ds overprints the mineral fabrics developed during Mr and is concentrated in the lower greenschist zone of distributed deformation along the base of the Altkristallin. The principal reason

for most of the previous workers considered that deformation within this basal zone of the Altkristallin was the same as that associated with the amphibolite metamorphism further south, was the similar radiometric ages found in both areas. The similar cooling ages and the structural evidence for two distinct events are compatible if Ds took place shortly after the cooling from peak Mr conditions.

3.9.2 Dating Mr

To the south of the zone of Ds, the Mr mineral assemblages are upper greenschist to amphibolite grade. The associated radiometric ages range from 110–75Ma. These are mainly K-Ar ages and in ideal case indicate the time at which the mineral cooled below its blocking temperature. In many samples there is a problem of excess argon giving falsely high ages (Brewer 1970). However, within this zone in the Polinik ridge of the Kreuzeck Group, detailed K/Ar radiometric studies give uniform results with the ages ranging from 76 to 83Ma (Oxburgh et al 1966). The differences between coexisting biotite and muscovite and between different samples are close to the analytical error. A comparable Rb-Sr whole rock age of 78 ± 1 Ma was obtained from the same area by Cliff (unpub.). The blocking temperatures vary for different minerals and isotopic systems and each method should date a different part of the cooling path. The consistency of the ages in this area implies rapid cooling of the rock at this time through the different blocking temperatures for the different isotopic systems. 83–76Ma is, therefore, a good estimate for the age of cooling for Mr in this area.

Waters has demonstrated that within the Altkristallin of the Kreuzeck Group the metamorphic fabric associated with Palaeozoic ages is progressively overprinted by the growth of new Mr metamorphic minerals, including staurolite. The resetting of the Palaeozoic isotopic systems, therefore, corresponds to a phase of Alpine mineral growth and not

simply a phase of reheating. An indication of the time at which peak T conditions were reached is given by one sample from which Brewer (1970) measured an unaffected Palaeozoic age in white mica but found the biotite age had been reset during the Cretaceous metamorphism. The preservation of a Palaeozoic age in the white mica is interpreted as being due to its higher blocking temperature (Brewer 1970). The difference in these blocking temperatures for the two minerals is not very great (Dodson 1973, 1979) and such conditions could not have lasted very long without the Palaeozoic isotopic system in the white mica also being reset. Brewer, therefore, suggests that the age of the biotite (103 ± 2 Ma) may date the peak T of the Mr metamorphism in the area.

The age of Dr could not be estimated from dating in the Kreuzeck Group, since the deformational fabric has been overprinted by the growth of Mr metamorphic minerals. However, in area 1 Mr probably did not exceed greenschist conditions. Under low-grade conditions recrystallization is required to reset the Rb-Sr systematics, they should not be affected by low-temperature reheating alone. The sample K3 comes from an area dominated by Ds deformation and contains mainly large (<2mm) micas, which have been partially overgrown by albite and subsequently deformed. These are interpreted as relic large micas from Mr metamorphism (p.51). This sample gives an age of 109 ± 1 Ma which could be related to Dr deformation. In the Altkristallin to the S.W. of the Tauern Window, Stöckhert (1984) suggests a similar 110 Ma date for the timing of the oldest Alpine deformation.

3.9.3 Dating Ds

Ds has not previously been recognized in the Schober or Sadnig Groups as a major deformation and no dates have been related to it. For this reason several samples were selected from known structural positions where Ss was the dominant fabric for Rb-Sr dating. This

method avoids the problems encountered with excess argon using the K-Ar method (Brewer 1969).

The Rb-Sr method dates either a time at which the ambient temperature fell below the blocking temperature for the particular mineral or the cessation of recrystallization caused by deformation. So far two new Rb-Sr results from area 1 can be added to the list of mainly K-Ar dates from the same area. Samples K1, K2 have similar fabrics to K3 but the micas are finer-grained ($<0.5\text{mm}$) and a lower proportion are enclosed within albite porphyroblasts. These fabrics are interpreted as a Ds deformational fabric and the samples give Rb-Sr ages of 40 ± 1 Ma and 50 ± 1 Ma.

In this area the Tauern Metamorphism reached a maximum of 400°C at the base of the Altkristallin (Bickle & Powell 1977) and without an accompanying deformation should not have affected the Rb-Sr isotopic system, which for white mica has a blocking temperature of c. 550°C (Dobson 1973, 1979). These post-Mr dates are, therefore, best explained as dating the recrystallization of micas during Ds.

Bickle (1973) reports a Rb-Sr age of $35.6\pm 0.3\text{Ma}$ on the platy minerals forming the gefüllte core of a feldspar in area 1 and relates this to the effects of the Tauern Metamorphism. The gefüllte texture of the feldspar is normally interpreted as ^{the} effect of retrograde reactions on feldspar formed at a higher grade (Frasl 1953, Cliff et al 1971). The minerals defining this texture could, therefore, have developed either during the dynamic recrystallization associated with Ds or during the Tauern Metamorphism.

A less reliable way of dating Ds comes from K-Ar studies in strongly deformed material. Deformation along the Teuchl Shear Zone in the Kreuzeck Group can be related to Ds (see 3.4.4). Waters (1976) records a K-Ar date of $61\pm 2\text{Ma}$ on strongly recrystallized white micas from this zone. The area of retrogression to the south of the T.S.Z., within the

Kreuzeck Group is spatially distinct from the main zone of Ds deformation along the base of the Altkristallin, but its relationships with the metamorphic history and the comparable mesoscopic features suggest deformation within this zone is also related to Ds (fig.3.16, Hoke 1987). Hawkesworth (1974) has obtained an isochron from within this zone, by plotting $\text{Ar}^{40}/\text{Ar}^{36}$ against $\text{Ar}^{39}/\text{Ar}^{36}$. The straight line obtained has a slope corresponding to $71 \pm 3 \text{ Ma}$. This line does not, however, pass through the origin and if it does represent a true isochron then the initial argon pressure was not zero.

3.9.4 Comparison of Ds and Mr

Brewer (1970) and Waters (1976) carried out most of the radiometric dating in the area and neither of them recognized a deformation comparable to Ds. Therefore, trying to relate the data to the effects of Mr and Ds necessarily relies on indirect comparisons. Even within area 1 in the Schober Group where detailed mapping was carried out in this work and some radiometric dates are available (Waters 1976, Bickle 1973), without the thin sections and field notes it is not always clear what the dominant fabric of an individual sample is.

There has long been a recognition that the ages tended to be younger nearer the margin of the Tauern Window (Brewer 1969, Bickle & Hawkesworth 1978). This has been explained as the diachronous cooling of the Altkristallin during the Cretaceous metamorphism (Mr). However, an alternative explanation is that two datable events closely followed one another with the youngest concentrated nearest the margin of the Tauern Window.

In the mapped areas in this study, the zone of penetrative Ds deformation corresponds to a zone of lower greenschist retrogression within the base of the Altkristallin. A very similar zone of retrogression associated with a NW-SE stretching lineation has been

mapped by Hawkesworth (1974) to the east of the Tauern Window and has also been recognized by workers in other parts of the Altkristallin (e.g. Bickle 1973). It is, therefore, possible to divide the Altkristallin around the S.E Tauern into two zones. The most northerly zone is associated with lower greenschist metamorphism and, by analogy with the mapped areas in this thesis, represents a zone dominated by Ds deformation. The second zone, further away from the base of the Altkristallin, contains relatively fresh amphibolite - upper greenschist grade Mr metamorphic rocks (fig.3.16.).

A statistical test was carried out on the age data grouped in this way to test whether there was a significant difference in the mean ages of the two zones. For this analysis, only white mica K-Ar ages were used and any samples with signs of excess argon were ignored (Brewer 1969, Brewer 1970). Most of the samples containing excess argon come from the zone dominated by unretrogressed Mr assemblages, and since these samples indicate a falsely high age, their inclusion in the analysis would enhance the distinction between the two zones. There is also a narrow zone along the base of the Altkristallin which gives anomalously high ages. These have also been interpreted as being due to excess argon (Lambert 1970). If these data had been included there would have been greater scatter in the ages within the retrogressed Ds zone, but the conclusions of the analysis would not have been altered.

The mean age for the zone of retrogression is $76.2 \pm 12 \text{ Ma}$ (2σ), and $83.1 \pm 4 \text{ Ma}$ (2σ) for the zone of relatively fresh amphibolite grade metamorphism. Using unbiased estimates for the population variance within the two zones it was found that the mean age from the zone of Mr is significantly higher than the mean age associated with Ds, at the 0.5% level (see appendix 1).

The suggestion that ages are generally lower in the northern zone is, therefore, a real phenomenon and very unlikely to be due to

statistical fluctuation. However, since the data have been divided into two groups to perform the analysis, the result cannot be taken as a priori evidence in favour of the existence of the two datable events as proposed in this thesis. The result does show, however, that the data are compatible with this assertion.

Summary

The interpretation of the radiometric data is not straightforward. However, circumspect use of the information does give some idea of the general age of Mr and Ds.

Cooling of Mr took place at around 80Ma and, at least for the Polinik ridge, was very rapid. The preceding peak T conditions may have been reached around 103Ma for one sample in the Kreuzeck Group. There is some indication that Dr deformation may have begun as early as 110Ma.

Microstructural evidence clearly establishes Ds as a major deformation which post-dates Mr. Rb-Sr dating along the base of the Altkristallin indicates that Ds continued until 40-50Ma. These ages cannot have been due to the Tauern Metamorphism, which is too low temperature in this area to reset the Rb-Sr systematics in white micas. The K-Ar dates at higher structural levels within the Altkristallin may be related to earlier stages in Ds, which suggests Ds began ≈ 70 Ma. The possibility that these represent mixed ages cannot be ruled out.

A crude bulk analysis of the white mica K-Ar ages grouped into two zones corresponding to the retrogressed basal zone dominated by Ds, and the unretrogressed Mr assemblages, suggests an average age associated with Mr = 83.2 Ma and Ds = 76.2 Ma.

METAMORPHIC AND STRUCTURAL EVOLUTION OF THE MATREI ZONE

3.10. MAIN PHASE, D_s DEFORMATION

3.10.1 Mesoscopic Structures

The dominant D_s fabric in the base of the Altkristallin can be traced continuously into the underlying Matrei Zone. At the base of the Altkristallin Ss is axial planar to D_s folds. The same relationships are seen in the Matrei Zone where D_s folds are only rarely preserved (fig.3.29). Where a D_s stretching lineation is observed, the D_s fold axes are dominantly oriented sub-parallel to this direction, although some of the D_s fold axes have a steep plunge. It is likely that fold axes in the Matrei zone were progressively rotated towards the stretching direction during D_s, and that the steeply plunging fold axes represent localized areas of relatively low D_s strain. Rotation may be in either direction and the sense should depend primarily upon the angle between the original nucleation orientation of the fold axis and the extension direction during deformation. Unless all the minor structures rotated with the same sense, changes in vergence of these small-scale structures may not be a reliable indication of the presence of large folds. In the Matrei Zone there are a number of systematic changes in the vergence of small-scale structures along N-S traverses. The inference, that these reflected the position of larger-scale structures is supported by identifying areas of inverted stratigraphy and occasionally by tracing individual units around the large-scale fold hinges.

Ss forms the dominant fabric of the Matrei Zone and dips moderately steeply to the south. Ss has an associated generally weakly-developed stretching lineation, which follows the same NW-SE trend as seen in the base of the Altkristallin. However, in the vicinity of the large,

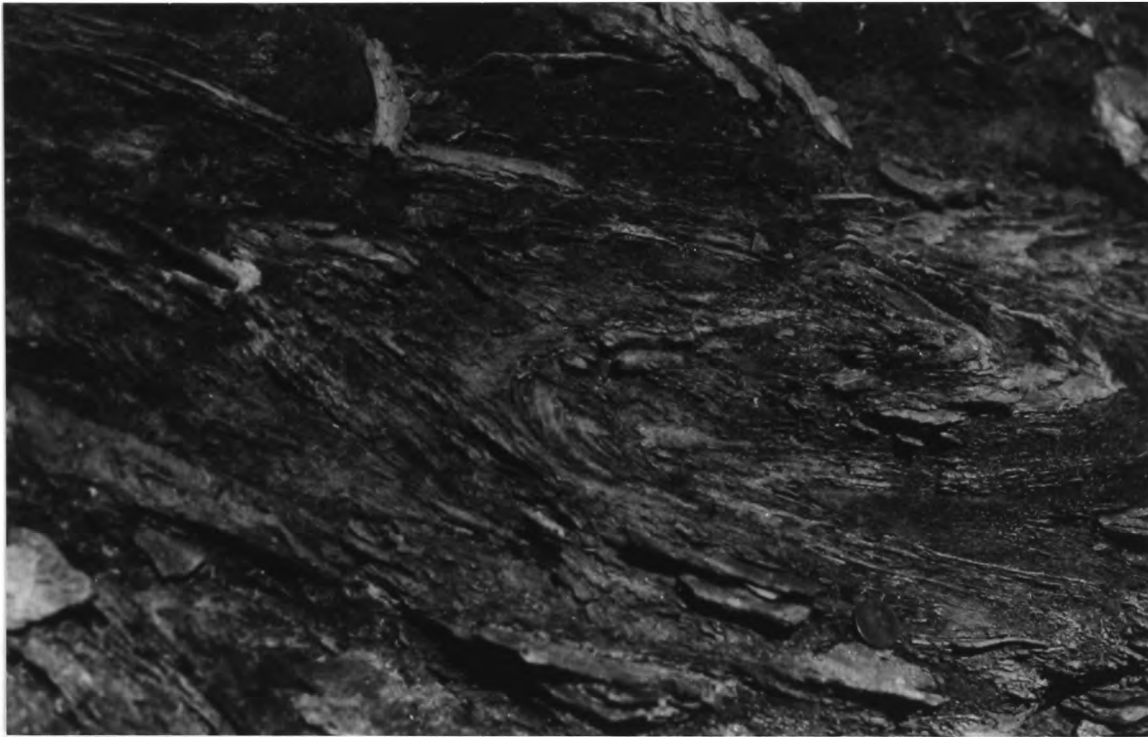


Fig.3.29. Tight Ds fold in calcareous schist of the Matrei Zone. The pre-Ds fabric is parallel to bedding. 100m N of Glorer Hütte, area 1.

competent dolomite lenses the mesoscopic features vary greatly in orientation, in particular the stretching lineation (fig.3.30). In area 3 north of the Bretterich Kamm, Ds is strongly overprinted by a later phase of deformation, Dt, which transposes Ss (section 3.13.2).

The Ds tectonites within the Matrei Zone are generally $S < L$ which might imply deformation in the flattening field, although the weak development of the stretching lineation could also be explained by deformation being concentrated along discrete high-strain horizons. The stretching lineation is best seen by the development of pressure shadows around rigid inclusions, e.g. opaques or feldspar. The prominent quartz-rodding present in Altkristallin is only rarely developed in the Matrei Zone. Extensional crenulations are only locally well-developed and their geometries correspond with those already described for the overlying Altkristallin (section 3.4.1.).

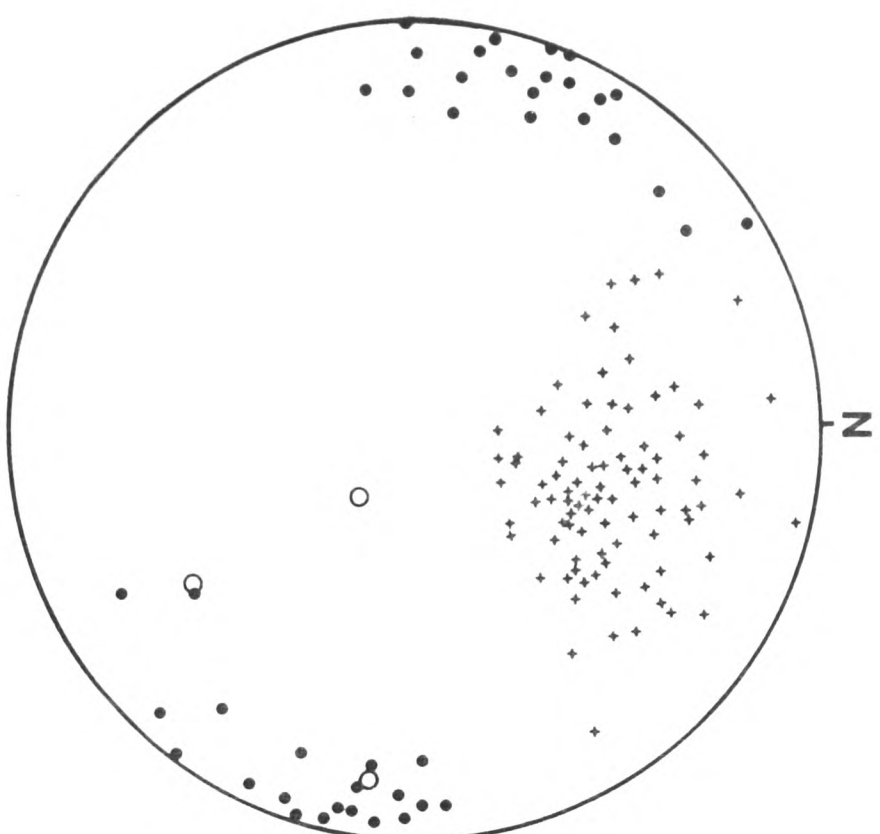
3.10.2 Ds Microstructure

Apart from a few relic minerals of higher grade metamorphism from the slivers of basement material, the Matrei Zone shows entirely lower greenschist facies metamorphism. The platy Ds foliation is defined by the preferred orientation of chlorite and white mica. Sense of shear indicators are rarely present but locally microshears cutting the Ss foliation indicate a general top to the NW movement. This sense of shear is supported by crystallographic preferred orientation patterns in deformed quartzites (fig.4.9, fig.4.11).

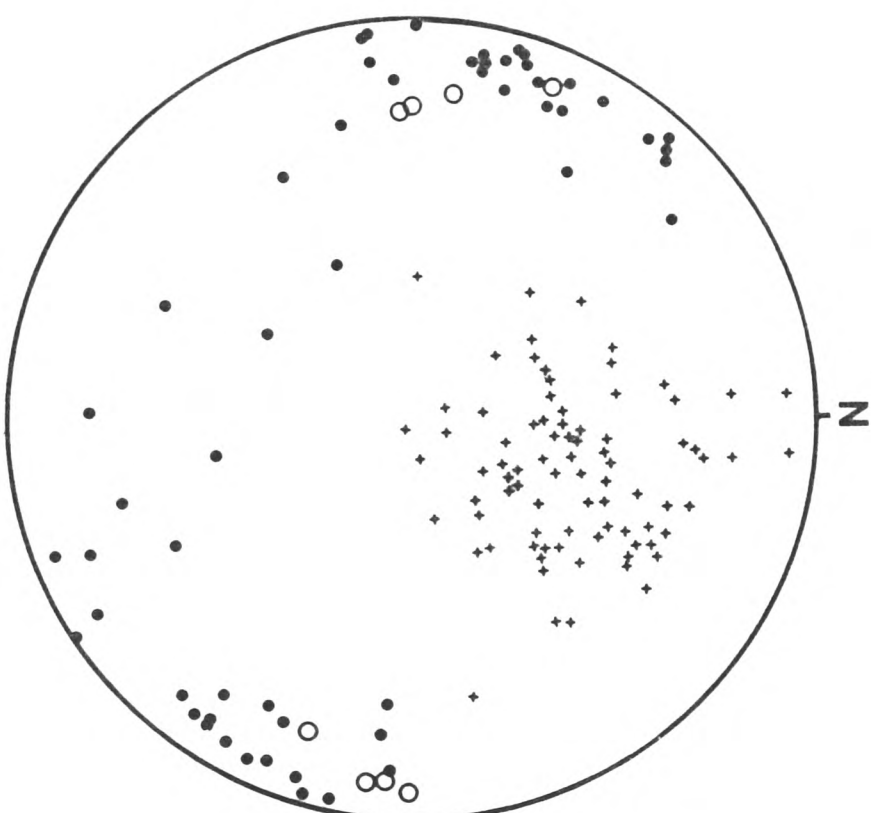
In the quartzites porphyroclasts of feldspar with inclusions of white mica and zoisite (gefüllte texture) are commonly fractured during Ds. The gefüllte texture probably indicates retrogression from a higher grade of metamorphism and may have been inherited from the erosion of a basement source area or developed during Ds. In places the feldspar has been overgrown by a rim of clear albite which can be related to an

Fig.3.30. Ds mesoscopic orientation data from the Matrei Zone plotted on Schmidt nets for map areas 1 and 3. In both map areas, subarea II is associated with large incompetent bodies of dolomite that cause large local variation in the mesoscopic orientation data, particularly the stretching lineation. + = pole to Ss, • = stretching lineation, o = fold axis. Each symbol represents an average of 3 to 8 readings for the schistosity and 2 to 4 readings for the stretching lineation.

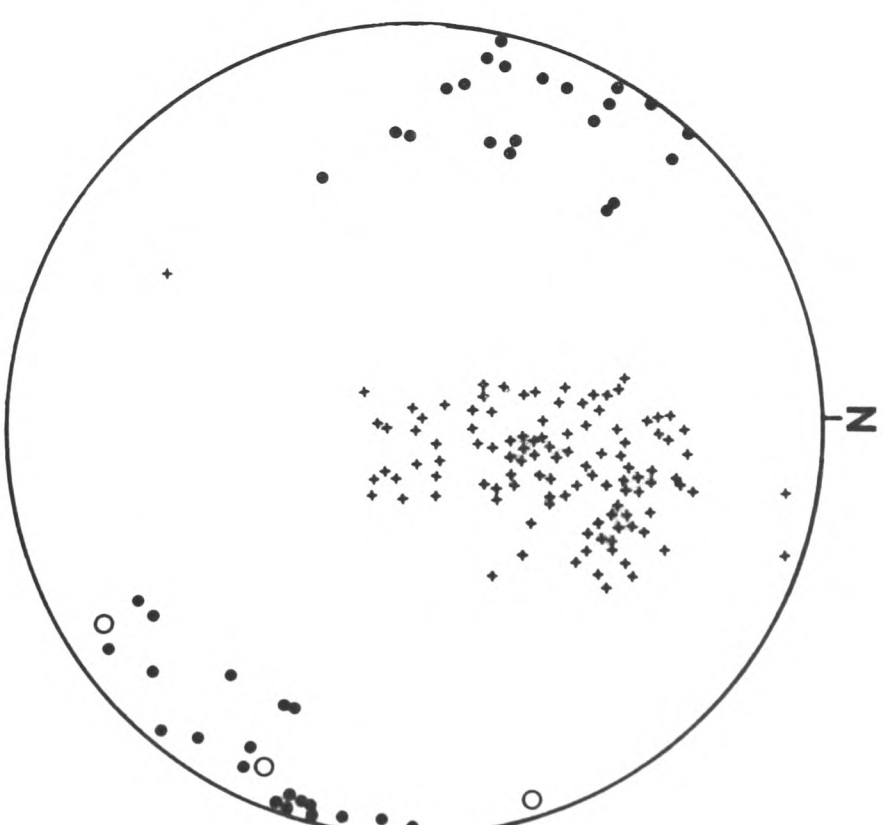
AREA 3



subarea I

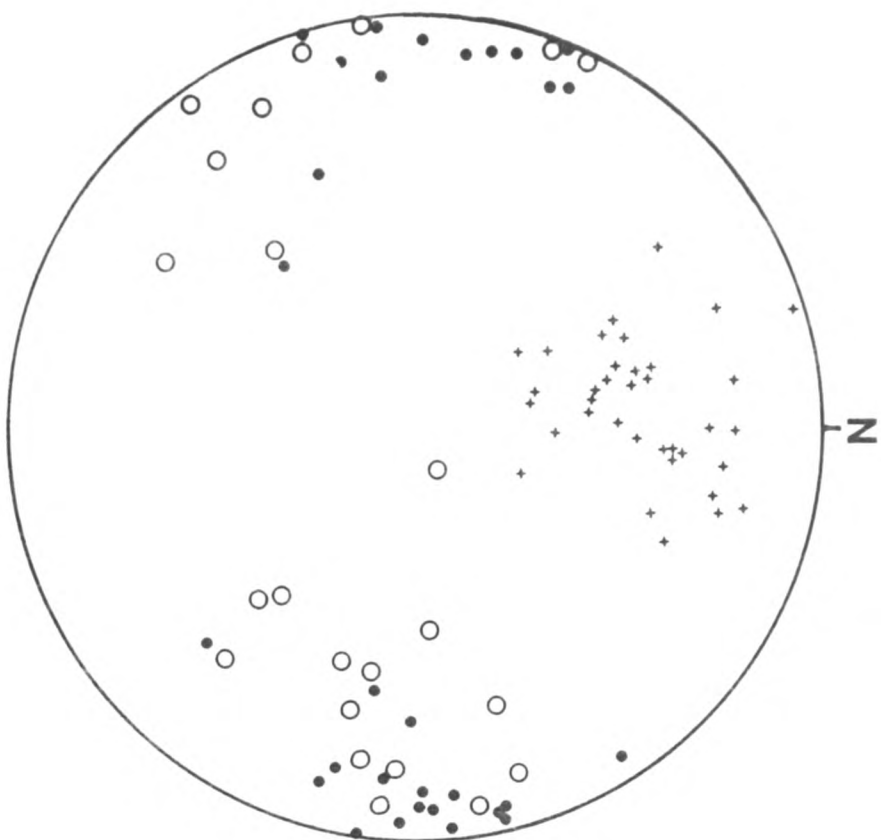


subarea II

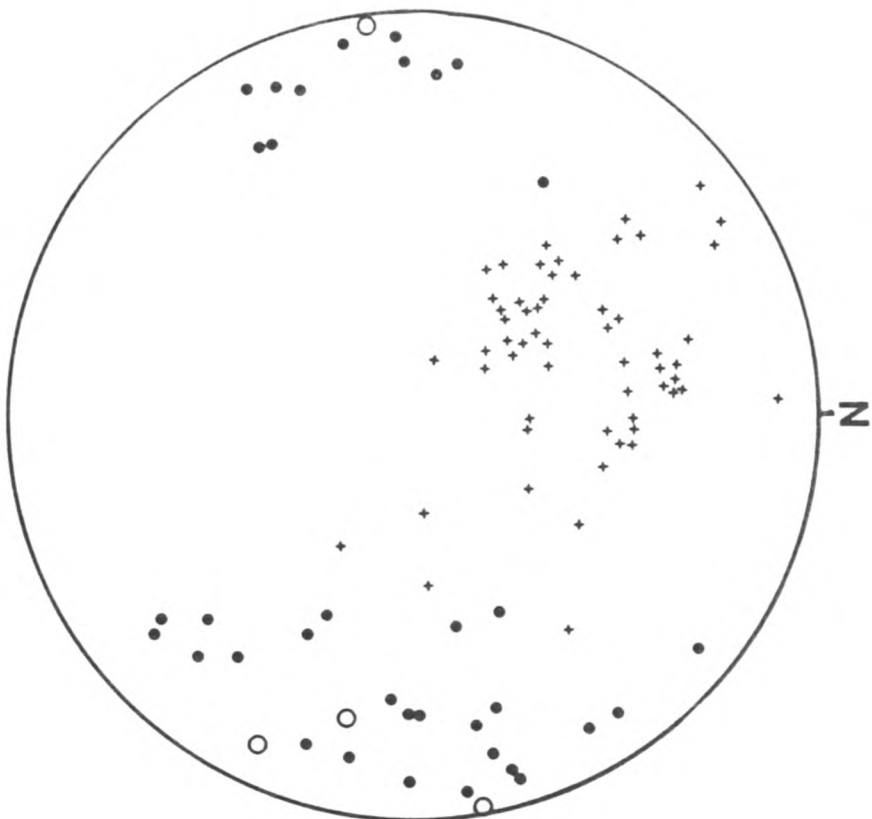


subarea III

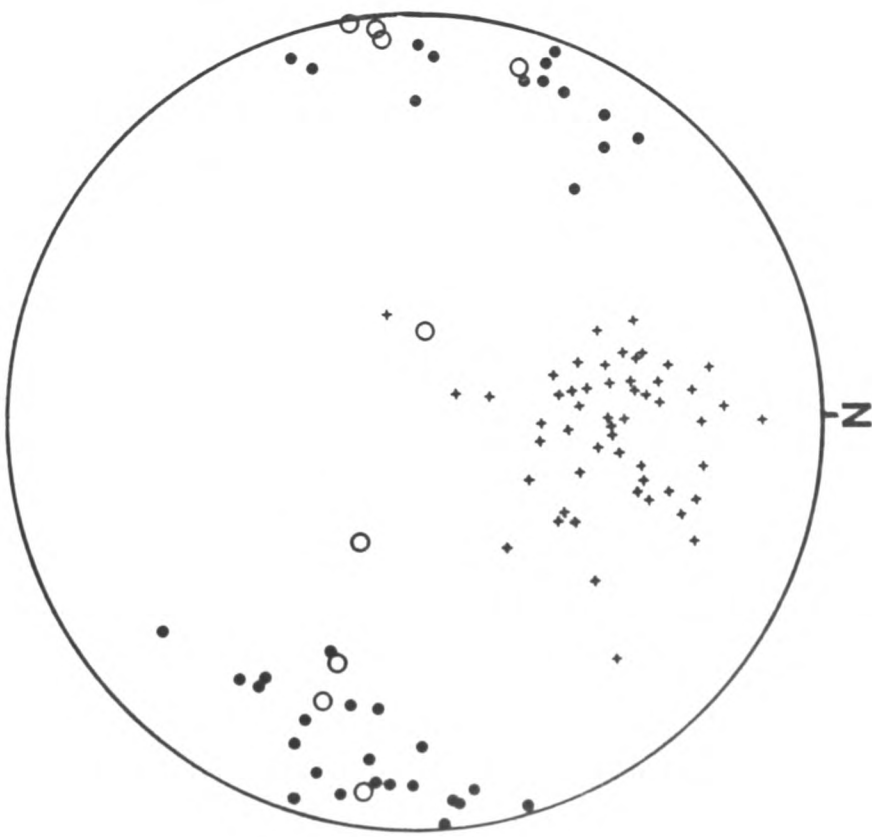
AREA 1



subarea I



subarea II



subarea III

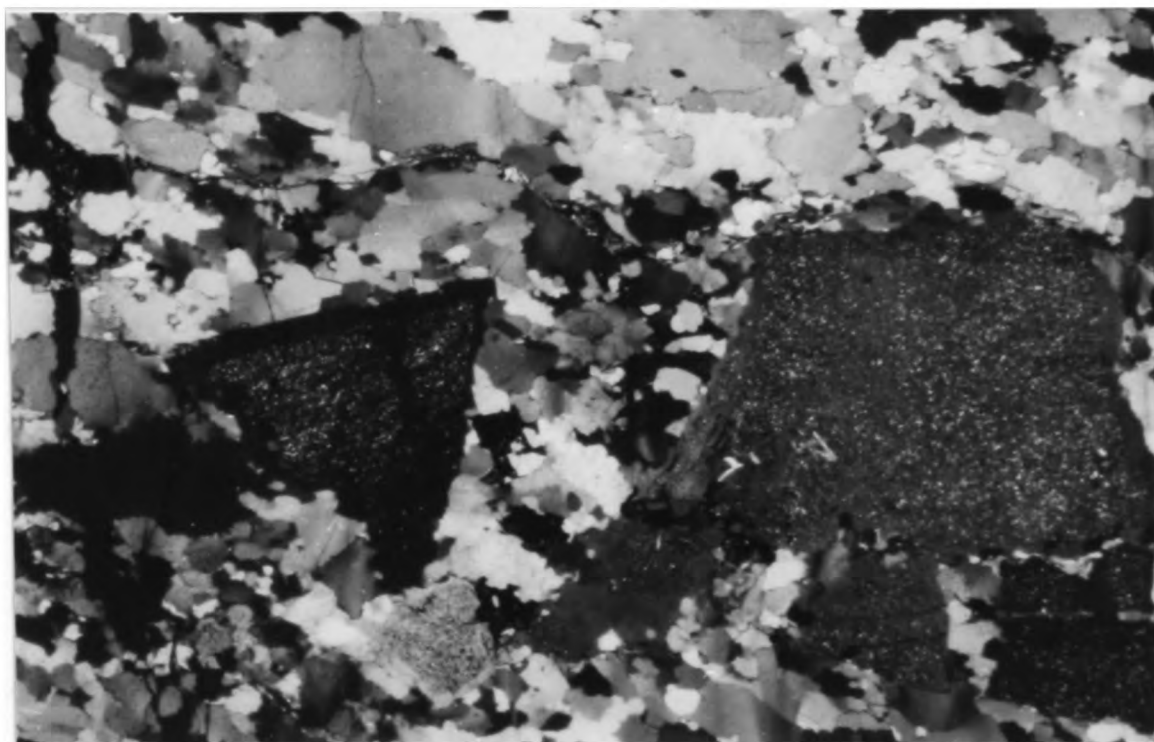
overgrowth during the Tauern Metamorphism (fig.3.31). In some cases feldspar that may have been fractured and then welded by newly grown albite can be identified.

In the Mafrei Zone the fabric deformed during Ds is always subparallel to bedding. This is a pre-Ds metamorphic fabric, which can be particularly clearly seen in the meta-greenschists, where layers of zoisite crystals are folded around tight Ds folds (fig.3.32). This early metamorphic fabric is unlikely to have formed parallel to bedding, and the two planar fabrics were probably rotated into parallelism with increasing strain.

3.10.3 Ds Deformation Mechanisms

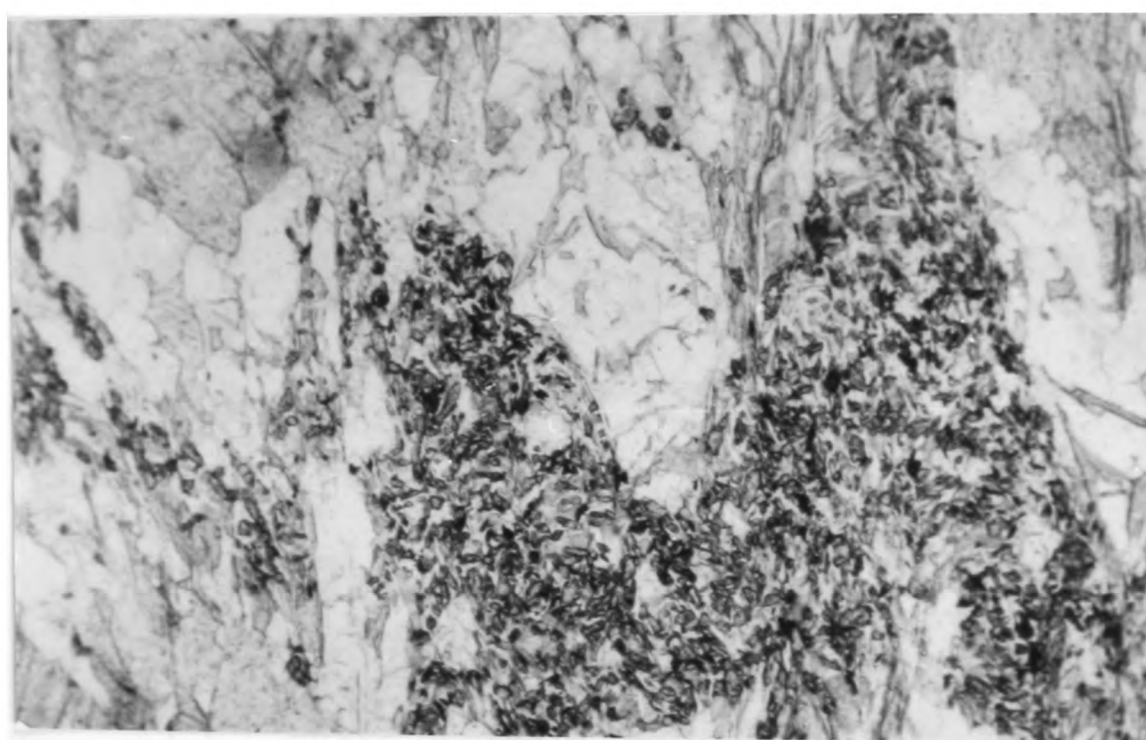
Most of the Ds microstructures in the Mafrei Zone were at least partially annealed during the Tauern Metamorphism, which obscures many of the original deformational features. However, within some of the quartz-rich rocks, large quartz porphyroclasts are present, which have strong undulose extinction and subgrain development (fig.3.33). This suggests the large grains, and by inference the surrounding smaller grains, were deformed by intracrystalline plasticity, and is supported by the presence of a strong crystallographic preferred orientation (e.g. B69).

The energy for recrystallization is mainly stored in the form of internal or grain-boundary dislocations. During dynamic recrystallization the intragranular energy in the form of unbound dislocations is an order of magnitude greater than grain boundary energy (Urai et al 1986). After deformation has ceased there is commonly a phase of static recrystallization (annealing). This can be divided into a primary phase which produces essentially strain-free grains, and a secondary phase which involves grain growth and the formation of 120° triple junctions. This division into two different stages of



2mm

Fig.3.31. Albite with a core containing abundant inclusions (gefüllte texture) and a clear rim. This can be interpreted as an overgrowth of clear albite on older retrogressed crystals during the Tauern Metamorphism. The clear seam dividing the triangular domains of feldspar with a gefüllte texture in the left-hand albite grain may represent a healed fracture. MS40, area 3.



1mm

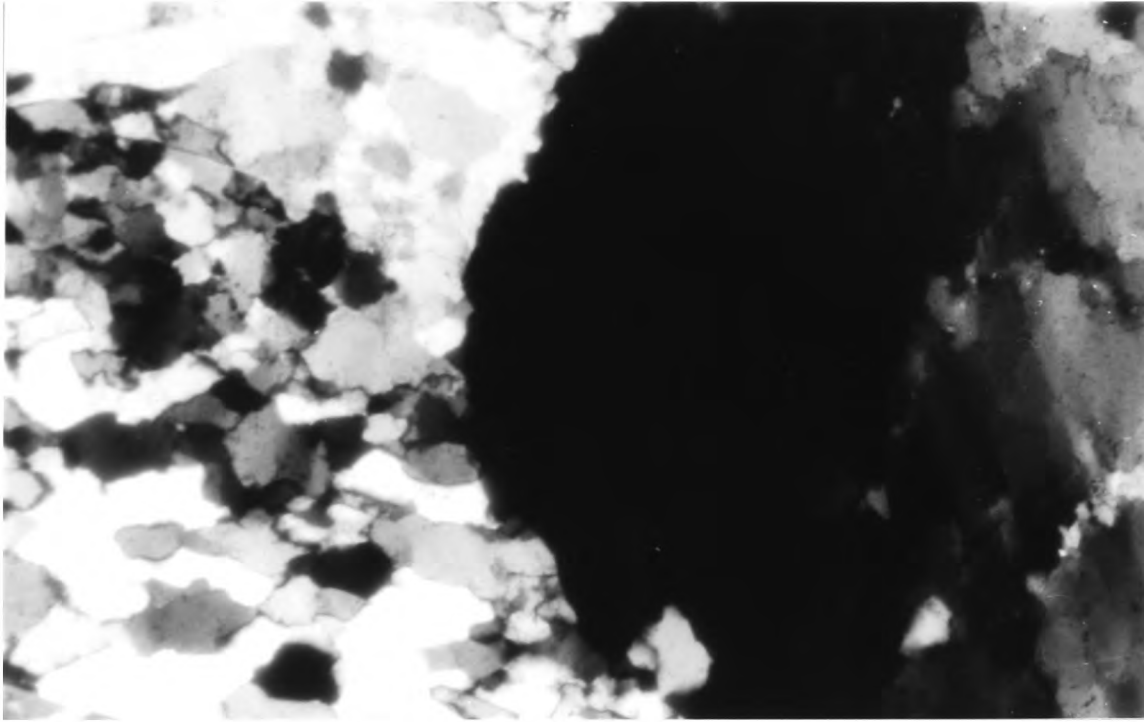
Fig.3.32. A tightly folded greenschist layer composed dominantly of chlorite and zoisite, within a quartzitic marble. The long axes of the zoisite crystals follow the folded bedding and suggest that this is a pre-Ds metamorphic fabric, which is sub-parallel to bedding. Bretterrich Kamm, area 3.

recrystallization corresponds to a change in the dominant driving force for recrystallization, from intragranular lattice defect energy to grain boundary energy.

Finer-grained material has a higher grain boundary to volume ratio. This is reflected in the finer-grained quartz-rich rocks of the Matrei Zone, by their greater susceptibility to secondary recrystallization. In the larger quartz porphyroclasts intragranular dislocations have migrated and become aligned to form well-defined subgrain boundaries enclosing relatively strain-free subgrains. There was, however, insufficient stored energy for new high-angle grain boundaries to develop within the porphyroclasts. The grain boundary energy stored around these large quartz grains was also insufficient to cause significant grain-boundary migration and the form of the old grains is still clearly visible (fig.3.33).

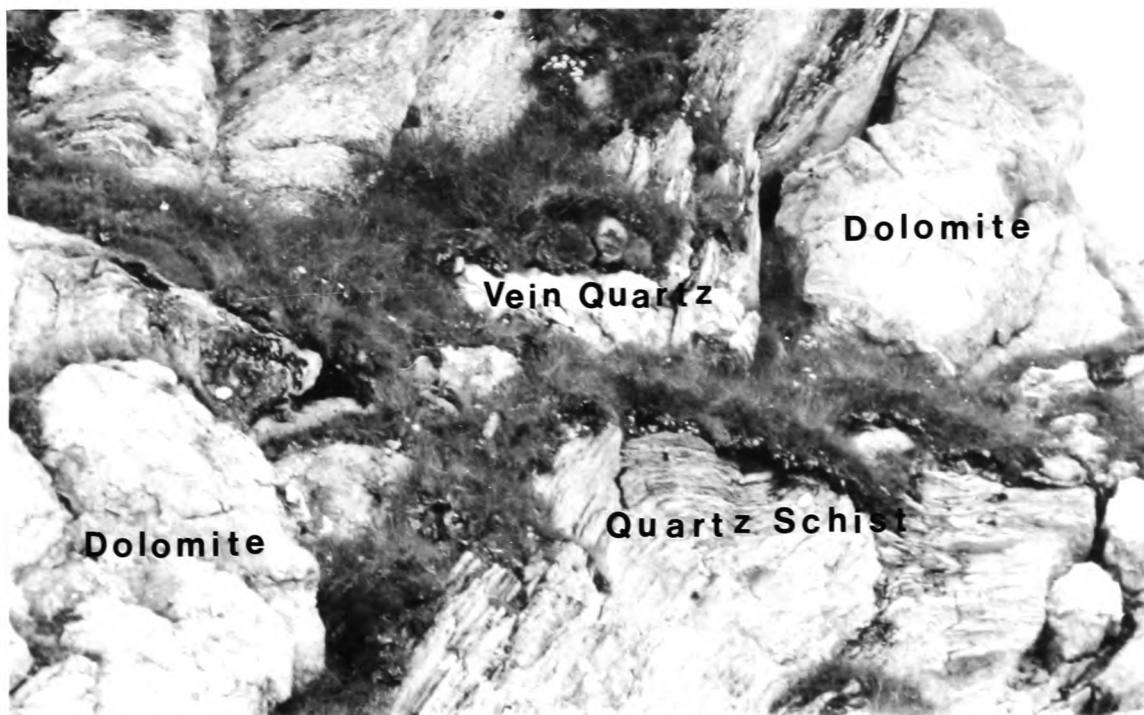
A particular feature of area 3, and to a lesser extent area 1, is the presence of strongly deformed dolomite. Some of the dolomite lenses have behaved competently, with fracturing and boudinage taking place during Ds (fig.3.34). In other cases dolomite has an obvious foliation (fig.3.35a) which is defined by a bimodal distribution of grain sizes preferentially partitioned into different layers (fig.3.35b.). The larger crystals of dolomite show few signs of twinning or other indications of crystal plastic deformation and no grain-shape fabric. Dolomite also occurs as thin bands of calc mylonite only a few centimetres thick. These strongly deformed layers consist of an equigranular fine-grained matrix ($\approx 10\mu\text{m}$), which only rarely contains larger dolomite porphyroclasts (fig.3.36).

The most likely explanation for these observations is that deformation occurred by a combination of diffusional transfer - either along grain boundaries or through the crystals - and grain boundary sliding. If diffusional transfer is the dominant strain accommodation



1mm

Fig.3.33. Quartz porphyroblast within a finer-grained matrix. Annealing due to the Tauern Metamorphism has produced straight extinction and relatively straight grain boundaries in the finer-grained quartz. In the porphyroblast, annealing has produced well-defined internal subgrains but only minor development of new grains. B69, area 1.



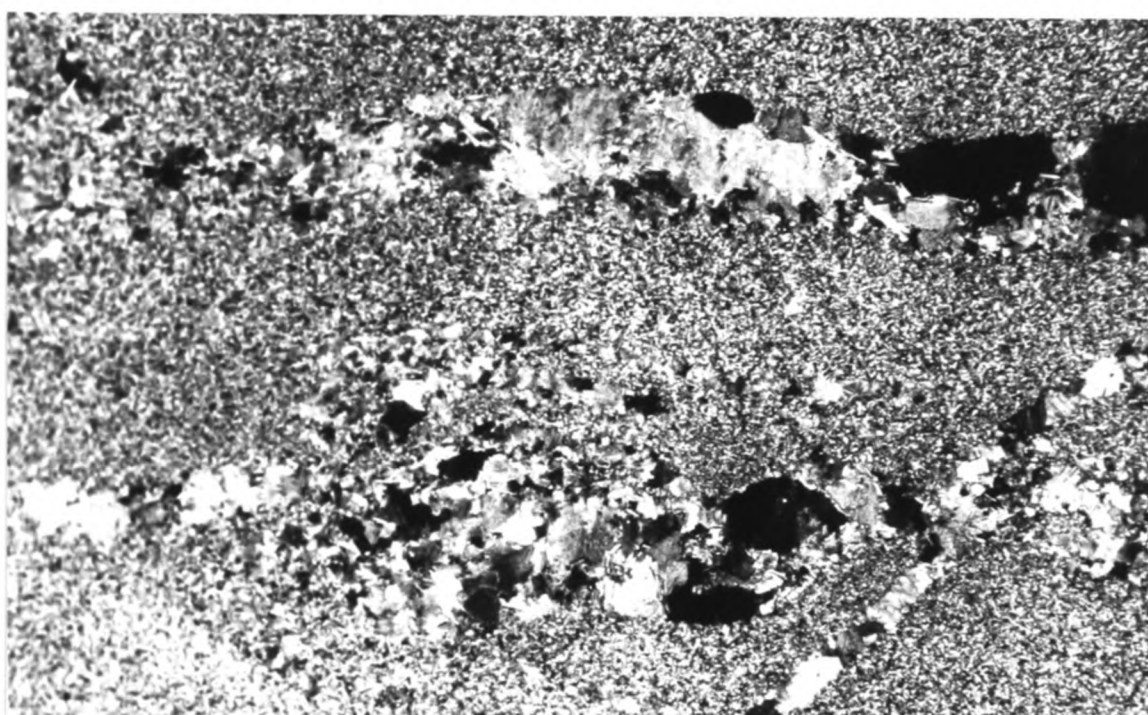
1m

Fig.3.34. Dolomite boudins surrounded by quartz phyllite with vein quartz in the zone of fracturing. Makerni Spitze area 3.



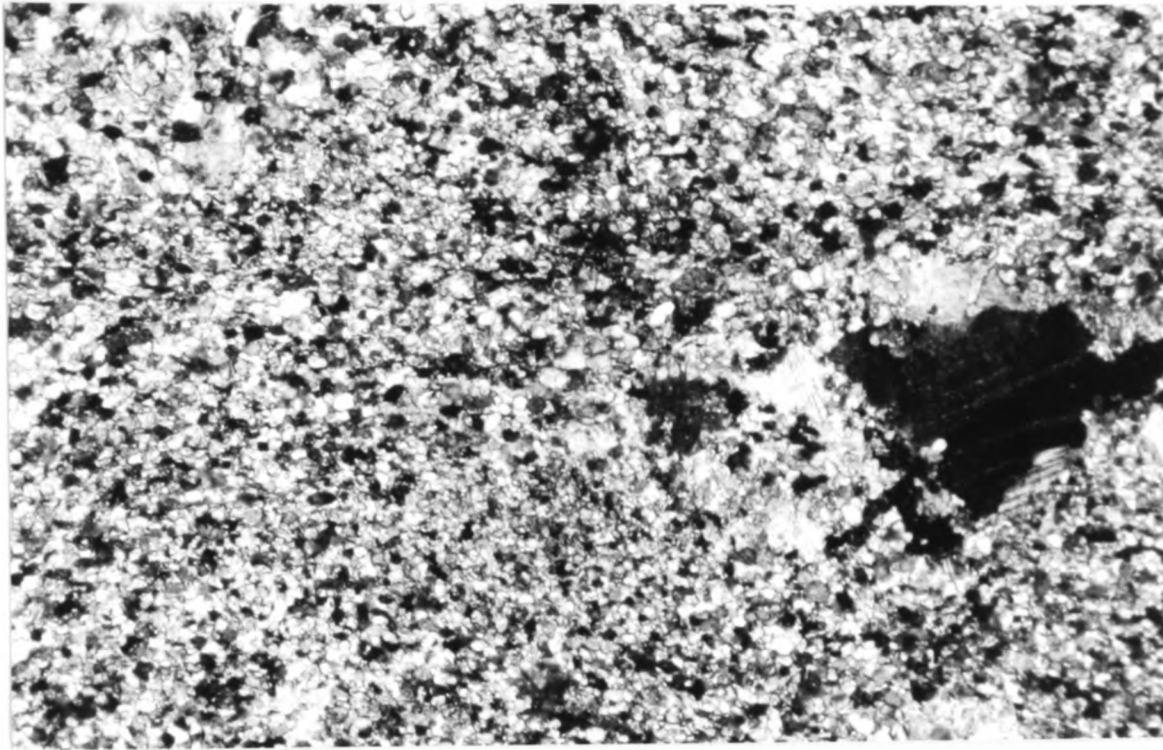
10cm

Fig.3.35a. Foliated dolomite, Makerni Spitze area 3.



1mm

Fig.3.35b. Foliation in the dolomite is defined by a bimodal grain size preferentially partitioned into different layers. Photomicrograph of sample from Makerni Spitze mountain side.



1mm

Fig.3.36. Strongly deformed dolomite with a well-developed mesoscopic foliation. The rock is dominantly fine-grained but locally contains porphyroclasts. These grains show little sign of crystal plastic deformation and grain-size reduction was probably achieved by diffusional processes. MS38, Makerni Spitze, area 3.

process the behaviour would be termed creep, if grain boundary sliding then superplasticity (Poirier 1987, ch.7). Since evidence for intracrystalline plasticity is lacking, the initial grain size reduction necessary to enhance the diffusional processes was probably achieved by recrystallization caused by pressure solution processes.

The marbles in the area have mostly been annealed during the Tauern Metamorphism and the presence of deformational features is usually related to deformation postdating Ds. However, in some layers a characteristic millimetre-scale banding is developed parallel to Ss. This banding is, like the foliated dolomites, related to a grain-size partitioning. The coarser layers are composed of calcite grains, which are interlayered with much finer-grained calcareous bands. Closer inspection reveals the finer layers are a two-phase mixture of calcite and dolomite. In polyphase aggregates boundaries between minerals of the same phase tend to have higher energies than those between grains of different phases. Therefore, the two phase mixture is less likely to be affected by secondary grain growth (Vernon 1983, p.139). The difference in grain size of the two layers is, therefore, probably a secondary effect produced by late stage annealing of an originally fine-grained tectonite.

A small grain-size implies there is a relatively large area of grain boundary per unit volume of rock, which is favourable for diffusional transfer of material. The relatively easy diffusion through the rock allows the material to stay in chemical equilibrium with changing metamorphic conditions. The presence of both dolomite and calcite in the finer grained layers and the favourable conditions for maintaining chemical equilibrium make these rocks suitable for calcite-dolomite geothermometry (Bickle & Powell 1977).

3.10.4 Ds Deformation in the Mallnitzer Mulde and Sonnblick Dome

Some attempt was made to follow the Ds deformation to the north of area 3 into the Mallnitzer Mulde, through the orthogneiss of the Sonnblick Dome. This involved following the Ds fabric through massive orthogneiss and no continuous profile could be traced. However, a NW-SE stretching lineation is developed throughout the profile and is associated with deformation that predates the Tauern Metamorphism. This deformation forms the dominant fabric of the Mallnitzer Mulde and I correlate this with Ss.

In the Mallnitzer Mulde the Ds folds are more open than in the Matrei Zone and the stretching lineation is obvious throughout the area. The Sonnblick Gneiss dome and the Mallnitzer Mulde are in the form of a large synform - antiform pair (Droop 1978, Exner & Prey 1964). The grade of the Tauern Metamorphism increases steadily from south to north across these structures and clearly overprints their formation (Droop & Cliff 1985). These could be large-scale Ds folds and there does appear to be a corresponding change in the vergence of small-scale Ds structures across the Mallnitzer Mulde. One large (30m) fold was found in the orthogneiss of the Sonnblick dome (fig.3.37) with a NW trending fold axis and stretching lineation. This is assigned to the Ds deformation and its presence suggests the Sonnblick gneiss also underwent strong Ds deformation. This deformation is, however, usually difficult to recognize due to the nature of the lithology.

The units either side of the Mallnitzer Mulde cannot be matched up exactly and this cannot represent a simple structure produced by one phase of deformation. An earlier phase has to be invoked to account for the pre-Ds geometry (Exner & Prey 1964). The fabric deformed by Ds in the Mallnitzer Mulde is an earlier metamorphic fabric and in places an north-south lineation is folded by Ds. This may represent a pre-Ds stretching lineation.



Fig.3.37. Large-scale Ds fold within the orthogneiss of the Sonnblick Dome. Steve Reddy for scale, Astrom Scharte.

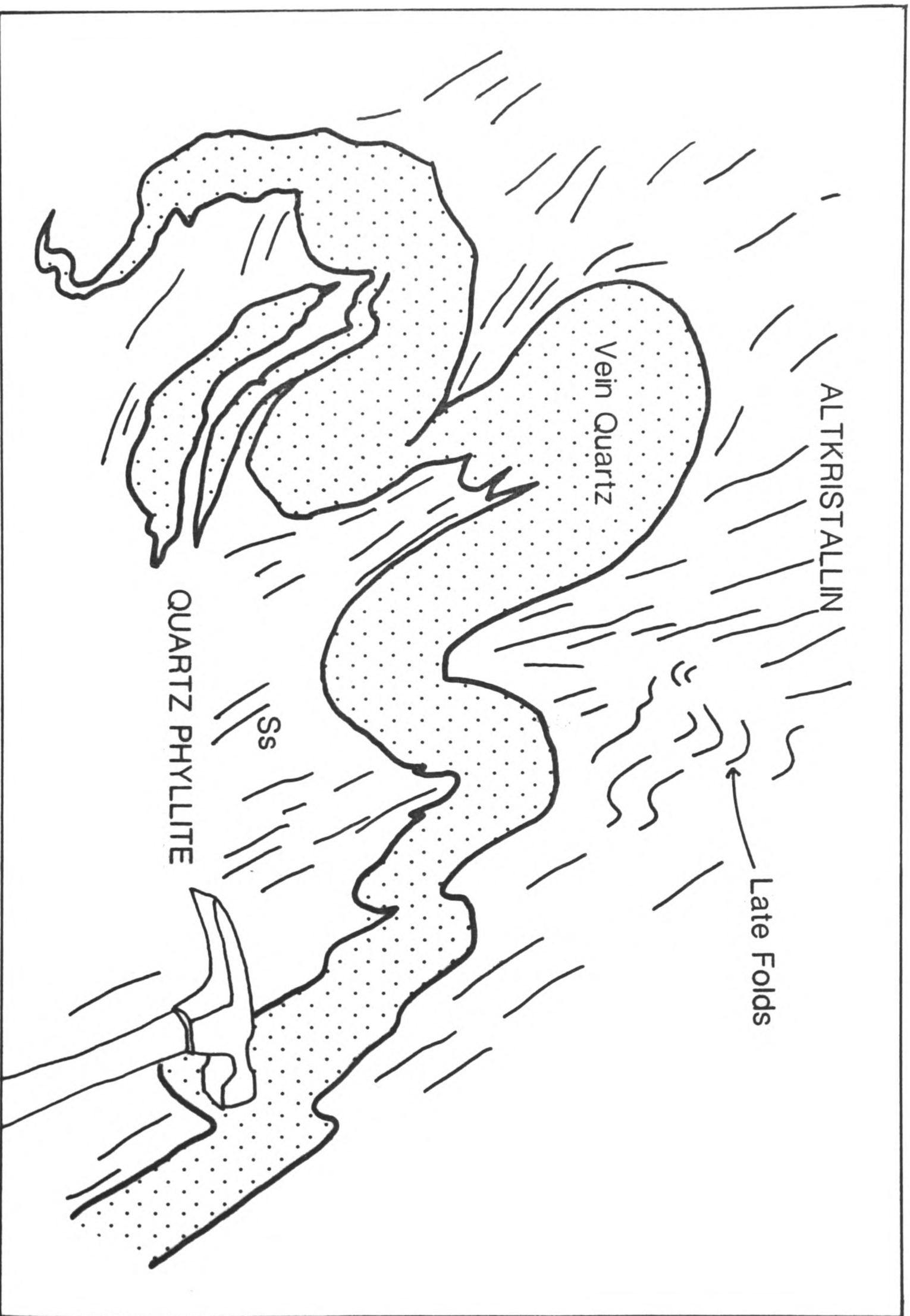
3.11 PRE-Ds DEFORMATION IN THE MATREI ZONE

3.11.1 Recognition of thrusts

Repetitions of the stratigraphic sequence reveal thrusts both within the Matrei Zone and along the contact between the Altkristallin and Matrei Zone sediments. In general, the thrust planes are subparallel to the main Ss schistosity. However, in a few areas it is possible to find overprinting relationships between the thrusts and Ds deformation. Thrusts at localities B114 and B61 separate Matrei Zone quartz phyllite from retrogressed Altkristallin. In both cases the tectonic contact has been folded during Ds (fig.3.38). This demonstrates that the thrusting predates Ds. Some thrust planes have high-strain Ds mylonites along them, these thrusts may have been reactivated or developed during Ds.

It is rarely possible to trace the thrust surfaces even a short distance into the Altkristallin and the relationship between the Mr mineral assemblages and the thrusts was not directly observed. However samples of Altkristallin taken from between the imbricates of quartz phyllite showed strong retrogression of the Mr metamorphic minerals. This suggests that thrusting postdates Mr. Bickle (1973) suggested that the relatively large grain size of the white micas in some of the white quartz schists imbricated with the Altkristallin were produced by an Alpine amphibolite grade metamorphism equivalent to Mr. Beyond this, however, there is no evidence to suggest that these rocks underwent the Mr metamorphism seen in the Altkristallin. The large micas in the white quartz schists can be readily explained as sedimentary grains derived from a pre-Alpine metamorphic basement or due to pre-Ds metamorphism within the Matrei Zone, implying that the imbrication took place after Mr.

Fig.3.38. Vein quartz marks the boundary between retrogressed Altkristallin above and quartz phyllite of the Matrei Zone below. This original thrust contact has been deformed during Ds, producing the tight folds and the dominant foliation. This clearly shows the relative ages of Ds and the brittle thrusting.



The strong shortening represented by the pre-Ds thrusting within the Matrei Zone almost certainly produced a mesoscopic fabric. Although no direct geometric relationships are seen between the imbricates and the pre-Ds metamorphic fabric in the Matrei Zone it is likely that the two are genetically linked. No critical mineral assemblages are preserved associated with the earlier metamorphic fabric. However, studies elsewhere in the Tauern Window have shown that the earliest metamorphism of the area was dominated by high P/T conditions (Droop 1983, Selverstone et al 1984, Holland & Ray 1985, Franz & Spear 1983). In the Matrei Zone there is very little evidence about the metamorphism prior to the development of Ds. One possible lawsonite pseudomorph was found (fig.3.39); MS43 may contain pumpellyite (within the black clasts of the white quartz schists), and Cornelius and Clar (1939, p.246) found one loose block of glaucophane + aegerine bearing metabasic. All these observations might indicate an early phase of high P/T metamorphism.

Convergence and subduction in the Eastern Alps probably began around 130Ma (see chapter 5). However, the arguments above suggest that the contact between the Matrei Zone and the Altkristallin was formed after Mr ($\approx$ 80Ma). Deformation within the Matrei Zone, therefore, probably began before the imbricates developed along the boundary with the overlying Altkristallin implying that more than one phase of early deformation is present within the Matrei Zone.

3.11.2 Thrust Geometries

In the Matrei Zone the thrusts are particularly clearly seen in the Makerni Spitze mountain side (area 3), where a dolomite-quartzite mass has been thrust and then tightly folded during Ds (fig.3.40c.). However, the thrusts are too poorly defined within the Matrei Zone to allow any reconstruction of the original three dimensional geometries. Along the contact between the Altkristallin and the Matrei Zone the

Fig.3.39. Rhombohedral pseudomorph possibly after lawsonite, composed of zoisite and minor amounts of calcite developed within a greenschist. Bl27, area 1.

5mm

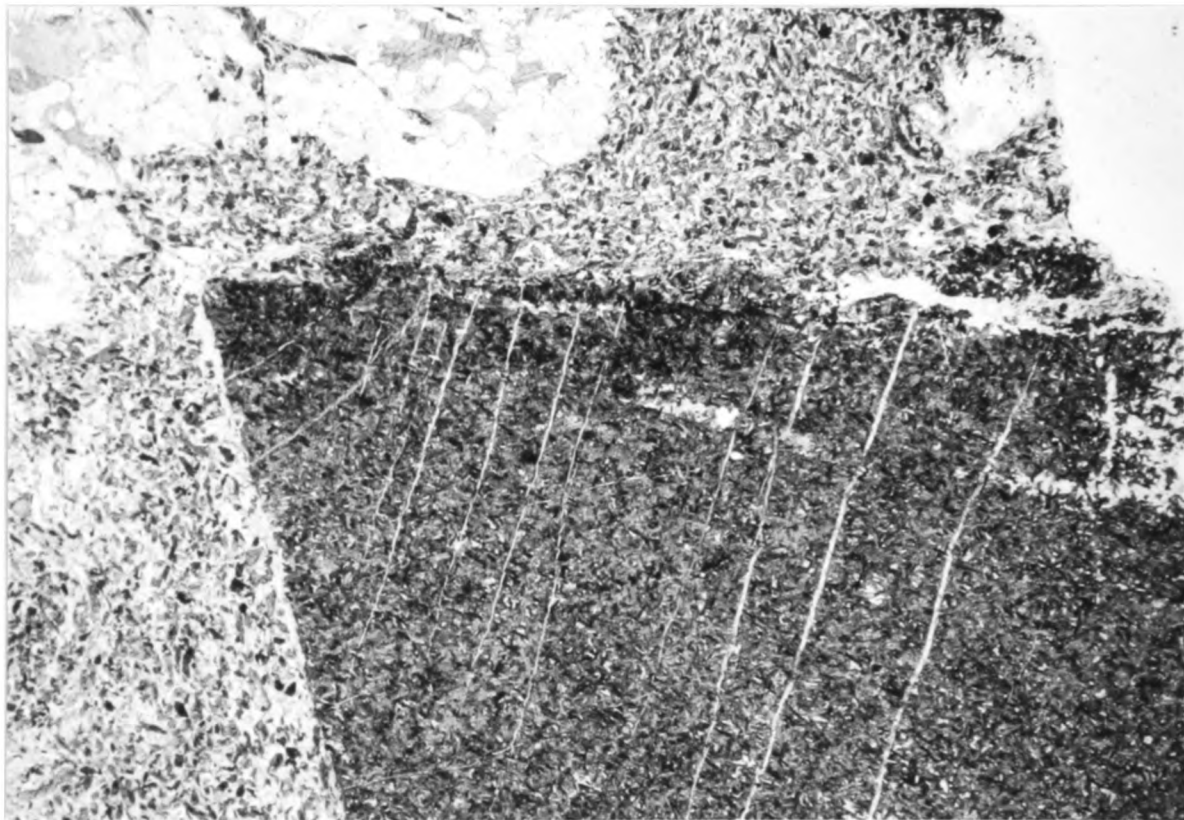


Fig.3.40. NNE-SSW cross-sections through the Matrei Zone in area 3, all to the same scale.

a) Imbrication of the quartz phyllite sheet with the overlying Altkristallin. The Bündnerschiefer and continental fragments between the Sonnblick gneiss and the quartz phyllite sheet, which are seen in section L - M, do not appear in this profile.

b) Representative section through the Makerni Spitze mountain, with the main thrusts marked on for reference with the structural section in fig.3.40c. The Matrei Zone can be divided into an upper Austroalpine quartz phyllite sheet and a lower Pennine unit, which is dominantly composed of Bündnerschiefer. These two units are separated by a major tectonic discontinuity. The upper contact of the quartz phyllite sheet with the Altkristallin also represents a major thrust contact. The complex structure of the Makerni Spitze mountain-side is shown in more detail in fig.3.41.

c) Mesoscopic structures of section L - M. The earliest structures which can be recognized in the area are thrusts which are present both within the Matrei Zone and along the contact between the Matrei Zone and the Altkristallin. The original imbricate stack has been deformed by a penetrative deformation Ds, which forms the main foliation of the section. This foliation, Ss, can be traced continuously from the Altkristallin into the Matrei Zone and is overprinted by three phases of deformation Dt, Du and Dv. Only Dt forms a persistent new foliation, St.

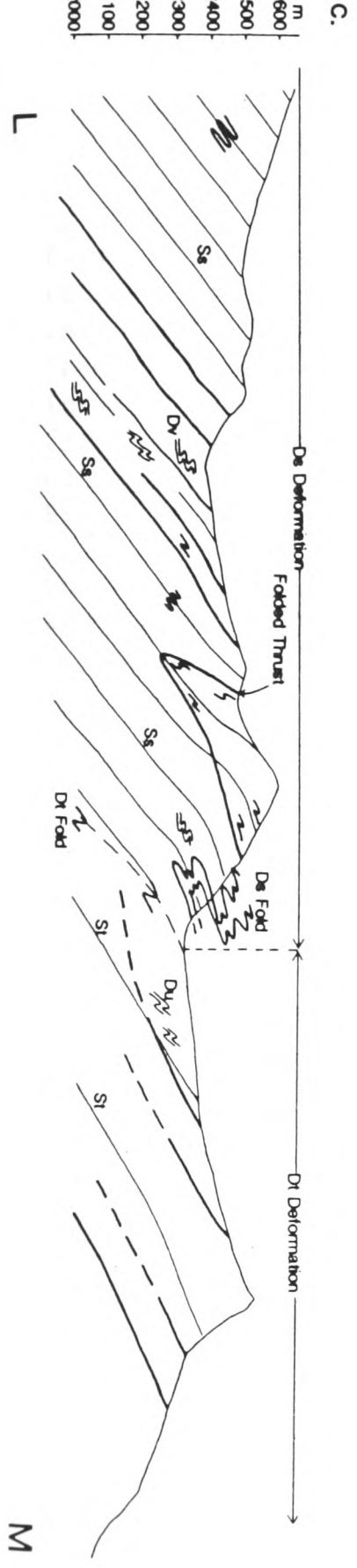
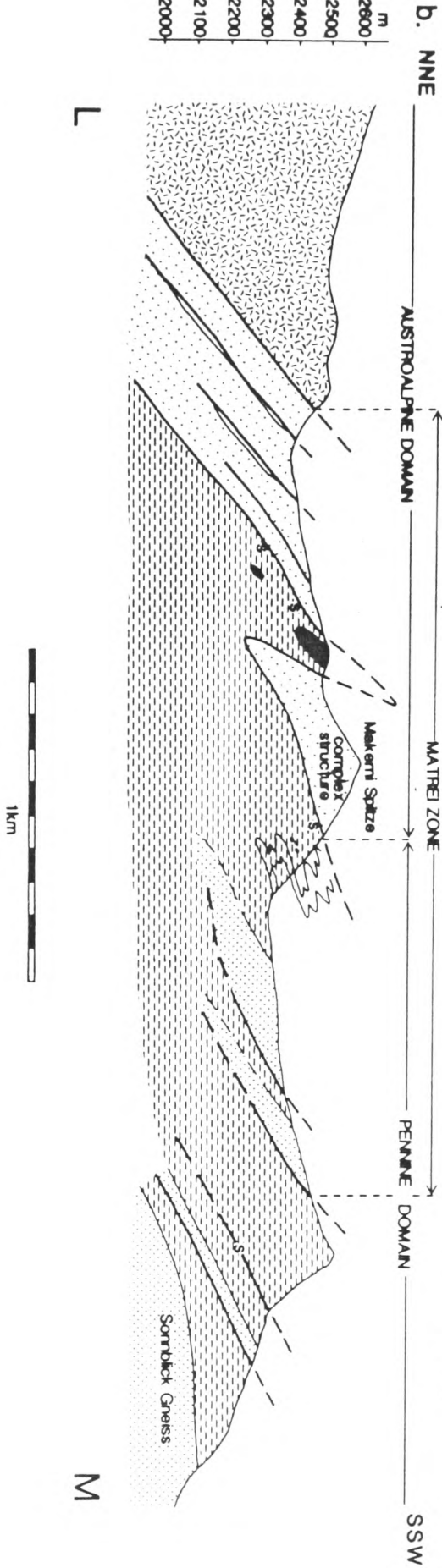
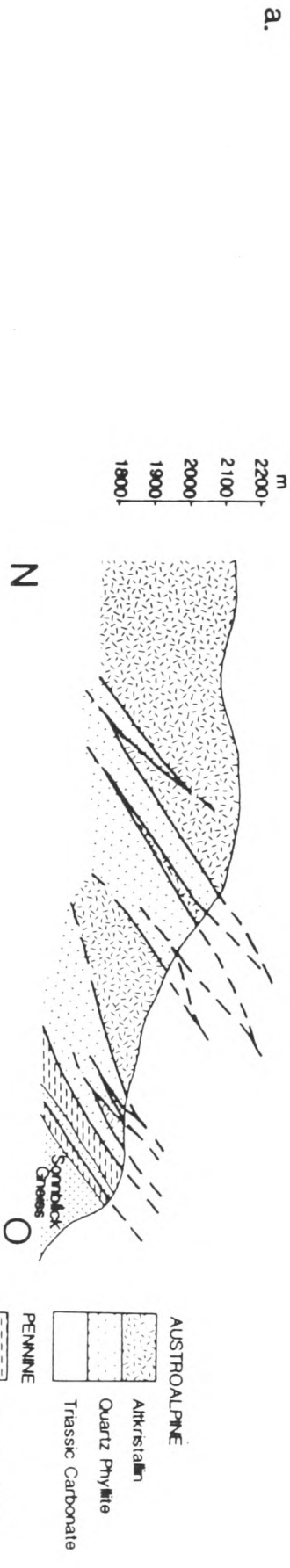
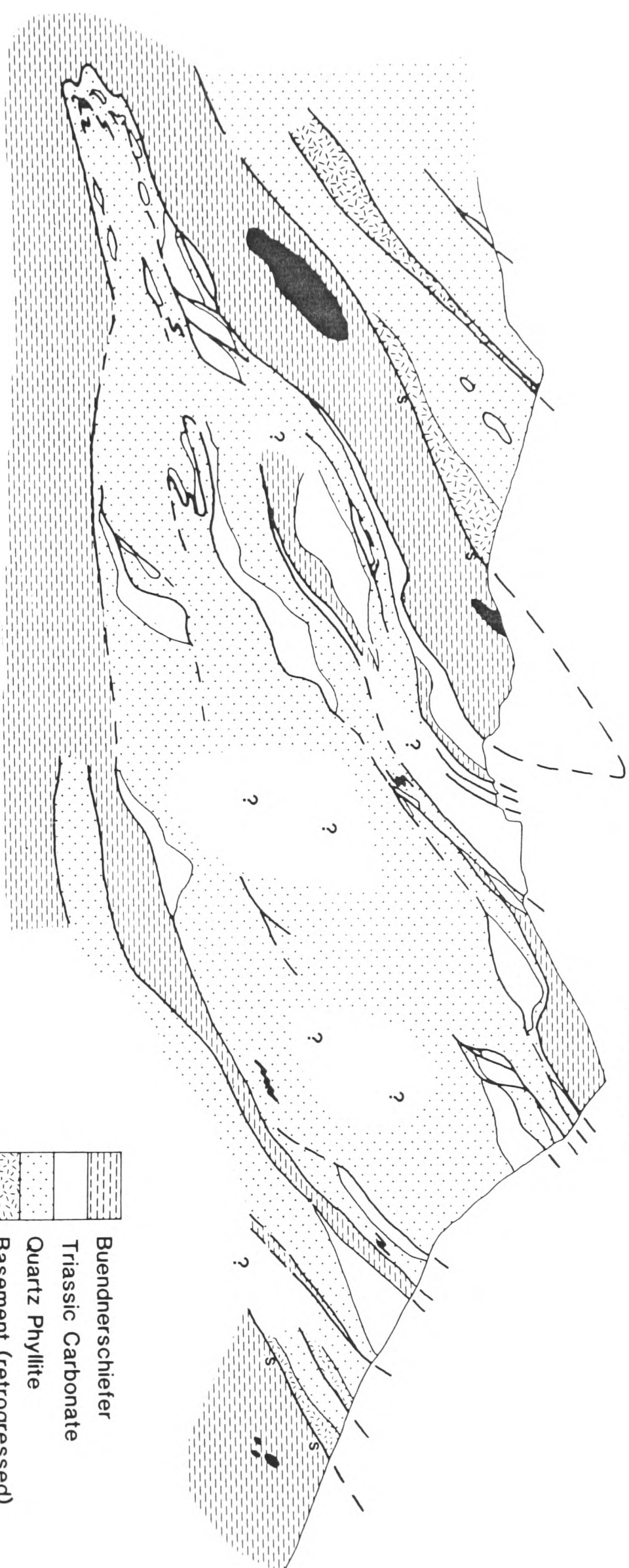


Fig.3.41. Detailed cross-section through the Makerni Spitze, oriented roughly north - south, based on mapping onto field photographs. The large dolomite olistolith has been included by projection along strike from the west. The main unit of dolomite and quartz pyllite is interpreted as a folded Klippe of the quartz phyllite sheet (see fig.3.40), whose base is a major tectonic boundary and is decorated with lenses of serpentinite. The two narrow imbricates of Bündnerschiefer along the northern slope of the mountain probably represent a late-stage imbrication of the original main thrust contact. The complex geometries of the thrusts within the main unit of dolomite and quartz phyllite may represent a movement direction oblique to the section and/or a polyphase thrusting history.

S

N

Makerni Spitze



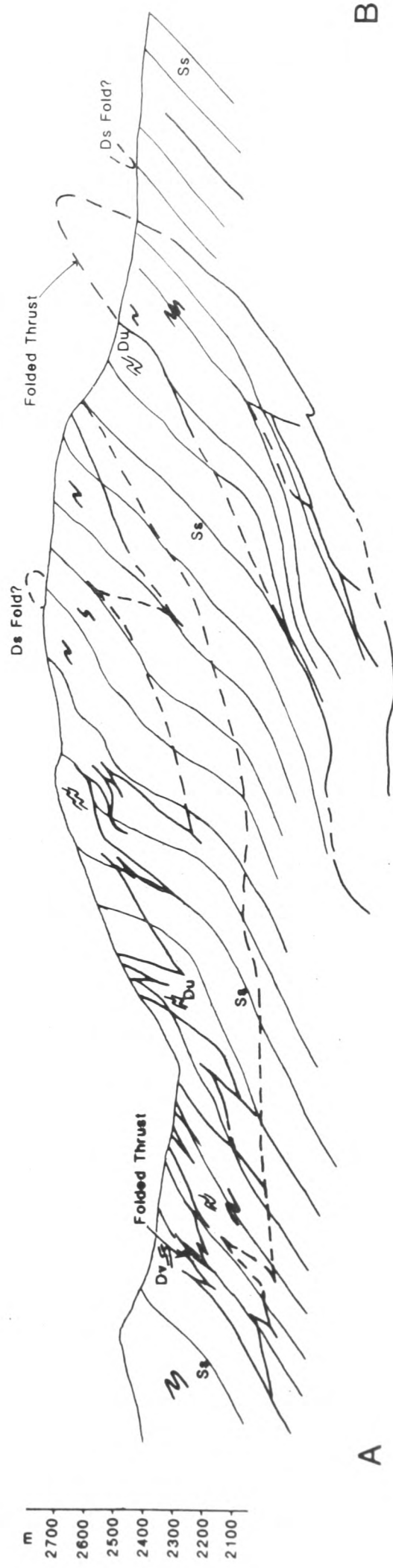
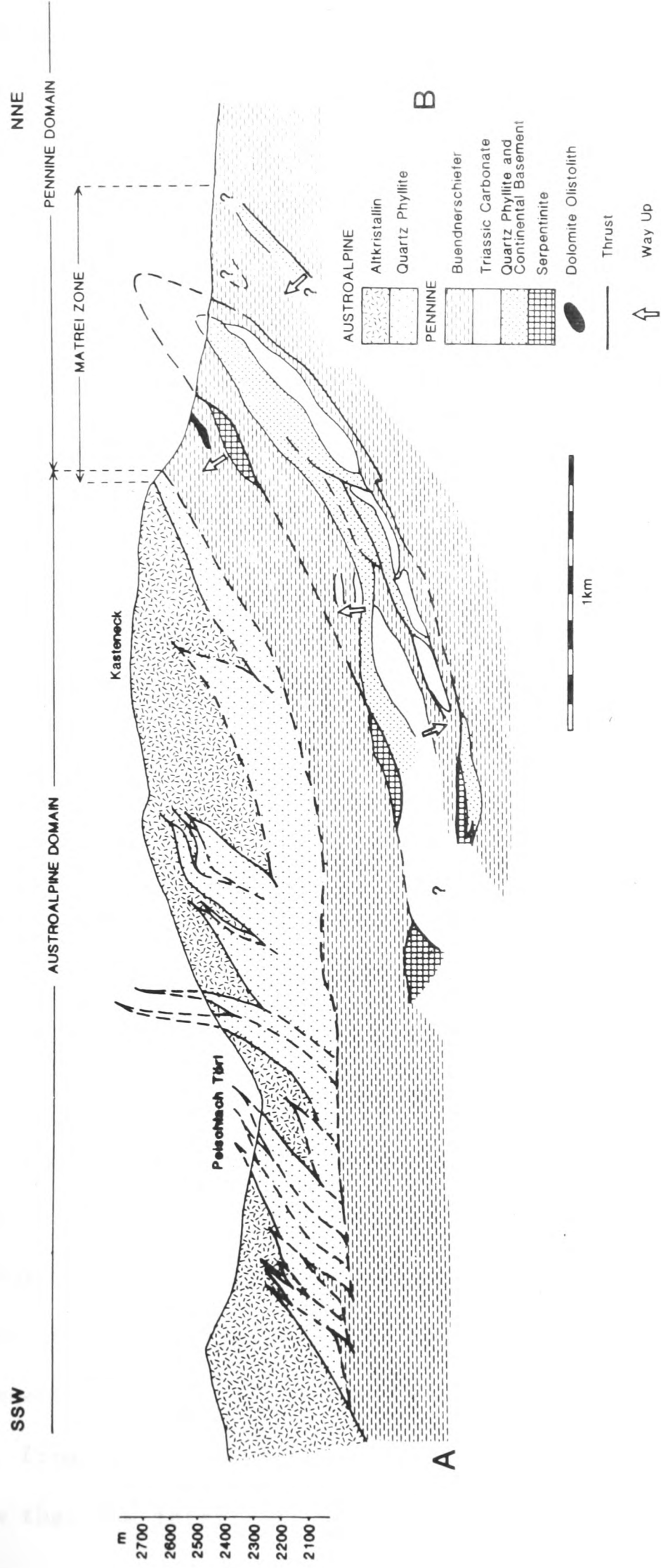
200m

- Buendnerschiefer
- Triassic Carbonate
- Quartz Phyllite
- Basement (retrogressed)
- Serpentinite
- Dolomite Olistolith
- Thrust
- Ds Fold

Fig.3.42. NNE-SSW oriented cross-sections through the Matrei Zone of area 1. As in fig.3.40 for area 3, the Matrei Zone can be divided into an upper Austroalpine quartz phyllite sheet and a lower Pennine unit dominantly composed of Bündnerschiefer. The quartz phyllite sheet has been imbricated with the overlying Altkristallin, however, the geometry of these imbricates changes rapidly along strike and the section can only represent their general relationships. The quartz phyllite sheet thins out to the west and locally Bündnerschiefer directly underlies Altkristallin. The imbricate marked (1) directly beneath Kasteneck, is included after the section drawn by Angel (1929).

The lower unit of the Matrei Zone also shows early imbrication, shown by repetitions of dolomite and quartz phyllite units. The imbricates have been deformed by Ds, can be related to a major fold within the Matrei Zone. This large-scale fold has been projected along strike of Ss into the section. The existence of this fold is indicated by way up indicators and to a lesser extent changes in the vergence of minor Ds structures. The fragment of continental material represented by the quartz phyllite and dolomite lenses, is interpreted as part of the thinned rifted continental crust which partly floored the Pennine ocean. The lenses of serpentinite form a semi-continuous layer, which is locally overlain by pelagic deposits. This clearly represents part of the original Pennine ocean crust.

The structural section is dominated by Ss, which can be traced continuously from the Altkristallin into the Matrei Zone. Later deformations only cause local overprinting of this main fabric. See fig.3.1 for position of section.



outlines of the pre-Ds thrusts are better defined and some idea of the original geometries can be ascertained.

Along the base of the Altkristallin the thrusts are brought out by the repetition of basement and quartz phyllites (usually distinctive clast-bearing white quartz schist of the Skythian). The repetitions of quartz-phyllite can be interpreted either as the imbricated original cover to the Altkristallin or as a part of the Matri Zone which has been imbricated with the overriding Altkristallin (fig.4.28).

No convincing original sedimentary contacts were found between the quartz phyllite and the Altkristallin and no signs of the Cretaceous *Mr* metamorphism were found in the quartz phyllite. Regional considerations also argue against the association of the two units prior to the emplacement of the Altkristallin over the Matri Zone; since there is no Permomesozoic sedimentary cover to the Altkristallin found anywhere else in the Schober or Sadnig Groups, and it would be a remarkable coincidence if the one place where this cover was preserved was exactly along the contact between the Matri Zone and the Altkristallin. All the available evidence, therefore, suggests that the imbricated white quartz schists are part of the Matri Zone and that the imbrication with the Altkristallin took place after *Mr*, i.e. after $\approx 80\text{Ma}$. The interpretation of the white quartz schists as part of the Matri Zone structurally beneath the Altkristallin, is in agreement with several other workers who have investigated this contact, e.g. Angel (1939 area 1); Senarclens Grancy (1964 area 1); Cornelius & Clar (1939 area 1); Exner & Prey (1964 area 3); Behrmann & Wallis (1987 area 1). However, Frisch et al (1987) consider the white quartz schist as the cover to the Altkristallin and Bickle (1973) mapped it as retrogressed basement.

Cross-sections through areas 1 and 3 (fig.3.40, fig.3.42.) both show that the interleaving of the Altkristallin with the Matri Zone

rocks is achieved by imbricating a pre-existing contact. Two phases of movement are required to produce this geometry both of which occurred after Mr and before Ds. These may or may not be related to one another.

No geometric analysis of the first formed contact was possible. However, the imbricate slices formed during the second stage can still be studied in some locations. In general, the imbricate slices of quartz phyllite pinch out to the east and the slices of basement pinch out to the west. This observation can be geometrically explained in two ways (see chapter 4). Either the quartz-phyllite overlies the basement and the cut-off lines plunge west, or the basement is structurally higher and the cut off lines plunge east. The evidence is in favour of the quartz phyllite being part of the Matrei Zone underlying the Altkristallin and, therefore, in favour of the cut-off lines having a general eastward plunge. This deduction can be used to suggest the kinematics associated with the imbricate formation.

3.12 PRE-Ds DUCTILE DEFORMATION IN THE PENNINE DOMAIN

Going north from area 3, deeper into the structural pile, Ds deformation weakens, the Ds folds become more open and it is correspondingly easier to recognize the earlier part of the tectonic history of the area. In the central part of the Tauern Window there is a general N-S/NE-SW trend of stretching lineations which have been reported as being overprinted by a later phase of NW-SE directed structures (Frasl & Frank 1964, Exner & Prey 1964). If the later NW-SE oriented stretching is related to Ds, the N-S to NE-SW stretching may be a ductile equivalent of the pre-Ds brittle deformation seen in the Matrei Zone. The lineation clearly represents a stretching direction as witnessed by the presence of mineral lineations and sheath folds (e.g.

between Fuscher Törl and Hoch Tor, Clar pers. comm.). At higher structural levels the evidence for this earlier phase of N to NE directed deformation becomes overprinted by the NW-SE trending linear fabric, which is probably related to Ds.

Within the south east Tauern there are a several other indications of this earlier deformation. The Sonnblick Gneiss body contains a number of dykes, recorded by Exner & Prey (1964), which have been deformed in shear zones with a sense of displacement to the NE. In other parts of the same Sonnblick area, shear zones were found with NE stretching lineations and associated tight folds with their axes parallel to the stretching direction (fig.3.43). One quartz c-axis fabric was measured from within from within one of these shear zones, which indicates a top to the NE sense of shear (fig.3.44).

3.13. STRUCTURAL DIVISIONS OF THE MATREI ZONE

A commonly-used restoration technique in thrustured terrains is balancing cross-sections parallel to the movement direction. The Matrei Zone is strongly deformed and contains several major tectonic discontinuities with unknown and probably large displacements along them. Such zones cannot be accommodated in balanced cross sections. Shortening estimates can, however, be obtained within a number of separate and internally more coherent units. However, even on this local scale the polyphase deformation with oblique movement directions introduces considerable uncertainty into the estimates. The estimates for shortening are derived from line balancing NNE-SSW sections, and are not extended over major tectonic boundaries.

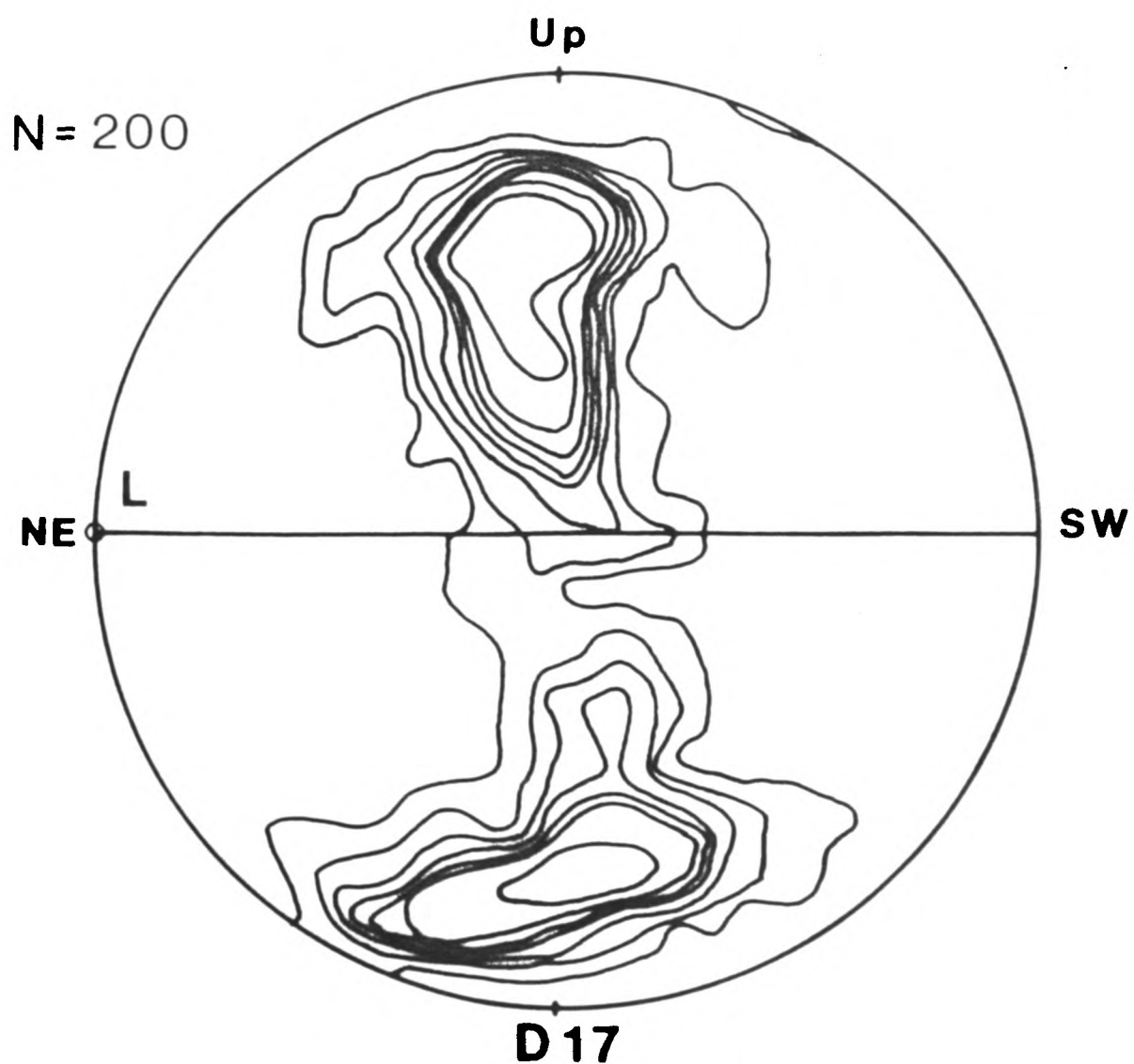
Generalized cross-sections are presented through the Matrei Zone for areas 1 and 3 in figs 3.40, 3.42) and a more detailed section is given through the Makerni Spitze of area 3 in fig.3.41. The upper



0.5m

Fig.3.43. Tight pre-Ds folds within the Sonnblick gneiss associated with NE-SW oriented stretching lineation parallel to the fold axes. D17, on the path from Fragner Hütte to Hagener Hütte.

Fig.3.44. Quartz c-axis preferred orientation pattern measured in a deformed metaquartzite, the asymmetry indicates a top to the NE sense of shear. D17, see fig.3.43 for location.



tectonic contact of the Matrei Zone is along the boundary between the quartz phyllite and the base of the Altkristallin, which represents a major tectonic break with an unknown amount of displacement along it. The initial thrust contact has itself been imbricated, causing further shortening of the area, $\approx 3\text{km}$ in area 1 and $\approx 1\text{km}$ in area 3.

In all three areas the dominantly quartz phyllite sheet immediately underlying the Altkristallin, represents a distinct tectonic unit, which is separated from the structurally lower parts of the Matrei Zone by a major discontinuity. In area 3 this contact is decorated with serpentinite lenses. The identification of an upper quartz phyllite sheet corresponds with the very first attempts to define laterally continuous structures within the Matrei Zone on the basis of associated lithologies (Schmidt 1950-52).

Underlying the quartz phyllite sheet to the north the Matrei Zone is dominated by Bündnerschiefer lithologies, which commonly contain clastic material including olistoliths. This clearly represents a very different palaeogeographic domain. Two distinct parts of this lower unit can be distinguished. Immediately beneath the quartz phyllite sheet there is generally a series of calcareous graphitic schists of the Bündnerschiefer, which in area 1 is associated with serpentinite and probable pelagic deposits (chapter 2). Several other large serpentinite bodies and related rock types have been mapped within the Matrei Zone (e.g. around Heiligenblut Exner & Prey 1964). Although discontinuous, these lenses are preferentially aligned along certain horizons which probably represent major tectonic contacts. The association with oceanic pelagic deposits suggests that these slivers of serpentinite can be interpreted as the disrupted remnants of oceanic crust. There is no direct evidence for this zone continuing to the east of area 3 and this may mark the original limit of the serpentinite in this area.

Structurally below this unit, material is also associated with

Bündnerschiefer lithologies, however, fragments of an incompletely preserved continental stratigraphy can still be recognized. In area 1 this is represented by the Medel Spitze dolomite quartz phyllite unit which probably includes some Palaeozoic pre-Alpine basement. Individual thrusts can be recognized within this unit, with a total shortening of $\approx 500\text{m}$ in a NNE direction. This unit is surrounded by Pennine deposits and is directly overlain by the graphitic calcareous schists typical for the Bündnerschiefer. I interpret this as a fragment of thinned continental crust which partly formed the floor to the Pennine ocean. In area 3 a similar sequence is present but the basement includes the distinctive orthogneiss of the Sonnblickⁿ dome and the Rotwand[^] Gneisslamella (Exner & Prey 1964). The gneiss lamella can also be interpreted as a fragment of the original continental crust which was thinned during rifting and formed part of the Pennine ocean floor. The quartz schists and associated rock types to the north of Makerni Spitze and south of the Sonnblick Gneiss are probably also fragments of thinned continental basement and their basal contacts also represent major tectonic discontinuities.

Regional arguments suggest that more than one phase of thrusting may be present within the Matri Zone and some of the complex geometries within the Makerni Spitze mountain side are not easily explicable in terms of a single phase of thrusting, but from the incompletely preserved imbricate stack, it is not possible to determine whether one or several phases of thrusting are required to produce the observed structures.

The imbrication of the Matri Zone was deformed by Ds, which forms the dominant deformational fabric of the area. In the Matri Zone, Ds produces large-scale tight to isoclinal folds, which caused further substantial shortening of the area. The striking dolomite-quartzite complex surrounded by Bündnerschiefer, which is exposed in the Makerni

Spitze is a tightly folded Klippe of the quartz phyllite sheet (fig.3.40). There is no evidence to suggest it is an olistolith as proposed by Frisch et al (1987). Ds folding is also present to the north of area 3, and the Sonnblick Dome - Mallnitzer Mulde synform-antiform pair are probably major Ds folds.

3.14. TERTIARY DEFORMATION AND METAMORPHISM

3.14.1 Tauern Metamorphism

The Cretaceous deformations are separated from a series of Tertiary deformations by the mainly static Tauern Metamorphism. This metamorphism is associated with cooling ages of around 40-35Ma near the margin of the Tauern Window (Cliff et al 1971), gradually decreasing to 25Ma at deeper structural levels (Cliff et al 1985). In the S.E. Tauern at the margins of the window, the effects of this metamorphism are relatively minor. In the Mafrei Zone chlorite and albite porphyroblasts locally cross-cut the Ss fabric (fig.3.45). Cornelius & Clar (1939 p.249) also report small undeformed rosettes of chloritoid growing at similar structural levels in the Mafrei Zone of area 1. Further north, deeper in the structural pile, higher metamorphic grades are attained. Just north of the Mafrei Zone in area 3 chloritoid and biotite occur (in separate samples). The orthogneiss of the Sonnblick Dome lacks a suitable chemical composition to show changes in metamorphic grade by the development of new minerals. In the Mallnitzer Mulde, however, garnet + amphibole + biotite assemblages occur, indicating amphibolite grade conditions (fig.3.46a,b.). One sample contains both staurolite and chloritoid and metamorphic studies show that the prograde $ctd \rightarrow gt + st$ reaction took place in this area at around 550°C, 7kb (Droop & Cliff 1985).

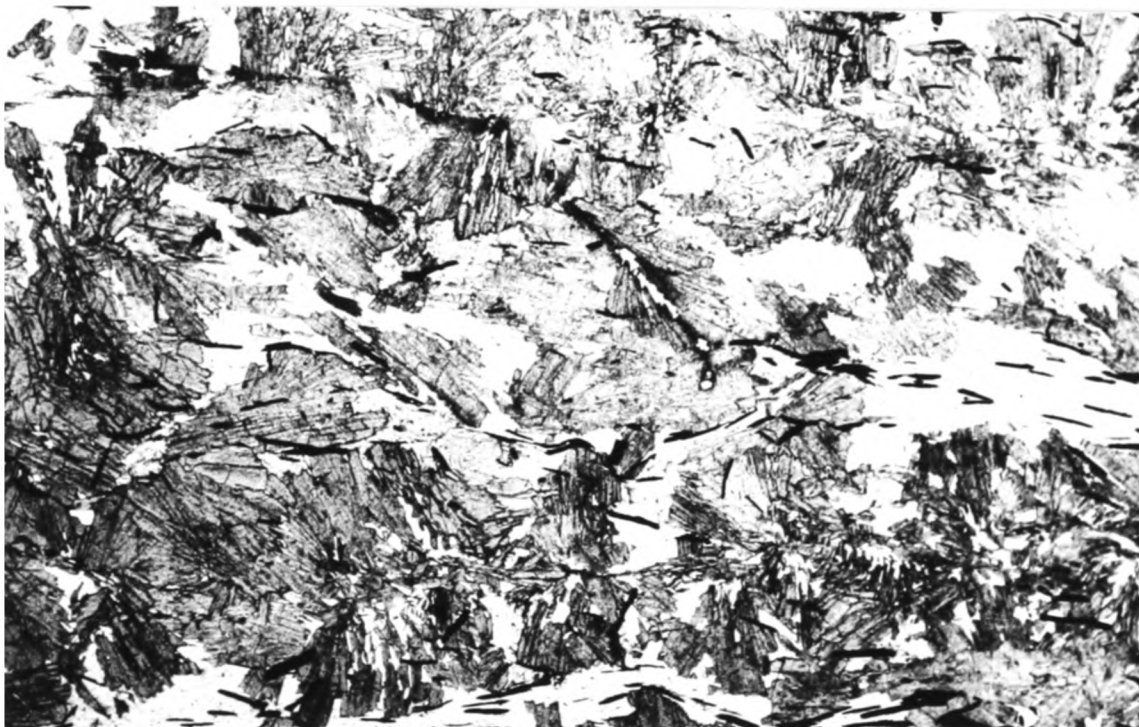


1mm

Fig.3.45. Chlorite porphyroblast developed during the Tauern Metamorphism, cross-cutting Ss. 1km to the north of the base of the Altkristallin in the Mafrei Zone, B113, area 1.



Fig.3.46a. Amphibolite grade Tauern Metamorphism in the Mallnitzer Mulde. The amphibole and garnet overgrow Ss and are slightly deformed by a later set of microfolds.



3mm

Fig.3.46b. Chloritoid schist from the Mallnitzer Mulde (see map by Droop (1978) for location of coarse chloritoid schist). Staurolite also appears in this sample. Chloritoid overgrows an earlier differentiated fabric defined in this section by the preferred orientation of opaques and platy minerals.

3.14.2 Dt Deformation

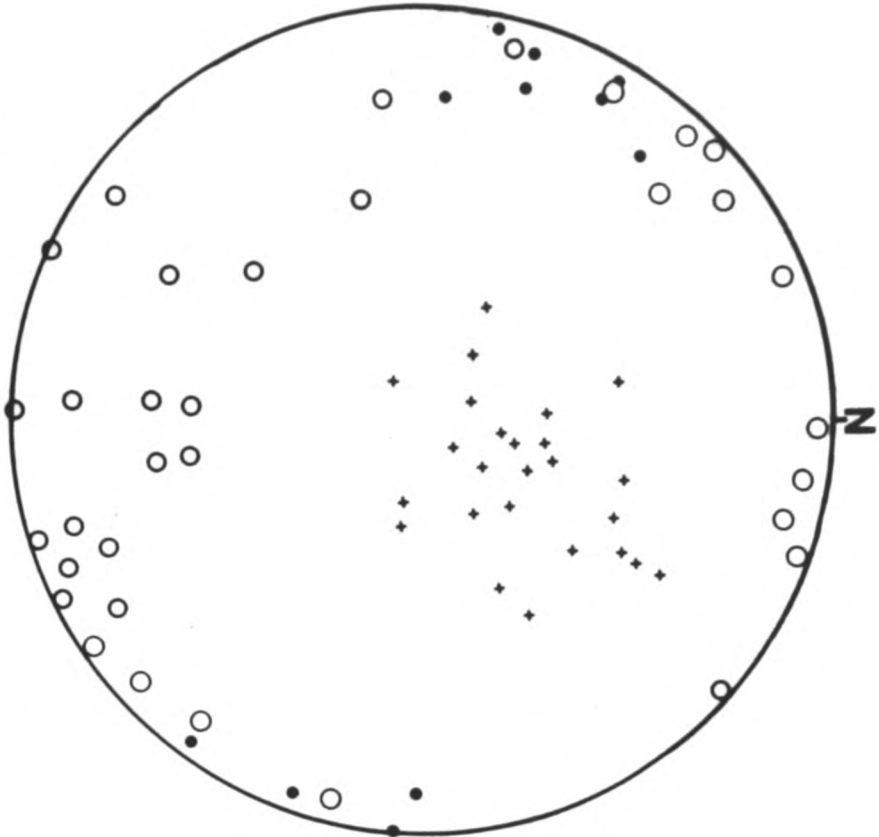
The first of the deformations which overprints Ds is labelled Dt. In the mapped areas Dt deformation is entirely restricted to the Matrei Zone where it clearly overprints Ds fabrics. In area 3, Dt forms a 500m wide discrete shear zone which occupies the north of the Matrei Zone and the section of the Pennine units up to the Sonnblick Gneiss Dome. In area 1 and the Mallnitzer Mulde there are also narrow shear zones, which show similar microstructural features and may correlate with the same phase of deformation.

Area 3 was divided into two different sub areas north and south of the Bretterich Kamm, which approximately marks the junction between the penetrative Dt deformation and the more open Dt folding (fig.3.1). The mesoscopic orientation data are plotted in fig.3.47. In low-strain areas Dt fold axes are oriented approximately N-S, at a high angle to Ls. In this area Ls can commonly be found folded around Dt folds. In the higher-strain area to the north, a good Dt stretching lineation is developed with a WNW-ESE trend. The change in the orientation of the Dt fold axes with increasing Dt strain is due to their rotation into the maximum finite stretching direction. This rotation can be seen particularly well in the area behind the Fragner Hütte and in area 1 south of the serpentinite body on Glatz Schneid, where open folds are associated with N-S oriented fold axes and tight high-strain folds have their axes trending sub-parallel to the stretching lineation. Direct evidence for the rotation of Dt fold axes during deformation comes from three dimensional exposures of curvilinear folds in area 1 (fig.3.48) and several 10cm circular outcrop patterns within the Bretterich Marble which I interpret as sheath folds.

In area 3 the sense of shear during Dt is given by extensional crenulations, quartz c-axis fabrics and asymmetric pressure shadows around rigid inclusions. These all indicate a top to WNW sense of

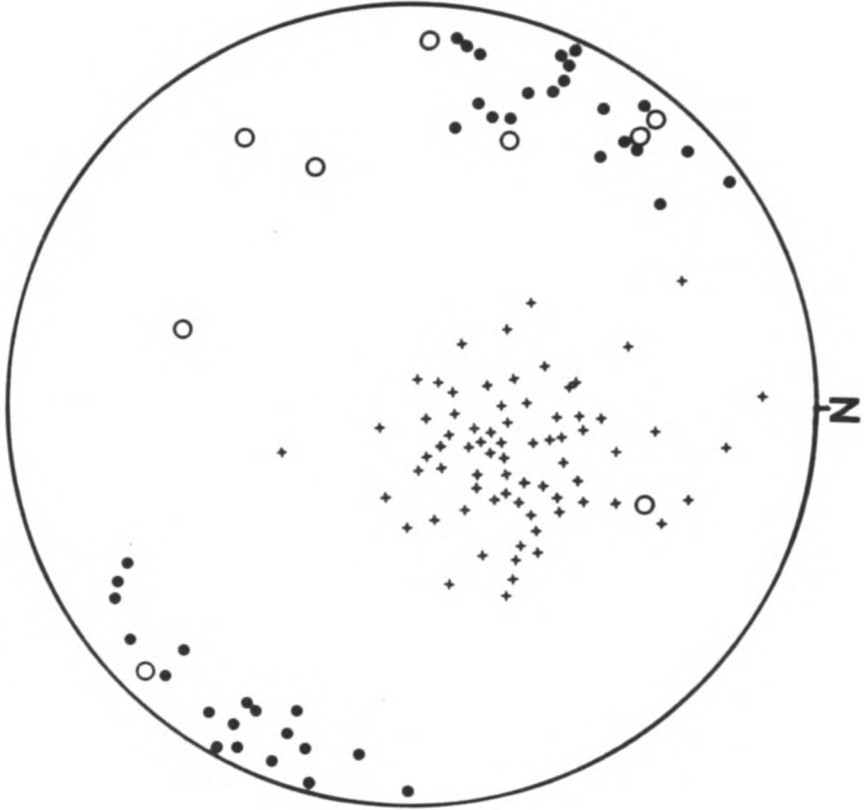
Fig.3.47. Mesoscopic structures of Dt deformation plotted on a Schmidt net in map areas 1 and 3. Dt is only an important deformation in area 3, where two subareas can be defined corresponding to a change in the associated finite strain. In the low-strain area to the south of the Breterrigh Kamm, fold axes vary greatly in orientation and are more readily measured. To the north there is a rapid change in finite strain which is reflected in the greater number of stretching lineations and the lower number of fold axes that were recorded.

MAP 3



South of Breterich Kamm

MAP 3



North of Breterich Kamm

MAP 1

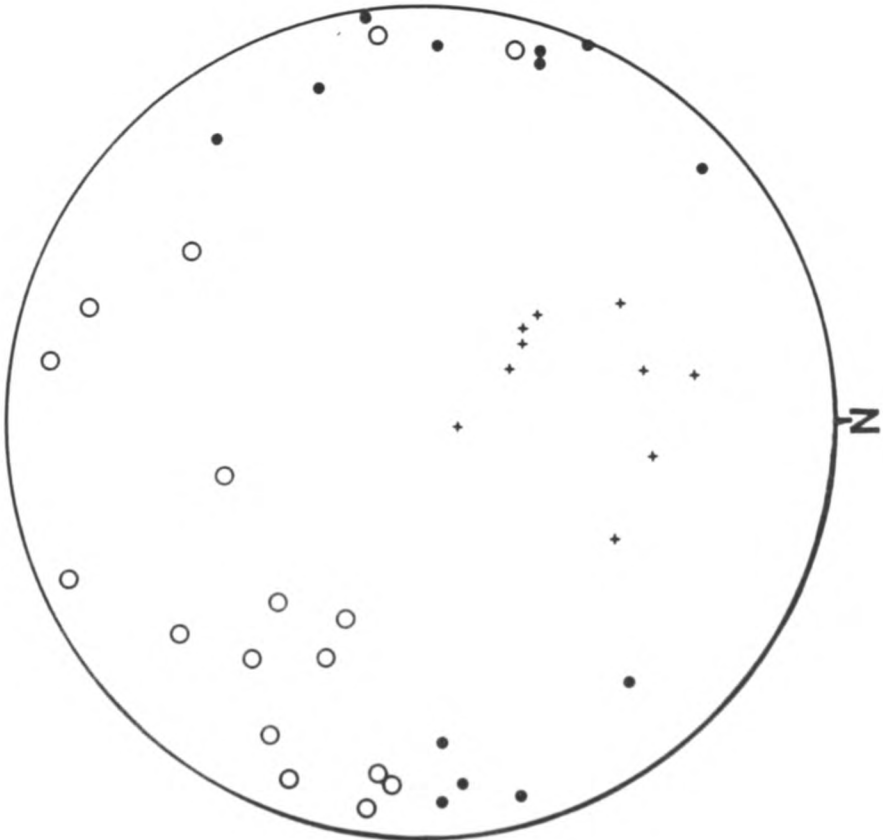




Fig.3.48. Three dimensional exposure of a curvilinear D_t fold. Looking south, B3, area 1.

shear. In area 1 the kinematic indicators are less clear. The stretching lineation is oriented E-W to ENE-WSW and ecc's suggest the sense of shear is top to ENE. The topology of the quartz fabrics suggests a relatively low rotational component of deformation and this is supported by the presence of globular quartz grains (Law et al 1984) in some samples. The dominance of L-S fabrics in area 3 suggests plane-strain conditions, which is supported by the cross girdle patterns of the quartz c-axis fabrics from samples 512, MS24 (fig.3.49a) and suggest deformation was between pure and simple shear.

To the south of the zone of penetrative deformation, Dt is concentrated in the relatively weak calcite-rich horizons. The calcite c-axis fabric of sample 412 is approximately symmetrical in both density distribution and topology (fig.3.49b), supporting the suggested low degree of non-coaxiality for Dt deformation. Sample 228 is a quartzitic marble offering the opportunity for both calcite and quartz fabric analysis in the same sample. The calcite c-axis fabric has a monoclinic symmetry that suggests a top to the W sense of shear (Schmid et al 1981), whereas the quartz c-axis fabric suggests non-plane strain coaxial deformation in the constrictional field (fig.3.49b). This implies that strain partitioning took place between the two minerals.

St tends to look rather more platy and fissile than Ss but the two cannot be distinguished on the basis of mesoscopic structures alone. Where obvious overprinting relationships are lacking, Dt can best be identified by its associated microstructure. Dt overprints the Tauern Metamorphism and is, therefore, microstructurally characterized by deformed Tauern Metamorphic minerals with unannealed fabrics. Some of the albite porphyroblasts affected during Dt contain slightly curved inclusion trails suggesting a synkinematic growth.

Fig.3.49a. Contoured quartz c-axis diagrams from Dt quartz mylonites. The contour interval is 1 to 6% per 1% area. • = orientation of a globular quartz grain (see text). A low degree of non-coaxiality is indicated by the presence of globular grains with a preferred orientation of the c-axis at a high angle to the flattening plane and the general orthorhombic symmetry of the c-axis fabrics.

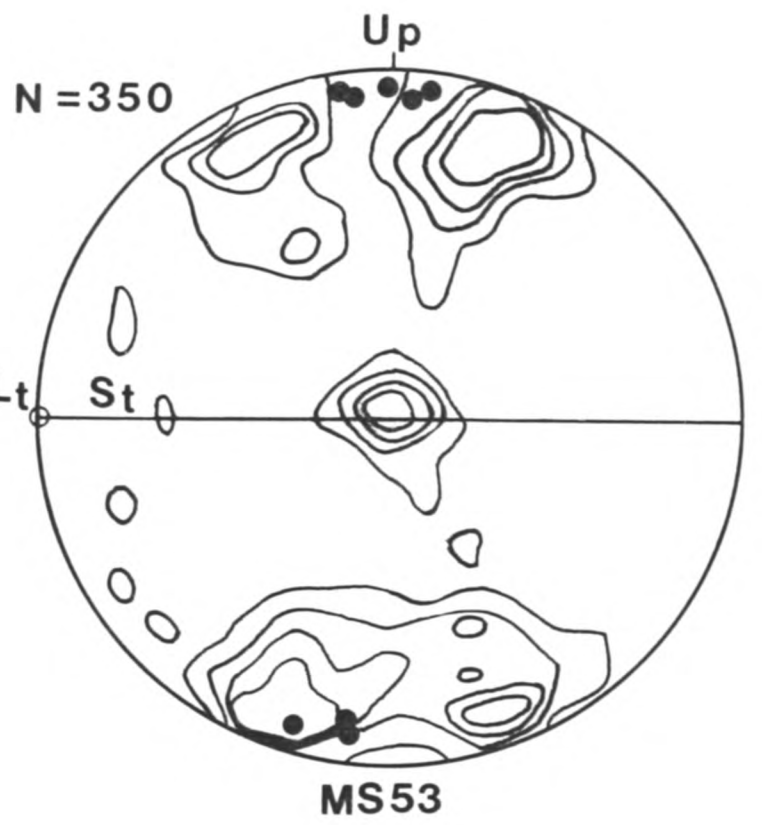
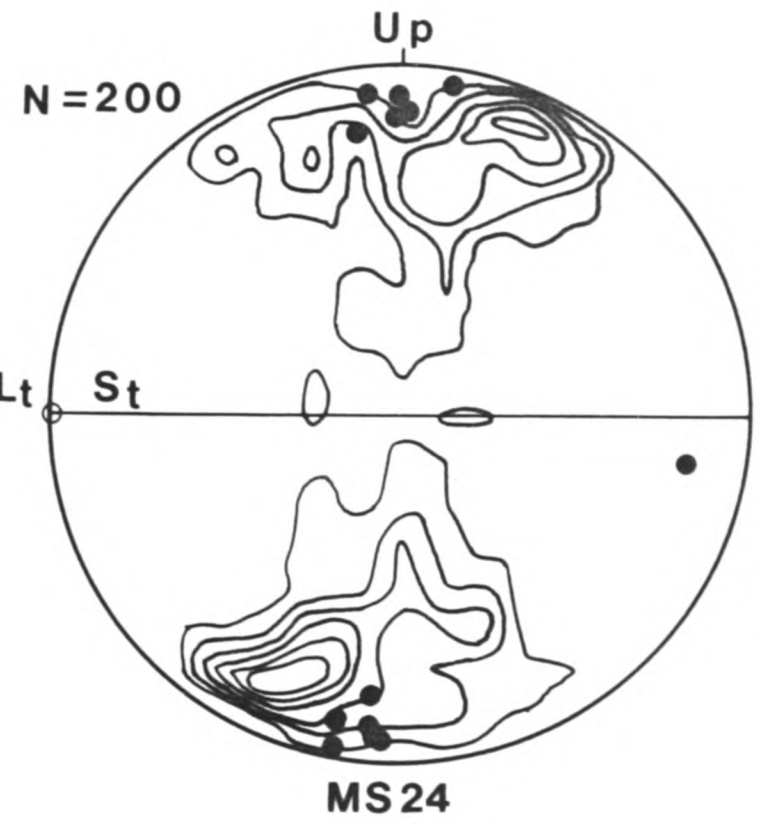
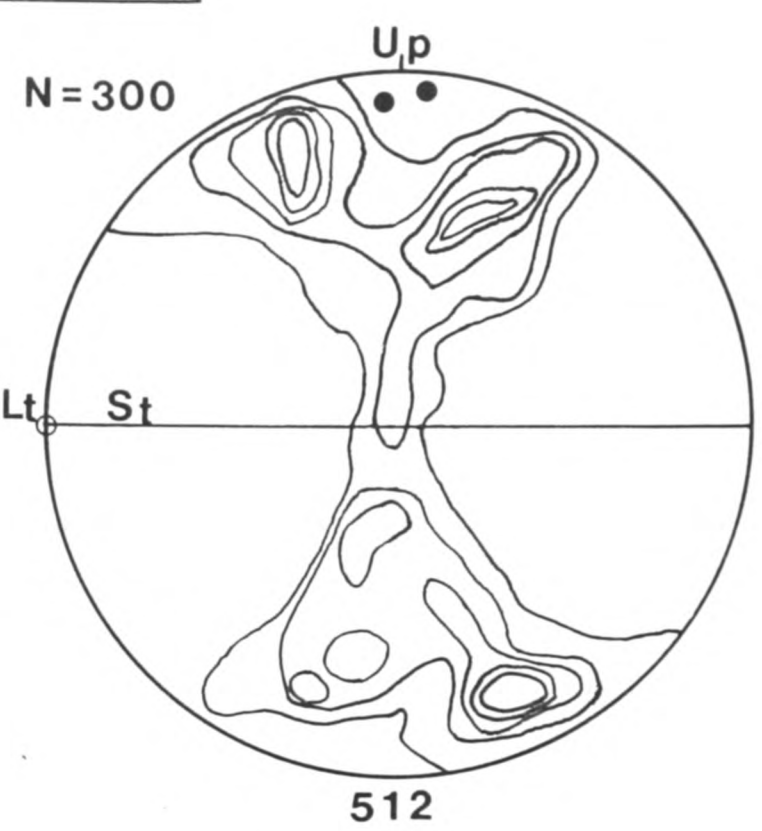
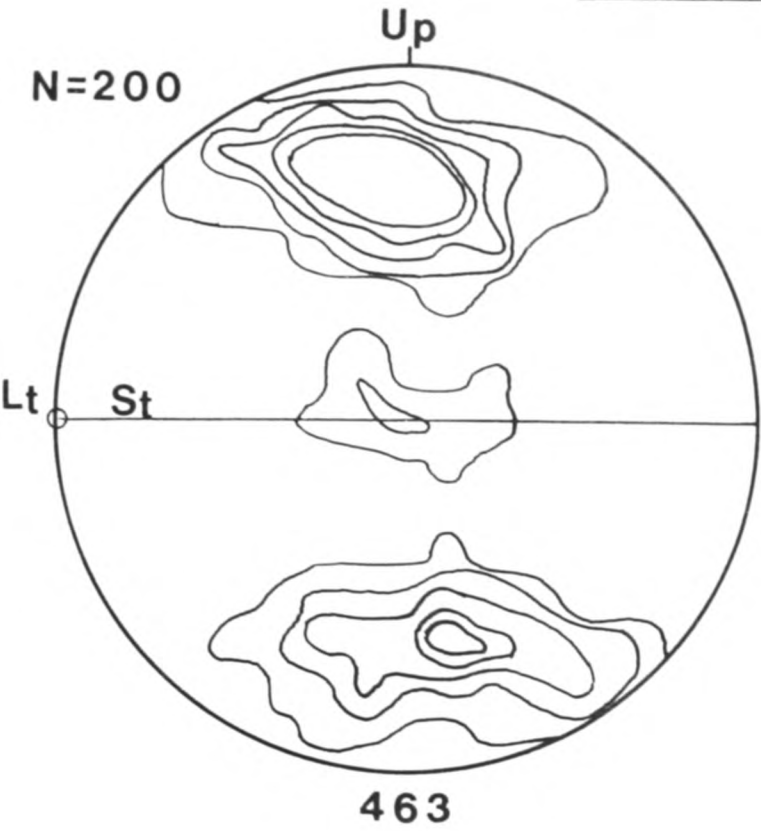
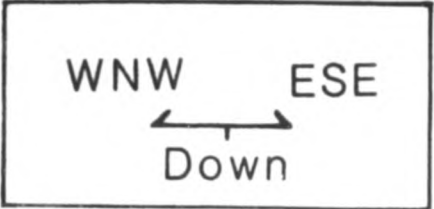
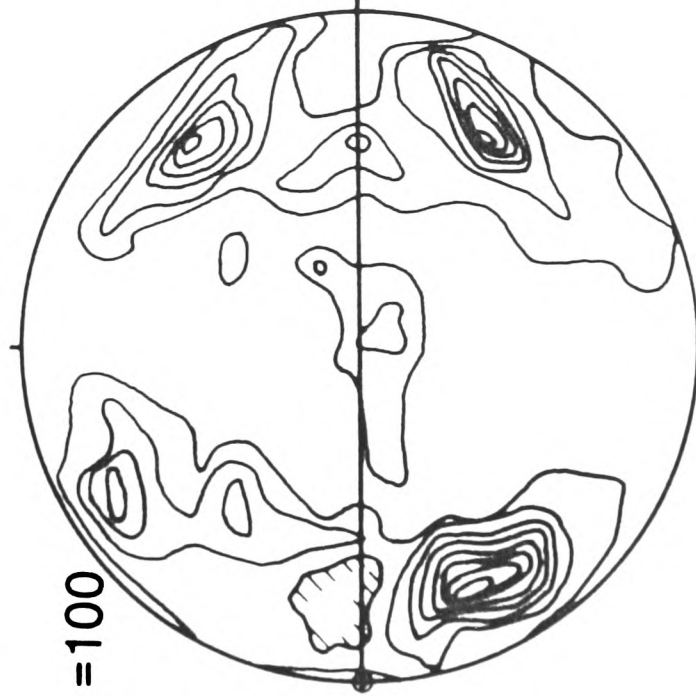
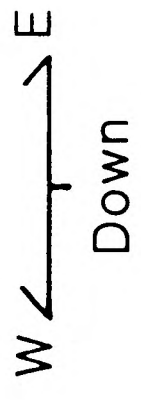
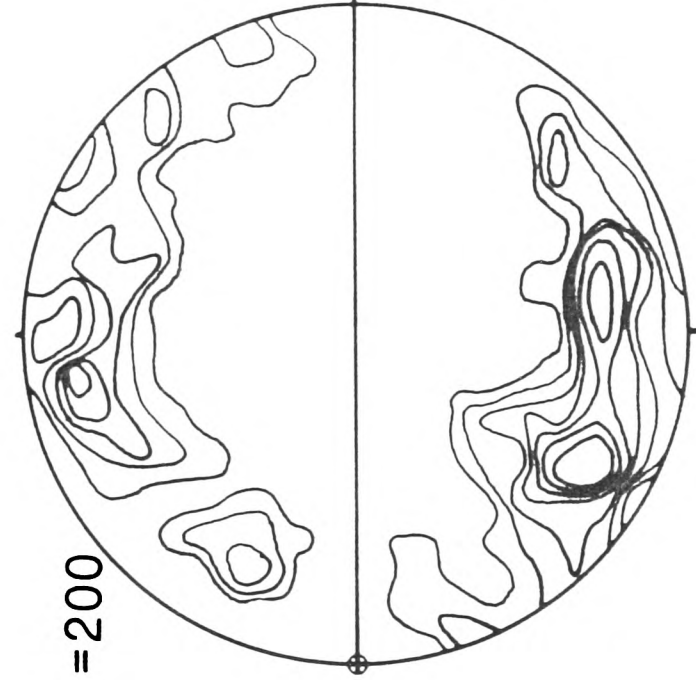


Fig.3.49b. Contoured calcite and quartz fabrics for samples 228 and 412. Contour interval for all diagrams is 1% per 1% area.



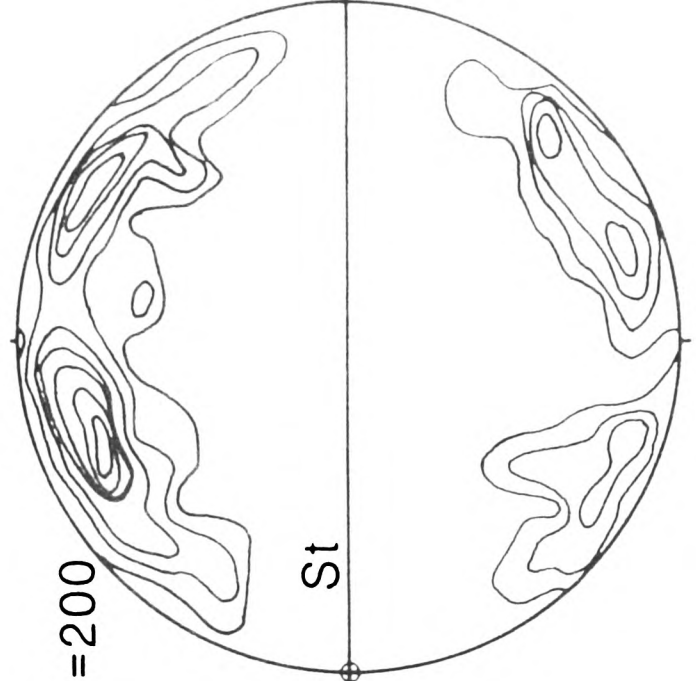
228

Quartz



228

Calcite



412

Tracing the Matrei Zone to the east it steepens and thins to a thickness of only a few metres (Exner & Prey 1964). Samples from this zone show post metamorphic deformation with a top to the NW sense of shear. It is likely that this phase of deformation is also related to Dt. Further to the east the steep zone passes into the Mölltal Line which is a late stage fault (Exner 1962). The discrete fracture along the Mölltal Line cannot be traced further W than Obervellach and the movement along it may pass into a zone of distributed shear. From the offset of the base of the Altkristallin, the sense of movement on the Mölltal Line is top to the NW and the post metamorphic ductile deformation within the Matrei Zone and Mallnitzer Mulde may be related to movements along this fault.

3.14.3 Later Deformations

Surprisingly good correlation between the different areas is found for the later, low-strain events. These can be divided into two main phases of folding Du, Dv and later joints and fractures.

The two fold phases are distinguished primarily on the basis of their geometries. In area 1 interference structures show that two distinct phases of folding were present (fig.3.50.). They are important to document since their fold axes are approximately colinear with both the high strain Dt and Ds fold axes. There is, therefore, clearly a potential for confusion.

a) Du Deformation

Du is a series of north vergent folds with steep axial planes which overprints both Ss and St, and occasionally a new schistosity, Su, is developed. The Du folds are usually open and may have a wavelength up to several metres. The fold axes have an average trend of 110°

Fig.3.50a. Interference of D_u and D_v folds in a strongly foliated quartz schist, Peischlach Törl, area 1. The dominant foliation is S_s .
Looking east.

Du
Folds

3cm

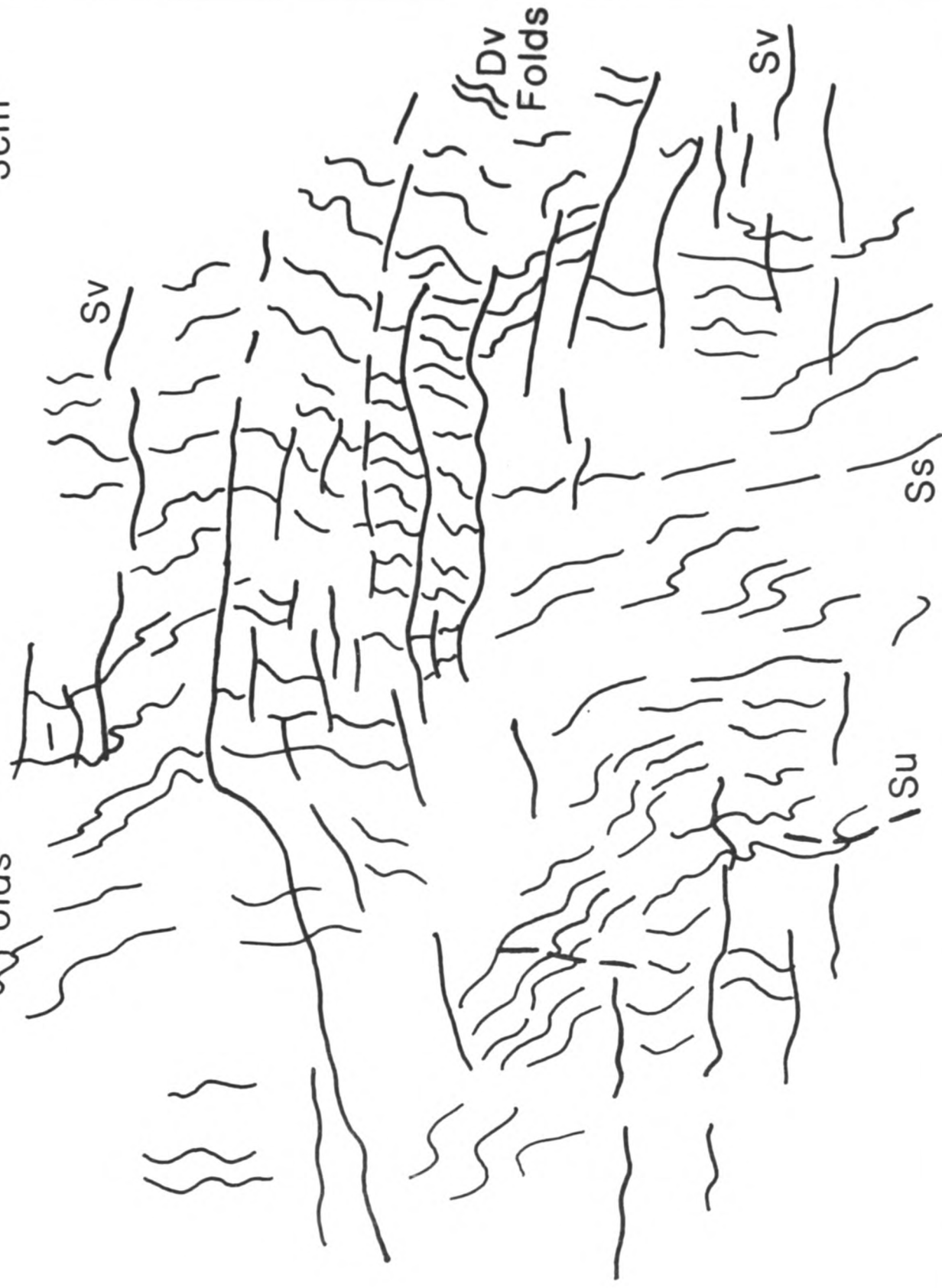




Fig.3.50b. Field shot of Du kinks in a strongly foliated quartz schist, Peischlach Törl, area 1.

(fig.3.51.). The Du deformation is well-developed to the north of area 3, where St is the dominant fabric. In the Altkristallin Du produces a series of open flexures which in one place in area 1 are associated with a penetrative Su fabric. In the field Su can be distinguished from Ss in most cases by its steep dip and lack of associated stretching lineation.

b)Dv Deformation

Dv is a south vergent phase of folding with gently south-dipping axial planes and fold axes trending about about 110° (fig.3.51). In many cases Du can only be distinguished from Dv on the grounds of style. Locally, overprinting relationships are present which show that Du was older than Dv. Near the Altkristallin - Matri Zone boundary in part of area 3, Dv becomes strong enough to begin to form a new differentiated fabric, Sv.

Du and Dv are only locally associated with high strains and are not an important feature of the Alpine structural development in this area.

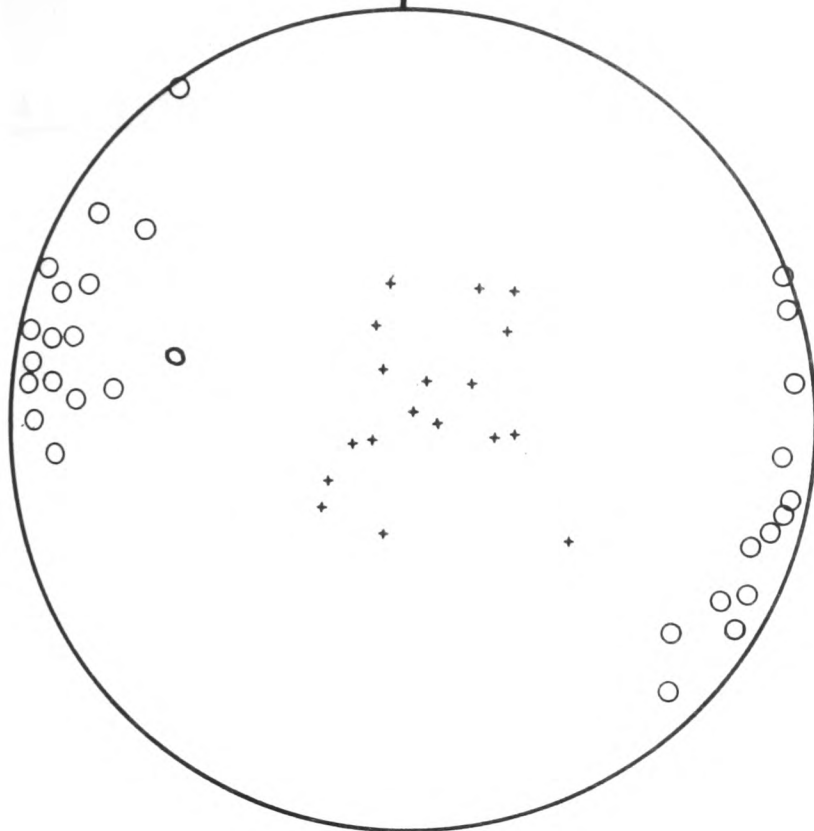
c)Joints and Fractures

Finally a series of joints is developed. The orientations were measured in area 3, where they show great variation. In the field joints oriented roughly N-S are usually the most prominent and these are commonly associated with west vergent kinks. In the Sonnblick Gneiss prominent NE-SW striking tension gashes are commonly mineralized and filled with vein quartz. These fractures represent a NW-SE extension of the Sonnblick Dome. A large-scale analysis of the joint patterns in the area has been carried out by Norris et al (in Cliff et al 1971) and nothing new can be added by the observations in this study.

Fig.3.51. Orientation of mesoscopic Du and Dv features. + = pole to foliation (each symbol average of 3 to 5 readings), o = fold axis (each symbol average of 2 to 5 readings).

MAP 3

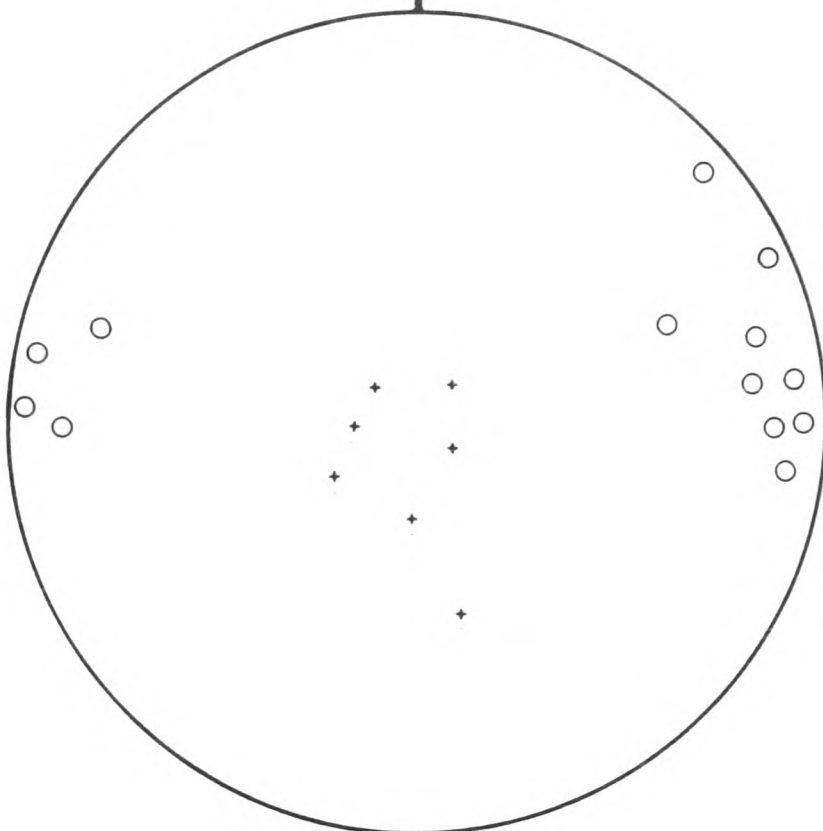
N



Dv Deformation

MAP 1

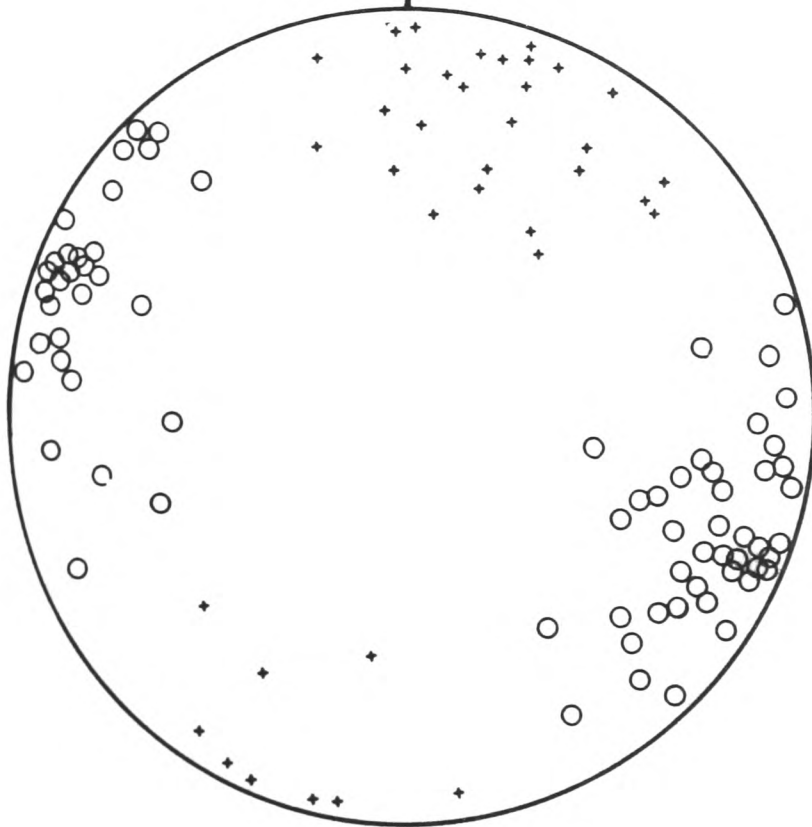
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Dv Deformation

MAP 3

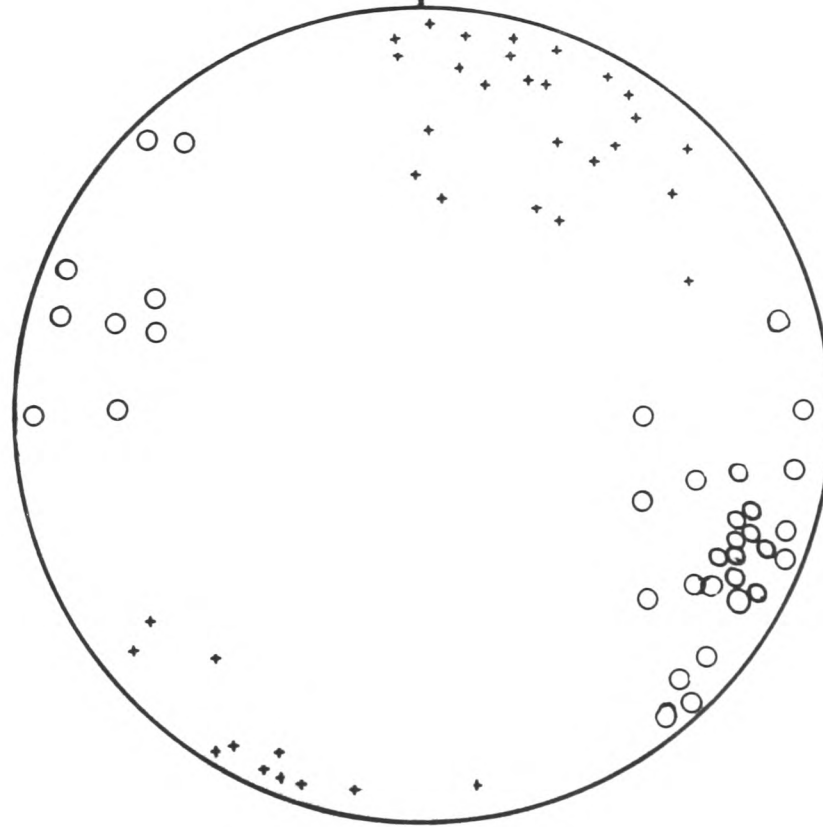
N



Du Deformation

MAP 1

N



Du Deformation

CHAPTER 4 THE KINEMATICS OF DEFORMATION

4.1. INTRODUCTION

Two of the principal goals in structural geology are to determine the large-scale movements that are responsible for deformation and to investigate the processes that cause them. In this chapter I examine the kinematic aspect of deformation, which is concerned with the motion of bodies without reference to the associated driving forces. Deformation across the Austroalpine - Pennine boundary can be divided into three main phases (chapter 3). The kinematic data for each of these phases are derived from a variety of mesoscopic and microscopic structures, concentrating on the relative movement direction during deformation and, for the dominant ductile deformation, D_s , the degree of non-coaxiality.

Most workers in the Eastern Alps have assumed that the Austroalpine domain was emplaced to the north or north east over the Matrei Zone and Pennine units (Kober 1955, Clar 1965, Tollmann 1962, Oxburgh 1968, Bickle & Hawkesworth 1978). This was largely based on the east - west trend of the mountain belt and the assumption that fold axes and 'b' lineations (Sander 1950), which some workers identified as stretching lineations, form and remain perpendicular to the tectonic transport direction throughout deformation. However, several studies have shown that fold axes may progressively rotate towards the stretching direction with increasing strain (Bell 1978, Williams 1978), and recent structural studies in the Eastern Alps suggest that deformation was associated with a substantial west-directed component of movement (Ratschbacher 1987, Brunel & Geyssant 1978).

To the south east of the Tauern Window three deformation phases can be defined which affect the Austroalpine - Pennine boundary (chapter 3).

The earliest phase, Dr, is restricted to the Altkristallin and took place before its juxtaposition with the Matrei Zone. The kinematic data for this deformation suggest it was associated with a top to the W/WNW sense of shear. The first deformation that is common to both the Matrei Zone and the Altkristallin is a phase of thrusting which took place $\approx 80\text{Ma}$. This caused imbrication both within the Matrei Zone and between the Matrei Zone and the Altkristallin. From a geometric analysis of the imbricate stack, at least two stages of thrusting can be identified. The second stage may have been associated with a top the NE movement of the Altkristallin over the Matrei Zone.

This early phase of thrusting is overprinted by a major phase of ductile deformation, Ds, which forms the main deformational fabric across the Austroalpine - Pennine boundary. In this chapter, I document the associated consistent top to the NW sense of shear. At higher structural levels the same deformation is expressed as discrete shear zones also with a top to NW sense of shear. A kinematic analysis shows that during this deformation the Austroalpine domain was thinned by at least 10km (see later for further details). The vast majority of the data is related to Ds and the study is much less complete for the other two deformation phases.

Pre-Ds ductile deformation is present in the Pennine domain at lower structural levels than the Matrei Zone, which is characterized by N-S to NE-SW stretching lineations. Insufficient data were collected for any detailed study of the associated kinematics. Deformation which post dates the Tauern Metamorphism, i.e. post-collisional, is not considered in this chapter.

PART A : THE KINEMATICS OF Ds DEFORMATION

4.2. INTRODUCTION

Ds is the dominant deformation across the Austroalpine - Pennine boundary, where it forms a broad, kilometre-scale zone of distributed shear. This zone can be traced from within the Altkristallin through the Matrei Zone and into the upper levels of the Pennine domain (chapter 3). At higher structural levels within the Altkristallin, above the zone of distributed shear, Ds is expressed as discrete shear zones (chapter 3). Deformation at these two structural levels is considered separately.

Assuming spatial continuity of deformation on the scale of the analysis, i.e. any discrete fractures are shorter than the scale of the analysis, a continuum mechanics approach can be used to analyse the kinematic development during Ds.

4.3. NON-COAXIALITY AND VORTICITY

In a homogeneously deforming medium the rate of displacement of a particle at $\underline{x}$ in a coordinate system $\underline{x}$, can be described by the equations

$$\dot{\underline{x}}_i = L_{ij} \underline{x}_j, \quad L_{ij} = \partial v_i / \partial x_j \text{ where } L_{ij} \text{ is the velocity gradient tensor.}$$

According to the fundamental theorem of continuum mechanics, L_{ij} can be decomposed into a stretching rate tensor D_{ij} , and a spin or vorticity tensor, W_{ij} , where $L_{ij} = D_{ij} + W_{ij}$ (Malvern 1969). D_{ij} represents the coaxial component of deformation and the eigen vectors of D_{ij} , d_1 , d_2 and d_3 are the orthogonal instantaneous stretching axes of the flow. The components of W_{ij} can be expressed in terms of the vorticity vector

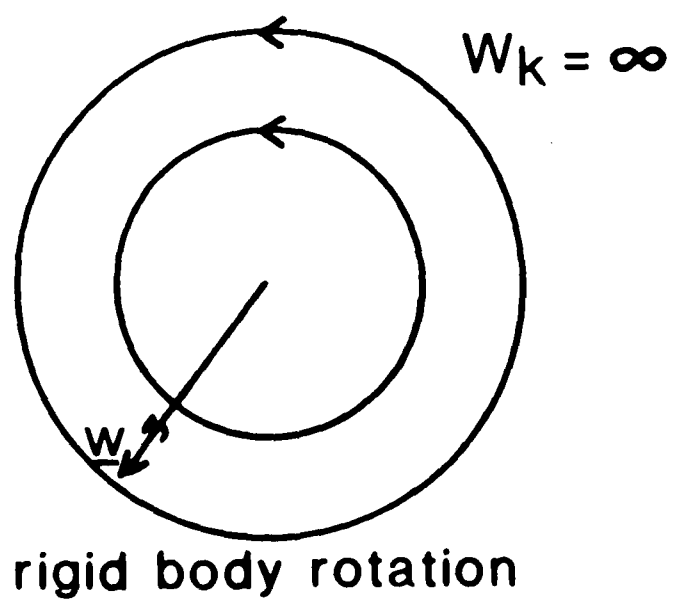
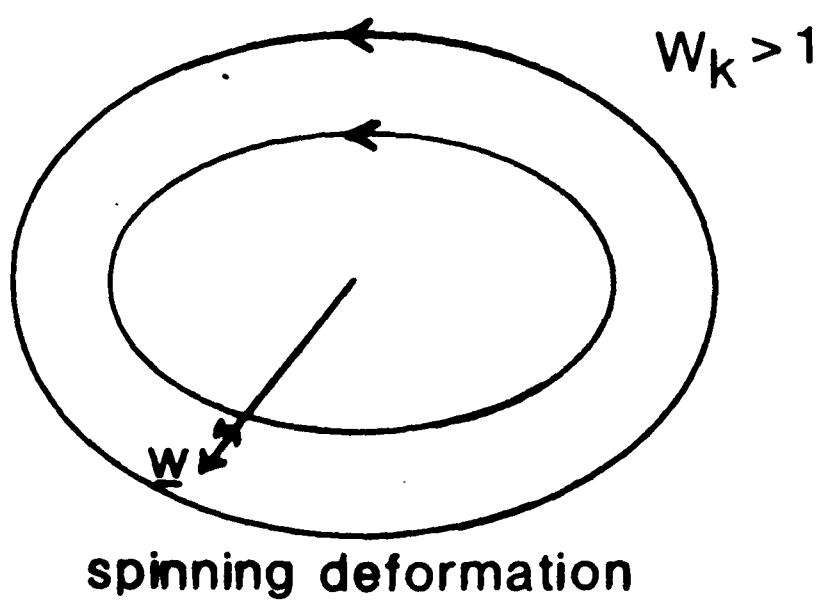
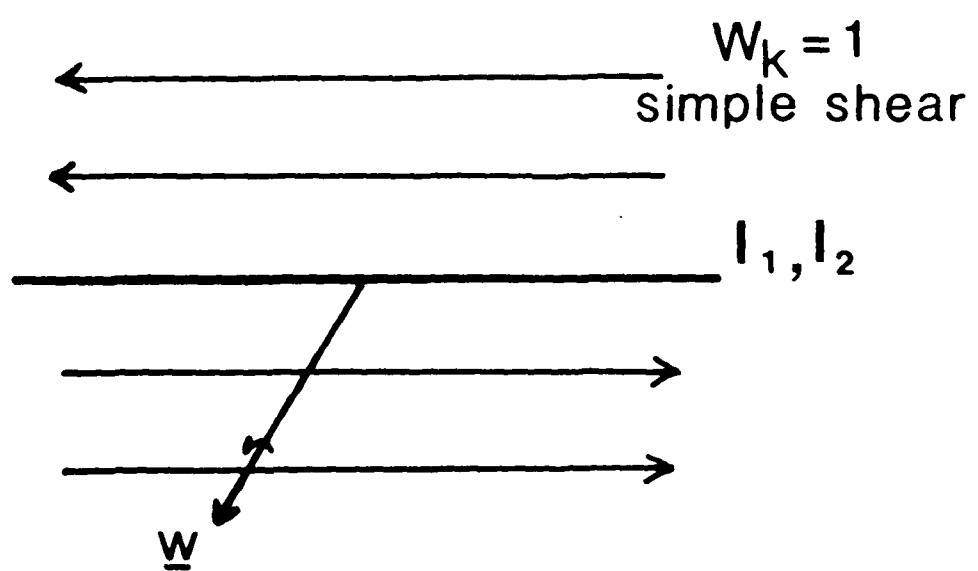
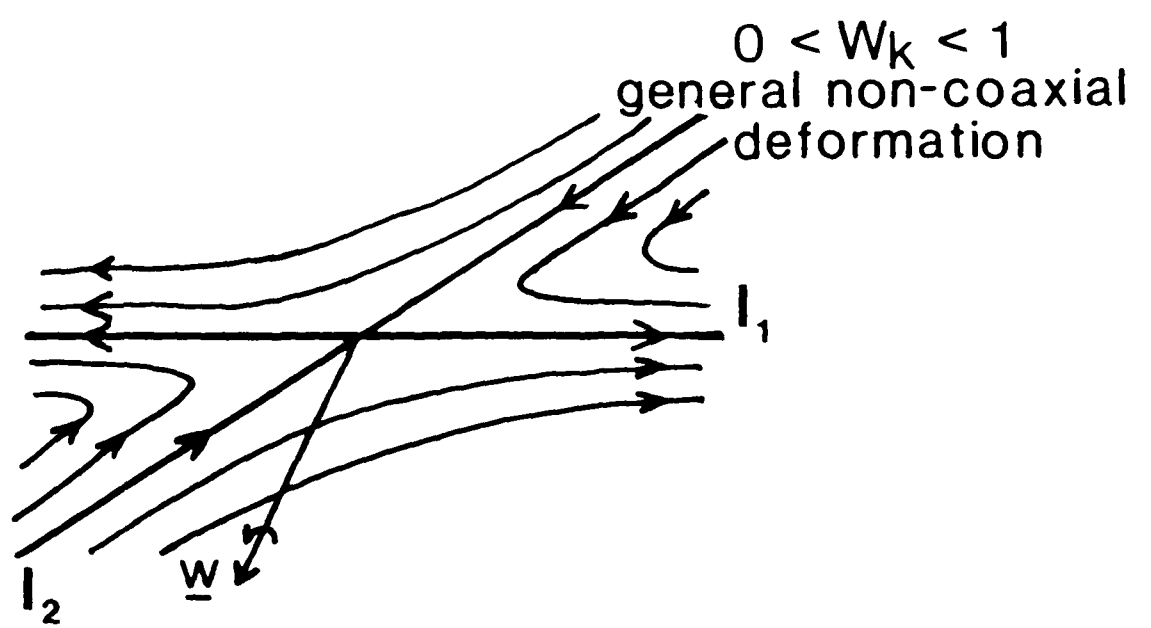
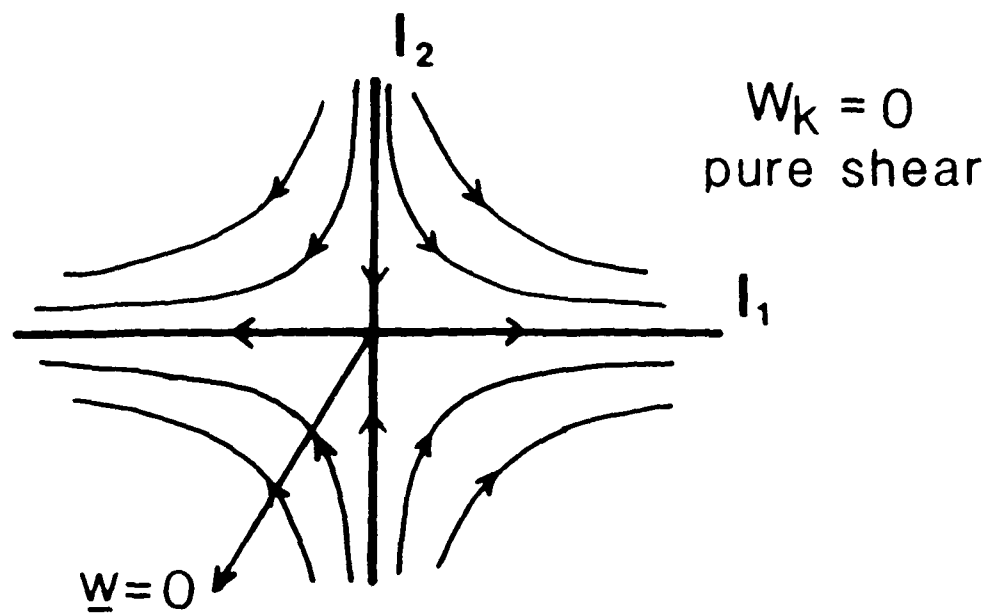
$\underline{w}$, or the angular velocity vector, $\underline{\omega}$, which have a magnitude and orientation characteristic of the rotational component of flow. The vorticity vector is defined as the curl of the velocity field, i.e. $\underline{w} = \nabla \times \underline{V}$, and $\underline{\omega} = \frac{1}{2} \underline{w}$ (Malvern 1969, p. 147, 428). In isochoric plane strain flow, $\underline{w}$ lies parallel to d_2 , the intermediate instantaneous stretching axis, which has a zero stretching rate and ^{lies} within the stream lines of the flow (fig.4.1). The sense of vorticity is the same as the sense of rotation of a rigid spherical object embedded in the deforming medium.

An important feature of flow is the degree of non-coaxiality, which can be expressed as a ratio of the magnitudes of $\underline{w}$ to the maximum principal stretching rates of D_{ij} . This is referred to as the kinematic vorticity number ($= w_k$) after Truesdell (1954), where

$$w_k = |\underline{w}| / \sqrt{2(d_1^2 + d_2^2 + d_3^2)}.$$

The different flow regimes implied by different values of w_k are most easily envisaged for isochoric plane strain deformation; for simple shear $w_k = 1$, for pure shear $w_k = 0$, and for rigid body rotation $w_k = \infty$ (fig.4.1). The vortical component of deformation is responsible for the asymmetric fabric development in rocks, and in principle the relative contributions of D_{ij} and W_{ij} can be determined by quantifying this degree of asymmetry.

Vorticity may be partitioned into a rotation of material lines relative to the d_1 , d_2 and d_3 axes of D_{ij} (internal vorticity of Lister & Williams 1983) and a rotation of these axes with respect to the external frame of reference (spin component of vorticity of Lister & Williams *ibid*). The frame of reference for most tectonic studies is the horizontal and vertical defined by the Earth's gravitational field. Due to the action of gravitational forces, bands of deformation will tend to rotate from steep to shallow inclinations with increasing strain. High-strain zones of distributed shear developed on a scale of several tens of kilometres are likely to have reached a stable horizontal



orientation, and the vorticity determined within such a zone is, therefore, likely to be representative of the bulk deformation on a large scale. However, the vorticity determined within smaller-scale shear zones may not be representative, since this excludes the rotation of the shear zone with respect to the Earth's gravitational field.

Deformation along major nappe contacts is commonly assumed to approximate to progressive simple shear (Ramsøy et al 1983). However, regions of thickened crust may become gravitationally unstable and deform under their own weight (England 1987, Platt 1986). The free unbounded surface represented by the surface topography will promote coaxial deformation (Lister & Williams 1983) causing extension and thinning. This implies that deformation driven by the action of body forces in a mountain belt is likely to be associated with a lower rotational component than simple shear, i.e. $w_k < 1$.

During progressive simple shear the shear zone boundaries remain rigid and at high strains the maximum finite stretching direction approximates to the tectonic transport direction. During progressive plane strain deformation histories with a lower rotational component the wall rocks to the shear zone must also deform. At high strain, however, the stretching lineation will still approximate to the relative displacement across the shear zone.

Deformation may be inhomogenous in both time and space causing local variation in the orientation of the kinematic axes. When considering deformation on a large scale, the problems of spatial variation can be partially overcome by considering the average properties of a regionally developed zone of deformation. The complications of temporal variation are less easily accommodated, since in geological studies only features related to the finite deformation are preserved. In general it is necessary to assume a time constant flow behaviour in order to derive the values of D_{1j} and W_{1j} . However,

in some special cases it is possible to identify structures that can be related to the kinematic development at different stages of deformation.

4.4. MESO- AND MICROSCOPIC OBSERVATIONS

The orientation of the Ds finite strain ellipsoid is reflected in the mesoscopic structures. Ss is axial planar to Ds folds, which I assume approximates to the X-Y flattening plane. The maximum principal extension direction is represented by the Ds stretching lineation, Ls, which in the Altkristallin has a consistent NW-SE trend (fig.3.3). In the Matrei Zone the orientation of mesoscopic features is generally the same but fluctuates greatly around large competent dolomite bodies (fig.3.30).

Extensional crenulations (eccs) are present in the higher strain, micaceous Ds tectonites. These are interpreted as Ds microshears which offset the Ss foliation (chapter 3). Empirical and theoretical work (Platt 1984, White et al 1980) suggests that these structures can be used as kinematic indicators, although in some cases they may be related to localized flow partitioning and be unrepresentative of the bulk flow (Platt 1984, Behrmann 1987a).

The numerical analysis of Platt & Vissers (1980) suggests that pure shear deformation is associated with a conjugate set of eccs striking at a high angle to the stretching direction, whereas progressive simple shear is associated with the development of a single set dipping in the direction of shear. Intermediate deformation histories between pure and simple shear would then be represented by the formation of conjugate sets with one dominant set indicating the bulk sense of shear.

In the study area the eccs are most clearly developed in area 1, where conjugate sets are developed, striking at a high angle to the

stretching lineation. In general one set clearly dominates over the other, suggesting a bulk non-coaxial deformation somewhere between simple and pure shear. The dominant set indicates a top to NW sense of shear (fig.4.2.a). In area 3 the eccs are less well-developed and have a more variable orientation. As well as some striking at a high angle to the stretching lineation, some of the eccs intersect Ss at a low angle to Ls. Simultaneous movement on these shears implies the extension of the Ss fabric in both the X and Y directions, i.e. non-plane strain in the flattening field. In this area, the eccs give no clear sense of vorticity (fig.4.2b), possibly reflecting a low rotational component to the deformation.

The amount of offset along each ecc is relatively small and the total strain that can be accommodated by movement on these micros shears is also small. These structures, therefore, probably develop during the final stages of deformation and their geometries reflect only this last strain increment. In both areas the eccs give results consistent with other kinematic indicators.

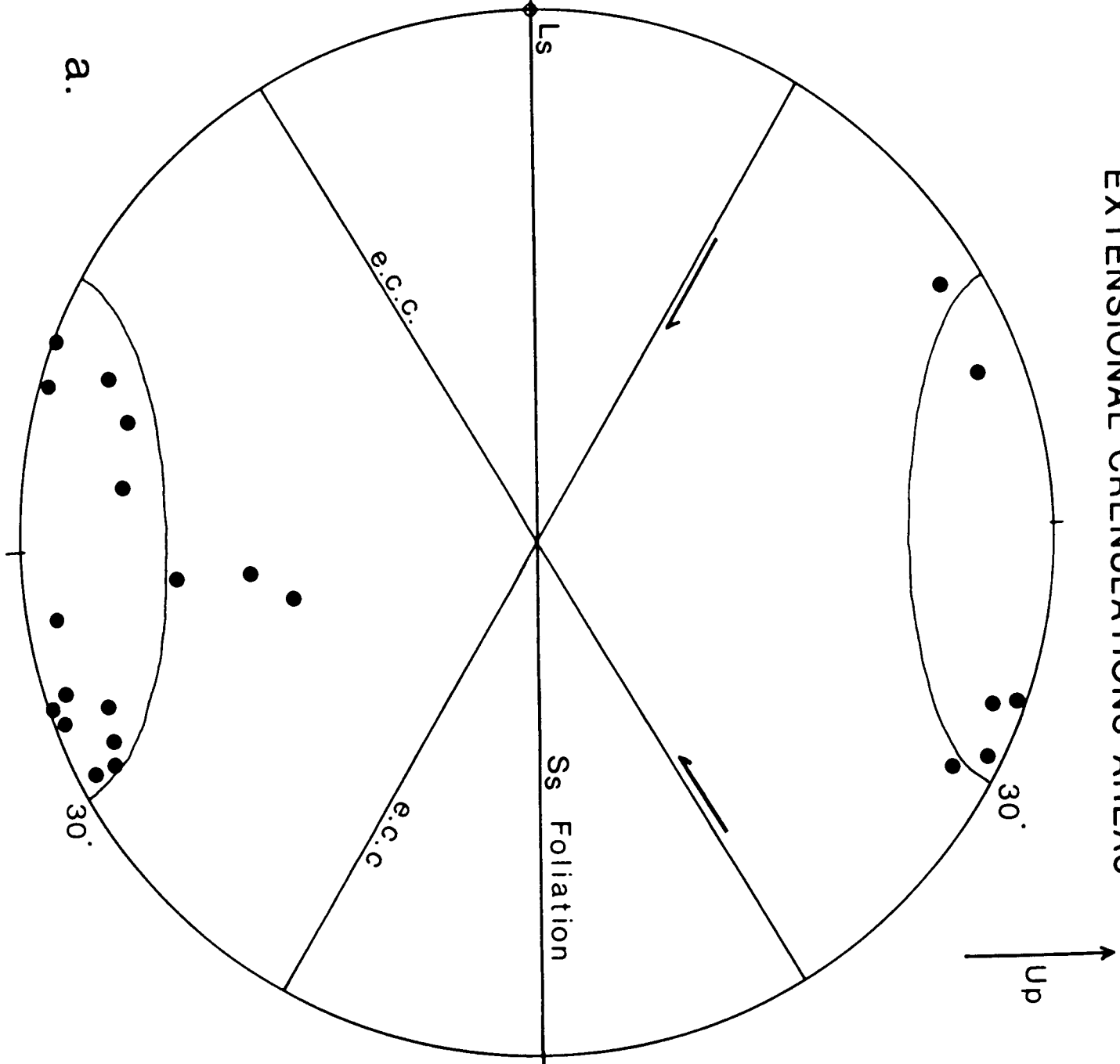
In thin sections cut parallel to the Ds stretching lineation, there are a number of asymmetric microstructural features, which can be used to determine the sense of vorticity.

Rigid porphyroblasts of garnet and albite locally develop asymmetric pressure shadows. These are rarely very pronounced but indicate a consistent top to the NW sense of shear (Passchier & Simpson 1987, fig.4.3). A more clear indication of the same top to the NW sense of shear is given by the rotation of the albite long axes with respect to the main Ss foliation (fig.4.4).

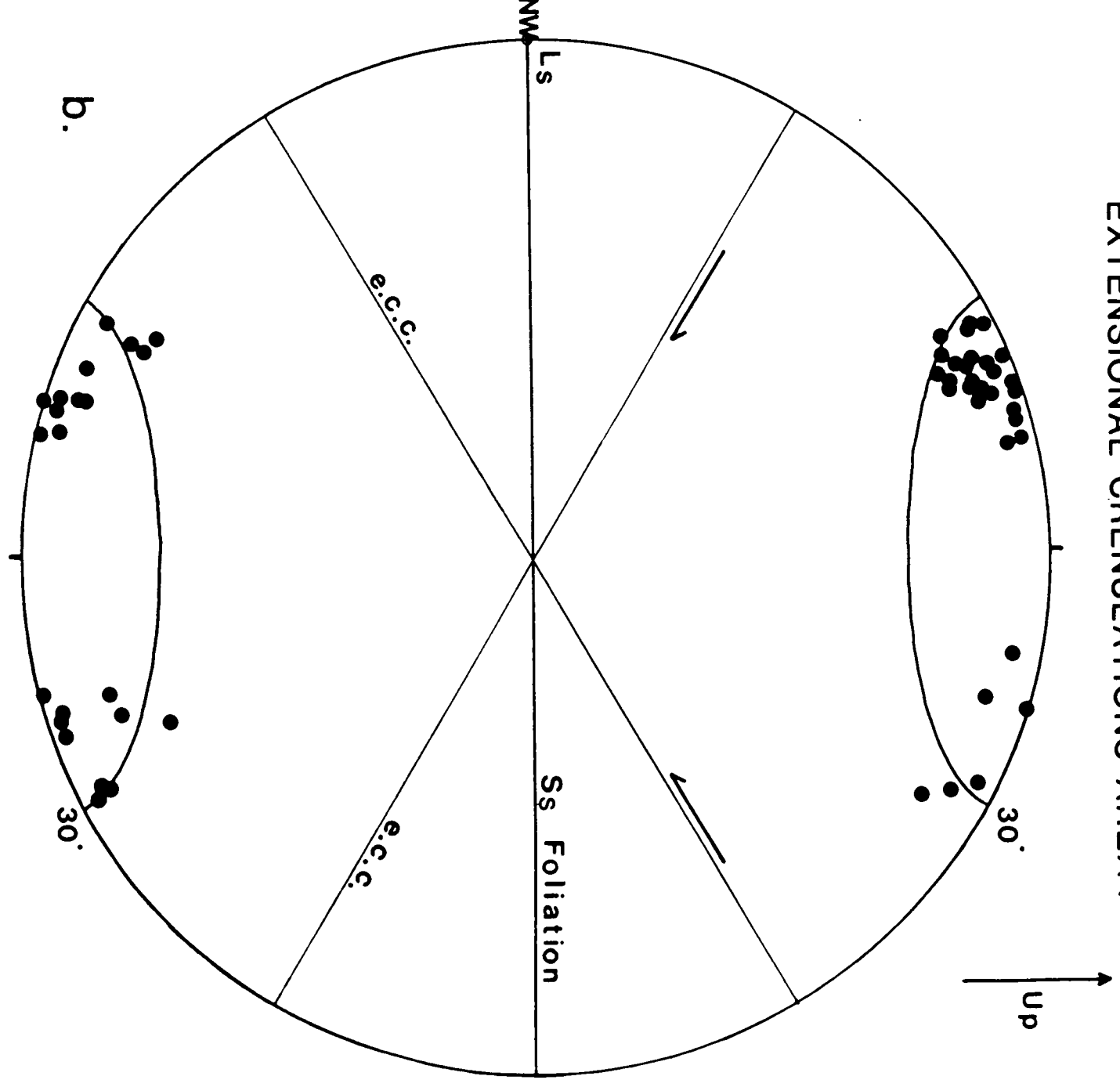
The quartz-rich Ds tectonites commonly have a subgrain fabric at a high angle to the foliation, which is locally accompanied by an oblique grain shape fabric (fig.4.5). Both these features indicate a top to the NW sense of shear (Simpson & Schmid 1983). A strong crystallographic

Fig.4.2. Poles to extensional crenulation cleavages plotted relative to Ss and Ls on a Schmidt net. Each dot represents the average of 3 to 5 measurements. 30° small circles and the implied sense of shear for two eccs at 30° to Ss are shown for reference.

EXTENSIONAL CRENULATIONS AREA 3

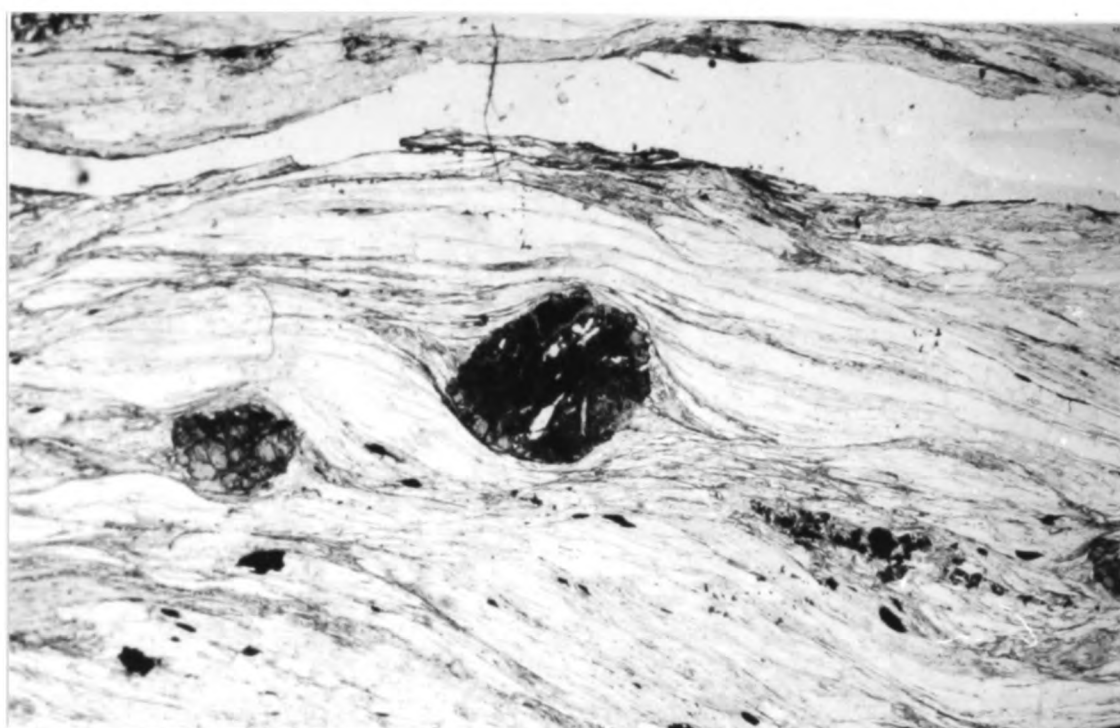


EXTENSIONAL CRENULATIONS AREA 1



NW

SE



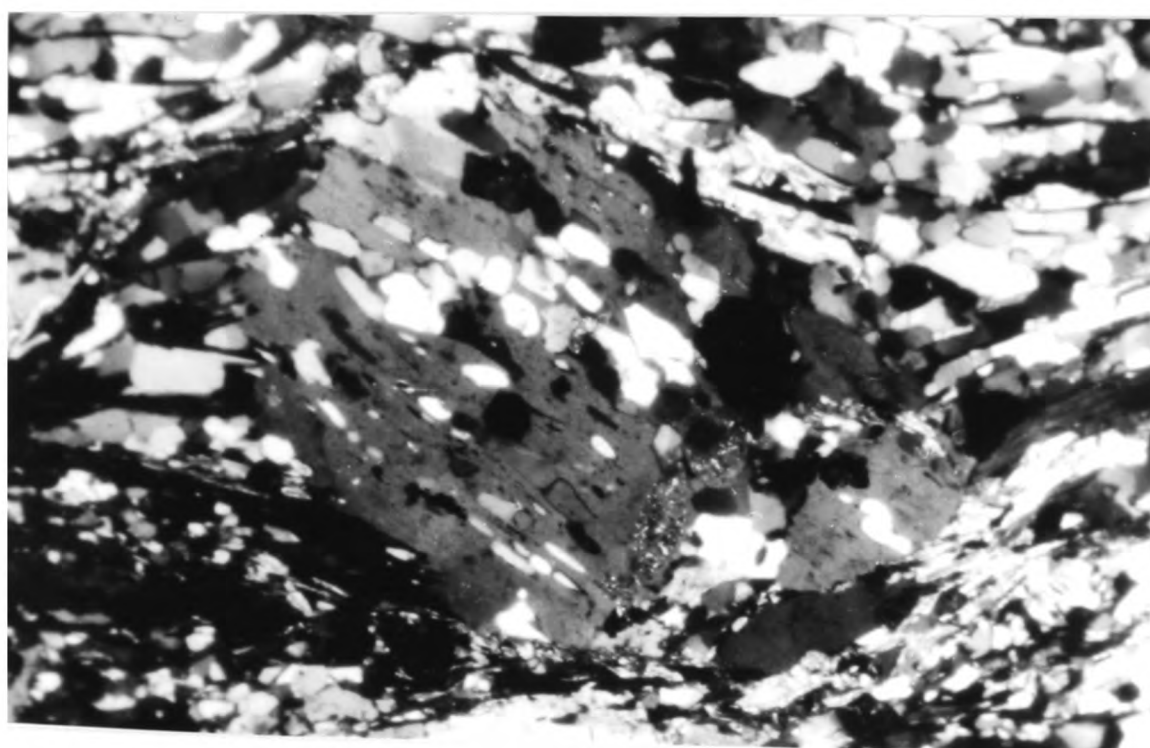
↑
Up

3mm

Fig.4.3. Rotated garnets with asymmetric pressure shadows. B129 area 1.

SE

NW



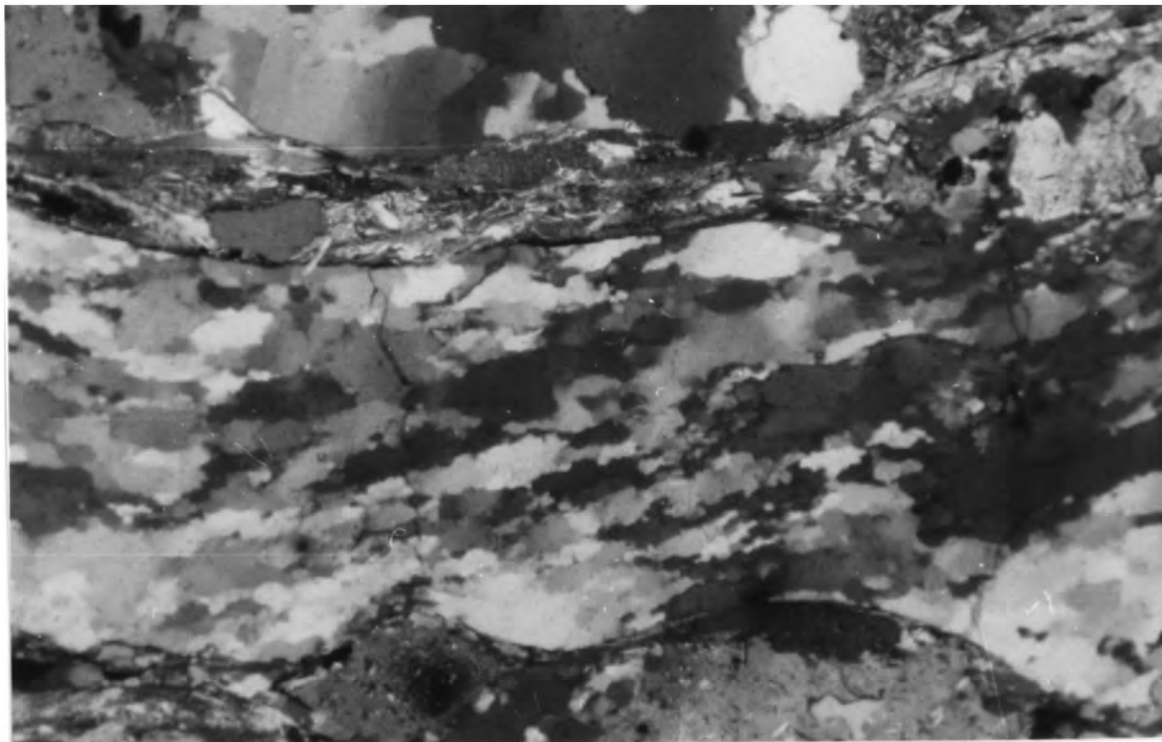
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Fig.4.4. Albite porphyroblast rotated with respect to Ss and containing inclusion trails of Sr. MS63 area 3.

SE

NW



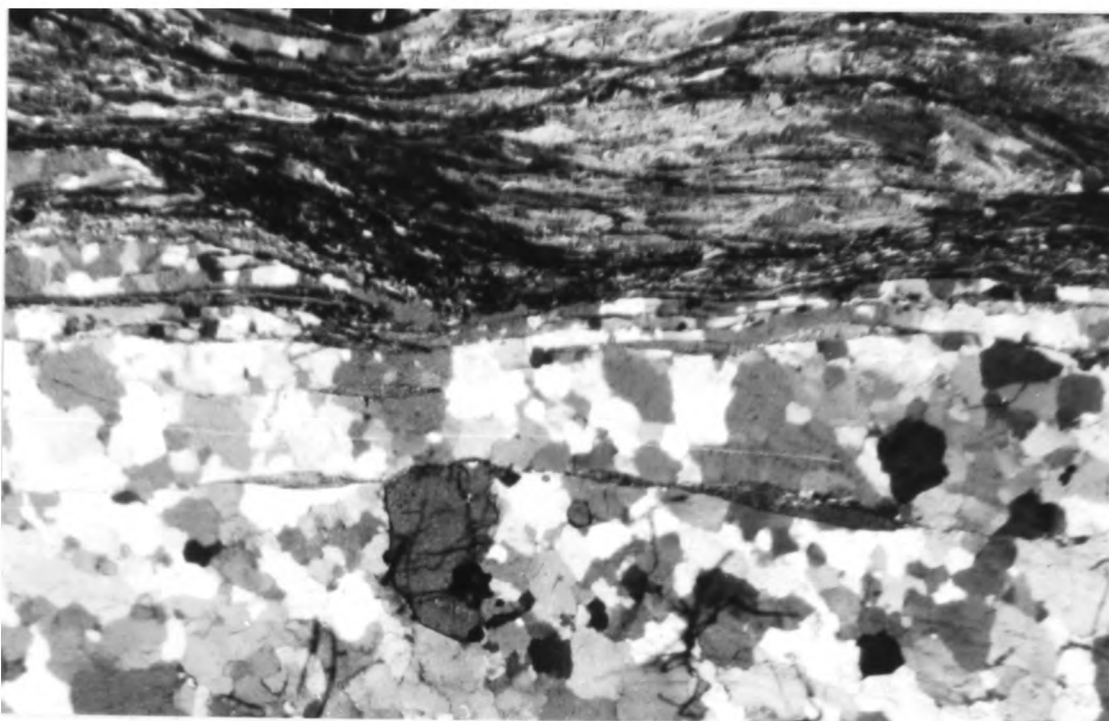
↑
Up

1mm

Fig.4.5. Oblique subgrain fabric in a deformed quartz vein, crossed nicols with a sensitive tint plate. K5 area 1.

NW

SE



↑
Up

2mm

Fig.4.6. Asymmetric micro-boudinage of white mica forming in conjugate sets, suggesting a top to the NW sense of vorticity.

preferred orientation is developed in many of these quartz-rich tectonites and the c- and a-axis preferred orientation patterns also indicate a top to NW movement (see 4.5).

Asymmetric micro-boudinage of white mica is present in some samples. This feature develops in conjugate sets, with the dominant set indicating a top to the NW sense of shear (fig.4.6).

The same top to the NW tectonic transport is indicated throughout the zone of distributed Ds shear. For all these kinematic indicators the sense of shear is more obvious in area 1 than in area 3, implying a greater magnitude of the vorticity vector in area 1 than in area 3. This suggestion is borne out by a qualitative study of the rotation of the albite porphyroclasts with respect to their aspect ratios (see 4.6).

4.5. QUARTZ FABRICS

4.5.1. Introduction

Quartz-rich tectonites commonly develop crystallographic preferred orientation patterns, as the result of slip along certain crystallographic planes. The nature of the resultant preferred orientation pattern (fabric) depends primarily on three factors (e.g. Schmid & Casey 1986):

- (i) the degree of non-coaxiality,
- (ii) the type of strain, i.e. flattening, plane or constrictional, and
- (iii) the slip planes that are operative through deformation.

The preferred orientation patterns are commonly used to determine the sense of vorticity shear zones. Quartz fabrics may, however, also be used to make qualitative statements about ancient flow regimes in naturally deformed quartz-rich tectonites.

A complete fabric analysis of quartz includes a, c, m, r, z axes (Schmid & Casey 1986, Law 1987), but c- and a-axis fabrics are the most commonly used diagrams. Only the c-axes of quartz can be measured optically on a universal stage, all other crystallographic orientations have to be determined by other methods, e.g. X-ray goniometry or back scatter S.E.M. (Lloyd et al 1987).

4.5.2. Interpretation of c- and a-axis fabrics

During non-coaxial deformation quartz fabrics generally develop asymmetrically with respect to the principal axes of finite strain (Lister & Williams 1979, Schmid & Casey 1986). The few exceptions that have been recorded (e.g. Garcia Celma 1983) are in relatively low-strain rocks and probably involve spinning deformation paths (Lister & Williams 1983). Both theoretical and empirical evidence suggests that under non-coaxial plane-strain conditions quartz c-axis fabrics develop a central girdle inclined in the direction of shear with respect to the axes of finite strain, X,Y and Z (Lister & Hobbs 1980, Carreeras et al 1979, Lister 1977). In most measured samples the density distribution of c-axes has a similar asymmetry, which can also be used to determine the sense of vorticity. The corresponding a-axis fabrics have maxima distributed within the X-Z plane, the two most prominent may be aligned with the slip direction during deformation (Bouchez & Pecher 1981) and the sense of vorticity can then be determined from the asymmetry of these maxima from the X-Y plane. The angle between the a-axis maxima and the finite strain axes is, however, commonly too great for this simple relationship to be true (e.g. Law 1987)

Topology and density distribution are two independent features of quartz fabrics, both of which can be used to determine the sense of vorticity. However, the computer modelling of Lister & Hobbs (1980) shows that every starting crystal orientation is associated with a

unique position in the final fabric. An initial preferred orientation within the deformed material will, therefore, complicate the relationship between the deformation and the density distribution within the fabric (Behrmann & Platt 1982). Fabrics developed from such material are likely to have an uneven distribution of points within individual fabrics, which are inconsistent from one fabric to the next. These effects may be progressively overridden by dynamic recrystallization.

The degree of non-coaxiality has a strong influence on the topology of both the c- and a-axis diagrams. Studies in shear zones with known geometries, demonstrate that simple shear is characterized by a single inclined girdle of c-axes associated with two dominant point maxima in the a-axis fabric (Schmid & Casey 1986). Computer modelling of Lister & Hobbs (1980) suggests that pure shear deformation produces symmetrical c- and a-axis fabrics and that the c-axes define a cross girdle pattern. Fabrics very similar to those predicted by Lister & Hobbs have been measured in quartzites of the Moine Schists in N.W. Scotland (Law et al 1986). Locally these quartzites contain porphyroclasts of quartz with their c-axes at a high angle to the foliation, termed globular grains by Law et al (1984). Under the lower greenschist conditions of these samples, the basal plane of quartz is the dominant slip plane (Schimid & Casey 1986). The lack of deformation within these grains can be explained if they are oriented so that the resolved shear stress along these slip planes was at a minimum. If deformation had been rotational they would have been rotated with respect to the stress field, eventually coming into a new orientation suitable for easy glide. Globular grains are, therefore, taken as independent microstructural evidence for a low degree of non-coaxiality. However, a complete fabric analysis of samples from similar structural positions reveals an asymmetry in the r- and z-axis fabrics (Law 1987), suggesting that the

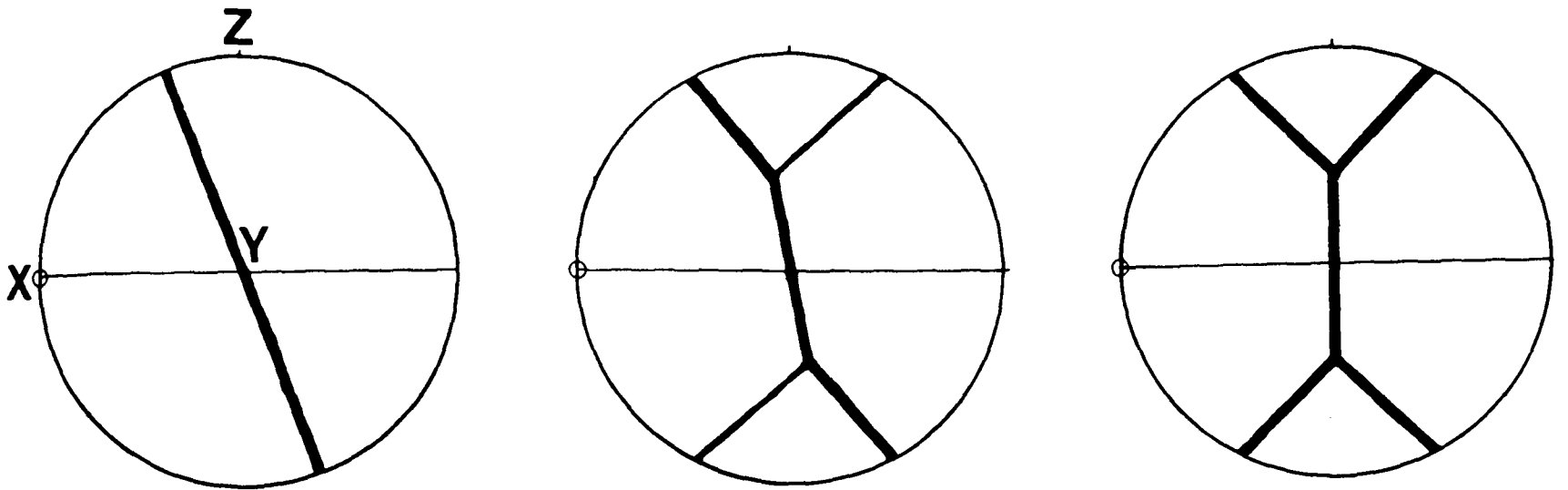
deformation in these samples cannot be properly described as coaxial. Orthorhombic quartz c-axis diagrams may not, therefore, be indicative of coaxial deformation, but the evidence suggests they are related to a low degree of non-coaxiality.

Many quartz fabrics have topologies intermediate between those predicted for simple and pure shear. These can be interpreted as the result of general non-coaxial plane strain deformation (fig.4.7, Schmid & Casey 1987, Platt & Behrmann 1986), the asymmetry indicating the sense of vorticity. There has as yet been no solution to the problem of quantifying the degree of non-coaxiality from the quartz fabrics alone.

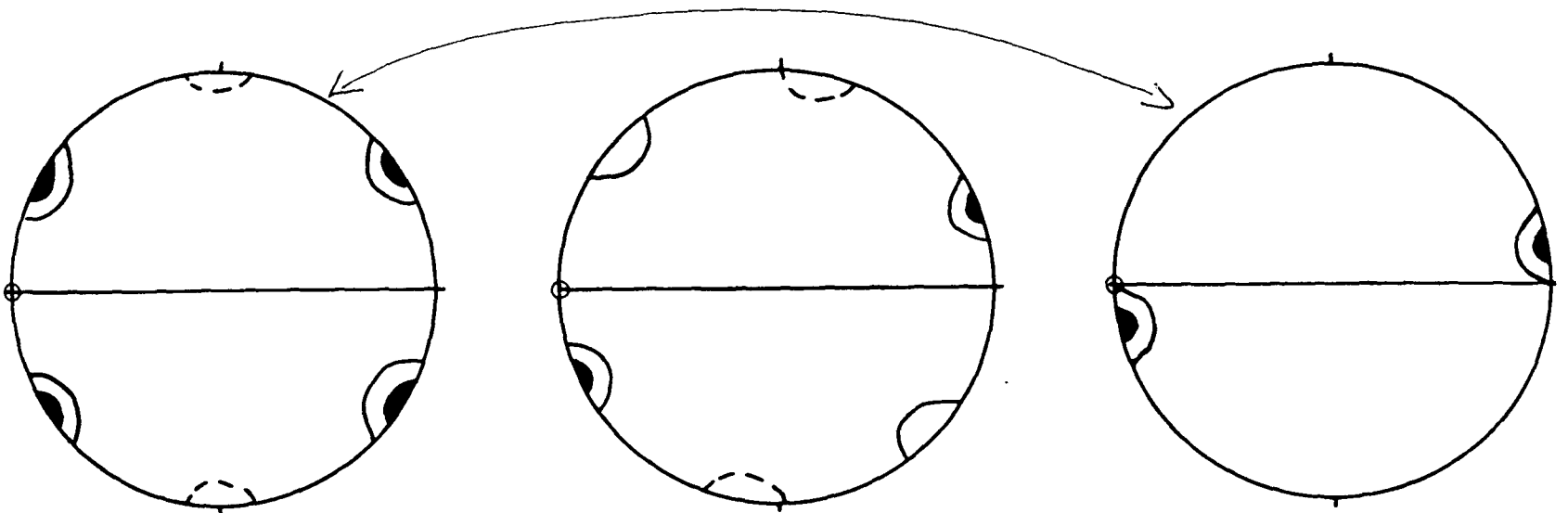
The computer modelling of Lister & Hobbs (1980) shows that the central girdle of a c-axis fabric forms perpendicular to flattening plane in pure shear and the shear plane in simple shear, both of which represent planes of zero instantaneous angular velocity (l_1 in fig.4.1). By analogy Platt & Behrmann (1986) suggest that during progressive plane strain deformation intermediate between these two extremes the central c-axis girdle may also form perpendicular to l_1 . The two planes of zero instantaneous angular velocity, l_1 and l_2 , will rotate with respect to the X - Y plane of the finite strain ellipsoid in a predictable manner; the angle between these planes is then a function of the kinematic vorticity number, w_k , and the finite strain (Passchier 1988). At high finite strain the central girdle should, therefore, be perpendicular to the flattening plane, irrespective of w_k . However, inclined fabrics are still recorded in high-strain shear zones (Carreras et al 1977), suggesting that dynamic recrystallization can maintain this angle at less than 90°. The obliquity of a central c-axis girdle should, however, give a minimum estimate for the angle between l_1 and the X - Y plane.

Non-plane strain conditions are reflected in the development of small circle distributions of both c- and a-axes (fig.4.8, Tullis et al

Increasing rotational component

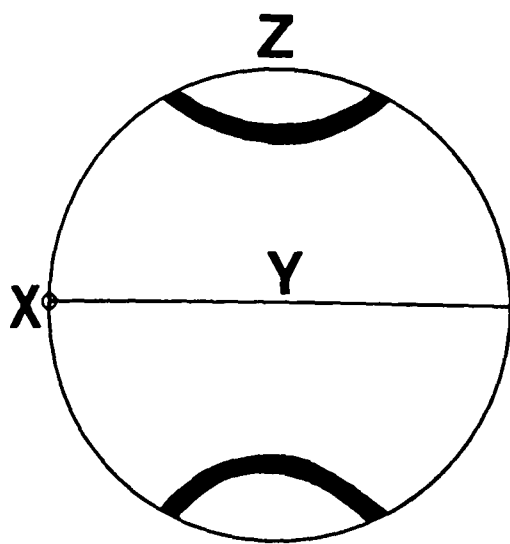


Plane Strain Quartz C-Axis Diagrams

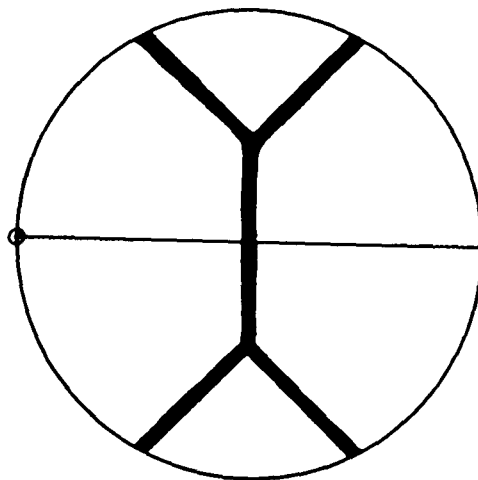


Plane Strain Quartz A-Axis Diagrams

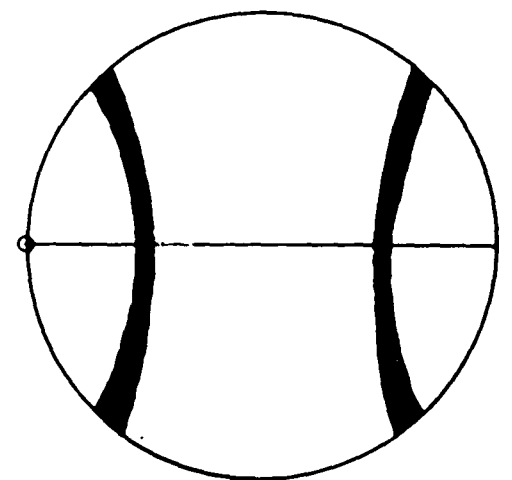
Fig.4.7. Plane strain quartz c- and a-axis fabrics. The c-axis diagrams are represented by the skeletal outline and the a-axis diagrams by schematic contour intervals (after Schmid & Casey 1986).



Flattening Strain

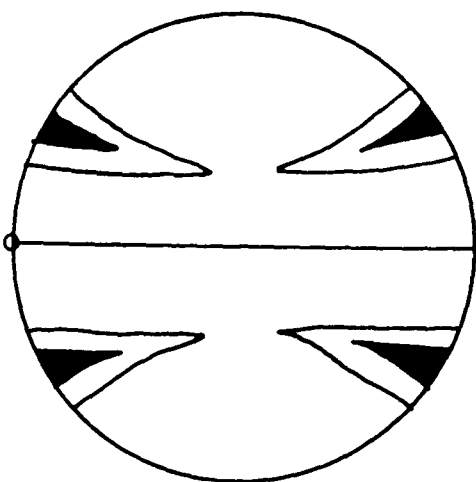


Plane Strain

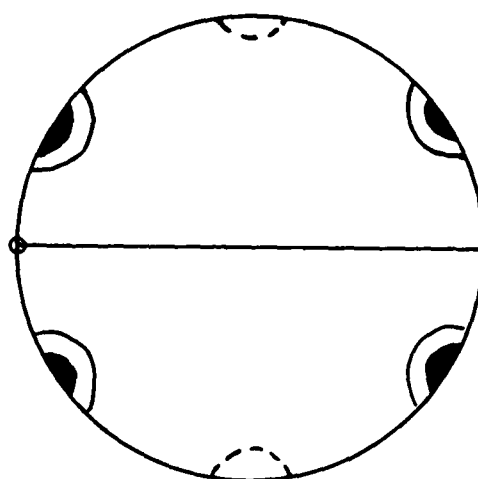


Constrictional Strain

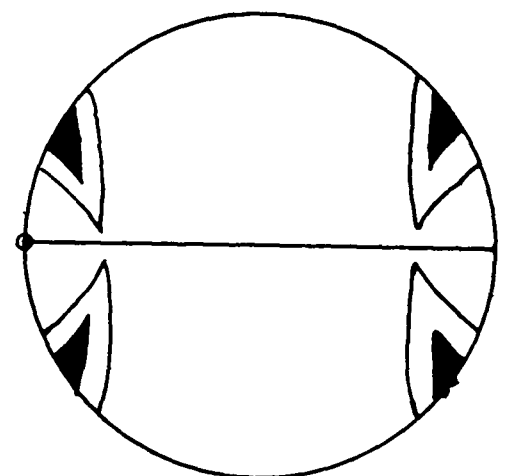
Quartz C-Axis Diagrams



Flattening Strain



Plane Strain



Constrictional Strain

Quartz A-Axis Diagrams

Fig.4.8. Coaxial quartz c- and a-axis diagrams for plane, constrictional and flattening strain. Quartz c-axis diagrams are represented by the skeletal outlines and the a-axis diagrams by schematic contour intervals (after Schind & Casey 1986).

1973, Schmid & Casey 1986). These small circles are centred about Z for coaxial non-plane strain in the flattening field and about X in the constrictional field. There has been no systematic study of the complex conditions represented by non-coaxial non-plane strain conditions. By analogy with the other examples, such a deformation history could be represented by small circle distributions of c- and a-axes asymmetrically arranged about the axes of the finite strain ellipsoid.

4.5.3. Sampling techniques

Quartz-rich tectonites were sampled in map areas 1 and 3 throughout the zone of distributed Ds shear in the base of the Altkristallin and the Matrei Zone. Most of the data are c-axis fabrics, although some a-axis fabrics were also measured courtesy of Dr S.Schmid, E.T.H., Zürich. The main aim of these measurements was to obtain a qualitative description of flow during Ds, including the sense of vorticity and the degree of non-coaxiality.

Samples were all >90% quartz and were dominantly deformed vein quartz material although metaquartzites were also sampled to test for systematic differences in fabric with rock type. The greater proportion of impurities in the metaquartzites reduced the intensity of the preferred orientation, but the topology and relative density distribution were unaffected. Some areas were sampled in several places in close proximity to one another to ensure the measured patterns were characteristic for a larger volume of deformed rock.

The thin sections used to measure the c-axes were cut parallel to the mesoscopic stretching lineation, Ls, and perpendicular to the schistosity, Ss. In most cases there was no direct evidence other than the quartz fabric itself, to suggest whether this planar fabric should be interpreted as an X-Y fabric or a fabric developed parallel to a plane of high shear strain.

The symmetry elements of the quartz fabric topologies are consistently oriented parallel to the principal axes of finite strain, defined by the Ds mesoscopic foliation and lineation. In thin section Ss is dominantly defined by the preferred orientation of platy minerals which is commonly accentuated by a grain-shape fabric in the quartz. The consistent quartz microstructures within both the quartz-rich and the platy mineral-rich Ds tectonites and the consistent relationship of the quartz fabric topology with respect to the Ds mesoscopic structures, both suggest that the quartz preferred orientation in the quartz-rich layers developed as a result of Ds deformation.

Near the base of the Altkristallin and within the Matrei Zone the Ds microstructure has been annealed during the Tauern Metamorphism. The quartz fabrics do not, however, show any systematic change with degree of annealing. The fabrics measured within the Matrei Zone are generally weaker than those from samples of the Altkristallin. This probably reflects both the greater impurity of the samples (Lisle 1985) and a change to deformational processes other than intracrystalline plasticity (see chapter 3). The interpretation of the quartz fabrics is, therefore, based mainly on the data from deformed vein quartz material in the Altkristallin.

The grains to be measured within a particular section were selected by taking the grain half-way along the cross-wires from the first measuring point. This avoids unconscious preferential selection of any particular crystal orientation. For a few samples, sections perpendicular to the stretching lineation were also measured and after suitable rotation, combined with the data from a section parallel to the stretching direction. In these cases the fabrics showed no significant differences, implying the cut-effect (Sander 1950) did not have a major influence on the fabrics.

The data is represented as contoured diagrams plotted with respect to the stretching lineation and the foliation.

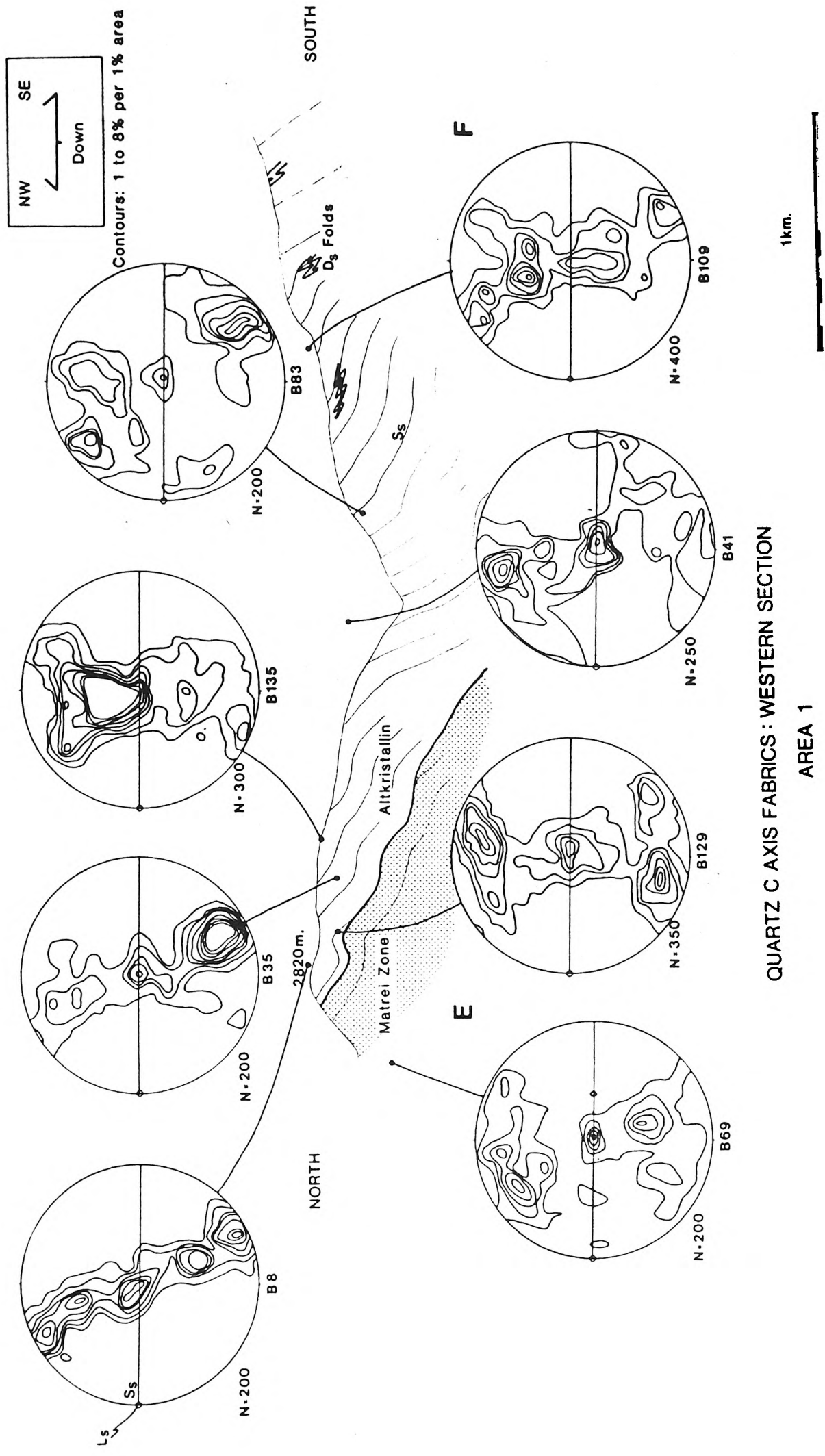
4.5.4. Map Area 1 - Western Area, (fig.4.9)

The fabrics all have an asymmetry in topology or density distribution, with a central girdle at a high angle to the foliation. There is a general increase in preferred orientation from high structural levels to low. The features of these diagrams can be readily interpreted as the result of bulk non-coaxial deformation with a top to the NW sense of vorticity, with increasing strain towards the base of the Altkristallin.

One fabric (B8) has a single inclined girdle, indicative of simple shear. However, all the other fabrics are kinked to a greater or lesser degree, and have cross girdle topologies. The generally well-developed central girdle implies that deformation was plane strain and comparing these to the fabrics shown in fig.4.7. suggests that they represent a general non-coaxial deformation with a kinematic vorticity number $0 < w_k < 1$, i.e. between simple and pure shear. The conclusion that Ds in this area was associated with a kinematic vorticity number less than 1, is supported by a quantitative analysis of the rotation of albite porphyroblasts and the orientation of eccs (sections 4.6, 4.4).

The quartz c-axis fabrics suggest that one sample, B8, is associated with simple shear, and it is surrounded by material characterized by deformation with a lower rotational component. Deformation within a shear zone can be described as the sum of a simple shear deformation parallel to the shear zone boundaries and a homogenous deformation which affects both the shear zone material and the wall rocks (Cobbold 1977). The deformation within B8 approximating to simple shear is compatible with the inference that the surrounding material underwent deformation with a lower rotational component if B8 is

Fig.4.9a. Representative quartz c-axis diagrams for area 1 presented on a cross-section through the area. The imbrication of the Matrei Zone quartz phyllites and the Altkristallin is not shown for the sake of clarity. Fig.3.40., 3.41., 3.42. show the structure in more detail. See fig.4.9c. for location of section and samples.



QUARTZ C AXIS FABRICS : WESTERN SECTION
AREA 1

Fig.4.9b. Ds quartz c-axis data not represented on the cross-section. Contours 1 to 8% per 1% area, shaded area <1% contour within the diagram. See fig.4.9c. for location of samples.

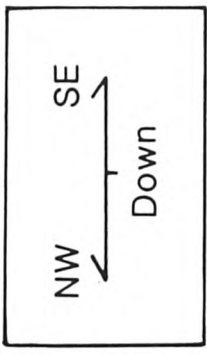
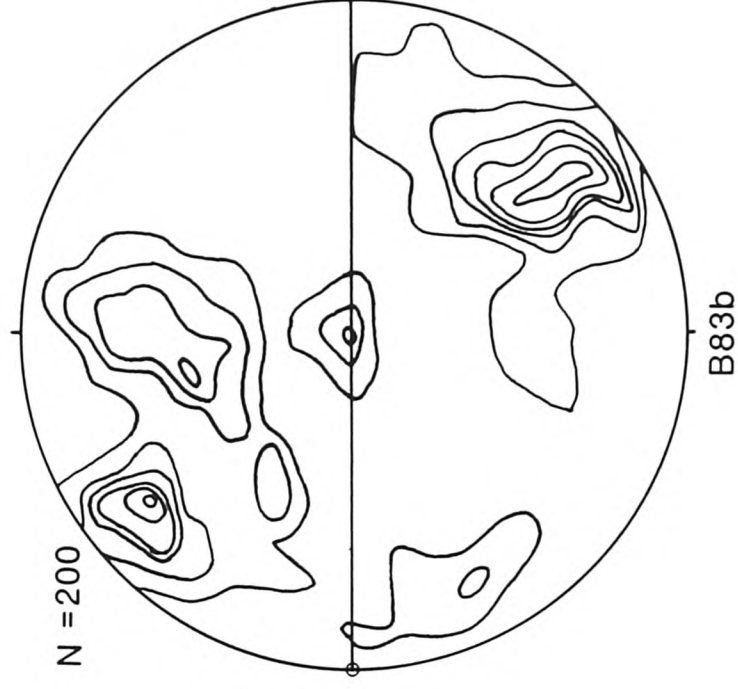
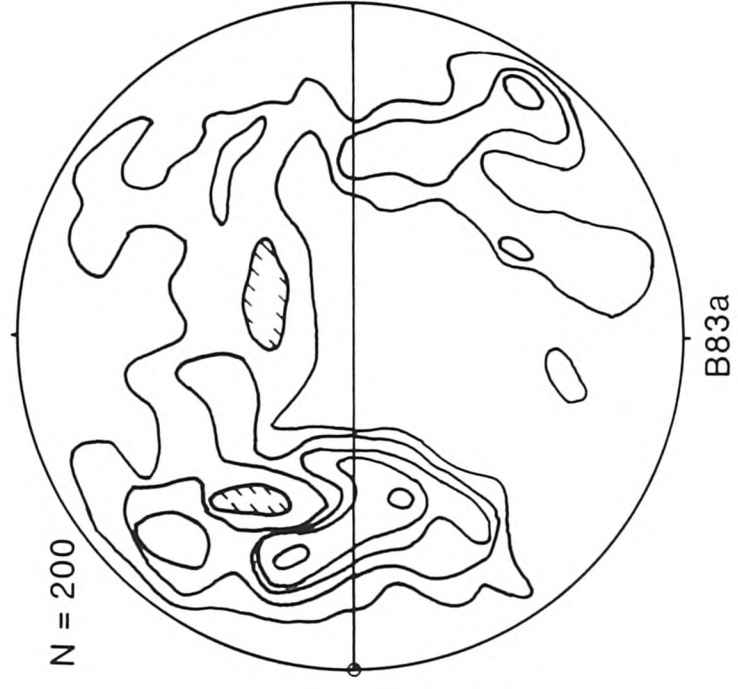
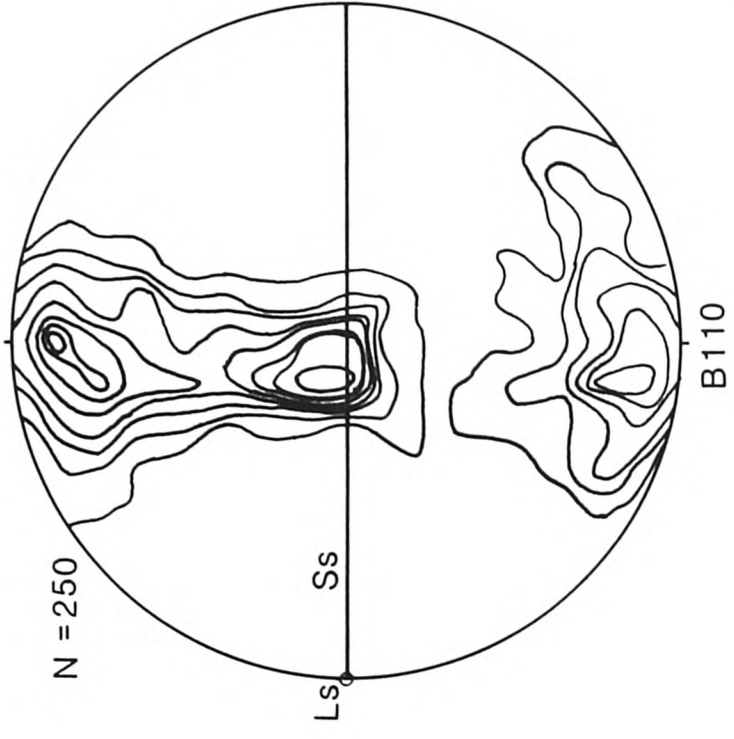
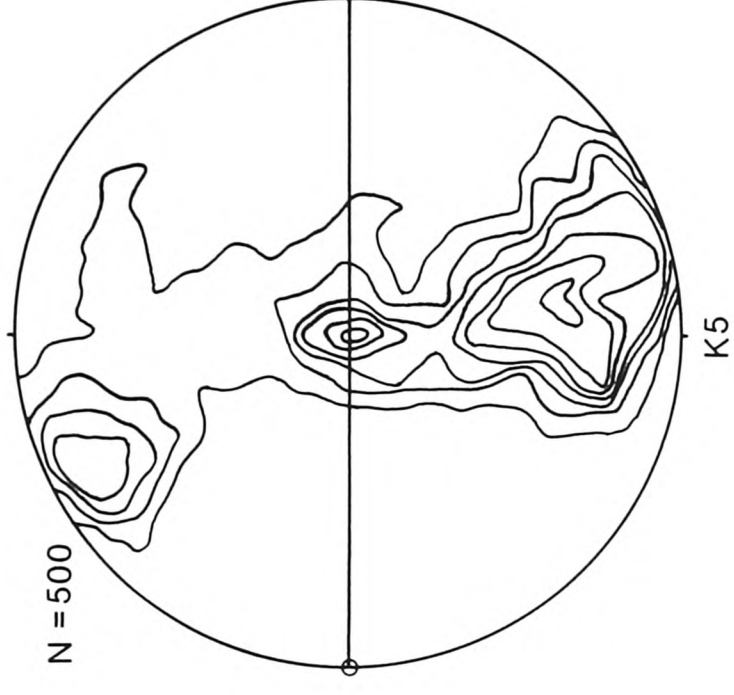
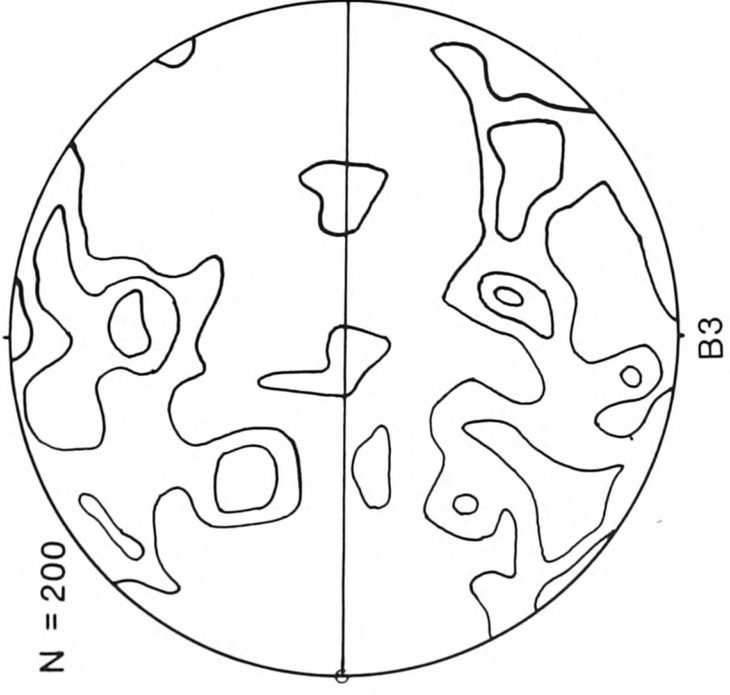
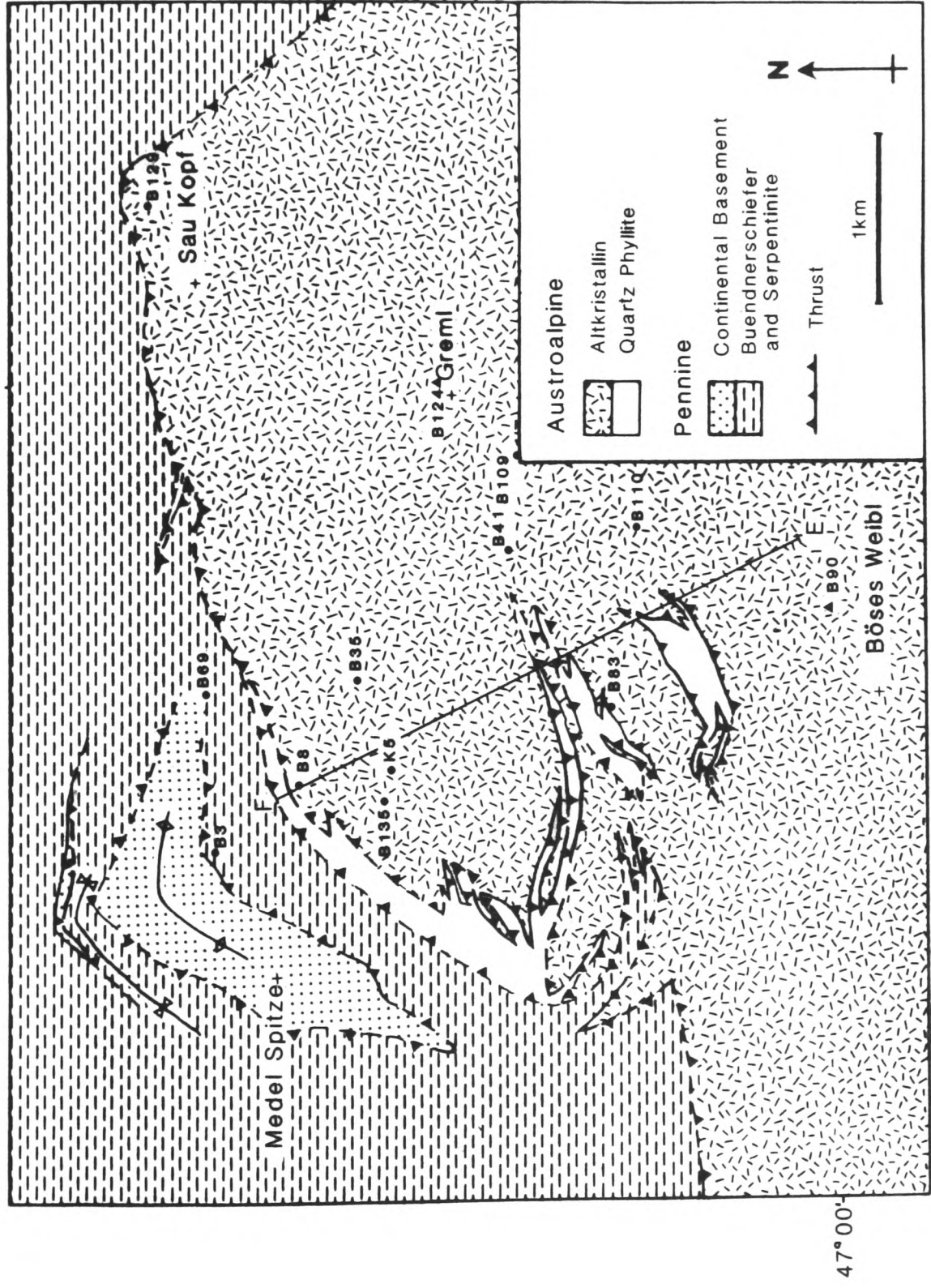


Fig.4.9c. Simplified geological map of map area 1 showing the main structural divisions and the location of the samples for the quartz fabrics and the cross-section shown in fig.4.9a, fig.4.9b. ● = Ds fabric, Δ = Dr fabric.

12° 45'



associated with high strain. The homogenous component of deformation common to both B8 and the surrounding material may then represent only a negligible departure from the simple shear for this sample. The well-developed and evenly distributed c-axis fabric suggests that dynamic recrystallization has overridden the effects of an initial preferred orientation that is seen in the other samples (see below) and supports the suggestion that B8 is associated with high finite strain.

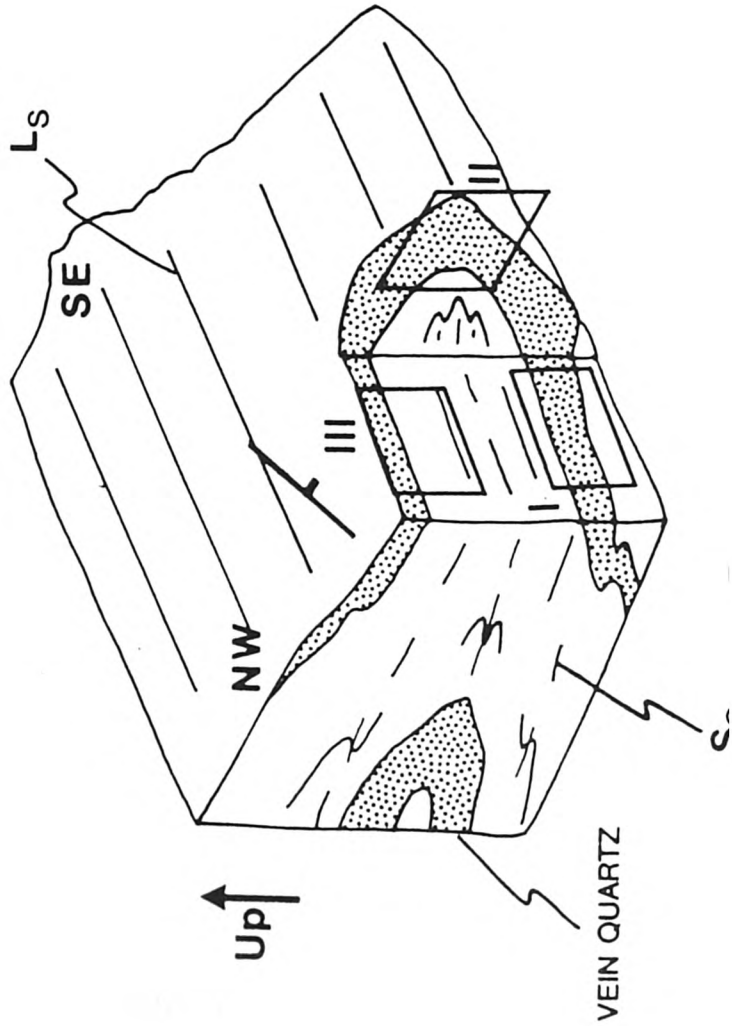
Many of the fabrics show a high concentration of points in one particular part of the diagram, e.g. B35. This can be explained as the result of an initial preferred orientation before Ds. Dr quartz fabrics can be directly measured within some porphyroblasts which contain Sr inclusion trails (samples K5, B90) and to the south of the zone affected by high strain Ds deformation (sample B124). The effects of the Dr preferred orientation were investigated by measuring several fabrics around a folded quartz-rich unit in sample B109. The fabrics from different parts of the fold have maxima in different parts of the final synoptic diagram (fig.4.10), demonstrating how a fabric with an asymmetric density distribution may be formed.

The two B83 samples have distinct and unusual quartz c-axis fabrics. These samples come from adjacent layers of deformed vein quartz material both with a strong stretching lineation. These two layers are separated by a 1-2mm thick layer of chlorite .

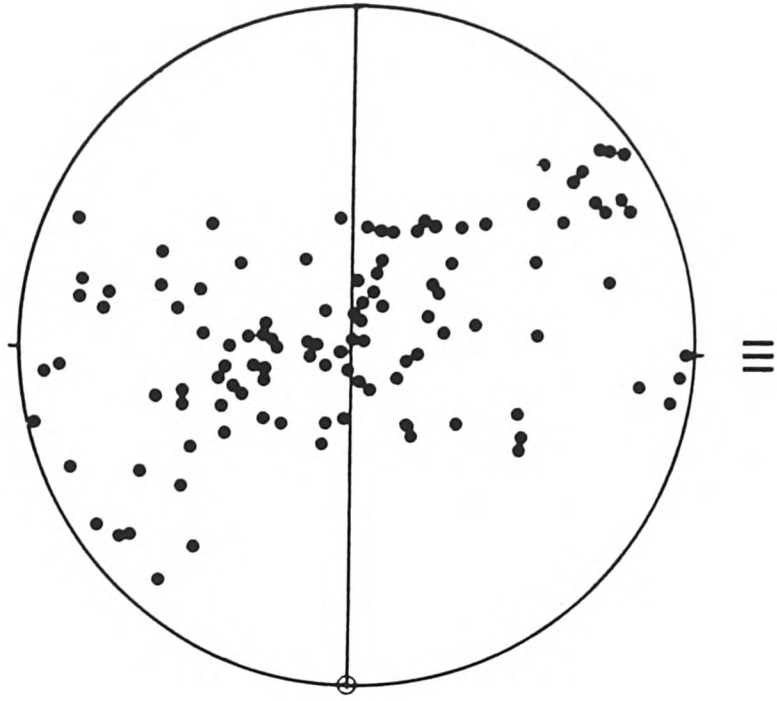
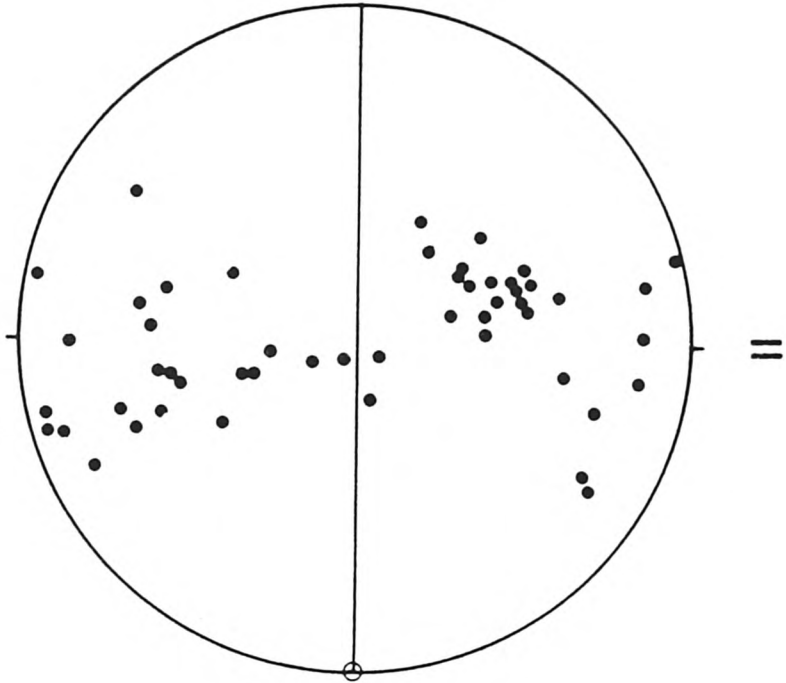
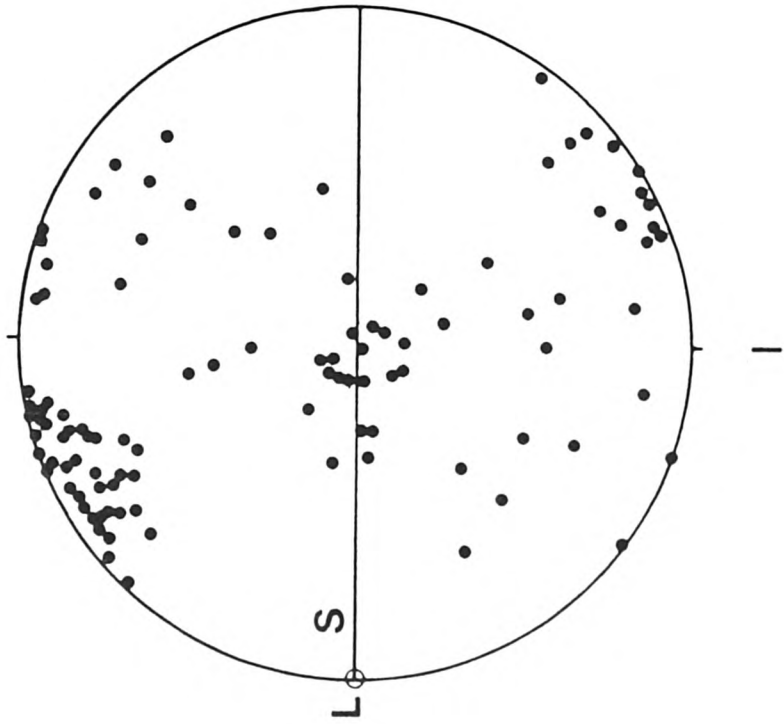
B83a has no central girdle, suggesting a departure from plane strain conditions. The main concentrations of c-axes lie on small circles symmetrically arranged around X, which implies coaxial deformation in the constrictional field. B83b has a concentration of c axes around Y but also lacks a central girdle. The c-axes define part of a small girdle distribution asymmetrically arranged around Z. This may suggest a non-coaxial deformation in the flattening field. The

Fig.4.10. Quartz c-axis fabrics measured in a deformed quartz vein around a tight Ds fold. Three samples from different parts of the deformed vein have concentrations of c-axis in different parts of the final diagram, suggesting that fabric development is controlled to an extent by the pre-Ds preferred orientation. The synoptic diagram is given in fig.4.9a with more data points collected in sections parallel to Ls.

QUARTZ C-AXIS DIAGRAMS MEASURED AROUND A D_s FOLD



NW SE
DOWN



sense of asymmetry implies a top to NW sense of shear, which is consistent with other kinematic indicators for Ds.

Deformation in the other samples in area 1 shows no significant departure from plane strain conditions. The close spatial association of the two samples suggests that they could represent localized strain partitioning into two distinct non-plane strain deformation histories, which combined represent the same bulk deformation recorded elsewhere in the area. This interpretation suggests that deformation may have been heterogeneous on the local scale, in a similar way to that suggested by Law (1987) on the scale of a thin section. In neither case is there any indication how the resulting strain incompatibilities might have been overcome.

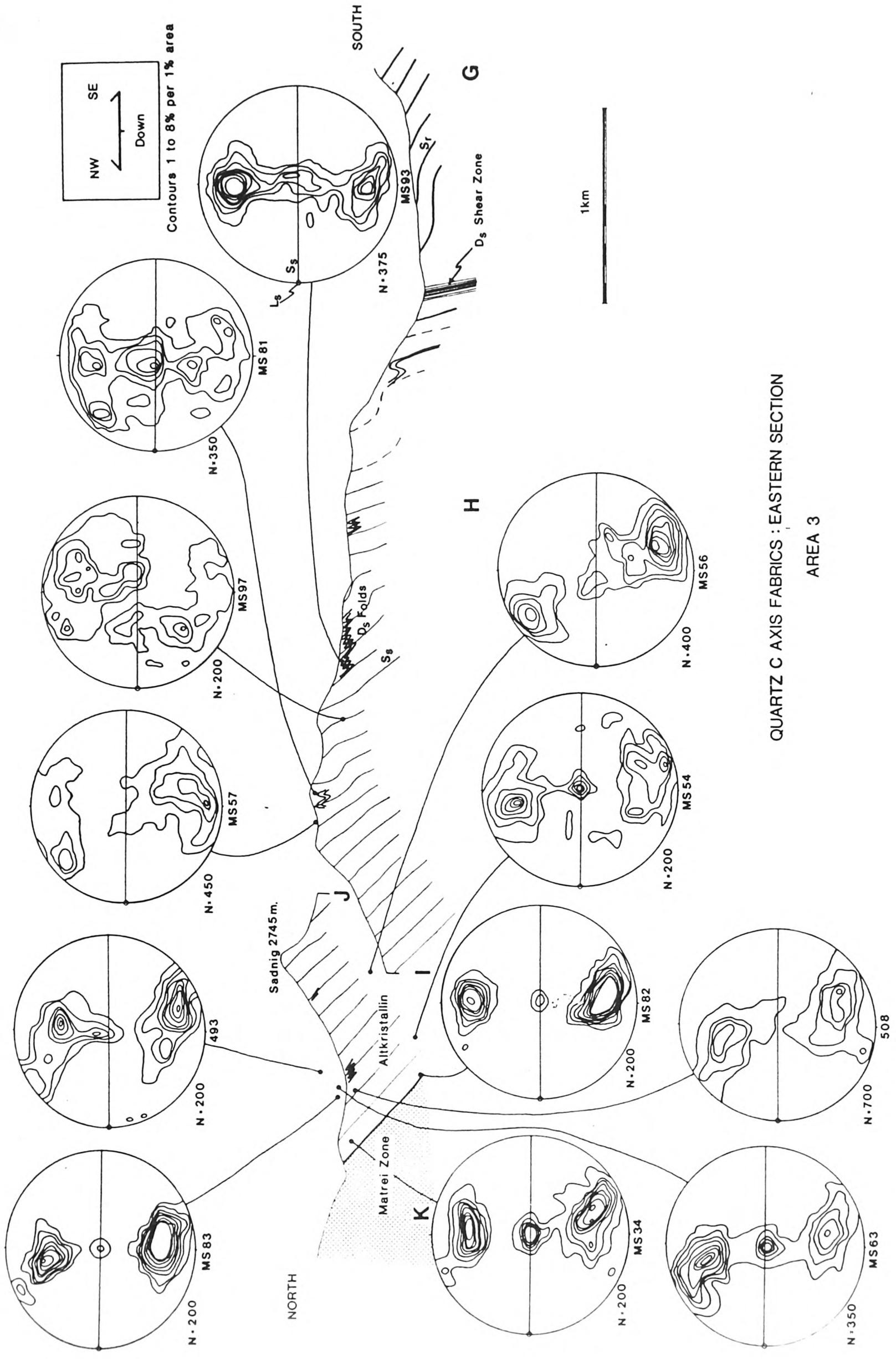
4.5.5 Map Area 3 - Eastern Area (fig.4.11)

In common with the western area, these quartz fabrics show an increase in intensity towards the base of the Altkristallin reflecting a corresponding Ds strain gradient. However, both the fabric topologies and the nature of density distribution are quite distinct from the western area.

To the south of the zone of Ds deformation MS97 has a random fabric. MS93 shows no obvious asymmetry in either topology or density distribution, which suggests deformation with a low degree of non-coaxiality. Closer to the basal contact MS81 and MS57 both have asymmetric fabrics which could represent incomplete cross-girdle fabrics, and suggest top to the NW sense of vorticity. These fabrics are associated with relatively low Ds strain and Ds microstructures suggest that they may be related to an early stage in the Ds deformation (section 3.4.3).

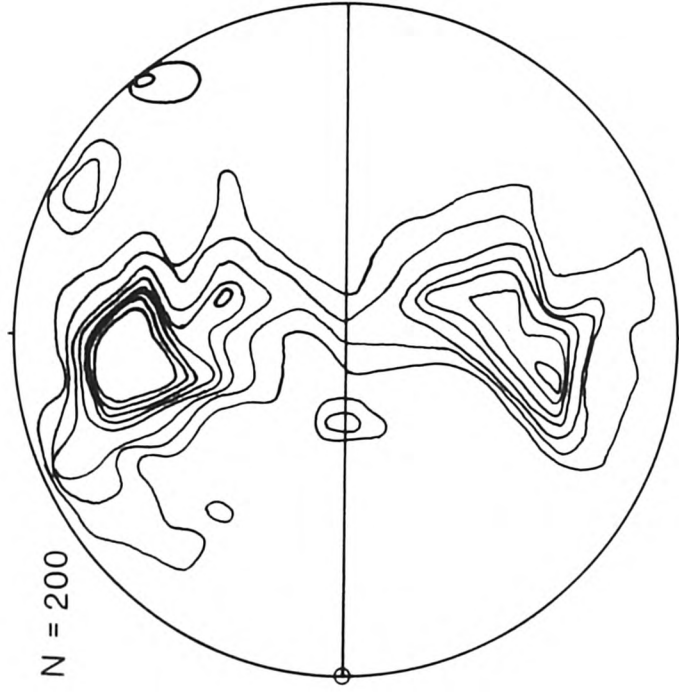
Fabrics from the most highly-deformed part of the sequence have a strongly asymmetric density distribution but a central girdle which lies

Fig.4.11a. Representative Ds quartz c-axis diagrams plotted on a cross-section through area 3. See fig.4.11c. for location of samples and section.

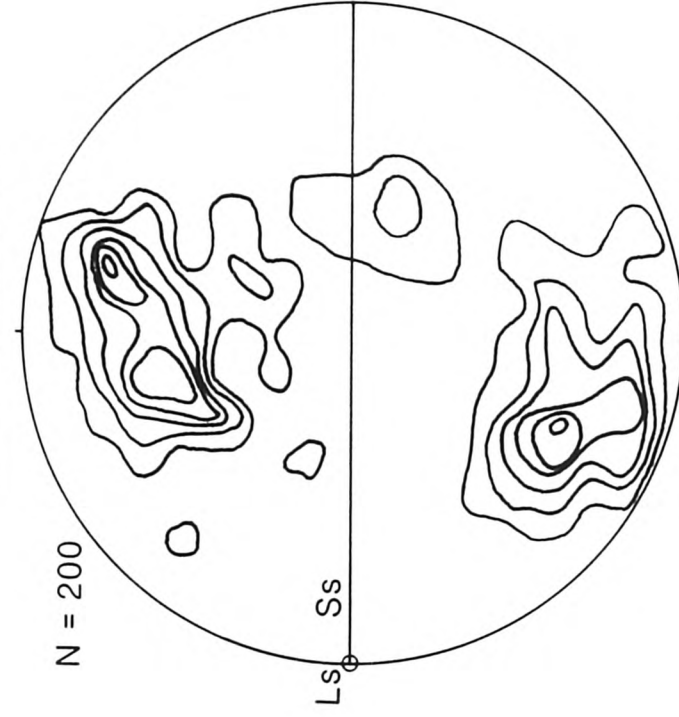


QUARTZ C AXIS FABRICS : EASTERN SECTION
AREA 3

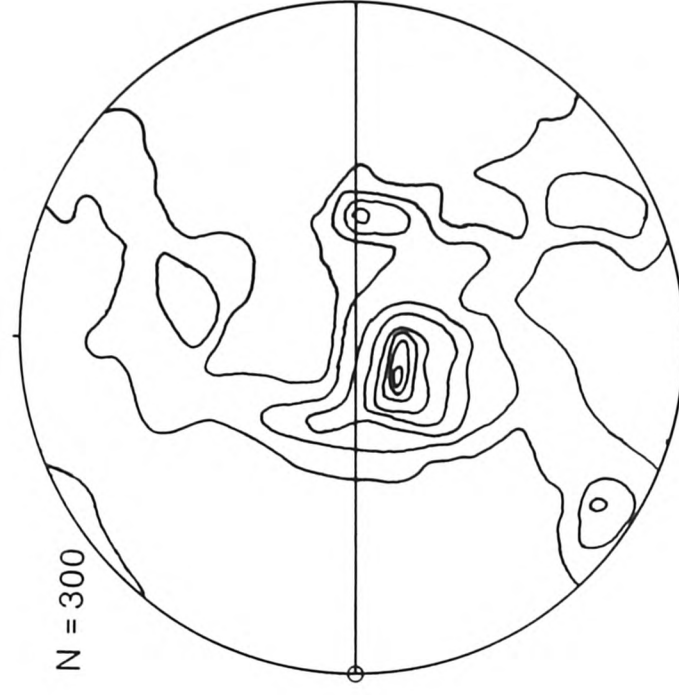
Fig.4.11b. Ds quartz c-axis data not included in the cross-section of fig.4.11a. Contours 1 to 8% per 1% area, shaded areas <1% contour within the diagram. See fig.4.11c. for location of samples.



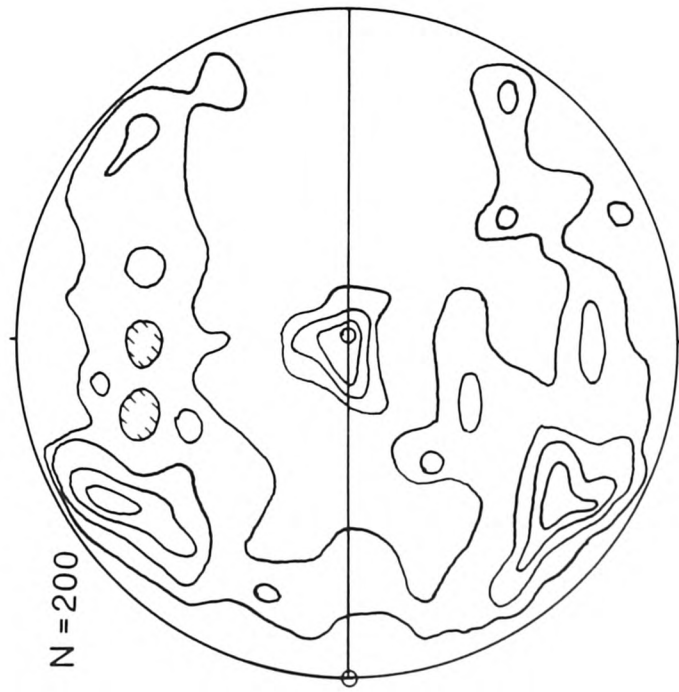
433



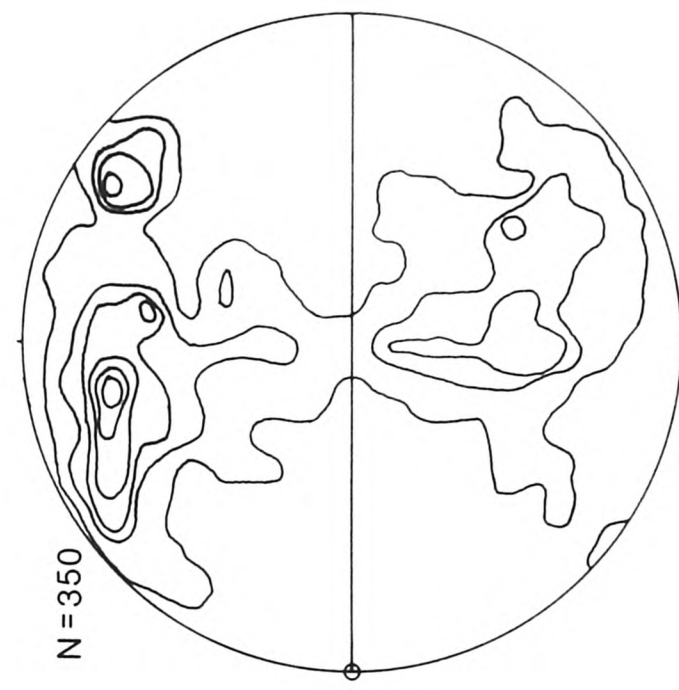
MS40



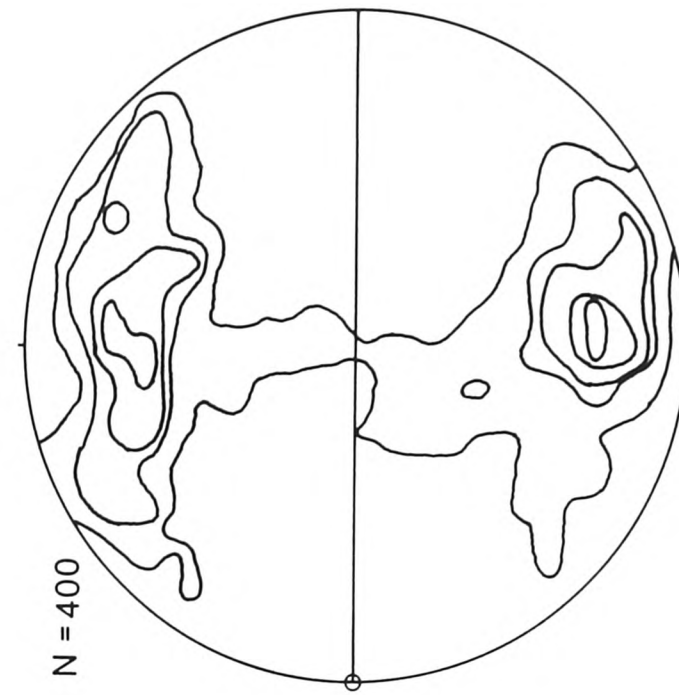
MS49



MS38



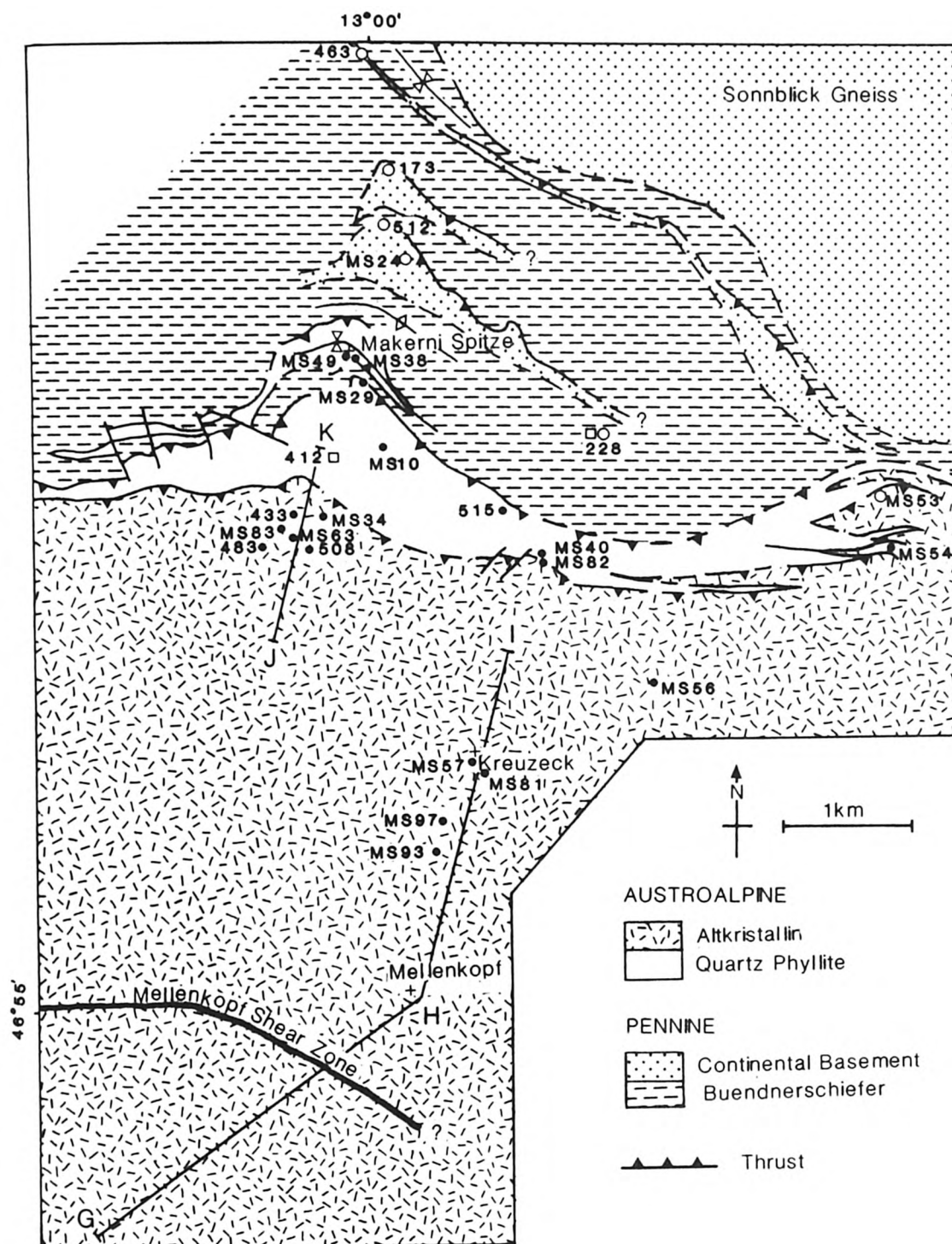
MS29



MS11

Fig.4.11c. Simplified geological map of area 3 showing the main structural divisions, the location of the section in fig.4.11a. and the samples used for quartz and calcite fabric analysis.

● = Ds quartz c-axis fabric, □ = Dt calcite c-axis fabric, o = Dt quartz c-axis fabric.



sub-perpendicular to the schistosity. The asymmetric density distribution is consistent from one fabric to the next and suggests a top to the NW sense of vorticity.

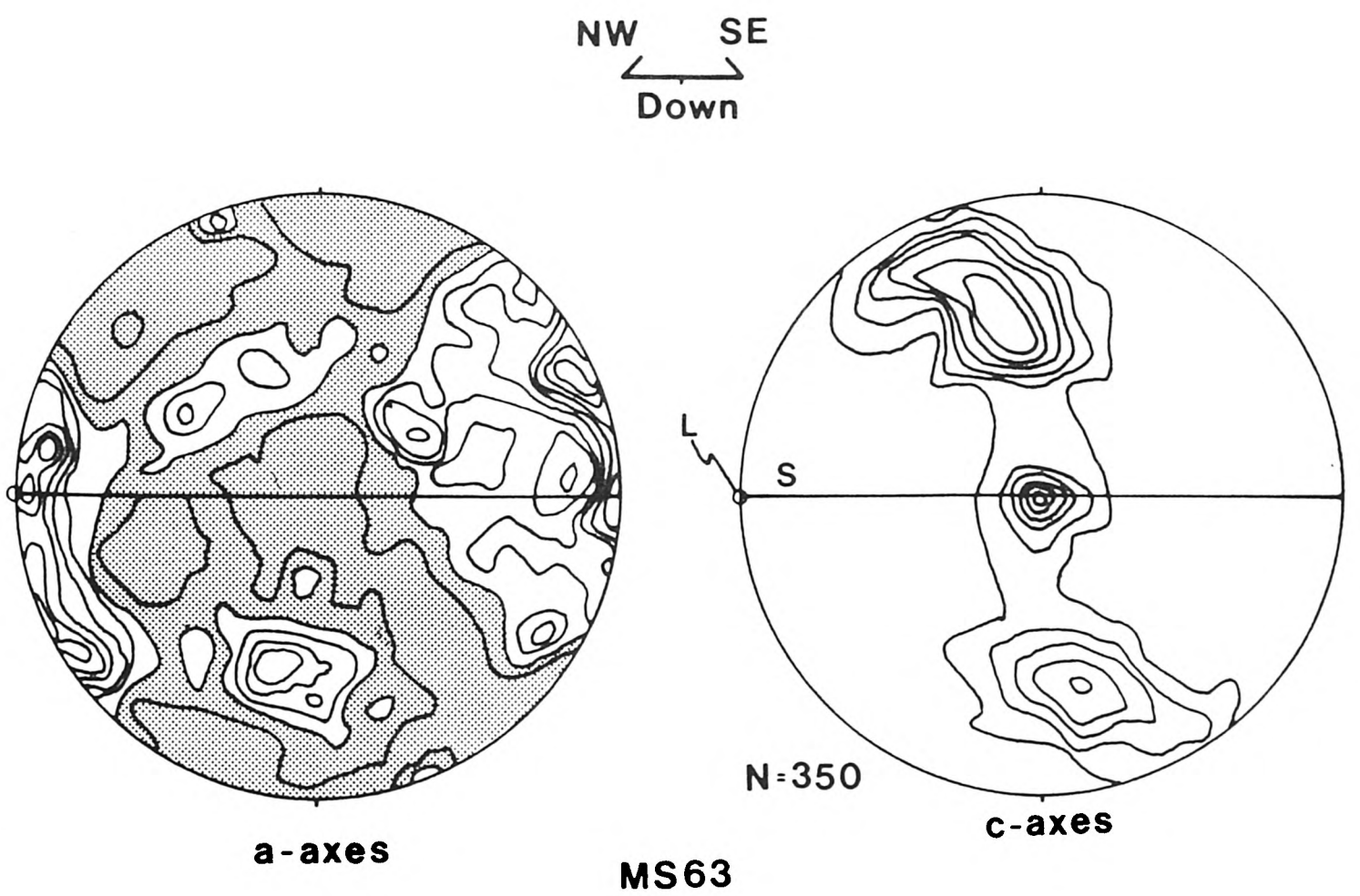
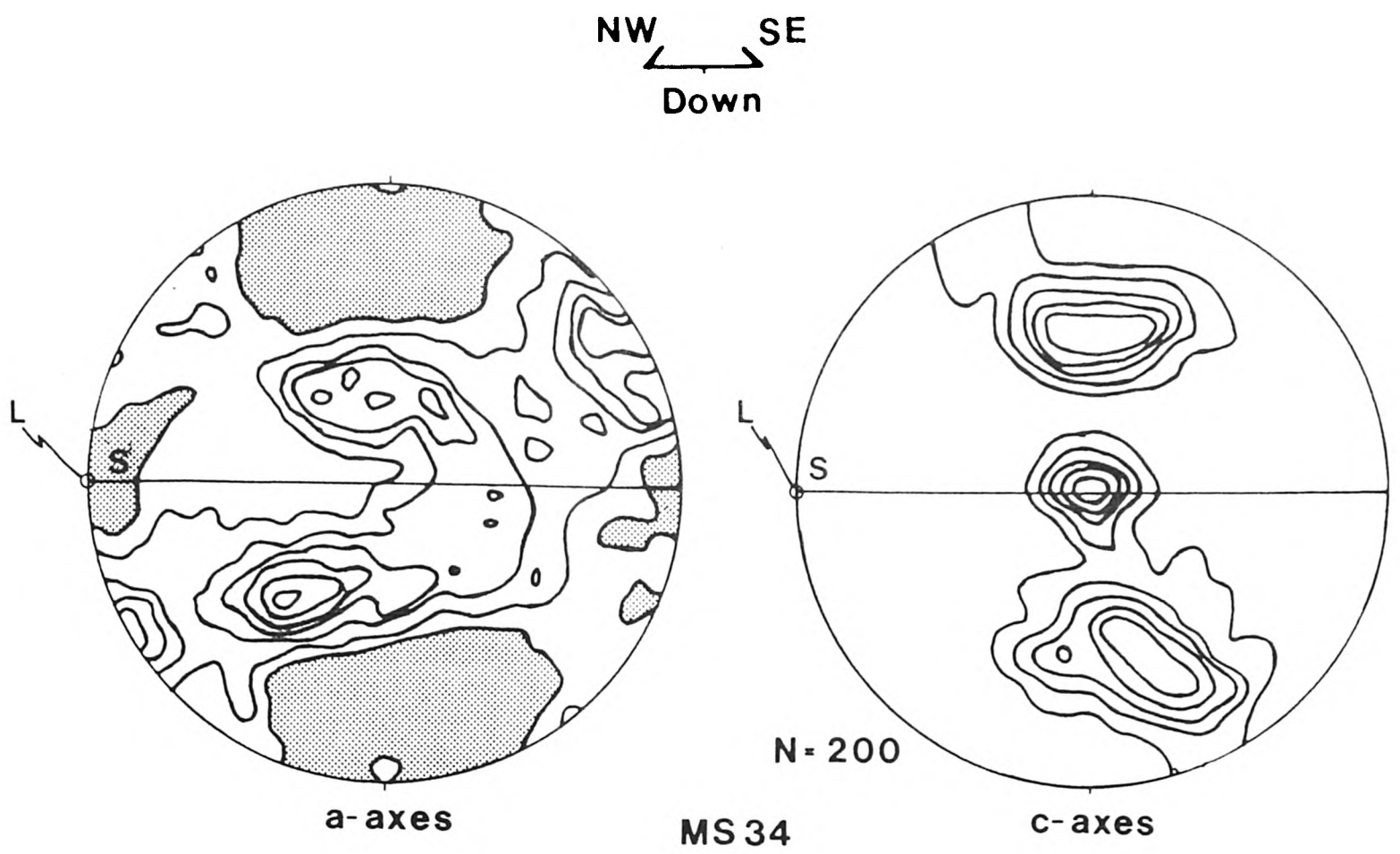
Unlike area 1, the majority of the fabrics have an even internal distribution of c-axes. This could reflect either an initially random distribution of quartz c-axes or a high Ds finite strain, which has overridden the effects of a Dr preferred orientation. MS97 has a random pre-Ds fabric, but Sr inclusion trails within albites from sample MS63 still preserve non-random c-axis fabrics, which demonstrates the existence of a Dr preferred orientation in this area. The influence of the preferred orientation has, therefore, probably been eradicated by progressive dynamic recrystallization. This implies that deformation within area 3 is generally higher strain than in area 1.

The fabric topologies from the high-strain zone show only a weakly developed asymmetry. The increase of Ds strain towards the base of the Altkristallin, indicated by the intensity of crystallographic preferred orientation and mesoscopic structures, suggests that the greater angle between Ss and the central girdle in the high strain fabrics, compared to MS81 and MS57, reflects a higher Ds strain. This change in angle could, however, also be explained as a decrease in the kinematic vorticity number.

In the basal high-strain part of the Altkristallin and within the upper parts of the Matrei Zone, there is a gradual transition from cross girdle fabrics indicating plane strain, to small-circle distributions indicating deformation in the flattening field. Intermediate fabrics have a cluster of c-axes around Y with only a very weakly developed central girdle.

The fabrics reflecting the flattening strain define a zone approximately 1km wide. The development of this zone was studied more closely with a-axis fabrics for samples MS63 and MS34 (fig.4.12a.).

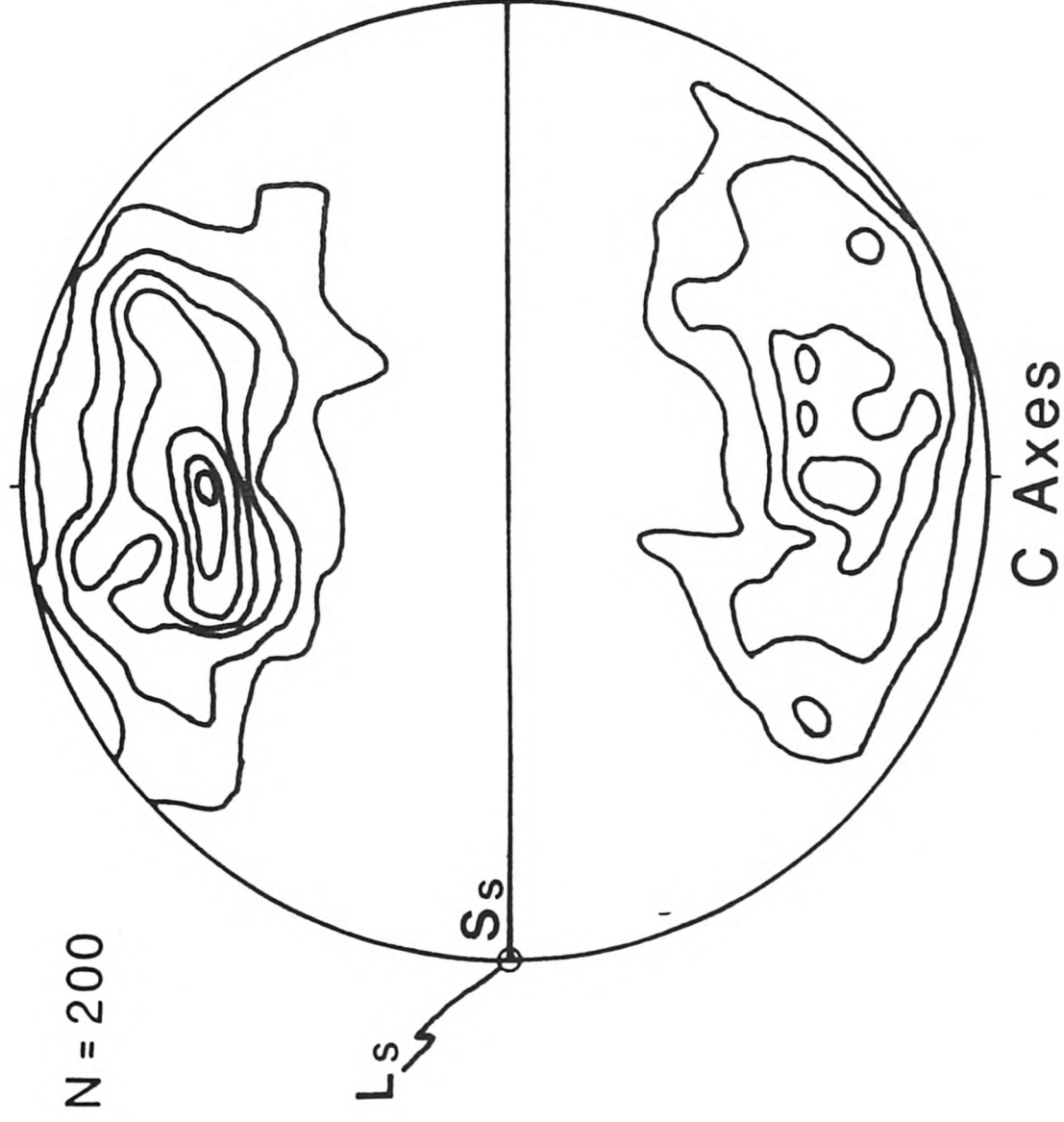
Fig.4.12a. A- and c-axis fabrics for samples MS63 and MS34. The contours for the a-axis diagrams are in 0.5% intervals times uniform. The shaded area is < 0.5% for MS34 and < 1% for MS63. Contour intervals for the c-axis diagrams are 1 to 6% per 1% area.



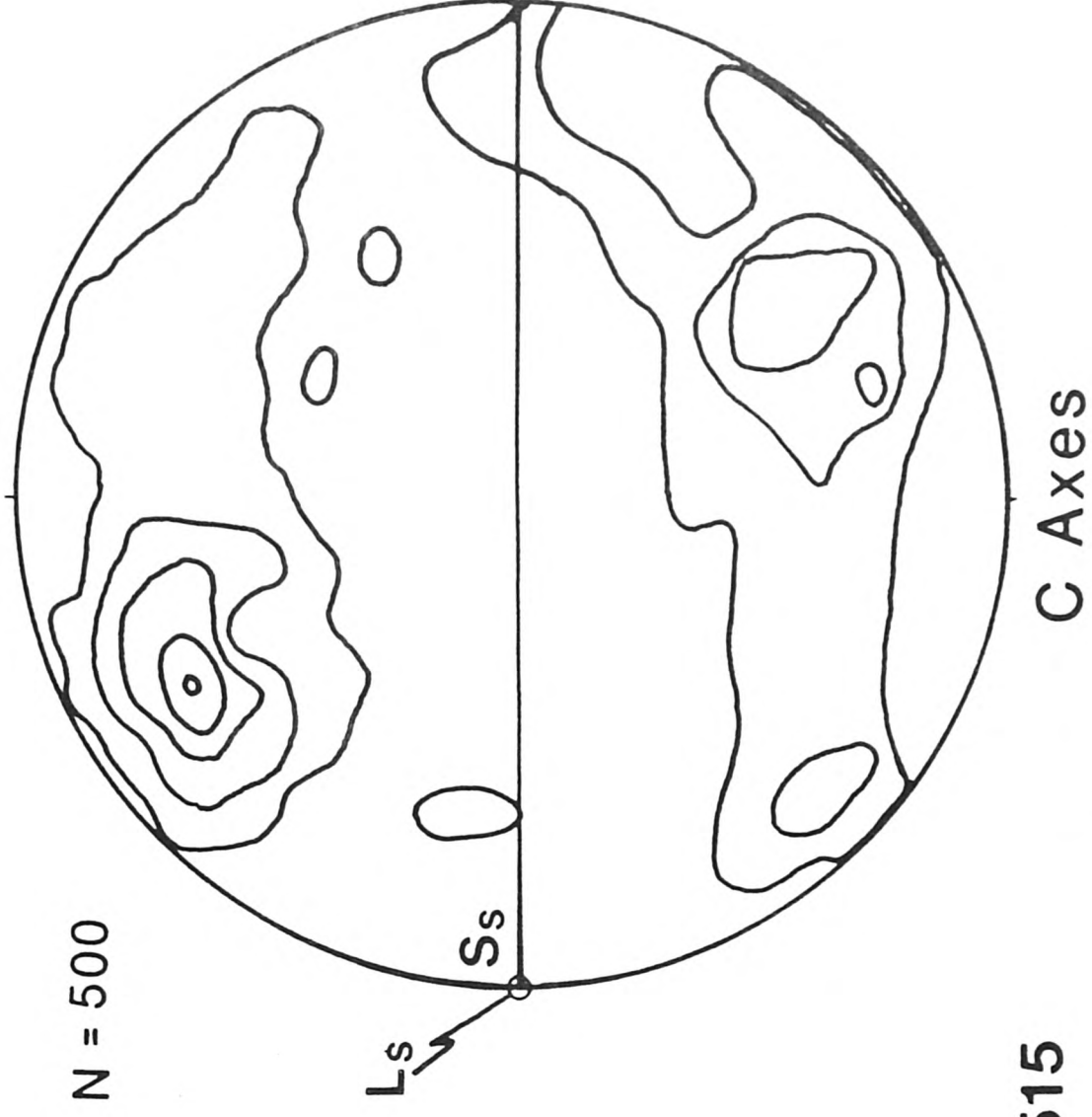
QUARTZ C AND A AXES DIAGRAMS

Fig.4.12b. Quartz c-axis diagrams measured in sample 515. The pattern in the matrix has a well-developed small circle distribution with an asymmetric density distribution suggesting non-coaxial deformation possibly associated with a top to NW sense of vorticity. The quartz c-axis fabric from the synkinematic veins have a similar, although less well-defined, small circle distribution suggesting coaxial deformation in the flattening field but with no obvious asymmetric density distribution suggesting deformation with a low degree of non-coaxiality.

NW SE
 $\swarrow$
Down



Synkinematic Quartz Veins



Quartz Phyllite Matrix

515

These show concentrations of a-axes distributed around small circles centred on the Z axis of the finite strain ellipsoid. For MS63 this is clearly asymmetric with respect to the strain axes and indicates a top to the NW sense of vorticity. The small circle distribution implies a non-plane strain deformation in the flattening field. MS34 shows more ambiguous results. The strong asymmetry of the a-axis concentrations suggests a top to the NW sense of vorticity, however, their symmetrical arrangement indicates a coaxial deformation history. This ostensibly contradictory evidence could be explained as a late-stage coaxial overprint on a fabric which already had a strong asymmetry developed during Ds (e.g. Lister & Williams 1979). However, there is no microstructural evidence of there being a separate later deformation which affected MS34 and not the other samples. Microstructures suggest that deformation is entirely related to Ds. An alternative possibility is that Ds itself had a late-stage increment of strain, which was coaxial for sample MS34.

A further piece of evidence for coaxial deformation taking place at a late stage of Ds within the zone showing non-plane strain, comes from measuring quartz fabrics in deformed synkinematic vein quartz material. In sample 515 the c-axis fabric for the matrix (excluding the veins) has a small circle distribution of c-axes around Z, indicating deformation in the flattening field, and an asymmetric density distribution suggesting a top to NW sense of vorticity (fig.4.12b). The deformed vein quartz material reflects only a late stage of deformation and has a similar symmetric small circle c-axis distribution around Z but does not have the asymmetric density distribution. The fabric measured in the veins is clearly different from that of the whole rock and indicates a dominantly coaxial deformation history in the flattening field.

This final increment of deformation cannot have been coaxial throughout the zone of non-plane strain, since the a-axis fabric of MS63

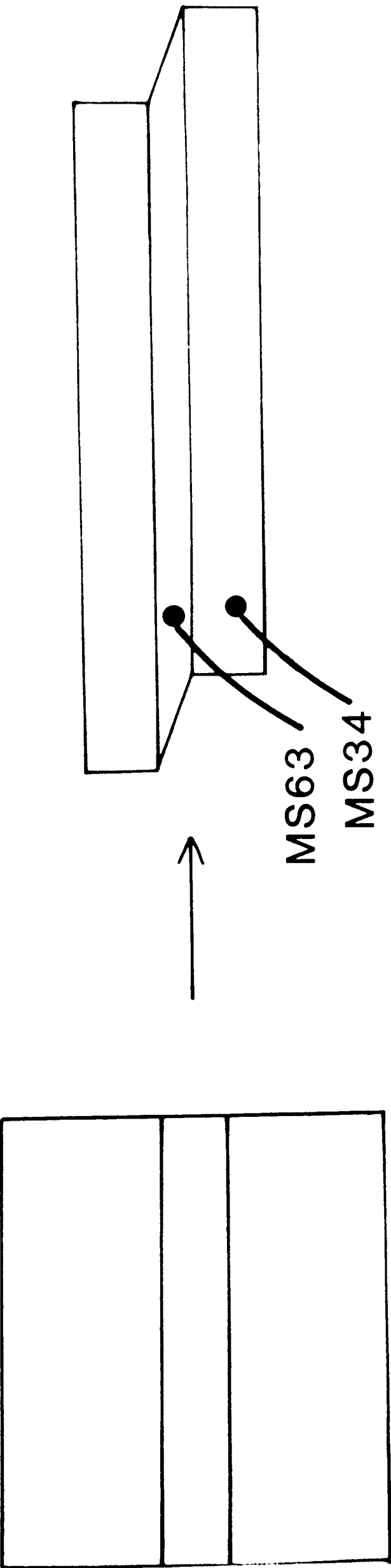


Fig.4.13. Two dimensional representation of the suggested strain partitioning during the late-stage increment of Ds deformation.

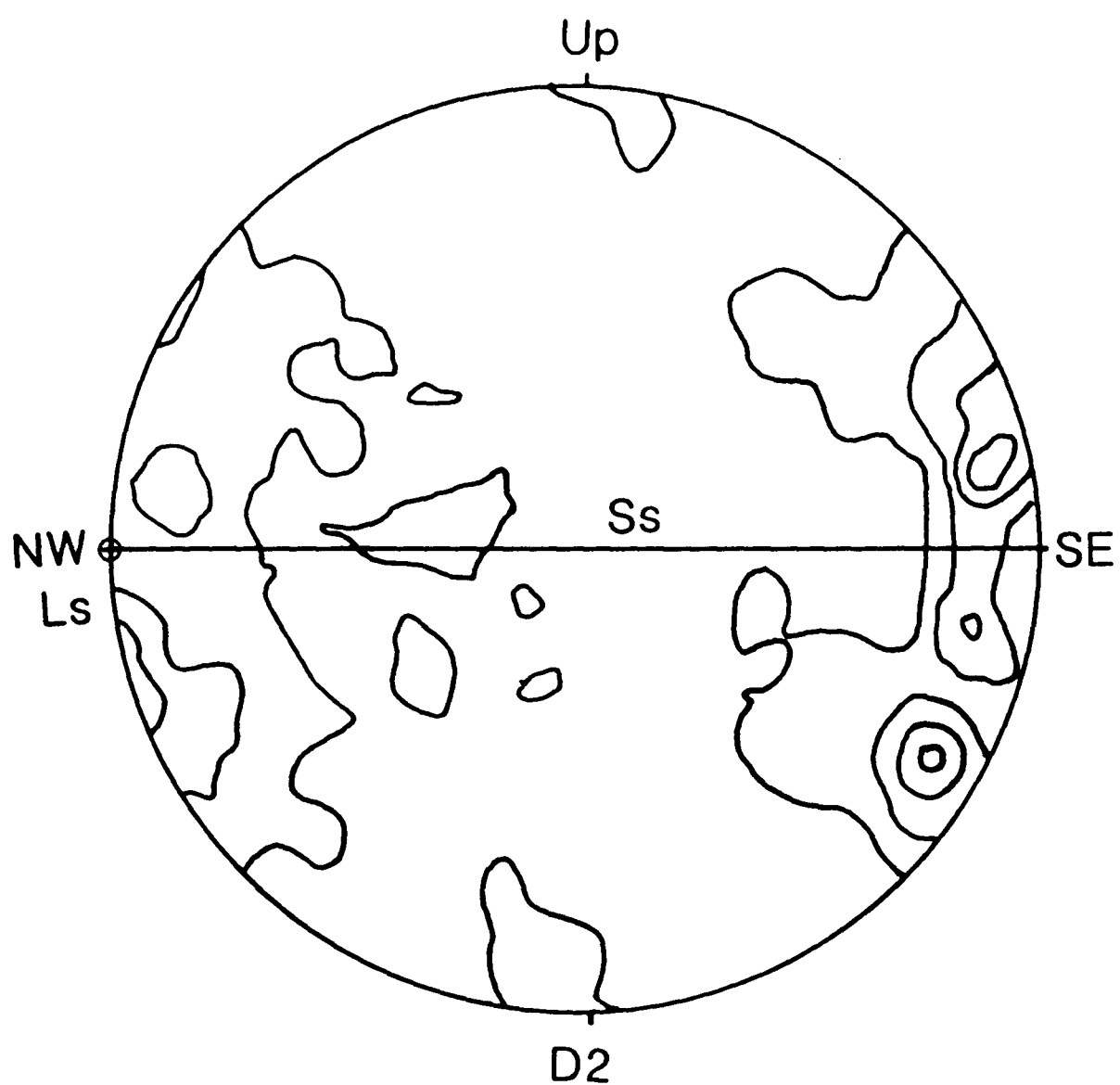


Fig.4.14. Ds a-axis fabric measured in metaquartzite from the Mallnitzer Mulde (sample D2). The asymmetric density distribution and slight asymmetry in the topology suggests a top to the NW sense of vorticity and a kinematic vorticity number, $1 > w_k > 0$.

has maintained an asymmetric topology. The difference between MS34 and MS63 can be explained by strain path partitioning (fig.4.13). Strain compatibility between adjacent domains undergoing coaxial and non-coaxial deformation requires that the coaxial component is the same in each domain. The vortical component is unrestricted.

The orientation of eccs in area 3 also suggest a similar late stage increment to Ds deformation in flattening field. However, both the quartz fabrics and the eccs may develop after relatively low strain. Within MS63 extension along the Y axis has had no appreciable affect on the orientation of the long axes of porphyroblasts (section 4.5.3). This is good evidence for the flattening in this sample being a low-strain effect.

To the north of area 3 Ds also forms the dominant fabric within the Mallnitzer Mulde (chapter 3). A deformed quartz tectonite with a strong NW-SE stretching lineation has a well developed a-axis preferred orientation, which indicates a top to NW sense of vorticity (fig.4.14).

4.6. ROTATION OF RIGID INCLUSIONS AS A MEASURE OF NON-COAXIALITY

4.6.1. Description of flow

A general non-coaxial two dimensional flow can be described by the velocity gradient tensor

$$L_{ij} = \begin{bmatrix} 0 & S/2+W/2 \\ S/2-W/2 & 0 \end{bmatrix}$$

where S is the sum of the maximum and minimum extension rates; W is the magnitude of the vorticity vector, and the reference frame, $\underline{x}$, is oriented at 45° to the directions of maximum and minimum instantaneous extension rate (Passchier 1987a). L_{ij} can be represented by a circle in Mohr space with axes of angular velocity and extension rate (Lister and Williams 1983, Means 1983)(fig.4.15).

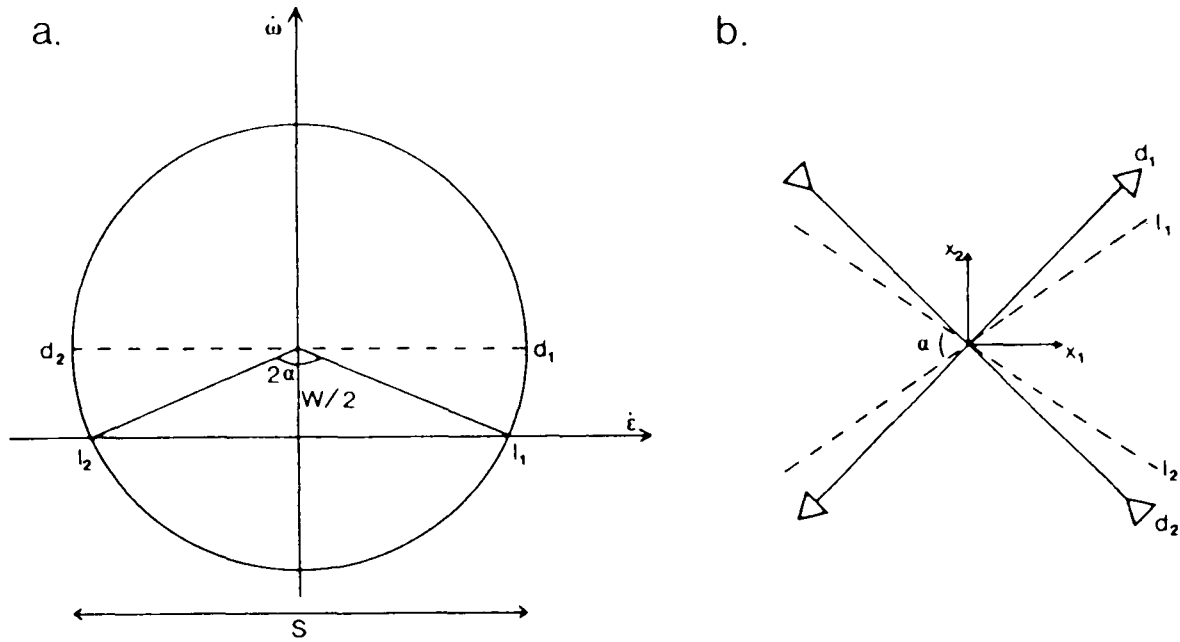


Fig.4.15a. Mohr circle representation of general two dimensional non-coaxial flow with a kinematic vorticity number $w_k = 0.4$. Points on a diameter of the circle have coordinates that give the instantaneous angular velocity and stretching rate of two instantaneously orthogonal material lines.

Fig.4.15b. Orientation of axes of zero instantaneous angular velocity l_1, l_2 , and the principal instantaneous stretching axes d_1, d_2 , with respect to the reference system x_1, x_2 . The angle between l_1 and l_2 ($=\alpha$), varies according to w_k (see fig.4.16).

$$l_1 \wedge l_2 = \cos^{-1} w_k.$$

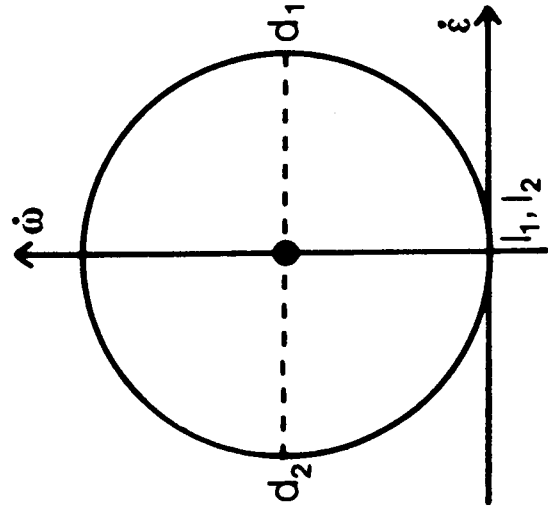
Points on a diameter of the Mohr circle have coordinates which are the instantaneous angular velocity and extension rates of two mutually perpendicular material lines. In fig.4.15, d_1 and d_2 represent the two axes of maximum and minimum instantaneous extension rate; and l_1, l_2 are the two lines of zero instantaneous angular velocity also termed flow apophyses (Ramberg 1975). The degree of non-coaxiality is determined by the ratio of the vortical component to the coaxial component of deformation. This is the kinematic vorticity number, w_k , and in two dimensions is defined mathematically as:

$$w_k = W / (2(d_1^2 + d_2^2))^{1/2} \quad \dots 1$$

which for isochoric flow simplifies to $w_k = W/S$, where $S = d_1 + d_2$. W , the magnitude of the vorticity vector, can be expressed as the sum of the angular velocities of two instantaneously orthogonal lines or $W = \omega_{12} + \omega_{21}$ (Lister & Williams 1983). In the Mohr construction, the position of the centre of the circle represents the average angular velocity of material lines in the deforming medium, or alternatively the average angular velocity of any two instantaneously orthogonal material lines, i.e. $(\omega_{12} + \omega_{21})/2$. Therefore, the distance between the centre of the circle and the $\dot{\epsilon}$ axis is $W/2$.

For pure shear $w_k = 0$; for simple shear $w_k = 1$, and the velocity gradients tensor for any general non-coaxial deformation with $0 < w_k < 1$ can be expressed as a combination of different proportions of the velocity gradients tensors for pure and simple shear (Bobyarchick 1986). This is represented in Mohr space by a circle with a centre located on the $\dot{\omega}$ axis between the positions for simple and pure shear (fig.4.16).

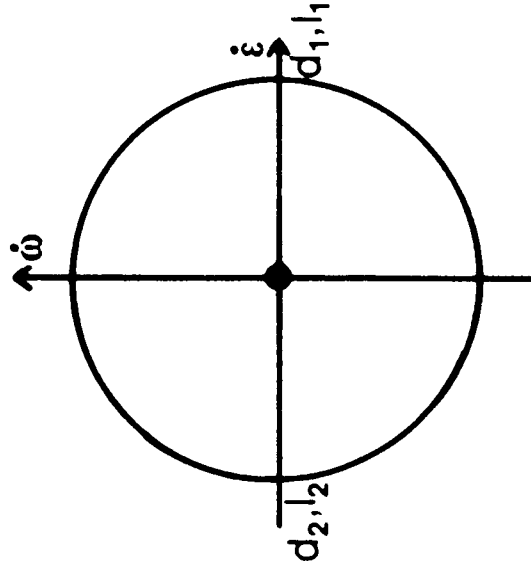
Fig.4.16. Mohr circle representation of two dimensional flow by simple shear, pure shear and general non-coaxial deformation with the corresponding velocity gradients tensors. See text and fig.4.15 for definition of symbols.



Simple Shear

$$L_{ij} = \begin{bmatrix} 0 & S \\ 0 & 0 \end{bmatrix}$$

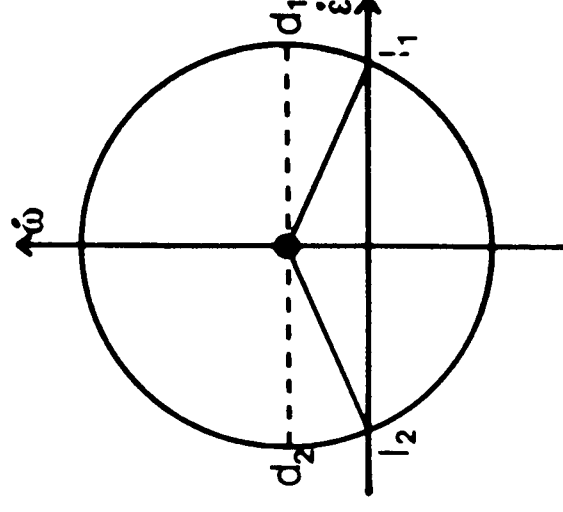
$$W = S, \alpha = 0$$



Pure Shear

$$L_{ij} = \begin{bmatrix} 0 & S/2 \\ S/2 & 0 \end{bmatrix}$$

$$W = 0, \alpha = 90$$



General Non-Coaxial
Deformation

$$L_{ij} = \begin{bmatrix} 0 & S/2 + W/2 \\ S/2 - W/2 & 0 \end{bmatrix}$$

$$W < S, W \neq 0$$

$$0 < \alpha < 90$$

4.5.2 Rotation of rigid particles in a flowing viscous matrix

Jeffery (1922) gives a complete solution for the rotation of rigid ellipsoidal particles in a viscous matrix undergoing flow by simple shear. In two dimensions this is:

$$\dot{\Phi} = \frac{\dot{\gamma}(R^2 \sin^2 \Phi + \cos^2 \Phi)}{R^2 + 1} \quad \dots\dots 2$$

where $\dot{\Phi}$ = rate of rotation of the particle with respect to the shear plane; $\dot{\gamma}$ = rate of simple shear; R = the axial ratio, a/b of the rigid particle and Φ is measured from the shear plane to the long axis of the particle (fig.4.17).

Ghosh and Sengupta (1973) give a similar equation for pure shear deformation

$$\dot{\Phi} = \dot{\epsilon}_x \frac{1-R^2}{1+R^2} \sin 2\Phi \quad \dots\dots 3$$

where R and Φ have the same definition as above and $\dot{\epsilon}_x$ is the rate of natural strain in the X direction (fig.4.17).

Combining L_{ij} for pure and simple shear can give any intermediate flow regime with $0 < w_k < 1$, and using combinations of the rotation equations for pure and simple shear, Ghosh & Ramberg (1976) derive equations for the reorientation of rigid bodies in general two dimensional flow for $0 < w_k < 1$. They show that the angle of finite rotation Φ is a function of the shear strain, γ ; the original orientation, Φ_0 ; the axial ratio, R and the flow regime or degree of non-coaxiality. Ghosh and Ramberg (ibid) define a measure of non-coaxiality $S_r = \dot{\epsilon}_x / \dot{\gamma}$. S_r is related to the kinematic vorticity number by the equation

$$S_r = \frac{(1 - w_k^2)^{\frac{1}{2}}}{w_k} \quad (\text{Ghosh 1987}). \quad \dots\dots 4$$

Three distinct but continuous solutions give $\Phi = \Phi(\Phi_0, \gamma, S_r, R)$. These three equations depend on critical relationships between R and

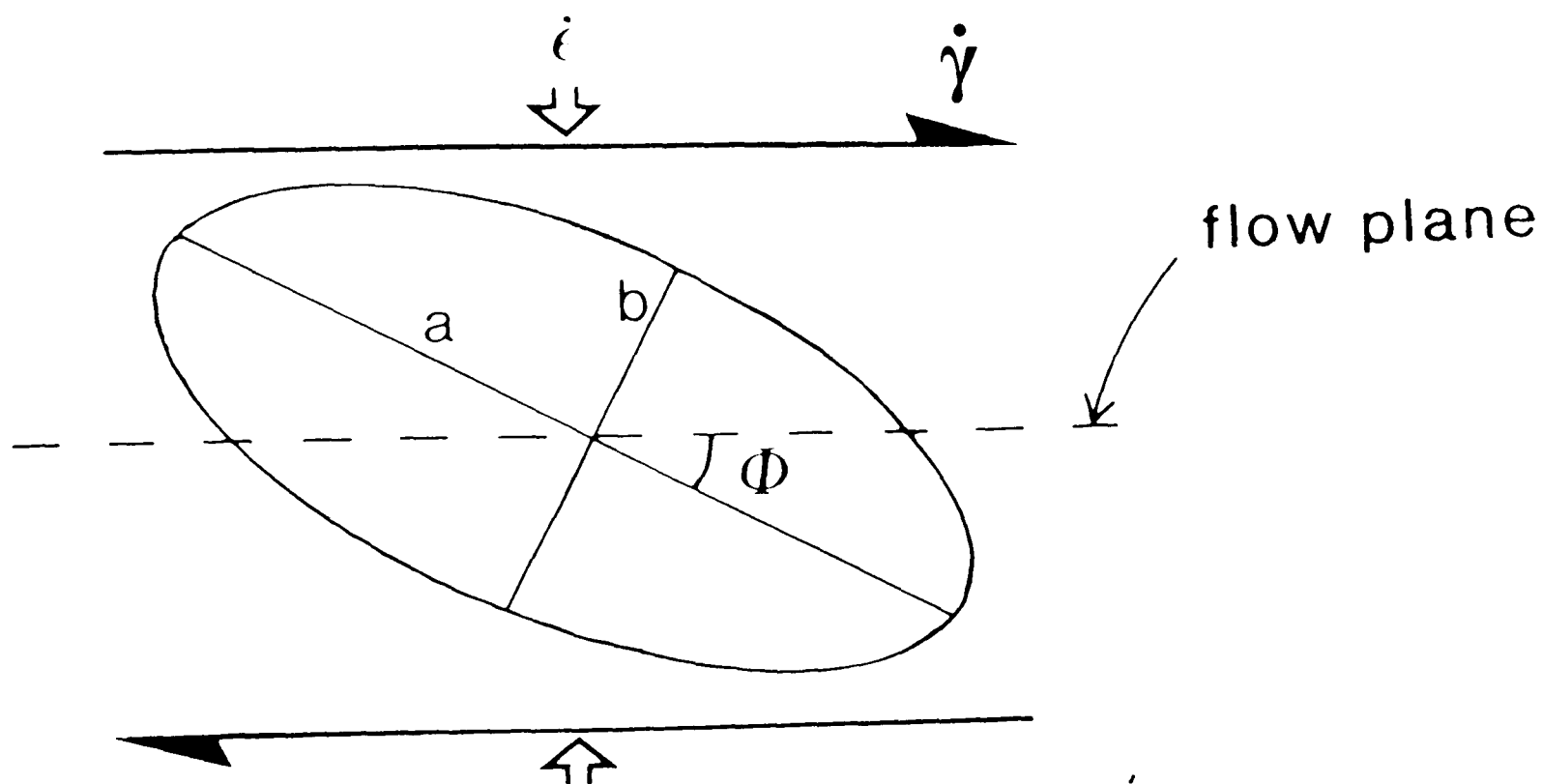


Fig.4.17. Section through an idealized porphyroclast perpendicular to the rotation axis and the parameters used in the kinematic analysis. $\dot{\gamma}$ = rate of shear strain, $\dot{\epsilon}_x$ = rate of natural strain along the plane of zero instantaneous angular velocity with a positive extension rate - the flow plane (only plane-strain flow with a kinematic vorticity number less than or equal to 1 is considered), Φ = the finite angle of rotation.

S_r .

Let $A = R^2/(R^2 + 1)$, $B = S_r(R^2 - 1)/(R^2 + 1)$, $C = 1/(R^2 + 1)$

$$P = \frac{B - C \cot \Phi_0 - \sqrt{(B^2 - AC)}}{B - C \cot \Phi_0 + \sqrt{(B^2 - AC)}} \cdot \exp(2\gamma \sqrt{(B^2 - AC)}),$$

then the three cases can be expressed as follows (modified after Ghosh & Ramberg 1976):

Case 1: If $S_r < R/(R^2 - 1)$, i.e. $AC > B^2$

$$\Phi = \arctan\left\{\sqrt{(AC - B^2)} \tan\left[\gamma \sqrt{(AC - B^2)} - \arctan\left(\frac{B - C \cot \Phi_0}{\sqrt{(AC - B^2)}}\right)\right] - \frac{B}{C}\right\} \dots 5$$

N.B. there is a mistake in sign, in the comparable equation given by Ghosh & Ramberg (1976) for this case.

Case 2: If $S_r > R/(R^2 - 1)$, i.e. $AC < B^2$

$$\Phi = \arctan\left\{\frac{P(B + \sqrt{(B^2 - AC)} - B + \sqrt{(B^2 - AC)})}{C(1 - P)}\right\} \dots 6$$

Case 3: If $S_r = R/(R^2 - 1)$, i.e. $AC = B^2$

$$\Phi = \arctan\left\{\frac{1}{C}\left[\frac{B - C \cot \Phi_0}{1 - \gamma(B - C \cot \Phi_0)}\right] - B\right\} \dots 7$$

Ghosh & Ramberg also show that for R above a certain critical value, particles will come to rest in a stable position, denoted by Φ_1 . The orientation of this position is a function of R and S_r alone, and is given by (after Ghosh & Ramberg 1976):

$$\Phi_1 = \arctan\{-S_r(R^2 - 1) - \sqrt{(S_r^2(R^2 - 1)^2 - R^2)}\} \dots 8$$

Particles with an aspect ratio less than the critical value will rotate continuously without reaching a stable position.

There are, therefore, two possible methods for determining the flow regime from the studies of rotated rigid particles. If the strain is high enough, then particles should come to rest in a stable orientation

(Φ_1), which from equation 8, is a function of the flow regime (S_r) and aspect ratio (R). Φ and R are measured in the sample and hence S_r determined (e.g. Passchier 1987b).

With an estimate of ϕ_0 and γ , the finite angle of rotation can also be used to determine S_r , even if the particles have not reached their stable position. This involves using the three equations 5, 6, and 7 (e.g. Vissers 1988).

4.6.3 Microstructure and Measuring Procedure

Two samples, MS63 and K5, of strongly deformed augengneiss from areas 1 and 3 were selected for a kinematic analysis using the rotation of albite porphyroclasts. In both samples the albites are set in a matrix which is approximately 90% quartz and 10% white mica. The preferred orientation of the mica defines the mesoscopic schistosity, S_s . Locally, deformed vein quartz material lies sub-parallel to S_s and has a strong crystallographic preferred orientation (fig.4.9, fig.4.11).

In general the long axes of the albite porphyroblasts are approximately parallel to their internal fabric suggesting that this foliation in the rock controlled the growth of the albites. The external S_r fabric in the two samples, K5 and MS63 has been tightly crenulated during D_s and transposed to give the new S_s foliation (fig.4.18). In sections perpendicular to L_s , the long axis of the albites follow the trace of S_r , which in the hinge zones of D_s crenulations is at a high angle to S_s . The external S_r foliation can commonly be traced into the albite porphyroblasts, suggesting that the internal fabric is S_r and that the albite porphyroblasts have been reoriented along with the S_r fabric during D_s folding. In sections parallel to L_s the trace of S_r and S_s are parallel. The long axes of

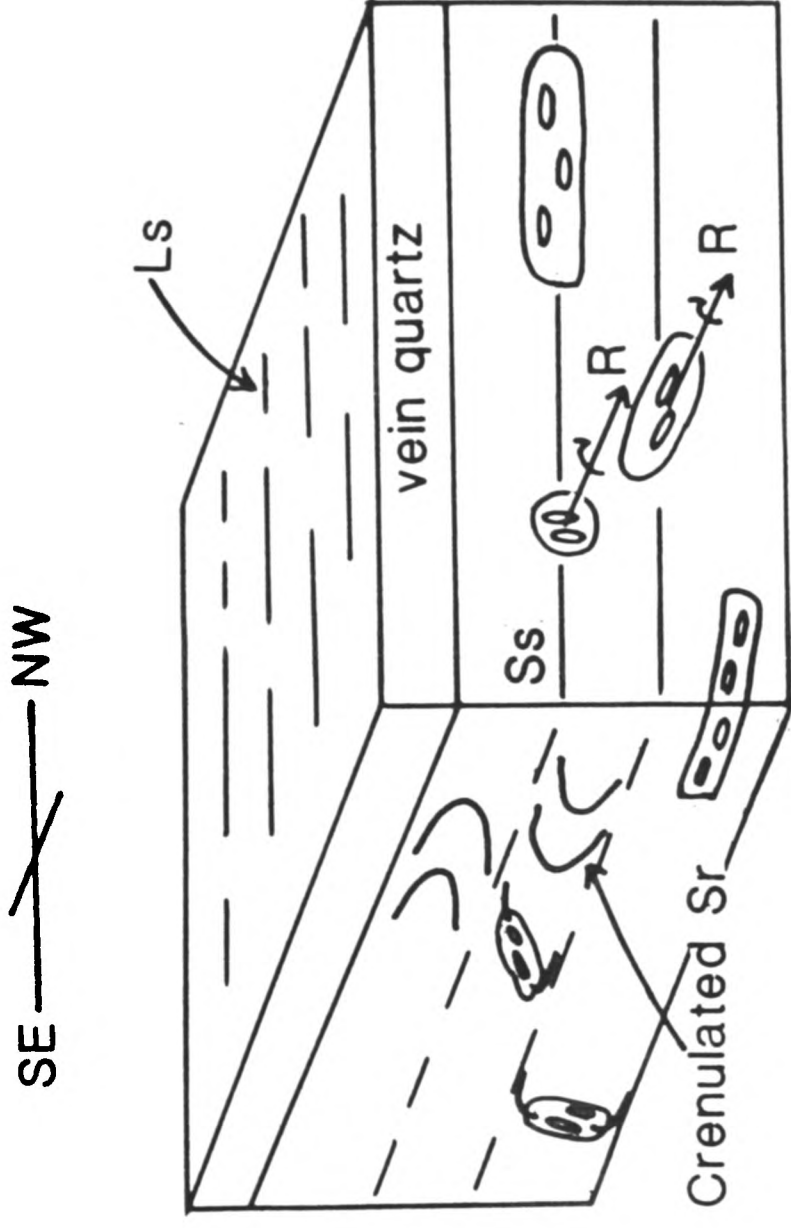


Fig.4.18. Generalized sketch of samples K5 and MS63. R = rotation axis of porphyroclast, R $\approx$ perpendicular to Ls. In sections perpendicular to Ls the internal fabric of the porphyroclasts is continuous with the external Sr fabric. In sections parallel to Ls, porphyroblasts have been rotated through varying angles dependent on their aspect ratio. Quartz fabrics were measured in the deformed vein quartz layers parallel to Ss, see figs.4.9, 4.11.

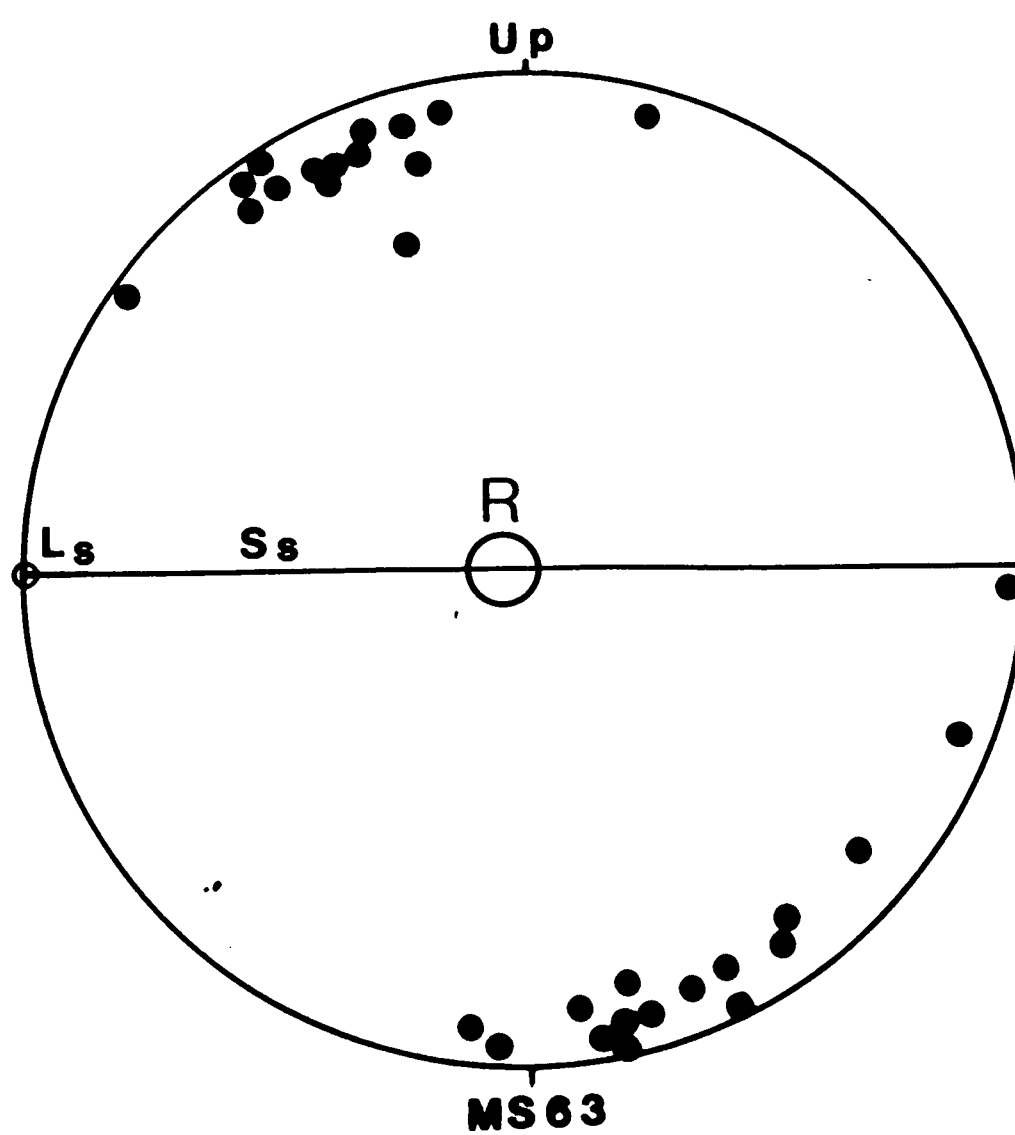
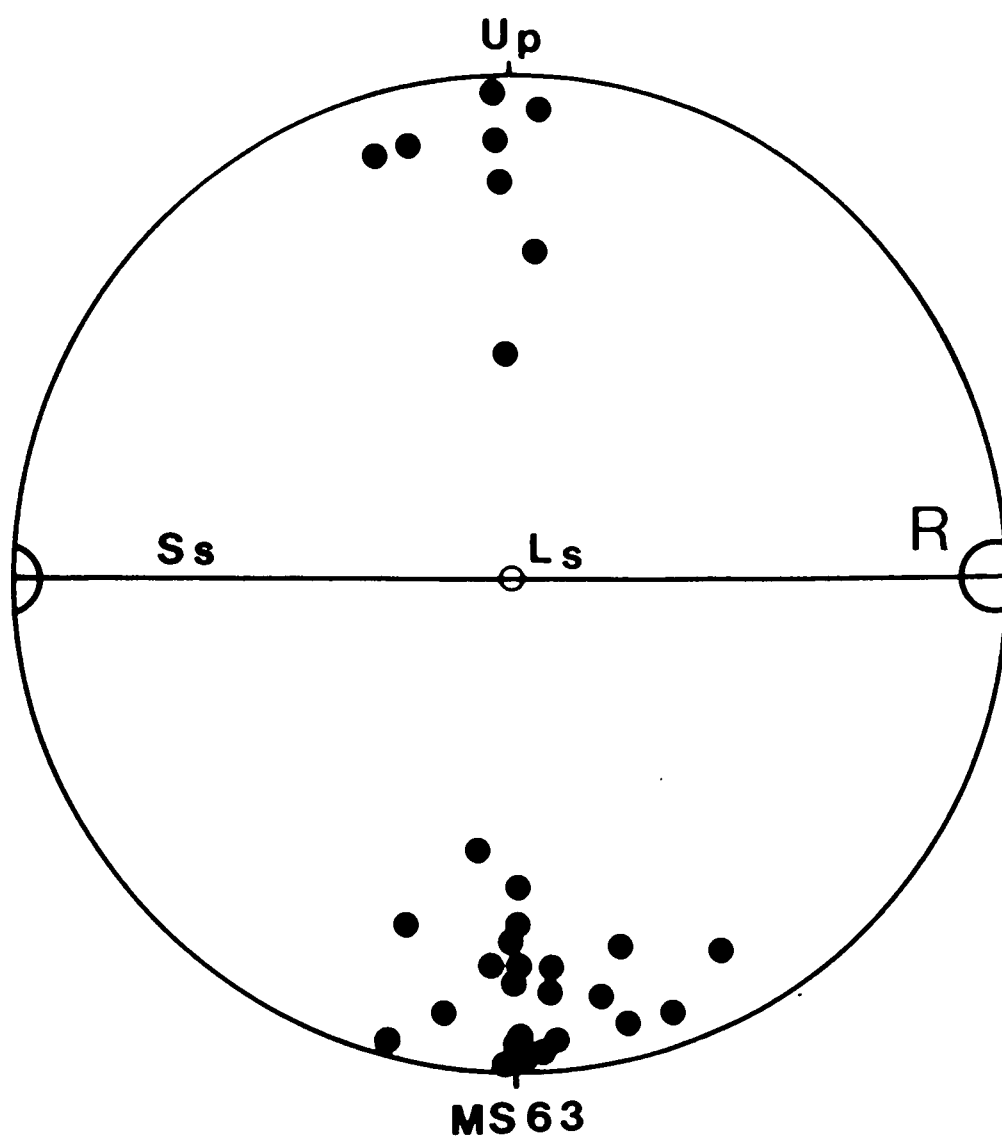
the porphyroclasts have been rotated with respect to both foliations due to the rotational component of Ds deformation.

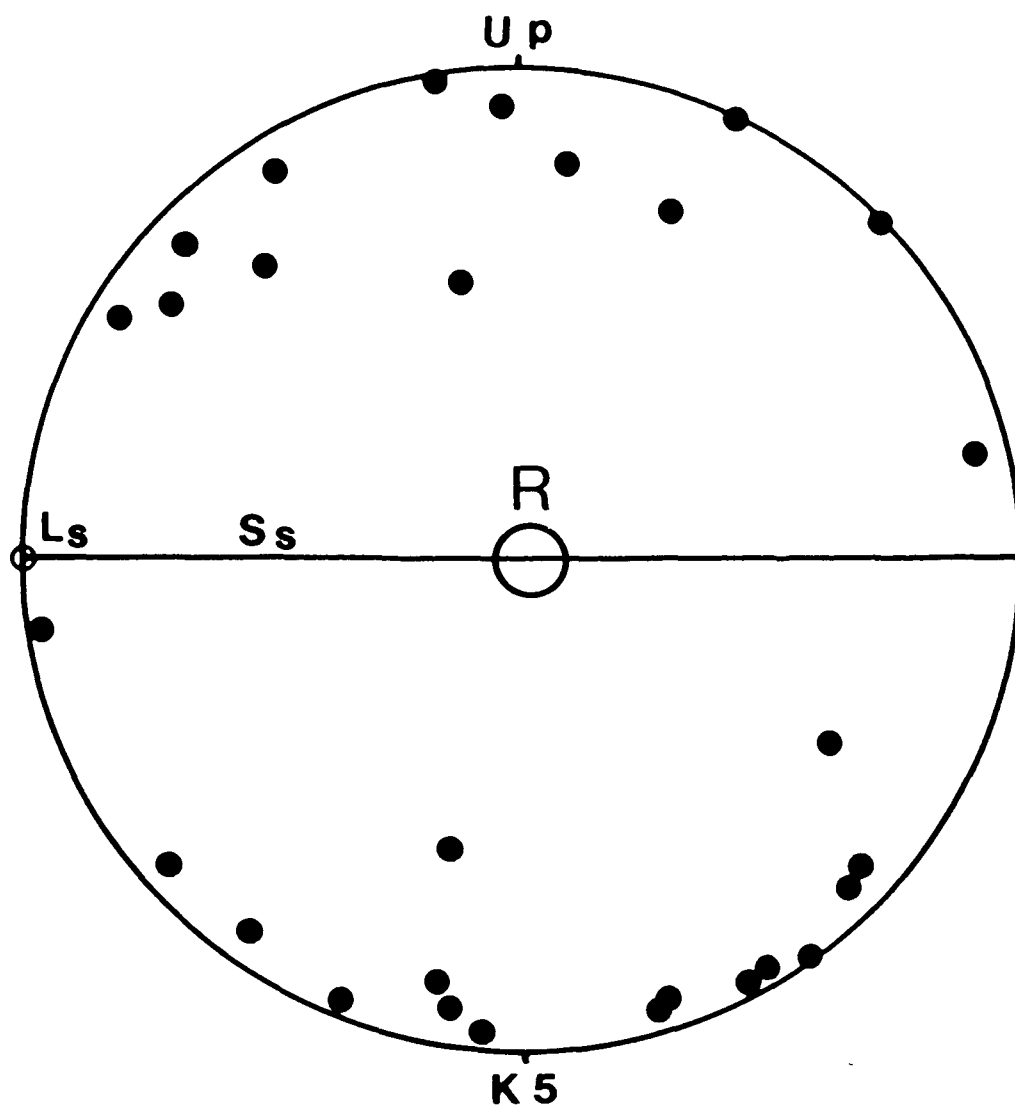
The continuity of the Sr from within the albites to the surrounding matrix suggests that the orientation of the internal fabric can be used to determine the rotation of the albites relative to Sr. In sections parallel to Ls the internal fabric of the albites is at various angles to Ss and in general the angle is greater for porphyroblasts with low aspect ratios. This suggests that the albites have rotated during Ds about an axis at a high angle to Ls. In sections oblique to the rotation axis, the porphyroblasts will exhibit lower angles of finite rotation than their true value (Vissers 1987). It is therefore important to determine its orientation and to cut sections perpendicular to this before measuring the rotation angles.

The orientation of the internal fabric of the albites can be measured on a universal stage. The poles to this fabric are roughly co-planar implying that the rotation of the albites is related to a common rotation axis, which lies within 15° of being perpendicular to Ls for both MS63 and K5 (fig.4.19a,b,c.). There is, however, an extra degree of freedom and it is conceivable that the albites rotated about an axis at a high angle to their included fabrics. Evidence that this was not a significant effect comes from the pre-Ds quartz c-axis fabrics still preserved in the inclusion trails within the albite porphyroblasts. These fabrics were measured in both MS63 and K5 by combining the data from several porphyroblasts, using the internal fabric as a reference (fig.4.20). Rotation about another axis due to ductile deformation would vary according to aspect ratio and cause differential rotation of the albites through the plane of the section. This would disrupt the angular relationship between the different albites and obliterate the quartz preferred orientation.

The rotation angle, Φ , should be measured between the long axis of

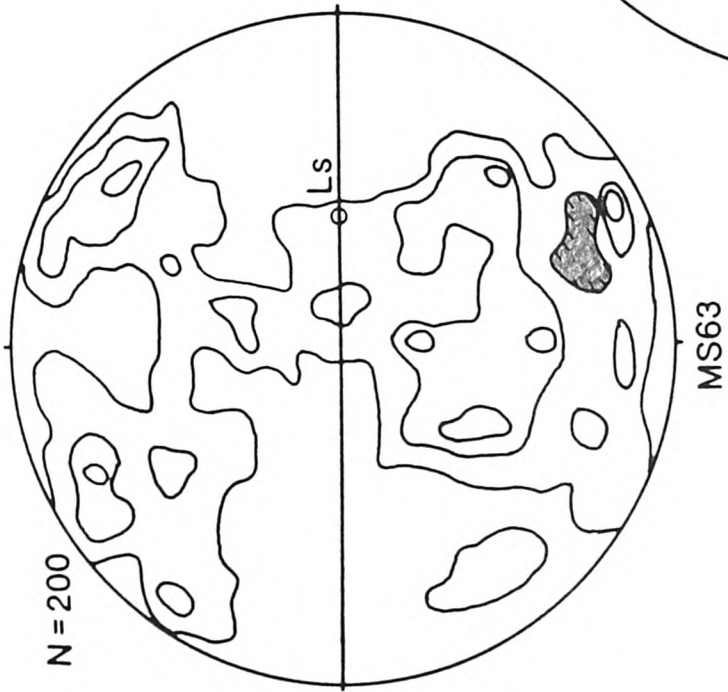
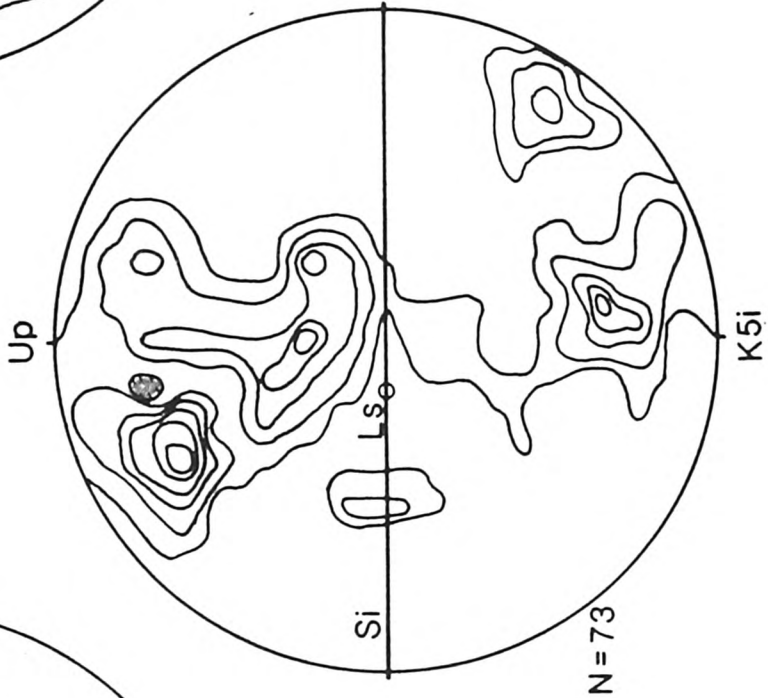
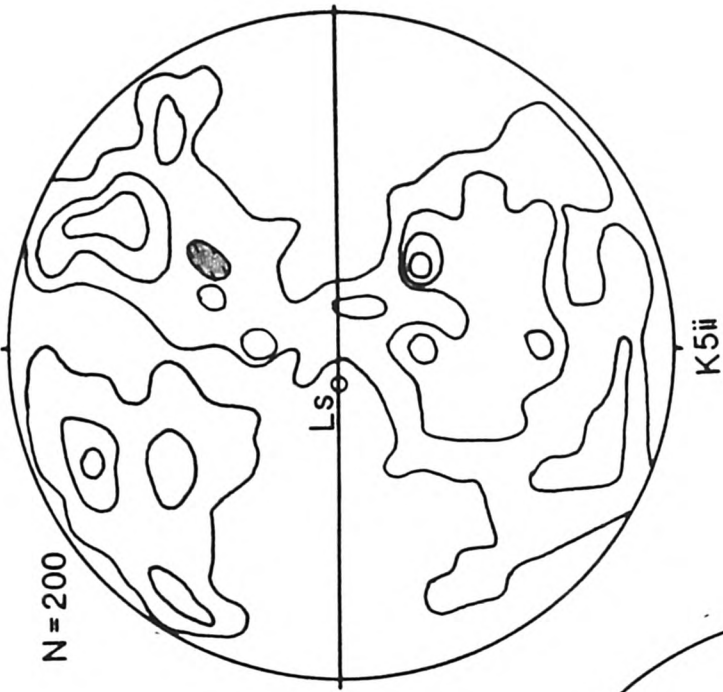
Fig.19. Poles to the internal fabric of the albite porphyroclasts plotted with respect to Ss for the two samples K5 and MS63; (a) section perpendicular to Ls for MS63 (b) section parallel to Ls for MS63 (c) section parallel to Ls for K5. The poles to Ss and the internal fabric of the porphyroclasts are approximately coplanar but lie at varying angles due to the rotation of the porphyroclasts. The inferred position of the average rotation axes is marked by R.





C.

Fig.4.20. Quartz c-axis diagrams measured from inclusion trails within albite porphyroclasts. Contour interval 1 to 8% per 1% area, shaded areas within the fabrics are <1%.



the rotating particle and the flow plane, defined as the plane of instantaneous zero angular velocity with a positive rate of extension (equal to l_1 in fig.4.1). A flow plane only exists for the special conditions of plane strain isochoric deformation and $w_k < 1$ (fig.4.1). In the samples used in this work the angle was measured between Ss and the long axis of the albites, which in general is parallel to the trace of the internal fabric. There is usually no direct microstructural evidence that would suggest whether Ss is an X-Y fabric related to finite strain or whether it developed parallel to the flow plane. The computer modelling of Lister & Hobbs (1980) suggests that the central girdle of a quartz c-axis fabric develops perpendicular to a plane of zero instantaneous angular velocity in pure and simple shear and several authors have suggested the same may be true for intermediate progressive plane strain deformation histories (Platt & Behrmann 1986, Schmid & Casey 1986, Vissers 1988). In samples K5 and MS63 the central girdle to the c-axis fabric lies subperpendicular to Ss, which suggests that the flow plane is subparallel to Ss. With increasing γ the angular separation between the flow plane and the flattening plane of finite strain becomes very small and the two are essentially identical.

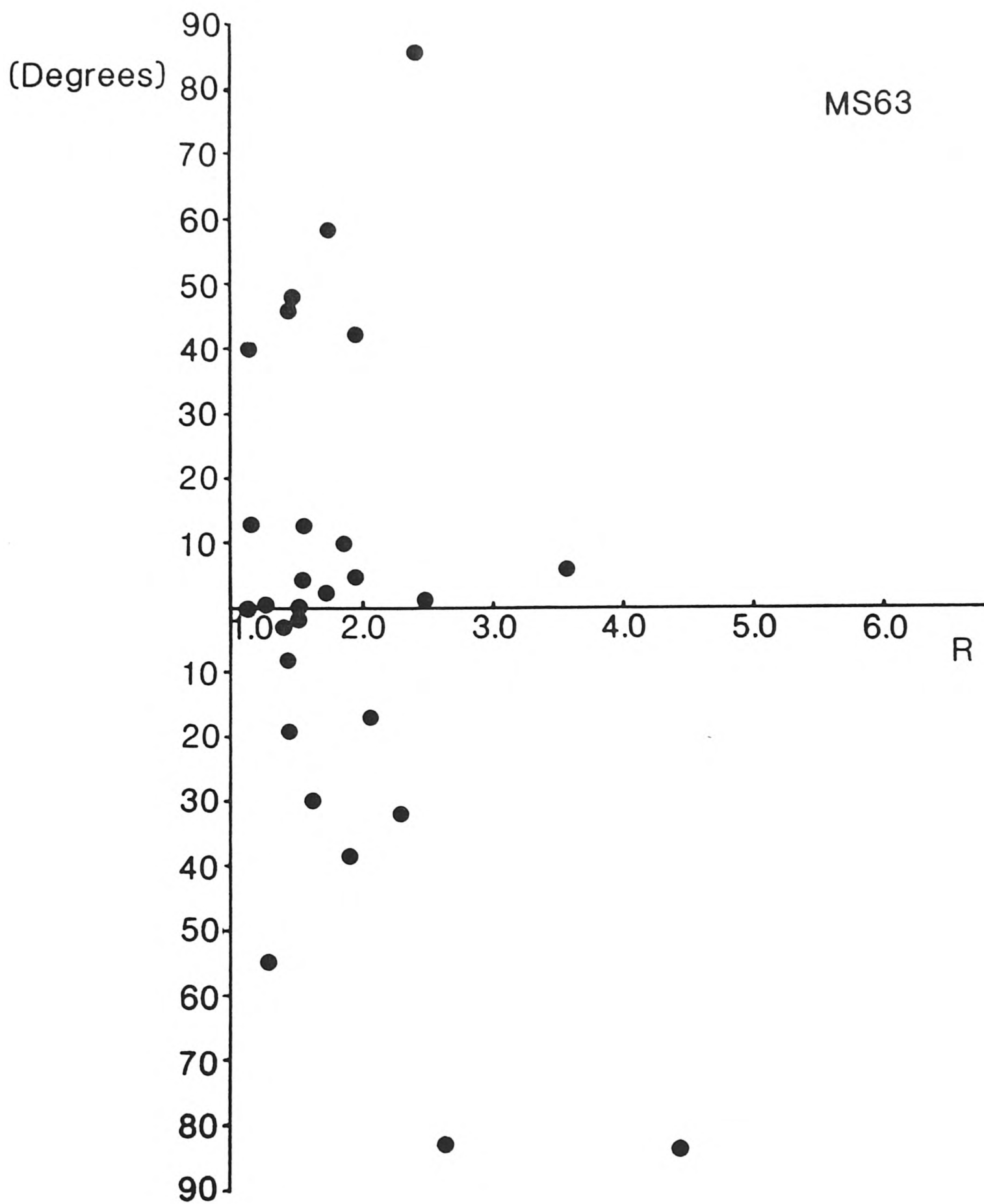
The albites in this study are all large compared to the grain size of the matrix, which implies that the deformation of the matrix can be considered continuous. There is only minor development of pressure shadows around the inclusions suggesting that diffusional mass transfer was not a major deformation mechanism. The rheology of naturally deformed polycrystalline aggregates is probably best described by a power law constitutive equation (Poirier 1985). Although in general different from the linear viscous behaviour assumed in the mathematical derivation of the formulae, this is not expected to have a significant effect on the results unless the strains are high (Ferguson 1979).

The mathematical analysis is only rigorous for rigid particles that are ellipsoids of revolution. However, Ferguson (ibid) suggests the results will not be markedly altered for objects with an orthorhombic symmetry. This theoretically derived conclusion is supported by the empirical work of Ghosh & Ramberg (1976) using rectangular blocks. In the two samples MS63 and K5 nearly all the albites have shapes which are approximately orthorhombic and any with obviously irregular shapes were not considered. Neither the albites nor the included quartz grains show any signs of internal strain (undulose extinction, subgrains etc.). In contrast the surrounding quartz matrix has strong undulose extinction and sutured grain boundaries, which are indicative of intracrystalline plastic deformation. This implies that the albites have behaved as rigid particles.

A further assumption made in this two dimensional analysis is that deformation was plane strain. The quartz c-axis fabric in K5 has a well-developed central girdle to the c-axis fabric, which implies there was no significant departure from plane strain conditions. However, MS63 has a concentration of c-axes around Y and a relatively weakly developed central girdle, which suggests non-plane strain conditions. This is supported by the point maxima of a-axes, distributed around small-circles about Z, which suggests deformation in the flattening field. The same inference can be made from the orientation of the eccs in the area (fig.4.2).

The effects of non-plane strain conditions on the rotation of rigid objects have been studied by Passchier (1987b), who shows that non-coaxial deformation in the flattening field causes a rapid rotation of the long axes of rigid inclusions away from the X-Z plane and into a stable orientation parallel to Y. A significant departure from plane-strain should, therefore, be readily identifiable by the preferred orientation of long axes perpendicular to the stretching direction.

Fig.4.21. Orientation of the long axes of porphyroclasts in a section perpendicular to Ls and Ss, with respect to Ss for sample MS63. The presence of porphyroclasts with high aspect ratios at high angles to Ss demonstrates that there was only minor deviation from plane strain in this sample.



Measurement of porphyroclasts in a section perpendicular to Ls for MS63 shows no preferred orientation of the long axes (fig.4.21). The presence of some albites with large aspect ratios at high angles to Ss clearly shows that extension parallel to Y, indicated by the quartz fabrics, was low-strain and its effects are ignored.

The angles of rotation were measured in sections parallel to Ls and perpendicular to Ss. Albites were not measured if their angle of rotation could have been influenced by movement on a nearby ecc or if the rotation of a porphyroclast brought it into contact with a neighbouring rigid inclusion.

4.6.4 Results

Two methods were employed for estimating S_r from the samples K5 and MS63.

(i) Estimate of S_r from stable positions of rotated porphyroblasts.

It is not possible to show that any particular rotating rigid object has reached its stable position. However, if an object has rotated through more than 90° in the direction of shear, then it is not restricted in its rotation. In this study a few porphyroblasts with low aspect ratios had their long axes and internal fabric oriented at a high negative angle to Ss (fig.4.22). I interpret these as objects which have rotated through an angle greater than 90° . This suggests that strain was relatively high and that porphyroclasts with higher aspect ratios have probably reached their stable positions. No clear microstructural division was observed between freely rotating porphyroblasts and those restricted in their rotation.

The stable orientation for rigid objects is given by equation 8. The minimum root mean squared difference was calculated between the data from the two samples and the theoretical curves for different S_r (fig.4.23). The higher the value of S_r , the greater the range of R

Fig.4.22. Porphyroclast orientation data for the two samples MS63 and K5. In general, low aspect ratios are associated with high rotation angles, see text for further discussion.

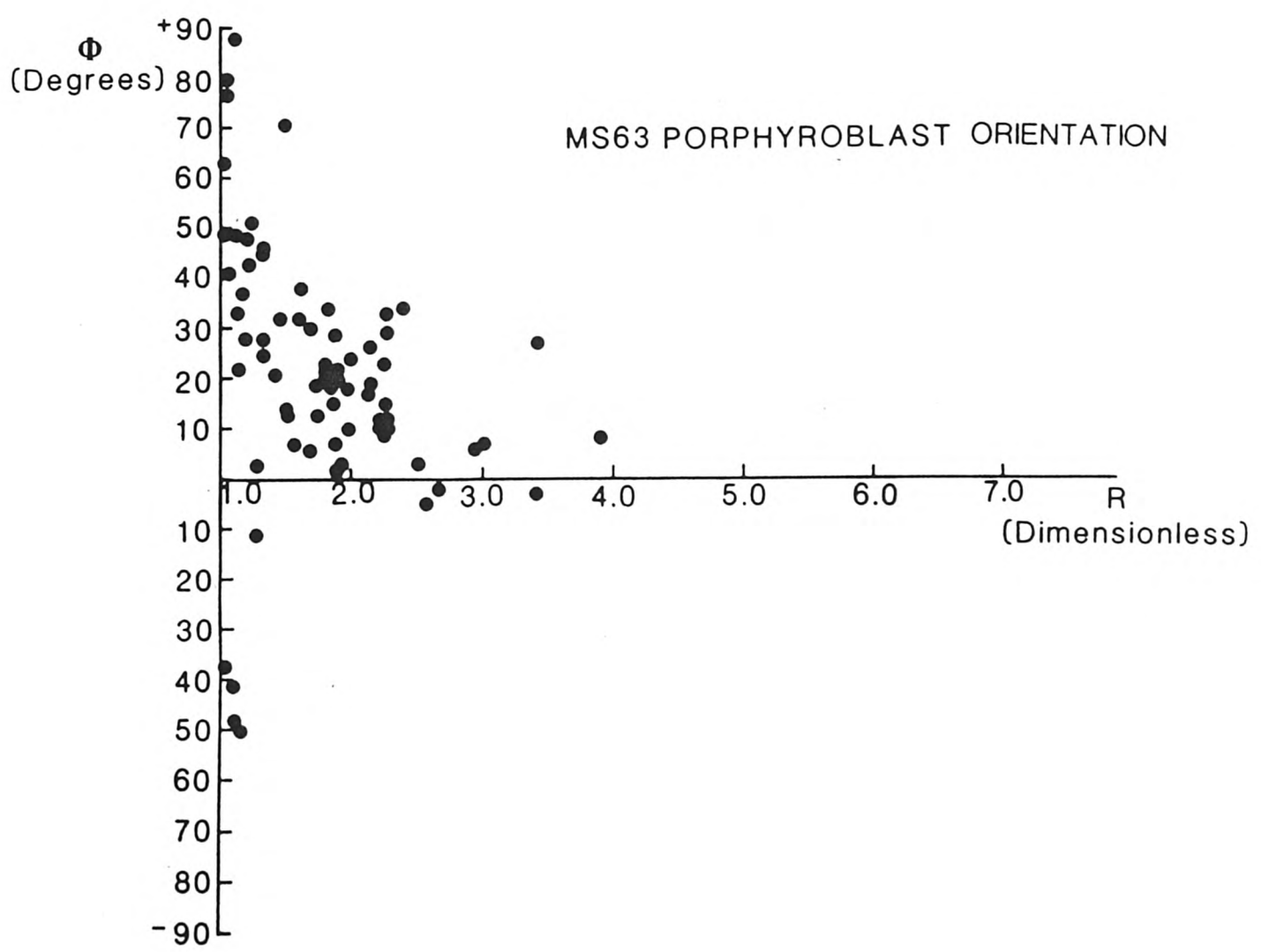
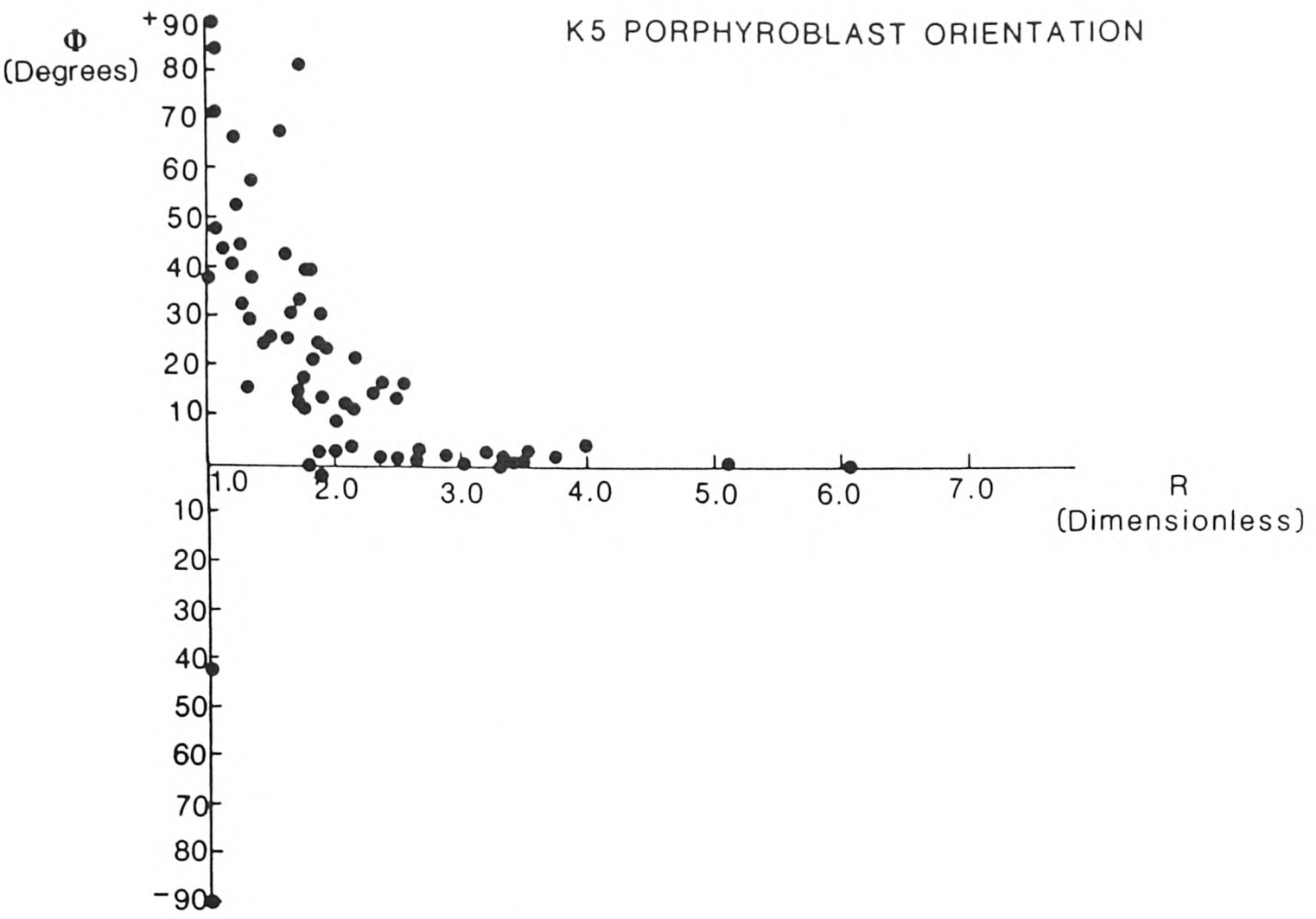
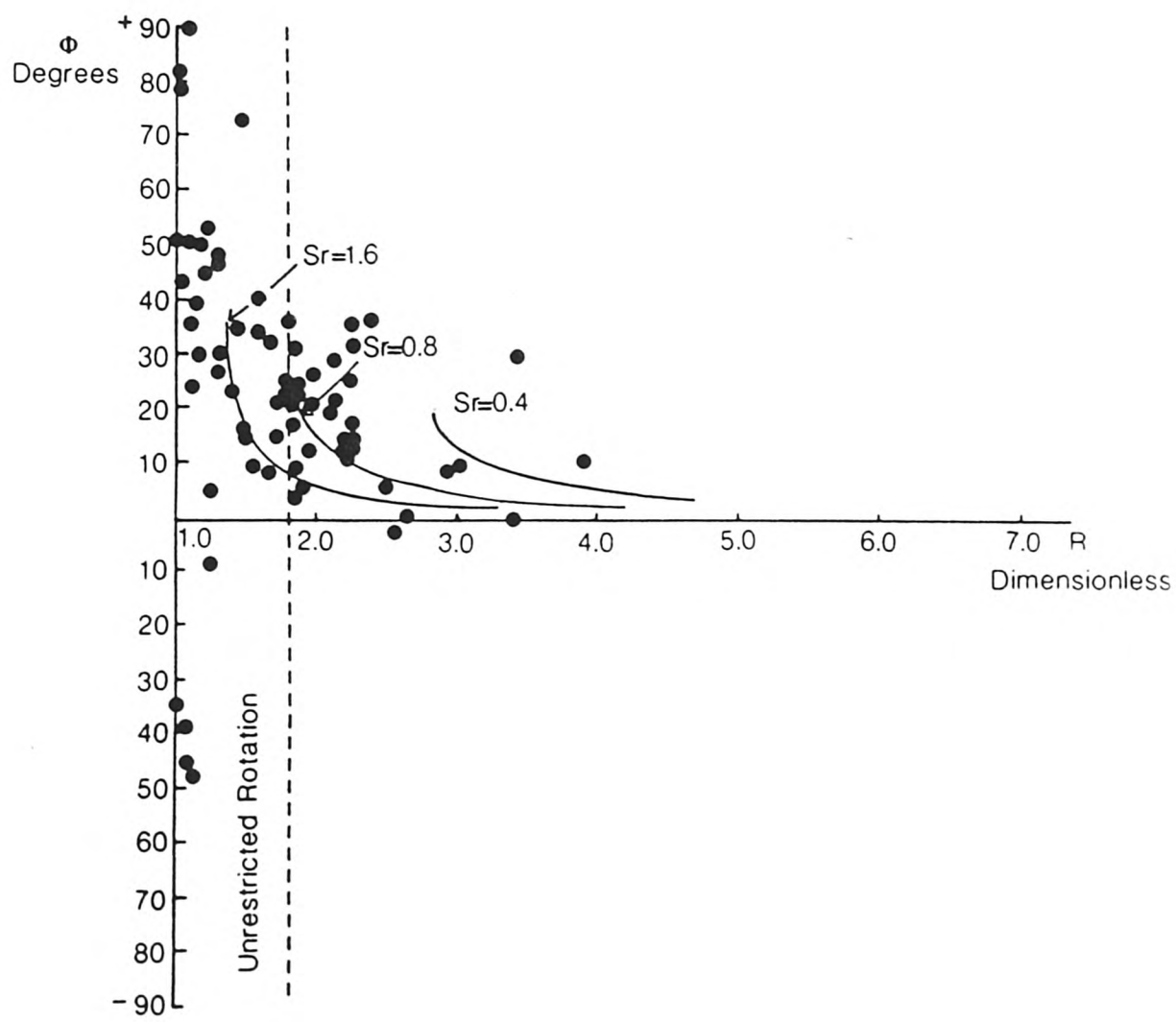
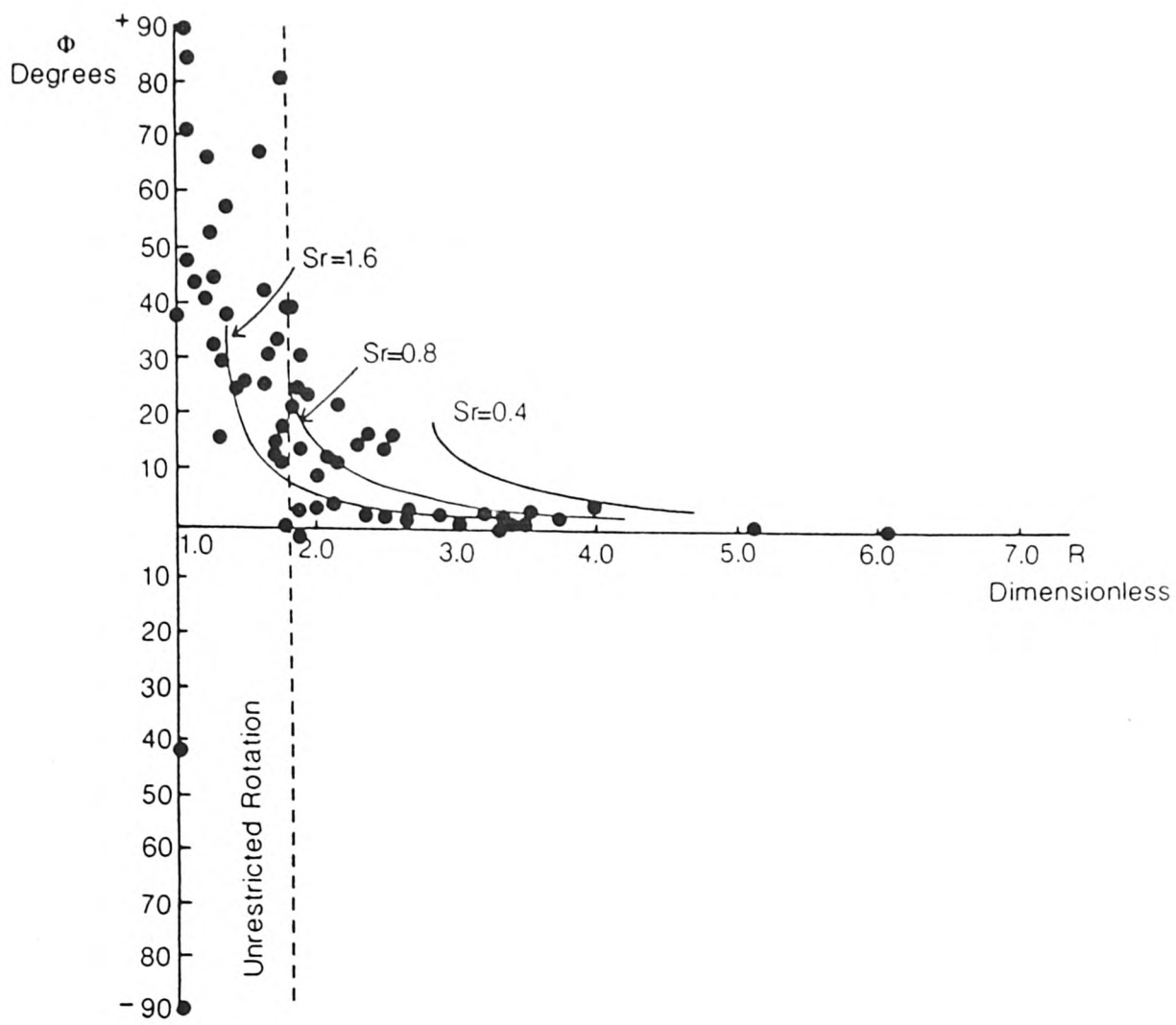


Fig.4.23. Porphyroclast orientation data for both samples with three different curves for three different values of S_r representing the theoretical stable orientations reached at high strain for rigid particles with different aspect ratios. Particles with aspect ratios less than a critical value, dependent on the value of S_r , will not come to rest in a stable orientation. The dashed line is the critical aspect ratio for $S_r = 0.8$.



for which there is a stable orientation, the R.M.S difference, therefore, includes more experimentally measured points for a higher S_r (fig.4.24). These results suggest a possible range of S_r for K5 (area 1) of $0.6 > S_r > 0.8$, and for MS63 (area 3) $0.6 > S_r > 0.8$.

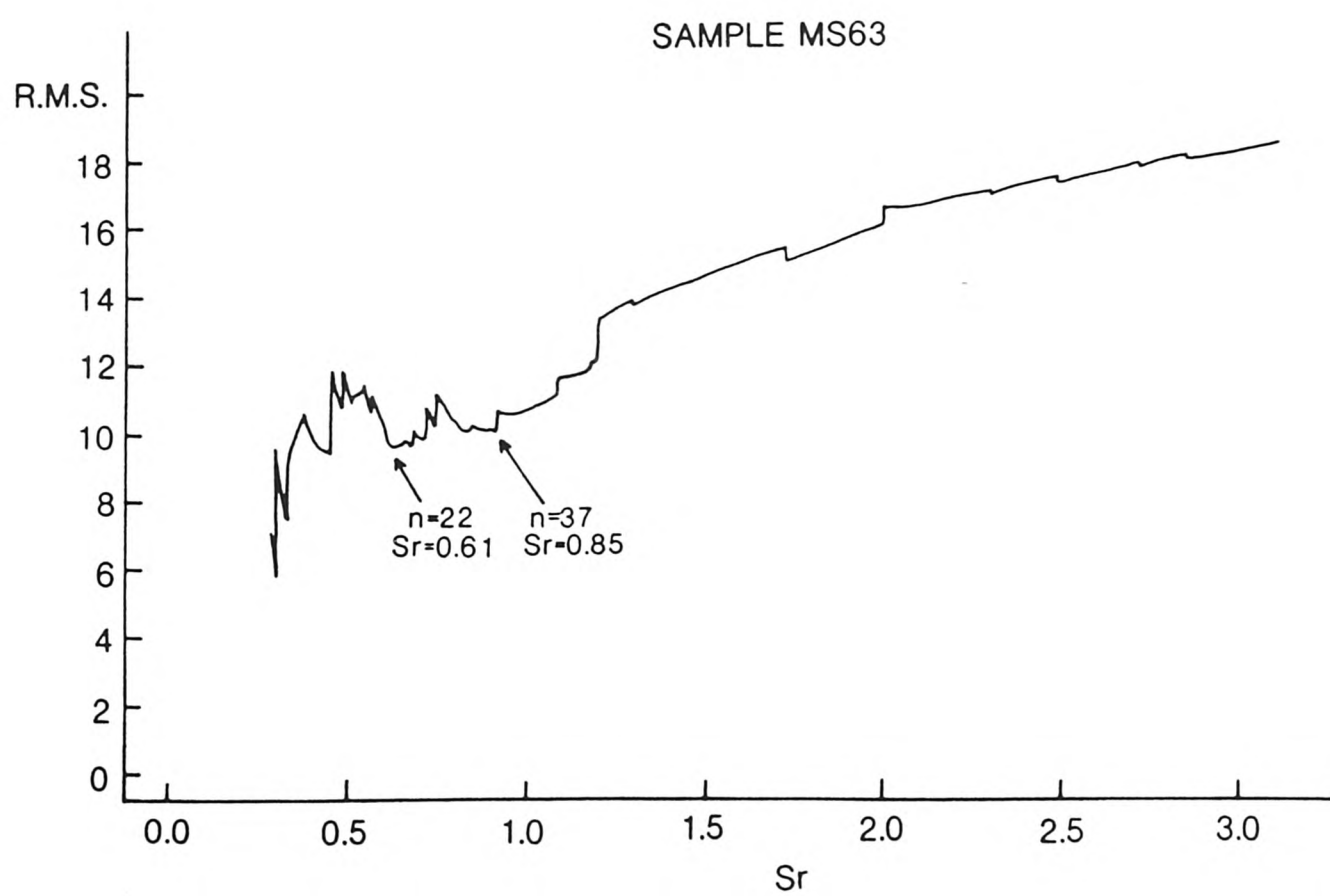
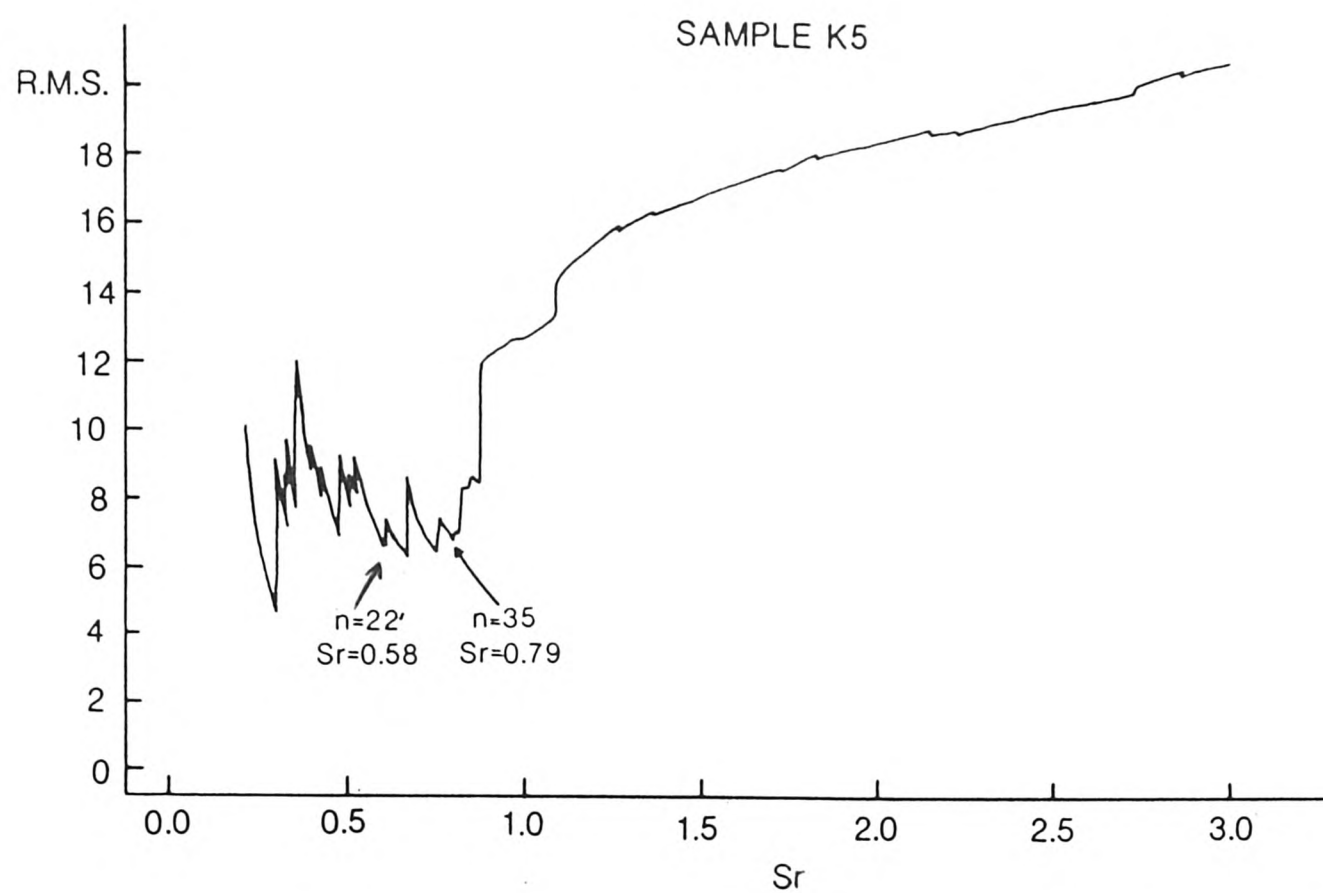
(ii) Estimate of S_r from the finite rotation angles of rigid inclusions.

The equations governing the finite angle of rotation, Φ , are more cumbersome than that governing the stable orientation. The three equations 5, 6, 7 give Φ as a function of Φ_0 , S_r , γ , and R . R and Φ are measured in the samples, but estimates of Φ_0 (the original orientation of the long axes of the porphyroblasts with respect to the D_s flow plane) and γ (the value of the shear strain) have to be made before S_r can be determined.

The continuity of the S_r fabric from the matrix of the tectonites into the porphyroclasts suggests that the albites have followed the S_r foliation during the major reorientation caused by D_s folding. S_r has been transposed during D_s and except for the hinge zones of crenulations, lies subparallel to S_s . If D_s had been purely coaxial this suggests that the long axes of the albites and their internal fabrics would now lie within S_s and it is the rotational component of D_s deformation that has caused a deviation away from this position, i.e. $\Phi_0 \neq 0$.

If Φ_0 is taken as 0° and the rotation of the internal fabric of the porphyroclasts with respect to S_s is due to the non coaxiality of D_s flow, the shear strain, γ , can be estimated from the angle of rotation of porphyroblasts with a circular cross section in sections perpendicular to their rotation axis. The coaxial component of D_s deformation will not cause these objects to rotate and their rotation is, therefore, a direct measure of the shear strain, γ . This rotation

Fig.4.24. Calculation of the value for S_r that gives the best-fit curve of the form given in equation 8 to the orientation data for MS63 and K5. The graphs are values of S_r plotted against the root mean squared difference between the theoretical curves and the data set. The value of S_r that gives the best fit curve is indicated where the R.M.S. difference is a minimum. The theoretical curves are undefined for R less than a critical value, and this value decreases for greater values of S_r . A greater number of points, n , is, therefore, included in the calculation at higher values of S_r and the fit is correspondingly better.



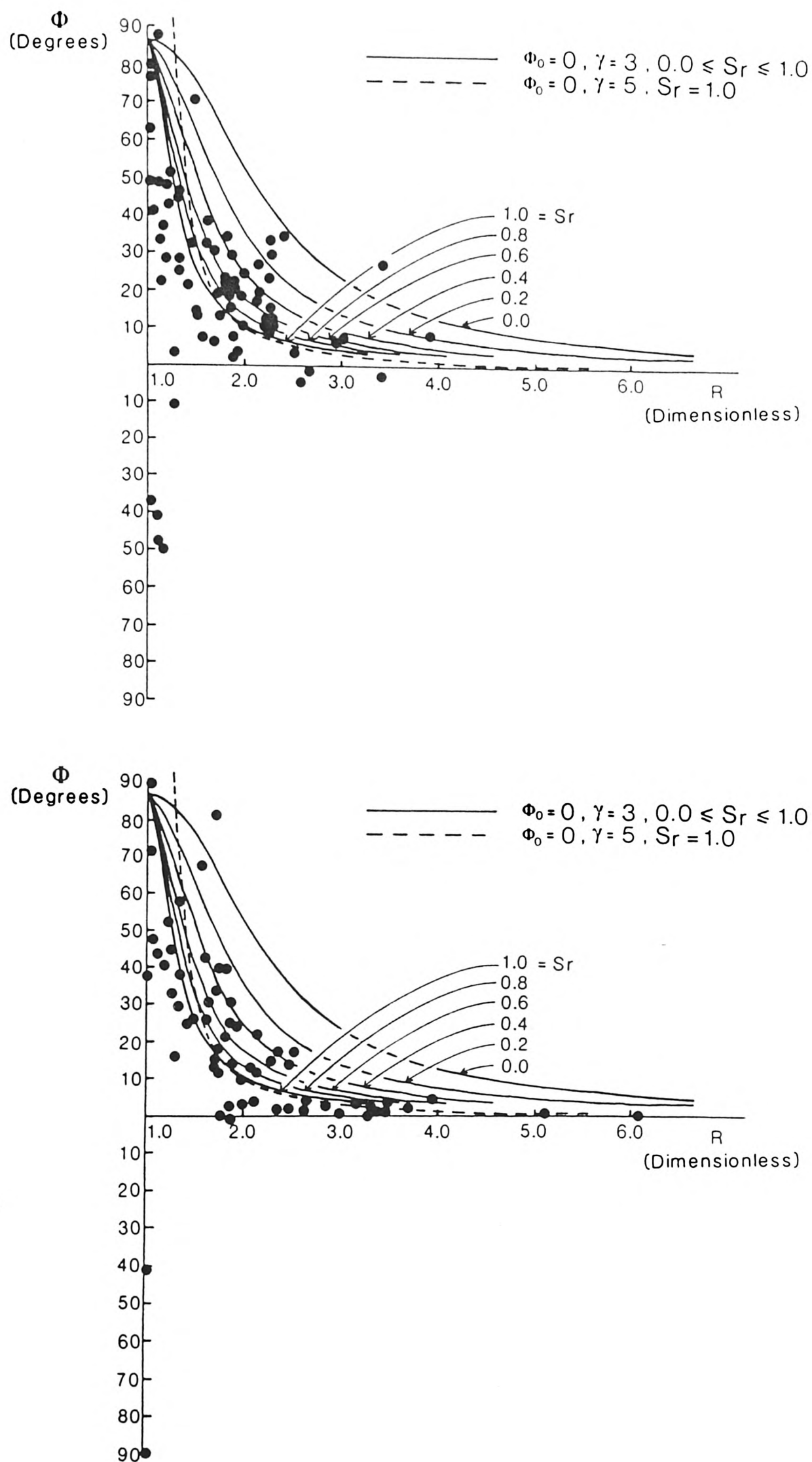


Fig.4.25. Estimation of S_r from the finite angles of rotation of the albite porphyroclasts. The curves are plotted using equations 5, 6 and 7 for $\gamma = 3$, $\Phi_0 = 0$, and different values of S_r . The dashed line is the theoretical curve for $\gamma = 5$, $\Phi_0 = 0$, and $S_r = 1$. See text for further discussion.

*

For a shear strain of 3 the difference between stable orientations attained at very high strains, and the theoretical position of porphyroclasts with aspect ratios $> R_{crit}$ is only a few degrees, except for values close to R_{crit} . It is therefore justified to use a stable orientation analysis to determine the kinematic vorticity number in these two samples.

can be estimated by the orientation of the internal fabric with respect to the external fabric (= Ss, the transposed Sr). In both samples some of the porphyroblasts with low aspect ratios have their internal fabrics and their long axes oriented at high negative angles to Ss. This is interpreted as their having been rotated through an angle greater than 90° . In both samples $\pi/2$ radians is, therefore, a conservative estimate of the angle of rotation of an object with an aspect ratio equal to 1. The rotation of a rigid sphere is related to the shear strain by $\omega = \gamma/2$ (Ghosh & Ramberg 1976), which implies a minimum shear strain for both samples of π , i.e. $\gamma \approx 3$.^{*} More of the porphyroclasts in MS63 are oriented at a negative angle to Ss than in K5, which implies a greater finite strain in MS63.

Using these estimates for Φ_0 and γ a series of curves were plotted for different S_r (fig.4.25). There is no clearly defined minimum R.M.S. difference but judging by eye, the best fit curves indicate, for K5 (area 1) $0.8 > S_r > 0.9$ and for MS63 (area 3) $0.6 > S_r > 1.0$.

Calculating the minimum R.M.S. difference without fixing γ , Φ_0 or S_r gives a range of possible best fit curves, which suggest: $-10^\circ < \Phi_0 < 10^\circ$; $\gamma > 5$, and $S_r > 1.0$ for both samples. The curve for $\Phi_0 = 0$, $S_r = 1.0$ and $\gamma = 5$ is plotted in fig.4.25 for reference.

Taking all the results together suggests that $\Phi_0 \approx 0$ and that minimum estimates for the shear strain and departure from simple shear are $\gamma = 3$, $S_r = 0.6$ for both samples.

4.7. ESTIMATE OF THINNING IN THE ZONE OF DISTRIBUTED Ds SHEAR

The kinematic vorticity number determined from the two samples, MS63 and K5, can be combined with the finite strain to obtain an estimate of the amount of thinning which has taken place perpendicular

to the shear zone boundaries. Both samples come from relatively high-strain Ds tectonites near the base of the Altkristallin which lie within a broad zone of distributed Ds shear affecting the basal 2km of Altkristallin and the upper parts of the Matrei Zone. Consistent quartz c-axis fabrics throughout this area suggest that deformation can be considered homogenous and that the two samples are characteristic for this deformation. The intensity of the quartz fabrics can be used as a qualitative measure of strain. There is no great difference between the fabric intensity of samples with a similar composition throughout the zone of distributed Ds shear. This implies that there is no significant change in finite strain.

Both quartz fabric analysis and porphyroclast rotation studies suggest that finite strain in area 3 is greater than in area 1. However, to obtain a minimum estimate for the thinning, a value of $\gamma = 2$ is taken for both areas as the average shear strain. The width of the zone of Ds distributed shear is determined by the extent of penetrative Ds fabrics associated with a strong quartz c-axis preferred orientation. Including the Matrei Zone, this is a width of 4km in area 1 and 2.5km in area 3. Taking an average map width of 3km and an average dip of 45° , this corresponds to a vertical thickness of $\approx 2.1\text{km}$.

The kinematic vorticity number, w_k is a measure of the instantaneous degree of non-coaxiality and can only be combined with the finite strain if a time constant behaviour is assumed. In the study areas there are several qualitative kinematic indicators which record the degree of non-coaxiality at different periods during Ds. Microstructures suggest that the area 3 can be divided into low-strain early-formed fabrics furthest away from the base of the Altkristallin and younger fabrics nearer the base (section 3.4.3). The quartz fabrics which were probably formed early in the development of Ds in area 3 are represented by the asymmetric cross girdle fabrics MS57, MS81, which

indicate a top to the NW sense of shear, and a symmetric fabric MS93. All three fabrics indicate a relatively low degree of non-coaxiality.

The last stage of deformation within shear zones is commonly recorded in the development of extensional crenulation cleavages (White et al 1980) although there is some evidence that these structures may form throughout deformation (Casa 1986). In area 1 the orientation of these structures suggests a final increment of deformation with a kinematic vorticity number less than 1 and a top to the NW sense of shear. In area 3 eccs suggest the same NW-SE extension direction but associated with a flattening strain. Therefore, the evidence from these various kinematic indicators suggests that the direction and general non-coaxial nature of the flow have remained roughly constant throughout the deformation. The late-stage flattening increment of deformation in area 3 is localized and low-strain.

A conservative estimate for the regional thinning is based on minimum values for S_r , γ and p_1 , the width of the zone of high strain distributed Ds shear. The calculation is done for $\gamma = 2$, $S_r = 0.6$, $p_1 = 2.1\text{km}$. The results are exponentially dependent on S_r and γ but linearly dependent on p_1 .

$$S_r = \dot{\epsilon}_x / \dot{\gamma}$$

for time constant behaviour therefore,

$$S_r = \ln(1 + e_x) / \gamma$$

Substituting the above values in the formula gives:

$$l_1/l_0 = e^{(0.6 \times 2)}$$

$$\underline{l_1/l_0 = 3.32}$$

Assume plane strain flow, i.e. $l_0 p_0 = l_1 p_1$ (fig.4.26)

$$p_0 / 2.1 = 3.32$$

$$\underline{p_0 \approx 7\text{km}}$$

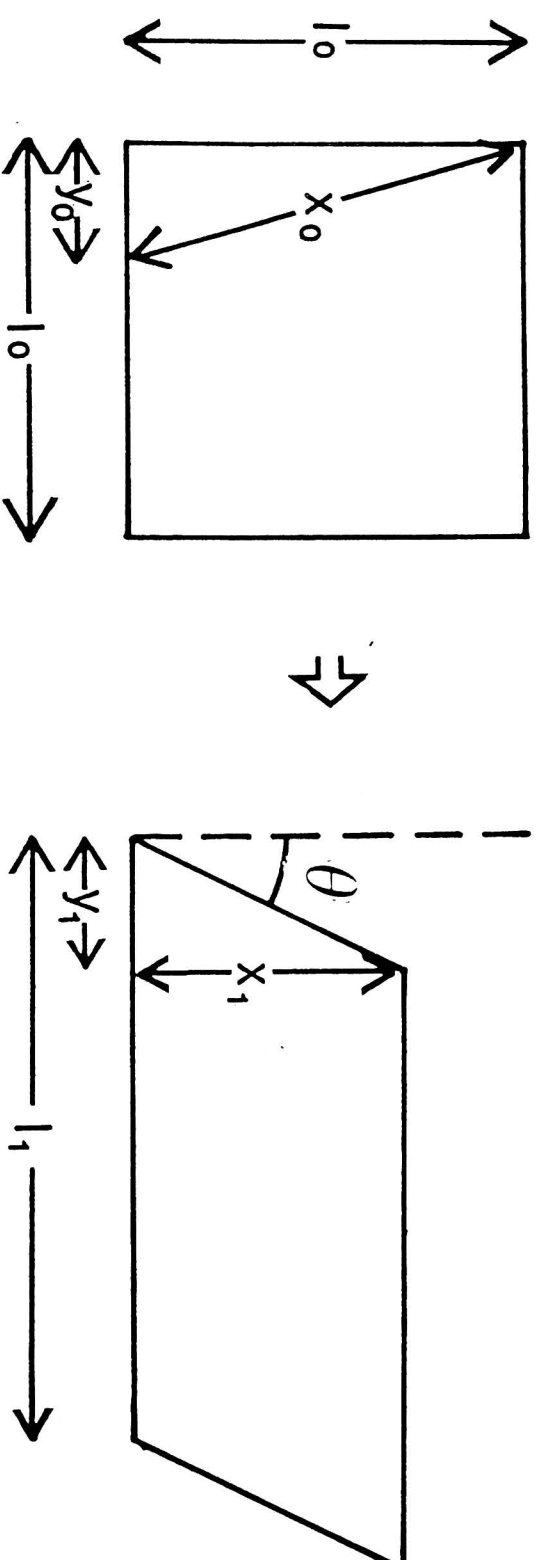


Fig.4.26b. Parameters used to determine the maximum principal strains, see text for further discussion.

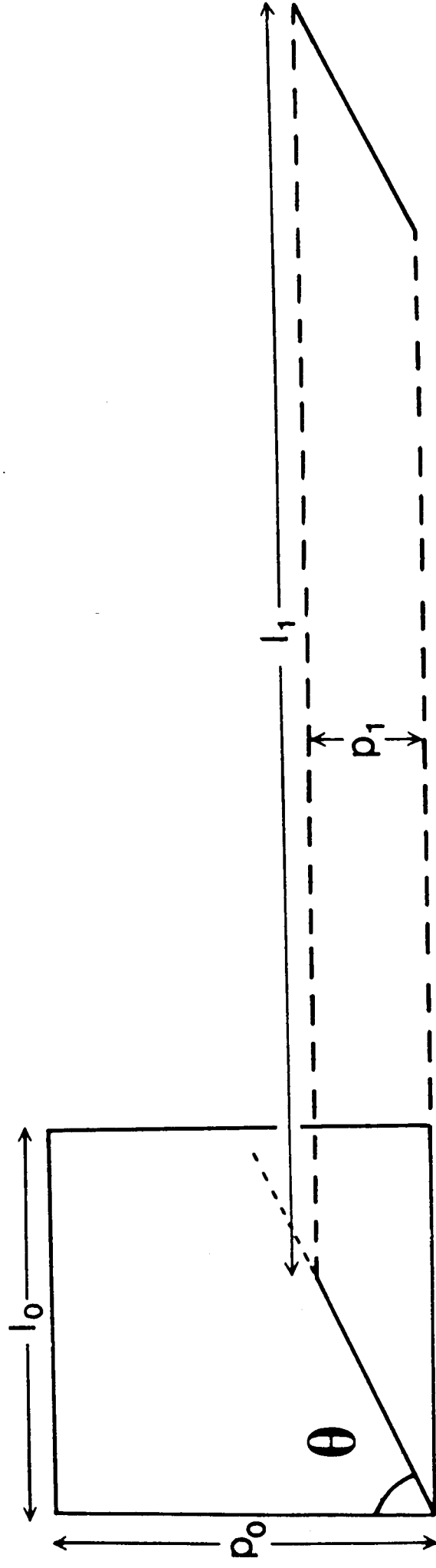


Fig.4.26. Schematic representation of Ds strain in the zone of distributed shear. The original dimensions of an arbitrary undeformed part of the shear zone are l_0 , p_0 , which have equivalents of l_1 , p_1 in the deformed state. Original thickness of shear zone is p_0 , final thickness is p_1 , $\tan^{-1}\theta = \gamma$ (shear strain). For plane strain $l_0 p_0 = l_1 p_1$. See text for further discussion.

Thinning is therefore

$$p_0 - p_1 \approx 5\text{km.}$$

Radiometric dating within the zone of distributed Ds shear suggests that this deformation was continuous from 70Ma to 40Ma (see 3.9.3). Given the estimates of thinning and the time it took, the average strain rate can be calculated.

First the maximum principal strain has to be found. This can be calculated from the elongation of two non-parallel material lines and the associated shear strain along one of them. Assuming plane strain deformation within the zone of distributed Ds shear and taking a unit square with sides of initial length l_0 , the lengths x_1 and l_1 can be calculated in the deformed state for different values of θ ($=\tan^{-1}\theta$) (fig.4.26). The principal strains can be calculated using the quantities $\lambda_1' = 1/(\lambda_1)$; $\lambda_x' = 1/(\lambda_x)$ and $\gamma_1' = \gamma_1/\lambda_1$, where λ_1 and λ_x are the quadratic extensions of lines l and x respectively and γ_1 the shear strain of line l .

For $\gamma = 2$, i.e. $\theta = \tan^{-1}2$ and $S_r = 0.6$

$$\begin{aligned} l_1/l_0 &= 3.32 \text{ (see above)} \\ x_1 &= 1/3.32 \ l_0 \text{ (since } x_1 l_1 = l_0^2, \text{ see fig.4.26b)} \\ &= 0.301 \ l_0 \\ y_1 &= 0.301 \tan\theta \ l_0 \\ &= 0.301 \times 2 \ l_0 \\ &= 0.602 \ l_0 \end{aligned}$$

$$\begin{aligned} y_0 &= 0.602/3.32 \ l_0 \\ &= 0.181 \ l_0 \\ x_0^2 &= l_0^2 + y_0^2 \\ x_0 &= 1.016 \ l_0 \\ \therefore x_1/x_0 &= 0.301/1.016 \\ &= 0.296 \end{aligned}$$

$$\begin{aligned} \lambda_x &= 0.0878 \\ \lambda_x' &= 11.385 \\ \lambda_1 &= 11.022 \\ \lambda_1' &= 0.091 \\ \gamma_1' &= -2/11.022 \\ &= -0.181 \end{aligned}$$

However, since x and y have been chosen such that they are perpendicular in the deformed state the values of the strain invariants J_1 and J_2 can be calculated :

$$J_1 = \lambda_1' + \lambda_2' = \lambda_1' + \lambda_x' = 11.476$$

$$\begin{aligned} J_2 = \lambda_1' \lambda_2' &= \lambda_1' \lambda_x' - \gamma_1'^2 && \text{(Ramsay 1967, equations 3-56, 3-57)} \\ &= 1.036 - 0.033 \\ &= 1.003 \end{aligned}$$

λ_1' and λ_2' can then be expressed in terms of the strain invariants:

$$\lambda_1' = \frac{J_1 - (J_1^2 - 4J_2)^{\frac{1}{2}}}{2}$$

$$\lambda_2' = \frac{J_1 + (J_1^2 - 4J_2)^{\frac{1}{2}}}{2} \quad \text{(Ramsay 1967, equations 3.60a, 3.60b)}$$

Substituting the above values for J_1 and J_2 gives

$$\begin{aligned} \lambda_1' &= 0.088 \\ \lambda_1 &= 11.364 \\ (1 + e_1) &= 3.371 \end{aligned}$$

$$\text{The average strain rate can then be calculated from the time taken} \\ \dot{e}_{av} = 3.371/30 \times 10^6 \text{ years} = 3.6 \times 10^{-15} \text{ s}^{-1}$$

This is the natural strain rate and does not remain constant through time for time constant flow (Pfiffner & Ramsay 1982). Finite strain is related to the linear strain rate for steady flow by :

$$\begin{aligned} (1 + e) &= e^{\dot{e}t} \text{ (Ramberg 1975).} \\ \ln(1 + e) &= \dot{e}t \\ \ln(3.371) &= \dot{e} \times 30 \times 10^6 \text{ years} \\ \dot{e} &= 1.3 \times 10^{-15} \text{ s}^{-1} \end{aligned}$$

The same calculations can be done for any value of γ . For $\gamma = 3$, leaving S_r fixed at 0.6, the thinning perpendicular to the shear zone boundaries is 10.5km; $(1 + e_1) = 4.55$; and $\dot{e} = 1.6 \times 10^{-15} \text{ s}^{-1}$.

This is within the bounds of generally accepted geologically reasonable strain rates (Pfiffner & Ramsay 1982, Price 1975).

4.8. DEFORMATION IN Ds SHEAR ZONES

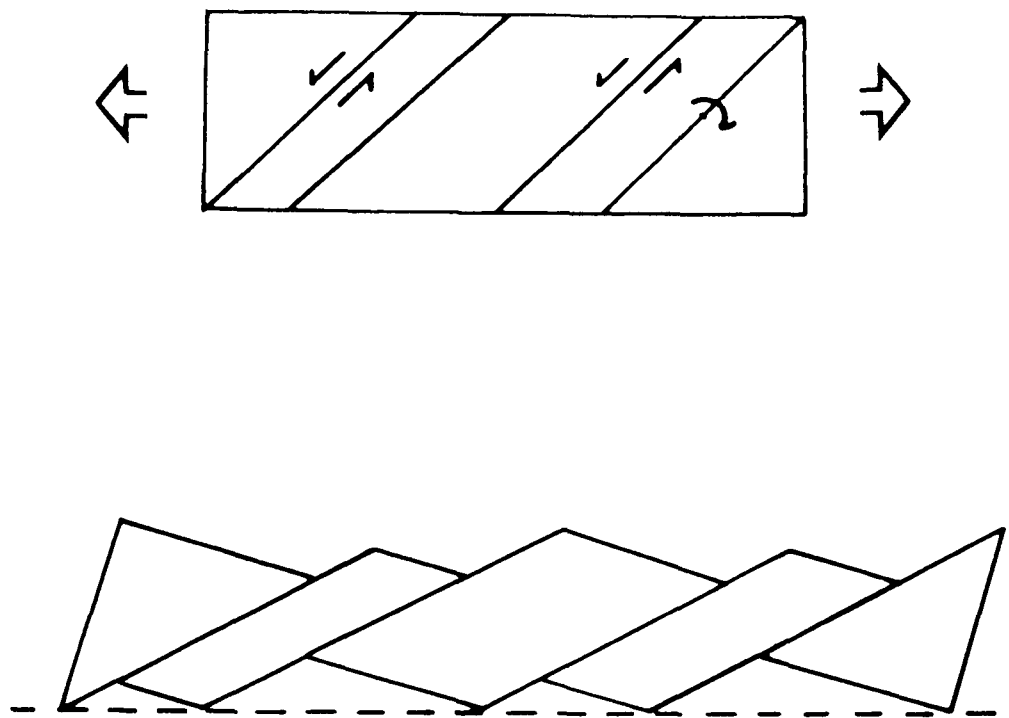
At higher structural levels within the Altkristallin above the zone of distributed Ds shear, Ds is also expressed as a series of discrete shear zones which are associated with a top to NW sense of shear (chapter 3). At least one of these shear zones cuts down section to the base of the Altkristallin. No suitable samples of shear-zone material were found in this study for either quartz fabric analysis or porphyroclast rotation studies. However, in a comparable shear zone in the Kreuzeck Group, Hoke (1987) measured an inclined single girdle quartz c-axis fabric indicative of simple shear with a top to the NW sense of shear.

The underlying zone of Ds deformation is associated with strong thinning perpendicular to the shear zone boundaries. To accommodate this extension the overlying material must also extend and thin. Deformation within at least one of these shear zones is probably simple shear with a kinematic vorticity number of 1. For lateral extension to occur, a simultaneous rigid body rotation has to be superimposed on the deformation, which acts to reduce the bulk vorticity (fig.4.27). Evidence for the rotation of the shear zone boundaries with respect to the flow plane may be recorded in the orientation of the finite strain ellipsoid (Passchier 1986). Due to the lack of suitable strain markers, this possibility could not be investigated. However, the extensional nature of these shear zones can be demonstrated in two instances. In the Kreuzeck Group the E-W trending Teuchl Shear Zone (TSZ) separates two distinct metamorphic zones. To the north, amphibolite grade Alpine metamorphics are preserved, which have an estimated equilibrium T of

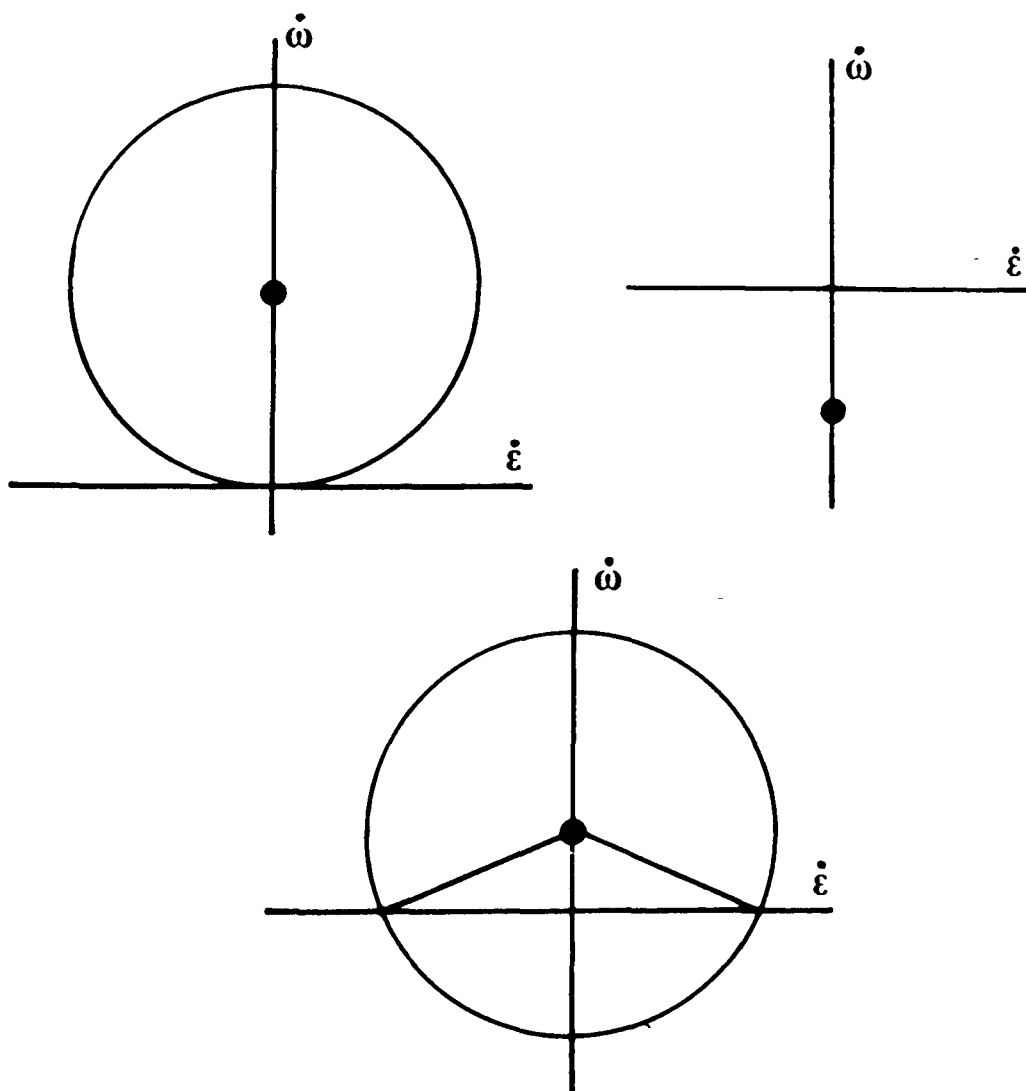
Fig.4.27a. Strain partitioning into shear zones undergoing simple shear whose boundaries are rotating in the reference frame thus reducing the bulk vortical component of deformation and allowing extension to take place.

Fig.4.27b. Mohr circle representation of deformation partitioned into simple shear and a rigid body rotation with an opposed sense of vorticity, which together constitute a general non-coaxial deformation with a lower rotational component than simple shear.

a.



b.



620±60°C (Hoke 1987). Immediately to the south of the shear zone, Variscan-aged metamorphic basement is exposed. This basement has been affected to some extent by the Alpine metamorphism, since the K/Ar radiometric ages in the biotites have been reset. However, the muscovites still preserve unaltered Variscan ages (Brewer 1970), which suggests that a temperature between the K-Ar blocking temperatures of biotite and muscovite was reached i.e. $\approx 400^{\circ}\text{C}$ (fig.4.28). A thickness of material corresponding to a jump of 200°C has, therefore, been removed from along this shear zone after the peak Alpine metamorphic conditions.

The Mellenkopf Shear Zone (MSZ) is a splay from the TSZ (fig.4.28), which can be traced towards the base of the Altkristallin (chapter 3). The displacement and thinning of the zone of distributed Ds shear in the area where the MSZ cuts through the Altkristallin into the underlying Matrei Zone shows that movement along this zone was top to NW. The offset of the base of the Altkristallin has a normal sense and acts to extend and thin the Austroalpine sheets. In the Altkristallin immediately overlying this shear zone the pre-Ds fabric is deformed into an open fold, which can be interpreted as a roll-over structure related to the listric geometry of the shear zone.

Estimates of thinning are difficult to obtain and rely on recognizing and quantifying jumps in metamorphic grade across the shear zones. Even at elevated geothermal gradients of $50^{\circ}\text{C km}^{-1}$, the sudden increase in the Cretaceous metamorphic temperature from S to N across the Teuchl Shear Zone from $\approx 400^{\circ}\text{C}$ to 600°C suggests that at least 4km of metamorphic stratigraphy have been removed from along this zone. This estimate is comparable with the minimum estimate of 5km thinning within the basal zone of distributed Ds shear.

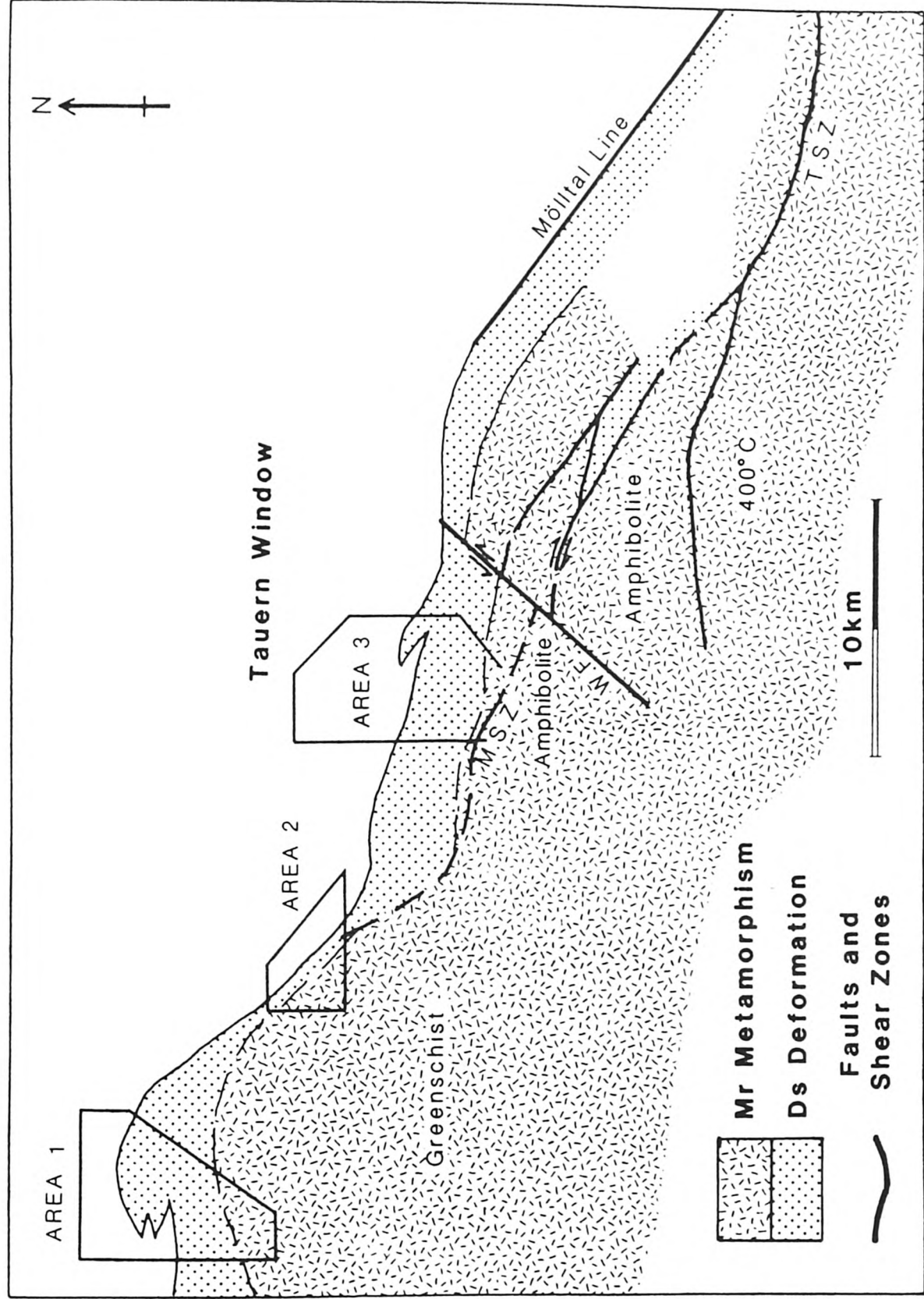


Fig.4.28. Cretaceous metamorphism within the Altkristallin to the south of the Tauern Window and Ds shear zones. The shear zones to the east of the Wöllratten Fault are mapped after Waters (1976) and Hoke (1987). T.S.Z. = Teuchl Shear Zone, M.S.Z. = Mellenkopf Shear Zone, W.F. = Wöllratten Fault.

4.9. REGIONAL THINNING OF THE AUSTROALPINE NAPPES

This is derived from adding the two estimates of thinning from both structural levels. Assuming a time-constant flow, the zone of Ds along the base of the Austroalpine is associated with a minimum of 5km thinning. Metamorphic evidence suggests a similar estimate of 4km thinning from the Teuchl Shear Zone at higher structural levels. There may be further evidence for thinning at deeper levels than the Matrei Zone and at higher structural levels than the Teuchl Shear Zone. This gives a minimum estimate for the regional thinning south of the Tauern Window of $\approx 10\text{km}$, which is in very good agreement with the estimates of regional thinning from metamorphic evidence (chapter 5).

PART B

4.10. KINEMATICS OF Dr

On average the Dr stretching lineation trends E-W (fig.3.32). However, within area 1 Dr stretching lineations are preserved in an area of relatively low Ds strain, which have an anomalous N-S to NE-SW orientation. These can be explained as the result of folding and rotation during Ds (fig.4.29b). A quartz c-axis fabric measured in strongly lineated quartz vein material folded around a NE-vergent Ds fold, indicates a top to the N sense of shear (fig.4.29a). Throughout the Altkristallin the Ds folds predominantly verge to the NE-quadrant. The paucity of S-vergent Ds folds suggests that the Ds fold axes generally rotated with the same anticlockwise sense in map view. The top to the N sense of shear determined from the Dr quartz fabric may, therefore, correspond to an original top to the NW quadrant sense of shear. Similar quartz fabrics indicating a NE-SW oriented stretching

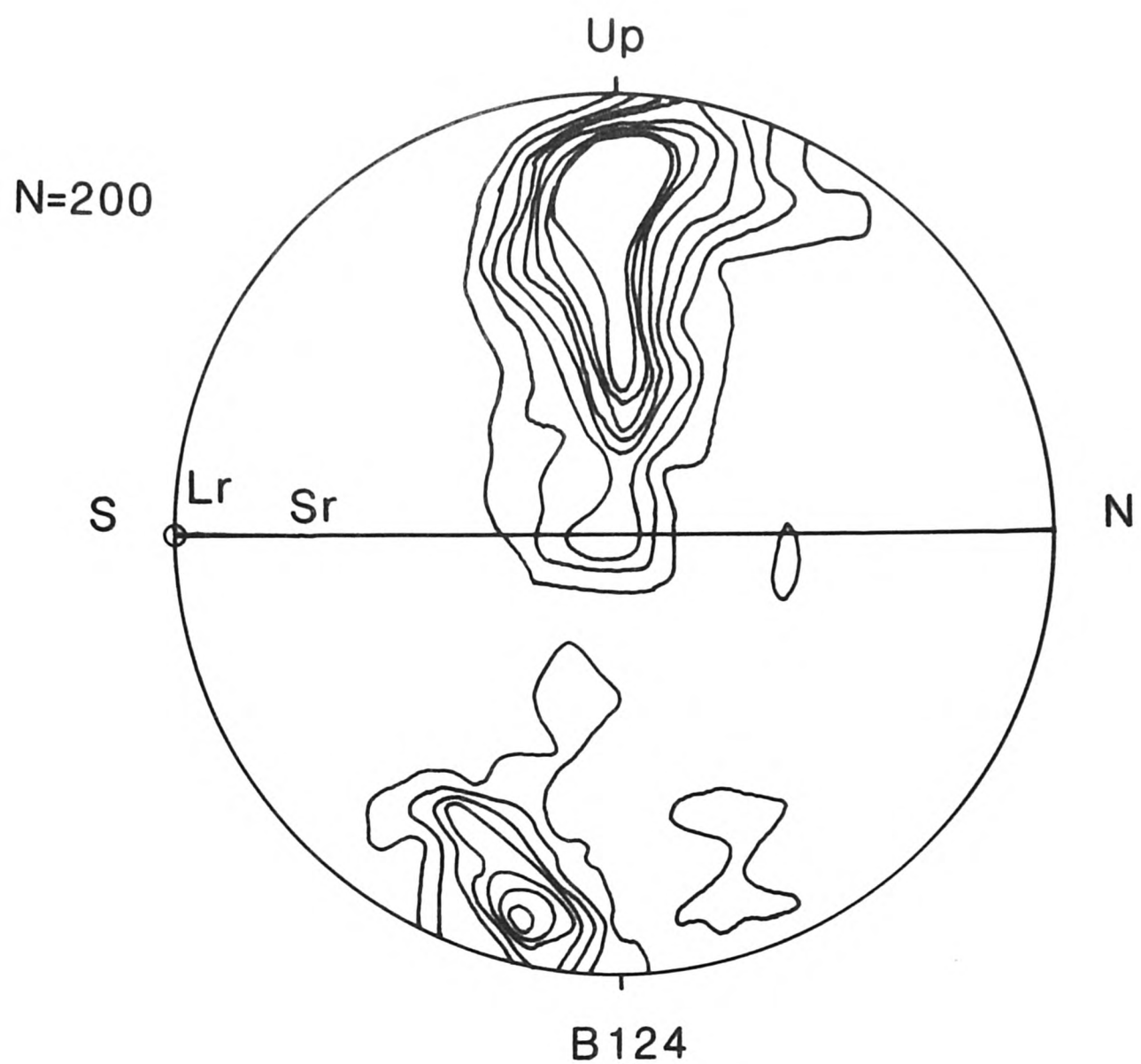


Fig.4.29a. Dr quartz c-axis fabric measured in a quartz vein, which had been folded by Ds. The present-day orientation of the Dr stretching lineation is N - S.

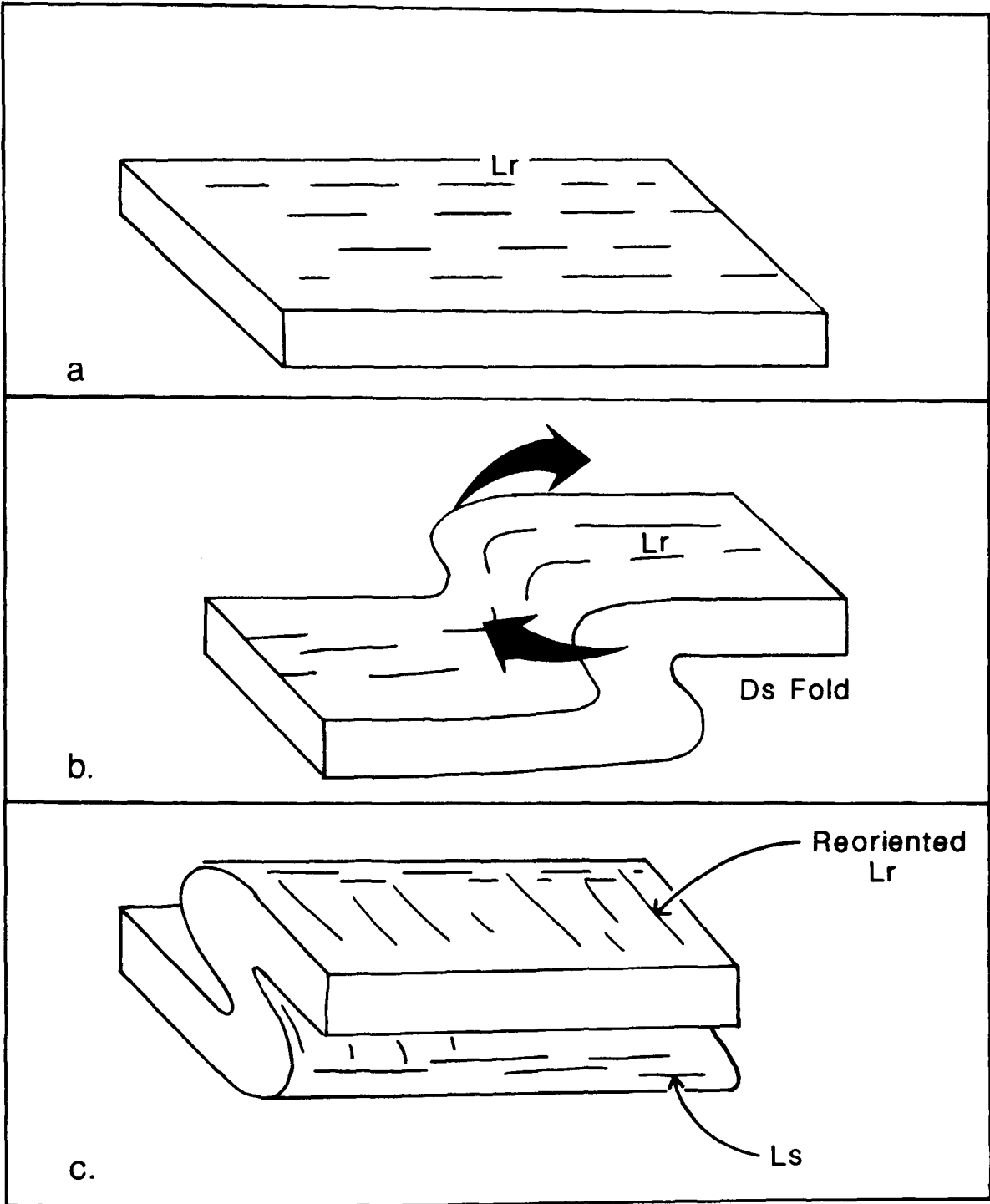


Fig.4.29b. Reorientation of Lr by folding and subsequent passive rotation during Ds.

direction were measured from the Sr inclusion trails within porphyroblasts in samples MS63 and K5 (fig.4.20).

The rotated garnets found in area 3 associated with Sr, could in principal be used as kinematic indicators but these are only well-developed in the zone strongly affected by Ds and the uncertainties about the effects of this subsequent deformation are too great for any reliable sense of shear to be determined.

South of area, 1 Dr is the dominant deformation, where it can be directly compared to Dj defined by Behrmann (1987b). Dj is also associated with E-W stretching lineations and quartz fabrics indicate a top to the W sense of shear (Behrmann 1987b).

PART C

4.11. THRUSTING ALONG THE AUSTRALPINE - PENNINE BOUNDARY

Within the study area, part of the deformation history is characterized by thrusting along discrete discontinuities. No kinematic information is preserved for movement along any individual thrust plane. However, the geometry of a linked system of thrusts can be used to determine the approximate tectonic transport direction. Various linear elements can be defined in the geometric analysis of thrust systems, e.g. cut off lines and branchlines. These dominantly form at a high angle to the thrust direction but may also be a lateral feature parallel to the movement direction (Butler 1982).

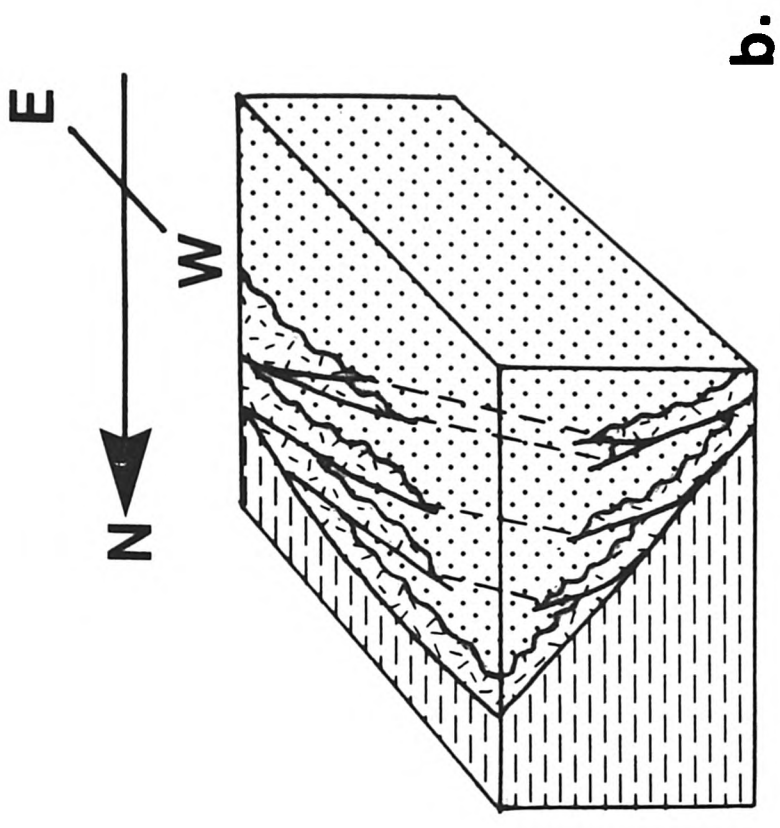
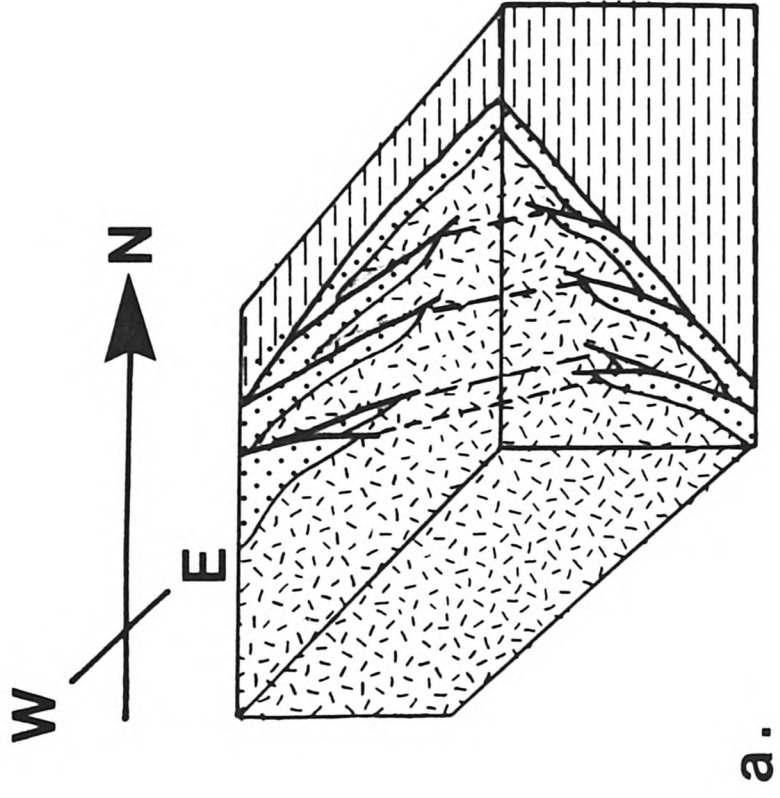
The imbrication within the Matrei Zone is too poorly preserved for a full geometric analysis. However, along the contact with the Altkristallin the imbricates can be mapped out. These imbricates are the result of two phases of thrusting. No kinematic data are available for the first phase.

In general the imbricates along the boundary between the Altkristallin and the Matrei Zone have slices of basement which pinch out to the W and slices of quartz phyllite which pinch out to the E (fig.3.1a,c.). This can be geometrically explained in two ways; either the basement overlies the quartz phyllite and the cut off lines plunge to the E, or the quartz phyllite is the structurally highest unit and the cut off lines plunge W (fig.4.30). Sections across this boundary show that the basement overlies the quartz phyllite and therefore suggests that the cut off lines plunge E.

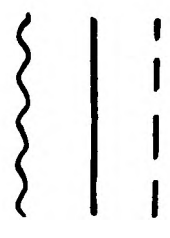
In area 1 the plunge of the cut-off lines for one group of imbricates can be directly constrained by constructing an E-W section parallel to Ss and using the topography. This shows that the cut off lines plunge to the E (fig.4.31). However, their original position will have been altered by the subsequent Ds deformation, by being rotated into the stretching direction. In this area the average plunge of Ls in the plane of the section is $\approx 0^\circ$. This suggests that the cut off lines originally had a steep eastward plunge, and have been rotated through an unknown angle during Ds into their present orientation.

The consistent geometry of the imbricates along the boundary between the Altkristallin and Matrei Zone over a wide area suggests that these are frontal features, which formed at a high angle to the movement direction. The general eastward plunge, therefore, implies a NE movement of the Altkristallin over the Matrei Zone.

Fig.4.30. Schematic block diagrams representing two possible geometric interpretations of the observed map relationships between the quartz phyllite and the Altkristallin along the southern margin of the Tauern Window. All the available evidence suggests that the quartz phyllite underlies the Altkristallin and, therefore, suggests that the cut-off lines have a general eastward plunge.



Buendnerschiefer
Quartz Phyllite
Altkristallin



Unconformity
Thrust
Cut off Lines

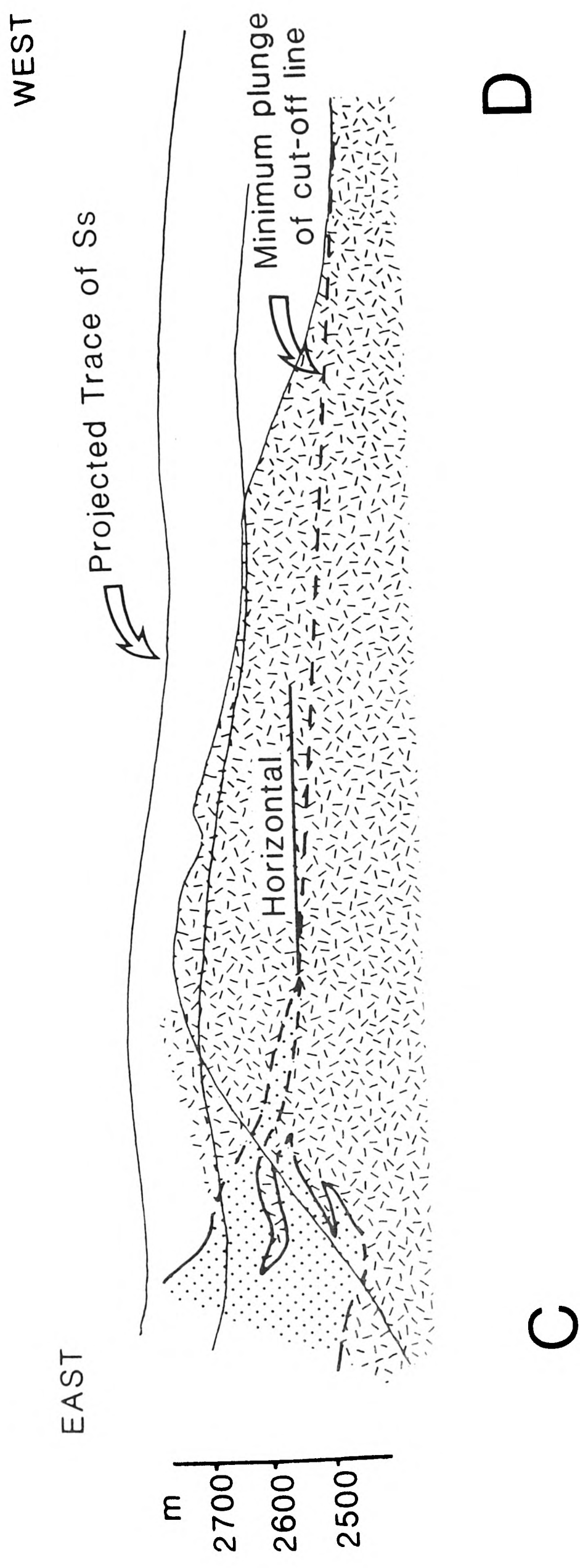


Fig.4.31. East-west cross-section through the imbricate stack in area 1. The trace of Ss has been projected into the plane of the section. This must also contain the projected orientation of the Ds stretching lineation. Ss has an approximately horizontal trace in this section and the minimum plunge of the imbricate cut off line is 5° E. See fig.3.1b. for location of section.

CHAPTER 5 CONVERGENT TECTONICS IN THE EASTERN ALPS.

5.1 INTRODUCTION

In this chapter I divide the convergent tectonic history of the Eastern Alps into two main tectonic episodes, both of which were associated with a top to the NW quadrant sense of shear. The first manifestation of convergence was a phase of shortening and crustal thickening which probably began around 120-130Ma and affected both the Austroalpine and the Pennine domains. The resulting Cretaceous burial metamorphism is characterized by high P/T conditions in the Pennine domain and Barrovian facies in the Austroalpine domain. Postdating crustal thickening, a major phase of crustal extension took place. This affected both the Austroalpine domain and the upper Pennine units, and caused substantial thinning of the Cretaceous metamorphic sequence. Radiometric dating suggests this deformation began around 70Ma and may have continued after collision at around 40Ma.

Along the southern margin of the Tauern Window, between these two dominant phases a large-scale out of sequence thrust juxtaposed the Austroalpine and Pennine units with their very different Cretaceous metamorphic histories. At deeper structural levels in the Pennine domain, deformation is of a similar age to the thrusting and associated with a NE-SW stretching lineation but the kinematic data are insufficient to speculate on the causal tectonic mechanisms.

The Austroalpine and Pennine domains have both been strongly deformed and neither represents a rigid plate. Crustal thickening is the direct result of the convergence between Adria and Europe, however, the evidence for contemporaneous regional extension also taking place in this overall convergent tectonic setting implies that gravitational

forces acting on the thickened crust of the destructive margin were a second major cause of deformation. Throughout convergence in the Austroalpine domain, the dominant sense of shear is to the NW. This constrains the large-scale plate convergence vector to the NW quadrant from the Lower Cretaceous onwards. This is incompatible with the most recent Africa - Europe motion paths and implies that Adria moved as an independent microplate from the onset of subduction in the Lower Cretaceous.

5.2 CRUSTAL THICKENING

5.2.1 The evidence from metamorphism

A prima facie case for Alpine crustal thickening can be made in the Eastern Alps by considering the early Alpine Cretaceous metamorphic history. This can be divided into two characteristic facies associations:

- (i) a high P/T facies caused by subduction, mainly restricted to the Pennine domain; and
- (ii) a Barrovian-type metamorphism found within the Austroalpine domain.

The Barrovian type metamorphism in the Austroalpine domain is associated with widespread 80Ma K-Ar cooling ages and is in part contemporaneous with the high P/T metamorphism seen in the Pennine units. Estimates for the peak metamorphic conditions (amphibolite grade) are given by Hoke (1987) for the area to the south of the Tauern Window as 6.25 ± 1.25 kb and $620 \pm 60^\circ\text{C}$. This is in broad agreement with estimates of 7 kb, 600°C (Frey 1984, Weber 1983) in the Koralpe and 5-7 kb, 600°C (Thöni 1986) in the Oetztal Alps for the peak conditions during the same metamorphic event. There have also been some suggestions that relic eclogite metamorphism found in the Altkristallin

may be partly an Alpine effect (Frank 1983, Hoke 1987), but this has yet to be substantiated. $P_{\max}$ during the Cretaceous metamorphism in the Altkristallin is, therefore, 6-7kbar which is equivalent to a depth of burial to $\approx 22\text{km}$ ($\rho = 2.8\text{gcm}^{-3}$).

To decide whether a phase of crustal thickening is required to account for these metamorphic conditions, it is necessary to know something about the crust prior to the Cretaceous metamorphism.

In places the original sedimentary contact between the Alpine sedimentary cover and the underlying Altkristallin is exposed. In some areas this erosion surface locally cuts into a series of Variscan-aged sillimanite-bearing gneisses, which ^{locally} show evidence of partial melting (Hoinkes 1973, Hoke 1987). In the Oetztal Alps Hoinkes (1973) estimates the conditions of this Variscan metamorphism as 680°C , 4kb ($\approx 15\text{km}$). Therefore, even before extension during the opening of the Tethyan region, the crust had already been thinned by erosion, which in places reached a depth of $>15\text{km}$.

From Permian to Triassic times only a relatively thin veneer of sediments was deposited above the region where the Variscan-aged high grade metamorphics were exposed ($<2\text{km}$) (Tollmann 1977a), reflecting the slow subsidence rate of the area. Further to the palaeogeographic south, subsidence was more rapid and a maximum sedimentary thickness of 6-7km is developed, e.g. Northern Calcareous Alps.

In the Jurassic there was a sudden increase in the rate at which the Tethyan Ocean was opening, which produced a regional and rapid increase in water depth throughout the Tethyan region. The contemporaneous sedimentation could not compensate for subsidence and the dominantly shallow-water limestones of the early Jurassic are overlain by a series of pelagic sediments including radiolarites and locally Aptychus and red nodular limestone (Tollmann 1977a). These deposits are all indicative of several kilometres water depth (Bernoulli

& Jenkyns 1974) and show that the continental crust must have been further drastically thinned at this time. Assuming an initial, pre-rifting thickness of 20km; $\rho_{\text{crust}} = 2.8\text{gcm}^{-3}$, $\rho_{\text{mantle}} = 3.3\text{gcm}^{-3}$; and water depth 2-3km, implies a Jurassic crustal thickness of 6-12km.

The Jurassic crust was, therefore, much too thin to account for the Cretaceous metamorphic pressures of 6-7kbar ($\approx 22\text{km}$ depth) found in parts of the Austroalpine nappes. A major phase of thickening is required to produce the recorded metamorphic conditions.

Two other possible tectonic settings have been proposed for the Cretaceous metamorphism in the Austroalpine domain:

(i) the high T zone commonly found above subducting slabs in circum-Pacific destructive margins (Hawkesworth et al 1975); and

(ii) the high T zone developed during crustal extension (Frank 1983).

In both settings the heating of the crust is due to heat transfer by igneous processes. Beyond the excess argon in some of the samples used for dating around the S.E. Tauern (Brewer 1970) and one dyke with a Mid Cretaceous cooling age (Waters 1976), there is no evidence of a contemporaneous magmatic arc or its associated zone of high T/P metamorphism. To circumvent this lack of evidence Roeder & Bögel (1978) propose the ad hoc hypothesis that the original igneous arc has been subducted away.

In summary, the Cretaceous metamorphism within the Austroalpine domain requires parts of the sequence to have been buried to depths of $>22\text{km}$. The evidence for originally thin Austroalpine crust and the lack of significant igneous activity, forms a compelling argument in favour of crustal shortening and thickening being necessary to account for these metamorphic conditions.

A similar argument can be made using the Cretaceous metamorphic conditions in the Pennine domain, which is characterized by high P/T metamorphism. Regional high P/T metamorphism is characteristic of

destructive plate margins and is formed where cool material has been rapidly buried to produce a crustal-scale thermal perturbation with anomalously low geothermal gradients (Oxburgh & Turcotte 1970).

Evidence for regional high P/T metamorphism is found mainly, but not exclusively, in the Pennine domain throughout the Alpine chain. In the Pennine units of the Tauern Window in the Eastern Alps, Selverstone (1985) has estimated regional $P_{\max}$ of >10kb (>35km depth of burial). In the Eastern Alps the Austroalpine crust was only around 10km thick before the onset of convergence. The metamorphic pressures recorded in the Pennine domain of the Tauern Window, therefore, require substantial tectonic thickening, which probably involved both the Austroalpine and the Pennine units. At the time of the peak metamorphism, the total structural thickness must have been at least 35km.

5.2.2 Structures related to shortening and their timing

The primary structural evidence for pre-metamorphic shortening and thickening in the Eastern Alps comes from the Austroalpine nappes. To shorten and thicken the stretched Austroalpine material sufficiently to cause the amphibolite metamorphism, requires at least a doubling of thickness, i.e. a shortening of 100%. Such a high value of shortening implies the Austroalpine domain underwent strong internal deformation. For structures to be related to this thickening, their formation must be synchronous with or predate the Cretaceous metamorphism. Radiometric dating shows that the Altkristallin cooled at around 80Ma from a thermal peak, which to the south of the Tauern Window may have been reached around 100Ma (Brewer 1970, chapter 3).

Within the cover nappes of the Austroalpine domain, deformation is present at two different structural levels both of which duplicate the stratigraphic sequence, and are therefore, related to thickening. At deeper structural levels, within the Austroalpine basement, early

ductile deformation may also be related to crustal thickening.

Deformation at each of these three levels is discussed briefly below.

(i) The pre-Gosau phase of deformation.

This is found at the highest levels in the Austroalpine domain, (fig.5.1) and is represented by a series of thrusts and folds which are unconformably overlain by the Gosau sediments (Oxburgh 1969, Tollmann 1977a).

This series of dominantly shallow-water clastic lithologies were first deposited in the Cenomanian, which gives an upper time limit for the pre-Gosau deformation. Sedimentation in the deformed underlying Austroalpine units, actually continued until the base of the Cenomanian in the western part of the Northern Calcareous Alps (Gwinner 1978). This is, however, only one local occurrence and elsewhere in the Austroalpine nappes no pre-Gosau sediments younger than Neocomian (120-130Ma) are found (Tollmann 1977a, Gwinner 1978). In the western part of the Northern Calcareous Alps, the pre-Gosau deformation can, therefore, be dated as Cenomanian (100Ma) but it may have begun earlier in other parts of the Austroalpine domain.

(ii) Emplacement of Palaeozoic phyllites onto Mesozoic cover rocks (UAA - MAA boundary of Tollmann).

In several parts of the Austroalpine domain in the Eastern Alps, Alpine shortening can be demonstrated where pre-Alpine Palaeozoic quartz phyllite is thrust onto a Mesozoic cover sequence above Austroalpine metamorphic basement. This contact has been studied at the base of the Gurktaler and Northern Calcareous Alps (Ratschbacher & Oertel 1987, v. Gosen 1982, fig.5.1). Alpine metamorphic mineral growth can be demonstrated in both areas, where greenschist minerals (including garnet and biotite) are developed in Mesozoic sediments. In the Gurktaler Alps mineral growth overprints the formation of the thrust contact and movement predates metamorphism (v. Gosen 1982). Hawkesworth (1976)

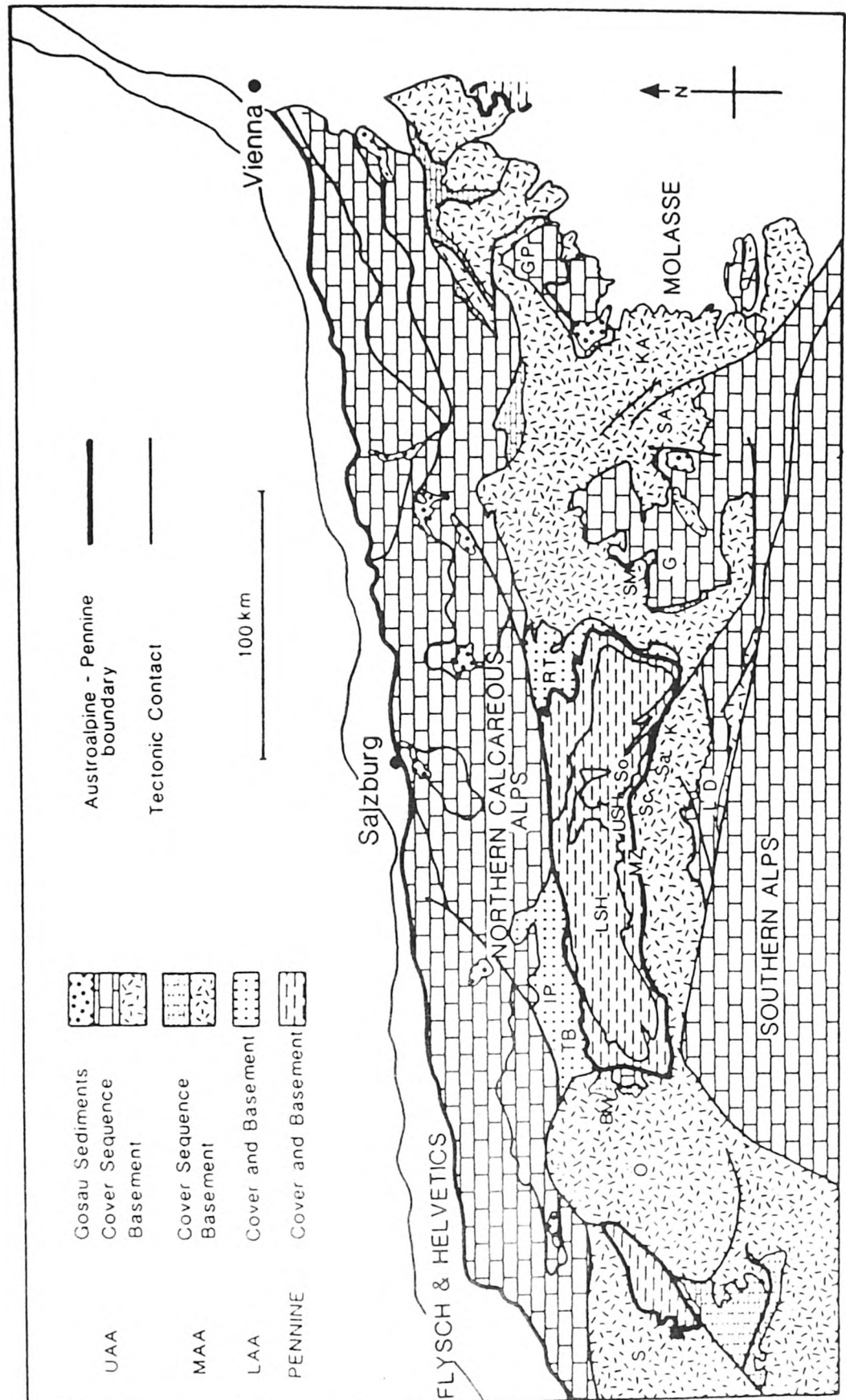


Fig.5.1. The principal tectonic divisions of the Eastern Alps (after Tollmann 1977).

UAA = Upper Austroalpine, MAA = Middle Austroalpine, LAA = Lower Austroalpine. O = Oetztal Alps, S = Silvretta, BM = Brenner Mesozoic, LSH = Lower Schieferhülle, USH = Upper Schieferhülle, Sc = Schober Group, Sa = Sadrnig Group, K = Kreuzeck Group, LD = Lienzer Dolomites, RT = Radstädter Tauern, MZ = Matriei Zone, TB = Tarntaler Berge, SM = Stangalm Mesozoics, G = Gurktaler Alps, KA = Koralpe, SA = Sau Alps, GP = Graz Palaeozoics, IP = Innsbruck Quartz Phyllite.

gives Rb/Sr mica ages of 99 ± 1 Ma and 92 ± 6 Ma from the Altkristallin immediately adjacent to the Gurktaler Alps, suggesting that movement was older than ≈ 100 Ma.

(iii) To the south of the Tauern Window, ductile deformation within the Altkristallin predates the peak of the Cretaceous metamorphism and may have begun as early as 120 Ma (see 3.9.2). Microstructural relationships show that deformation began before major growth of metamorphic minerals but had ceased before the peak conditions were attained (chapter 3). This suggests that deformation took place with an increasing grade of metamorphism, and may reflect a phase of ductile crustal thickening.

In all three cases, therefore, the formation of the structures predates the peak of the Cretaceous metamorphism and is probably older than 100 Ma. This suggests that these structures are all related to the early phase of crustal thickening responsible for Cretaceous metamorphism in the Austroalpine domain. K-Ar and Rb-Sr radiometric dating of deformed white micas in various parts of the Eastern Alps suggests that early Alpine deformation began around 110–120 Ma (Nowy 1977, Stöckhert 1984, Brewer 1970, Satir 1975).

The imbrication between the Matrei Zone and the Altkristallin occurred between 80 and 70 Ma and cuts rocks affected by the Cretaceous metamorphism (see 3.11.1). These structures are, therefore, too young to be related to the original phase of crustal thickening.

The Pennine tectonic units consist of sediments and basement fragments, which represent the floor of the Tethyan ocean basins. During convergence, these units also underwent strong shortening and thrust planes are commonly marked by horizons decorated with serpentinite lenses or gneiss lamellae (Exner & Prey 1964). Thrust planes have also been identified by the preferred alignment of dolomite lenses (Tollmann 1977a), but these features may also have an olistolithic origin (see chapter 2).

High P/T metamorphism in the Pennine units developed at depth along the destructive, northern margin of the Austroalpine domain. The greatest age of this metamorphism, therefore, gives a lower time limit for the onset of subduction. In the Western Alps radiometric dating shows that Penninic material was undergoing regional high P/T metamorphism by 100Ma (Hunziker 1974, Deutsch 1983, Frey et al 1974), which implies that subduction was already active at this time.

In the Eastern Alps the evidence is more cryptic but nevertheless points to the same conclusion. In the Northern Calcareous Alps a characteristic Cenomanian facies is locally developed called the Randcenoman. The dominant heavy mineral of this facies is chromite, which is associated with clasts of serpentinite, suggesting that oceanic crust was already exposed and being eroded at this time ($\approx 100\text{Ma}$). Palaeocurrent data indicates the source area for this detritus lay to the north of the Austroalpine units (Dietrich & Franz 1976). Detrital lawsonite and sodic amphibole, characteristic of high P/T metamorphic conditions occur in the Cenomanian - Turonian flysch deposits in western Austria (Oberhauser 1968, Winkler & Bernoulli 1986). At geologically reasonable strain rates of 10^{-14} s^{-1} , high P/T rocks may rise to the Earth's surface from a depth of 20-30km, some 10-20Ma from the time of subduction (Platt 1987). Making the not unreasonable assumption that the high P minerals and the erosion of oceanic crust is related to the formation of the Alpine subduction zone, this indicates that compressional tectonics and subduction had begun by 110-120Ma.

5.2.3 Estimates of orogenic shortening

Estimates of shortening are important for any reconstruction of the Eastern Alps. Although some attempts have been made to line balance sections across the Pennine units (Tollmann 1977a), the structure is too

complex and as yet poorly understood for any reliable estimates to be made by this method.

Within the Austroalpine domain there are two very different views of the large-scale structure, which are represented by the cross-sections of Clar (1965) and Tollmann (1977a).

Largely on the basis of sedimentary facies comparisons, Tollmann groups the Austroalpine nappes into three major tectonic units, the Upper, Middle and Lower Austroalpine Sheets (UAA, MAA, LAA), and suggests they represent laterally continuous thrust nappes. The highest of these thrust sheets (UAA) is now represented by a number of isolated Klippen bounded by tectonic contacts along the base. Tollmann links these together and suggests they represent a once continuous and extensive thrust sheet covering most of the Austroalpine domain. By line balancing N-S sections, Tollmann deduces a shortening of 500km within the Austroalpine domain, with an original N-S dimension of 600km. By postulating a once laterally continuous and relatively thick UAA thrust sheet, Tollmann explains the Cretaceous metamorphism by a phase of thickening within the Austroalpine domain, involving a 600km wide zone of inhomogenous deformation.

Clar has a very different approach, and considers the individual Klippen of Tollmann's UAA sheet as parautochthonous to the MAA basement, and that they have moved a relatively short distance to the north. In contrast to Tollmann, Clar postulates a N-S shortening of only 50km across an area originally 150km wide and his reconstruction largely eliminates the laterally continuous and far-travelled nappes suggested by Tollmann.

One of the major problems which Clar helped to solve by proposing a lesser amount of regional movement than Tollmann, was the lack of sufficient basement to accommodate the areal extent of the Austroalpine sedimentary units, in the Eastern Alps. At present the maximum N-S

dimension of the basement is 100km and Tollmann's reconstruction requires a N-S dimension of at least 600km. It was this lack of space that first lead earlier workers to suggest the now discredited existence of Unterströmungszonen or downwelling zones, where continental crust was absorbed into the underlying mantle (Ampferer 1906).

Clar suggests the different sedimentary sequences of the UAA and the MAA developed in close proximity to one another. However, this implies very rapid lateral variations in both sedimentary facies and thicknesses without any evidence of major synsedimentary faulting in the basement of this area. The amount of shortening proposed by Clar is also insufficient to account for the Cretaceous metamorphic conditions seen in many parts of the Austroalpine domain.

The two views of Clar and Tollmann represent extreme possibilities for the deformation within the Austroalpine domain and are equivalent to a N-S shortening of 50km and 500km respectively.

5.2.4 Kinematics of crustal thickening

In their reconstructions both Tollmann and Clar assume a general N-S movement of the Austroalpine nappes over the Pennine units. This is an assumption which permeates even the modern-day literature (Selverstone 1985). Apart from the general E-W trend of the Alpine chain, this conclusion is based on the false premise that fold hinges and stretching lineations form, and remain, perpendicular to the movement direction during deformation. However, with increasing finite strain linear elements such as fold axes are likely to rotate towards the stretching direction (Bell 1978, Williams 1978), which in many cases approximates to the direction of shear. A movement vector oblique to the trend of the orogen would allow the Austroalpine nappes to be derived from along strike of each other. The UAA tectonic units could then be far-travelled as suggested by Tollmann's sedimentary facies

arguments and the basement problem would be alleviated by relocating it to the Hungarian basin, now covered by Tertiary deposits. This leaves the mechanism open, by which the dominantly cover rocks of the Upper Austroalpine sheet were so neatly detached from their underlying basement.

There have been several suggestions based on sedimentological evidence, that convergence in the Eastern Alps was oblique to the present trend of the mountain belt e.g. Frisch (1979), Bechstädt (1978). A limited amount of structural data is now available from the Austroalpine domain that can be used to determine the movement direction. In the following, discussion will be entirely restricted to those structures that are related to the initial phase of crustal thickening. The kinematics of deformation related to crustal thinning will be discussed later.

Kinematic data for the early thrusting within the Pennine domain and the pre-Gosau deformation is non-existent. However, the thrust contact along the base of the UAA sheet where Palaeozoic sediments are emplaced over a Mesozoic sequence has been studied along the base of the N.C.A. (Ratschbacher & Oertel 1987) and the base of the Gurktaler Sheet (von Gosen 1982). Von Gosen suggests the first Alpine deformation along this contact is associated with an E-W/WNW-ESE stretching lineation. He does not record any reliable sense of shear indicators but the lateral cut-off features of the overthrust Stangalm Mesozoics on a map-scale imply a top to the NW sense of shear (fig.5.1, 5.2.).

Along the base of the Northern Calcareous Alps, Ratschbacher & Oertel (1987) have reported a deformation which is partly synchronous and partly postdates the Cretaceous metamorphism, which involves the transport of material to the WNW. It is unclear whether this should be related to the same phase of shortening and thickening or a later event.

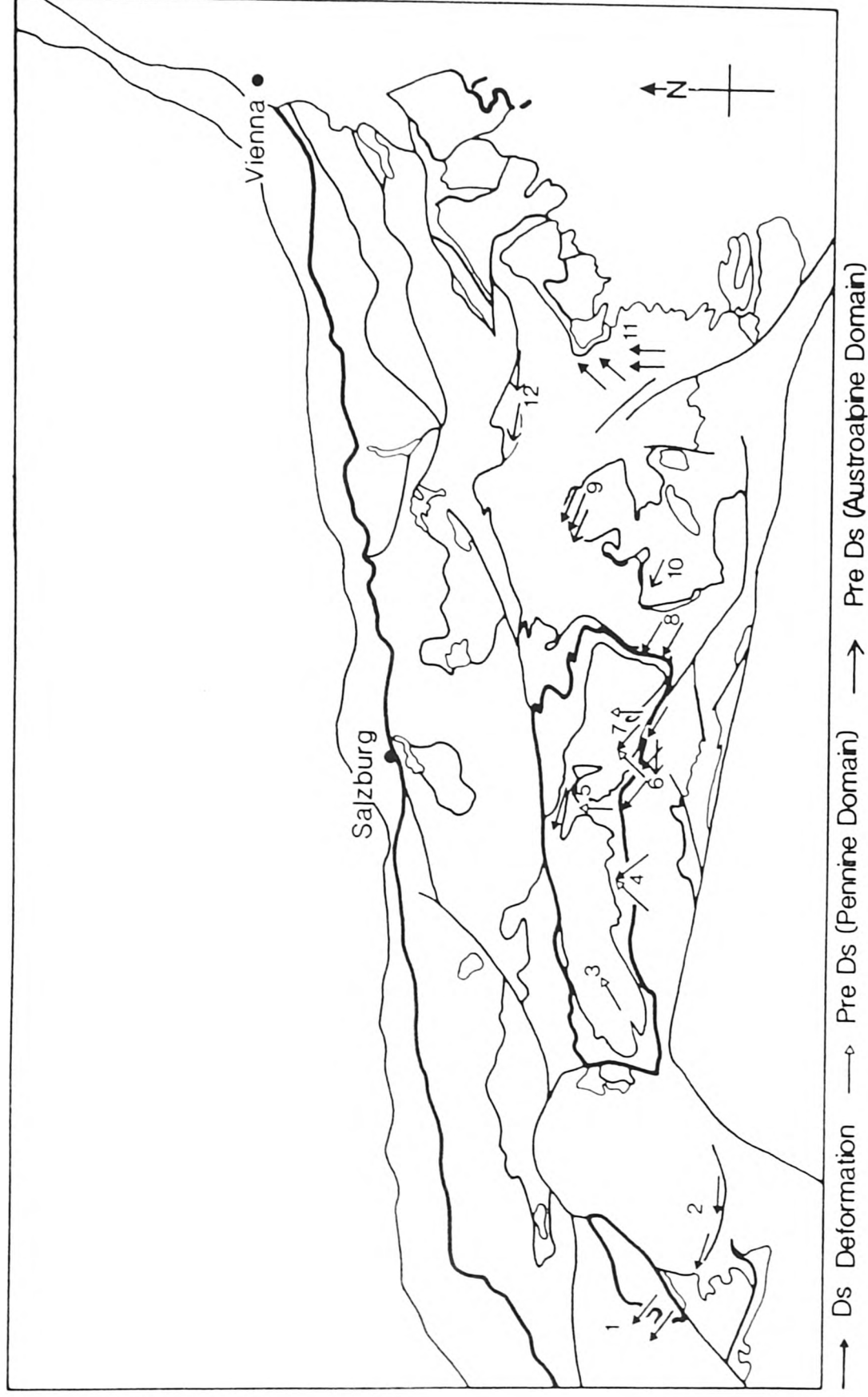


Fig.5.2. Kinematic data from the Eastern Alps for precollisional deformation. The main tectonic contacts are marked on for reference with fig.5.1. 1 = Masch, O' Shea, Koch (unpublished); 2 = Schmid & Haas (1987); 3 = Behrmann (1986), 4 = Zimmermann et al (1987), 5 = Frasl & Frank (1964), Clar (pers comm.), 6 = this work, Behrmann (1987b), Hoke (1987); 7 = this work, Exner & Prey (1964), Droop (1978); 8 = Hawkesworth (1974) - no sense of shear given; 9 = v. Gosen (1982) - stretching lineations, no sense of shear given; 10 = suggested movement from map-scale lateral cut-off features; Krohe (1986), Frank (1983); 12 = Ratschbacher & Oertel (1987), possibly related to Ds.

To the south of the Tauern Window at deeper structural levels, the first Alpine deformation took place in the ductile field. Stretching lineations and quartz c-axis data suggest an E-W stretching with a probable top to W sense of shear (chapter 4, Behrmann 1987b).

In conclusion, the rather sparse kinematic evidence associated with the early phase of tectonic thickening, suggests a general W to NW movement of the Austroalpine nappes over the Pennine units.

5.2.5 Tectonic setting during thickening

The thickening within the Pennine domain and its associated high P/T metamorphism can be explained as a predictable consequence of subducting relatively cool material at a destructive plate margin. However, the onset of convergence in the Eastern Alps is also expressed as a phase of shortening within the overlying Austroalpine domain. In plate tectonic reconstructions this domain represents the overriding plate. In many cases, however, the overriding plate above a downgoing slab undergoes extensional tectonics associated with the formation of a volcanic arc and back-arc basins (Karig 1971). This apparent anomaly has lead to a specific tectonic setting being proposed for the area.

Frisch (1983) suggests that the relatively new oceanic crust in the Tethyan Ocean was still warm and buoyant, which would make it difficult to subduct. This would imply that there was negligible slab-pull, which in many cases may be the main driving force for plate tectonics (Forsyth & Uyeda 1975). Therefore, the motive force for subduction has to come from the overall convergence between Africa and Europe, and to be transmitted through the Austroalpine crust. This would leave the Austroalpine domain in deviatoric compression and could explain the presence of compressional tectonics in the hanging wall of a subduction zone (enforced subduction of Frisch).

In the Eastern Alps convergence is expressed as a broad zone of shortening and thickening in both the Austroalpine and Pennine domains. This suggests the whole of the Eastern Alps can be viewed as an accretionary complex or orogenic wedge (Platt 1986) involving both the Pennine and Austroalpine units; and that there was no primary fundamental rheological contrast between the two domains as is implicit in nearly all previous tectonic reconstructions of the area.

5.3 CRUSTAL THINNING

5.3.1 Metamorphic evidence

The Cretaceous metamorphism in both the Pennine and the Austroalpine domains is the result of strong crustal thickening (see previous section). The high metamorphic pressures recorded in parts of the Pennine domain were produced by their burial in a subduction zone beneath the structurally higher Pennine units and the thickened Austroalpine domain. However, the depth of burial indicated by the metamorphic conditions in the Pennine units is far greater than the overburden that is preserved in the Eastern Alps. The overburden must, therefore, have undergone substantial post-metamorphic thinning.

The Pennine domain can be divided into two major structural units the Upper Schieferhülle (USH) and the Lower Schieferhülle (LSH) (Frisch 1980) both of which have undergone regional high P/T metamorphism. P_{max} estimates for the USH are $\approx 8\text{kbar}$ (Holland & Ray 1985, Selverstone & Spear 1985), and for the LSH $> 10\text{kbar}$ (Selverstone 1985). Along the contact between the two units there is a zone of very high P eclogites ($> 18\text{kbar}$, Holland 1979), which may be part of the LSH or form an isolated tectonic unit (Franz & Spear 1983). Assuming $\rho = 2.9\text{ gcm}^{-3}$ the metamorphic pressures indicate a depth of burial of $> 35\text{km}$ for the LSH, and $\approx 28\text{km}$ for the USH. The available overburden for the LSH consists of

the higher structural units of the Pennine and the full pile of Austroalpine nappes.

The Austroalpine domain in the Eastern Alps preserves an almost complete sedimentary sequence, developed from the onset of rifting through to the Tertiary. By analogy with the Western Alps, the metamorphism in the Pennine units of the Tauern Window began in the Early to Mid Cretaceous. This suggestion is supported by the K-Ar 78 ± 12 Ma age obtained on glaucophanes from the Central Tauern Window by Raith et al (1978).

The full structural thickness of the Austroalpine domain includes Upper Cretaceous and Tertiary sediments, which were deposited after the high P metamorphism in the Pennine had taken place, and therefore overestimates the potential mid Cretaceous overburden. Cross-sections of Tollmann (1977a), Clar (1965), and Kober (1955) all give similar thicknesses for the present day Austroalpine sheet around the Tauern Window of 10-12 km. This figure is considered too high by some workers (e.g. Roeder & Bögel 1978). The USH reaches a maximum thickness in the Glockner depression of 5 km (Cliff et al 1971). The total preserved overburden to the LSH is, therefore, 17 km (consisting of 5 km Pennine material and 12 km of Austroalpine nappes). This is 18 km too thin to account for the metamorphic pressures in the LSH. The maximum present-day overburden to the USH is also 17 km, which is ≈ 10 km too thin to account for the $P_{\max} = 8$ kbar metamorphic pressures.

Thickening of the Austroalpine domain produced a Cretaceous Barrovian facies metamorphism which locally reached amphibolite grade with a $P_{\max}$ of 6-7 kbar (fig. 5.3). Above the Oetzal basement nappe in the west of the Eastern Alps amphibolite grade conditions occur in Mesozoic sediments (Thöni 1986). These sediments overlie the Altkristallin and clearly, the associated metamorphic conditions underestimate the full structural thickness of the Austroalpine units.

Mid Cretaceous Metamorphism in the Austroalpine Nappes

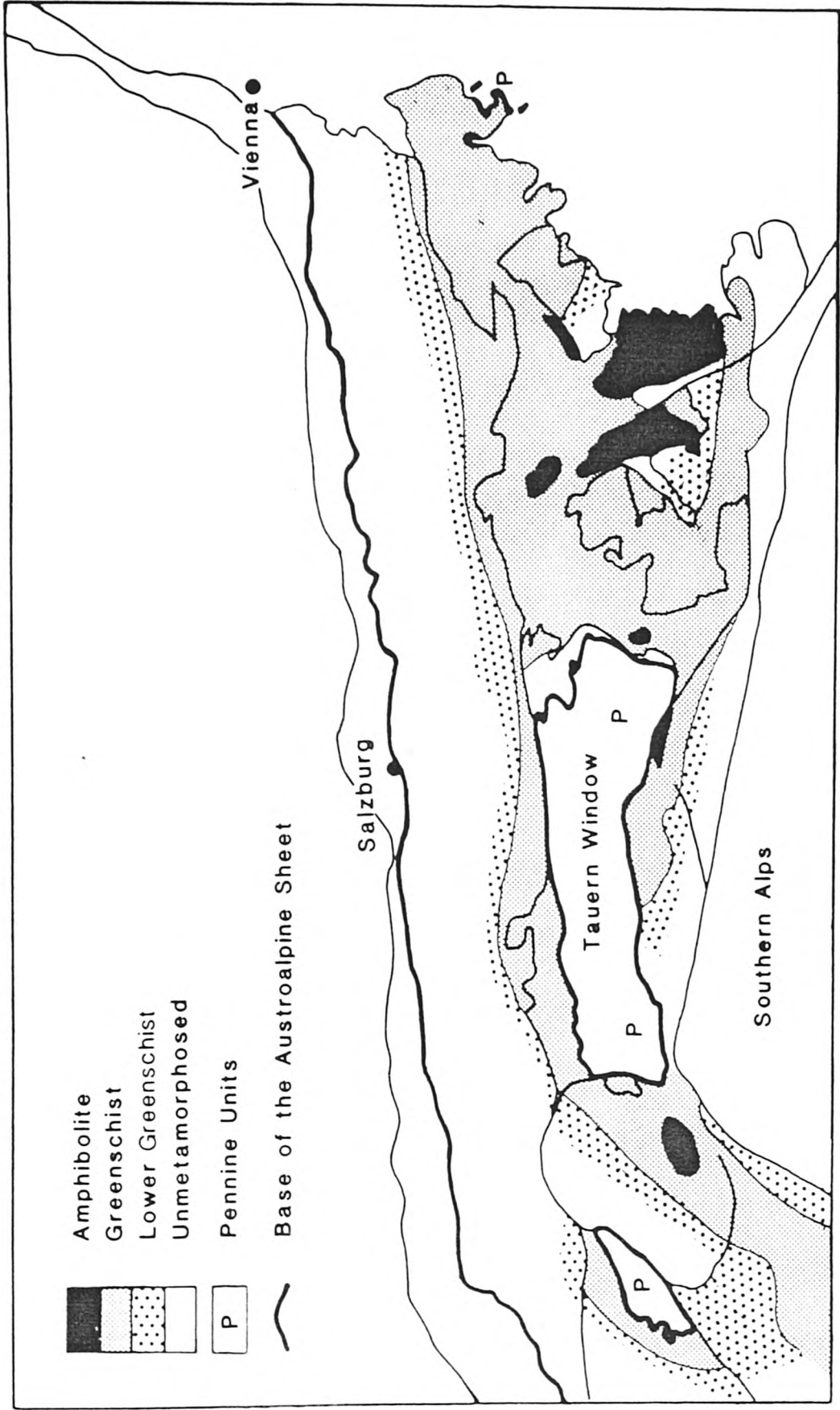


Fig.5.3. Mid-Cretaceous metamorphism in the Austroalpine domain (after Frank et al 1983). The main tectonic contacts are marked for reference with fig.5.1.

Taking $\rho = 2.8 \text{ gcm}^{-3}$ for the average density of mid-crustal rocks, P_{max} of 6–7kbar indicates a depth of burial to $\approx 22\text{km}$. The total structural pile is $\approx 12\text{km}$ thick and is, therefore, too thin by 10km, to account for the recorded metamorphic effects (fig.5.4).

Substantial post-metamorphic thinning of both the Austroalpine and Pennine units must have taken place in the Eastern Alps. The deficiency in thickness is greater for the LSH than the Austroalpine nappes, which suggests that part of the missing overburden has been removed from within the structural pile. Erosion from the contemporaneous topographic surface is, therefore, not a viable mechanism for thinning the sequence. The metamorphic pressure gradient in the Eastern Alps has been telescoped and oversteepened on a regional scale by tectonic thinning involving both the Pennine and Austroalpine units. This phenomenon has already been recognized by Selverstone (1985) and Platt (1986).

5.3.2 Structures related to thinning: Ds deformation

The metamorphic history indicates that substantial post-metamorphic thinning has taken place both within the Austroalpine nappes and the upper parts of the Pennine sequence. A detailed structural analysis carried out across the Austroalpine – Pennine boundary in the S.E. Tauern (chapter 3), showed this contact has been strongly deformed by a penetrative ductile deformation, Ds. This deformation forms a kilometre-scale zone of distributed shear, which can be traced from within the Altkristallin across the Mafrei Zone and into the structurally deeper Pennine units. Above this zone, at higher structural levels, Ds is represented by discrete shear zones (chapter 3). Kinematic and metamorphic studies in this area suggests that during Ds the Austroalpine sheets underwent a thinning of at least 10km (chapter 4). This minimum estimate of the local thinning of the

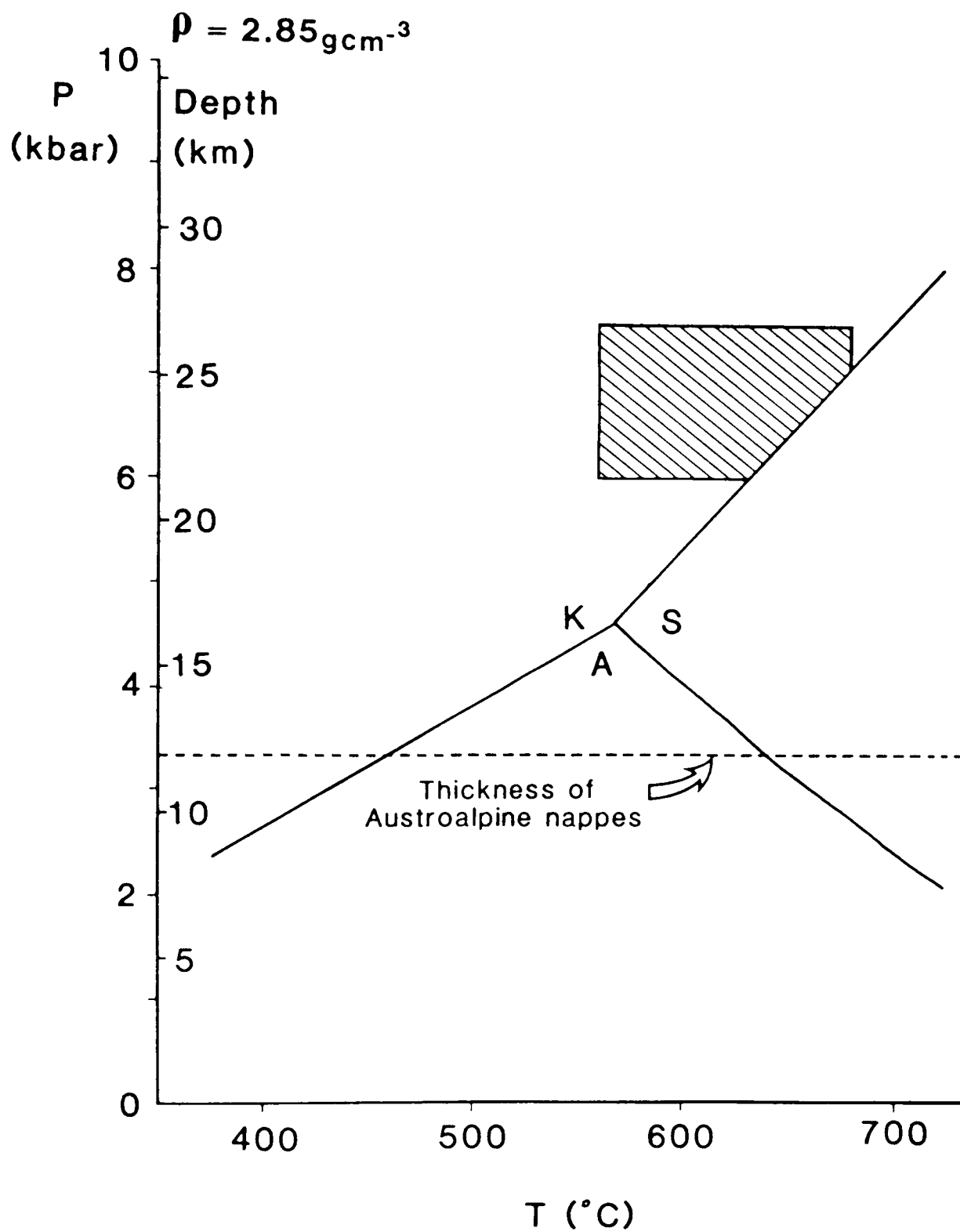


Fig.5.4. Estimated maximum metamorphic conditions at the base of the Austroalpine sheets during the Cretaceous metamorphism. Dashed line represents the thickness of the available overburden, allowing for erosion. The position of the Ky - And - Sill boundaries after Holland and Powell (1985).

Austroalpine domain is in very good agreement with the estimated shortfall in structural thickness required to explain the metamorphic conditions in the Austroalpine domain (see 5.3.1).

5.3.3 Regional correlation

To the S.E. of the Tauern Window, Ds is characterized by lower greenschist facies conditions associated with a top to NW sense of shear, which postdates the Cretaceous metamorphism in the Austroalpine domain. Correlating deformation phases over large distances is subject to considerable uncertainty. There is, however, a remarkable consistency about the direction and sense of shear associated with the first penetrative deformation found along the base of Altkristallin, where it overlies low-grade metasediments of the Pennine domain and the Matrei Zone, i.e. Silvretta basement, top to the NW (quartz c-axis diagrams; Masch, O'Shea, Koch, unpublished); Schober and Sadnig Groups, top to the NW (chapter 3), west of Tauern Window, NW-SE stretching lineations (Hawkesworth 1974). Deformation within all these areas is associated with lower greenschist facies metamorphic conditions and is directly comparable with Ds deformation found in the study area. Direct evidence for large-scale syn-convergent extension in these areas has not yet been documented.

At higher structural levels other major zones of deformation can also be correlated with Ds. In the Oetzal Alps, Schmid & Haas (1987) have investigated the Schlienig Line, which in part emplaces Altkristallin onto un-metamorphosed sediments, and in part is an intracrustal shear zone. Within the basement, deformation is associated with a top to the WNW sense of shear (Schmid & Haas 1987) and post dates an amphibolite grade metamorphism, which has associated cooling ages of 80-90Ma (Thöni 1986).

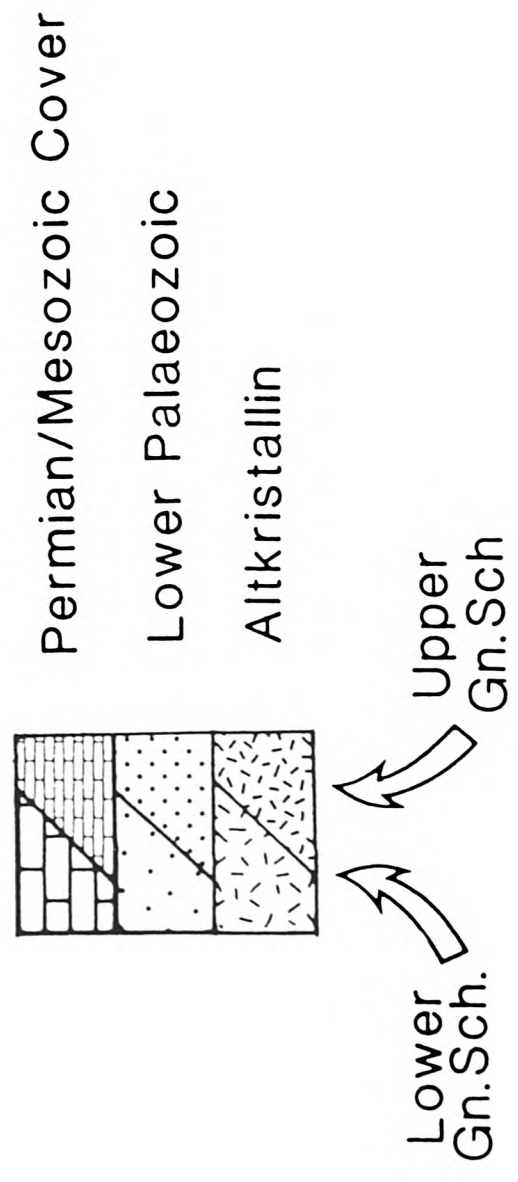
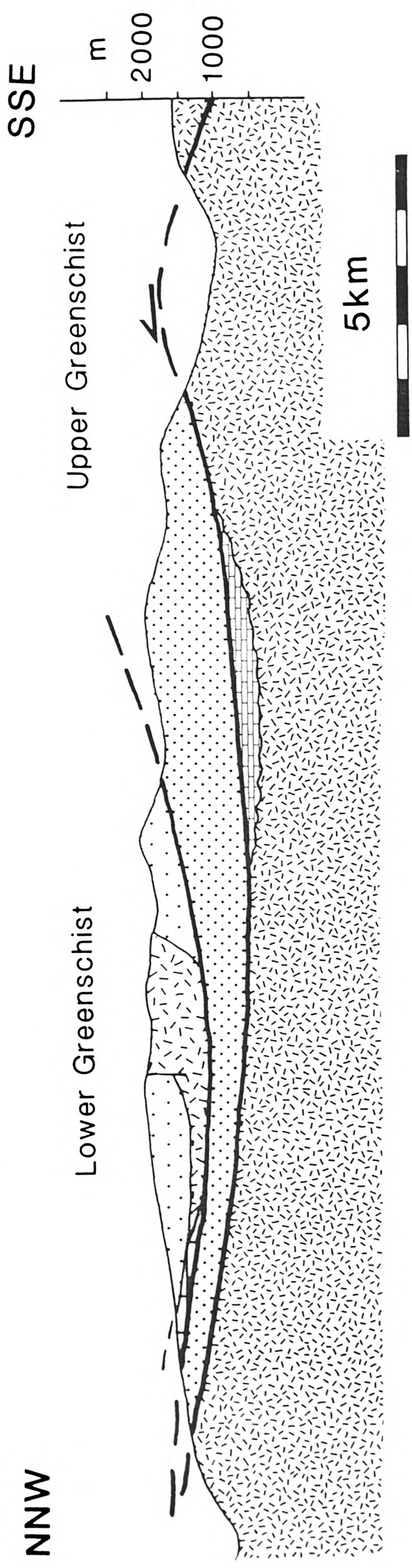
To the east of the Tauern Window, von Gosen et al (1985) have studied profiles through the Gurktaler Alps. The basal contact forms a zone several kilometres wide with penetrative deformation concentrated along a few discrete high-strain zones and is associated with WNW oriented stretching lineations (v. Gosen 1982). A cross-section (slightly oblique to the dominant WNW stretching direction) shows how one of these zones cuts down structure to the NW, through the Cretaceous metamorphic zonation (fig.5.5). Both timing and movement direction in this area are compatible with D_s and deformation is clearly extensional and acting to thin the metamorphic sequence in the Austroalpine sheet.

In the Koralpe region (fig.5.1) deformation is associated with a strong N/NE oriented stretching lineation (fig.5.2). Large parts of the area have been strongly affected by Alpine Mid Cretaceous metamorphism which attains a peak of amphibolite grade (Frank 1983). Syn- to post-metamorphic deformation is associated with stretching lineations which on the regional scale have a strongly curved trajectory; over a horizontal distance of 10km the orientation changes through 45° from N-S to NE-SW (Krohe 1986). The amphibolite grade Koralpe basement was emplaced onto Palaeozoic mica schists during the waning stages of metamorphism, but while the Koralpe unit was still hot enough to cause metamorphism beneath (Frank 1983).

Above the Koralpe basement, deformation is also associated with a top to N/NE movement but postdates metamorphism and is concentrated along narrow high-strain zones. These high-strain zones are flat-lying with sudden downward increases in metamorphic grade. This demonstrates that these tectonic contacts are low-angle normal faults which have excised part of the Cretaceous metamorphic sequence (Krohe 1986, Frank 1983).

The demonstrable existence of low-angle normal faults, is strong evidence in favour of extensional tectonics being involved in the

Fig.5.5.5. Cross-section through the Gurktaler Alps (after v.Gosen et al 1985). The sudden downward increase in metamorphic grade across the tectonic contact implies this is an extensional fault. See text for further discussion.



deformation and uplift of the area. To the S.W. of the Koralpe in the Saualpe, Neugebauer & Hammerschmidt (1986), report metamorphic evidence for post-metamorphic thinning of the Austroalpine structural pile, where amphibolite grade Cretaceous metamorphic rocks are separated from unmetamorphosed rocks by a structural thickness of only 3-5km. There is no indication of the associated movement direction.

There is, therefore, evidence throughout the Austroalpine domain for major syn-convergence deformation that overprints the peak of Cretaceous metamorphism, which locally is demonstrably associated with substantial post-metamorphic thinning. The associated transport direction is to NW in the west of the Eastern Alps and to the N/NE in the far east of the Eastern Alps.

At the highest structural levels within the Austroalpine domain, sedimentation was taking place at the same time as the postulated regional thinning. Further evidence for extension is preserved within the sedimentary record of the Mid-Cretaceous Gosau beds. These were deposited in fault bounded basins above the deformed and thickened Austroalpine nappes (Tollmann 1977a). Initially the sedimentary facies are shallow water brackish marine deposits. However, at $\Sigma 78$ Ma there is a sudden deepening of the sedimentary environment and sedimentation becomes dominated by calcareous turbidites (Tollmann 1977a). Some clay mineral rich horizons may have been deposited below the calcite compensation depth (Faupl & Sauer 1978). This very rapid deepening is seen over a wide area and can be related to the major phase of regional extension during Ds.

5.3.4 Timing of extension

To the south of the Tauern Window, radiometric dating suggests that the Austroalpine domain cooled from the Cretaceous Barrovian facies metamorphism around 80Ma, and that extensional tectonics began shortly

thereafter (see chapter 3). A similar age for deformation is suggested by radiometric dating along the Schlieren Line (Thöni 1986) and in the Koralpe region (Frank 1983, Krohe 1986), both of which can be correlated with Ds. A similar age for the onset of major extension is suggested by the dating of the Gosau sediments (see previous section).

The presence of detrital blue amphibole and lawsonite associated with clasts of serpentinite in Cenomanian to Turonian sediments (Winkler & Bernoulli 1986, Oberhauser 1968) shows that Pennine oceanic material had already been subducted, metamorphosed and uplifted by this time. Uplift of high P metamorphic rocks may be achieved by extensional tectonics (Platt 1987) and the presence of these deposits could be related to an early phase of extension. However, localized exhumation of high P rocks may also be achieved by other processes (Ernst 1984, Cloos 1982, Karig 1980).

The suggested correlation of Ds throughout the Eastern Alps implies that major extensional flow had begun by around 80Ma. This is long before collision took place and demonstrates that the extension was taking place during convergence. There is no clearly defined upper limit for the age of extension. Rb-Sr dating presented in this thesis suggests that Ds may have ceased around 40Ma to the south of the Tauern Window (see chapter 3), but there is some evidence for continued extension after the Tauern Metamorphism along the western margin of the Tauern Window (Behrmann 1987c, Selverstone 1988).

5.3.5 Extension and the consequences for estimates of orogenic shortening in the E. Alps.

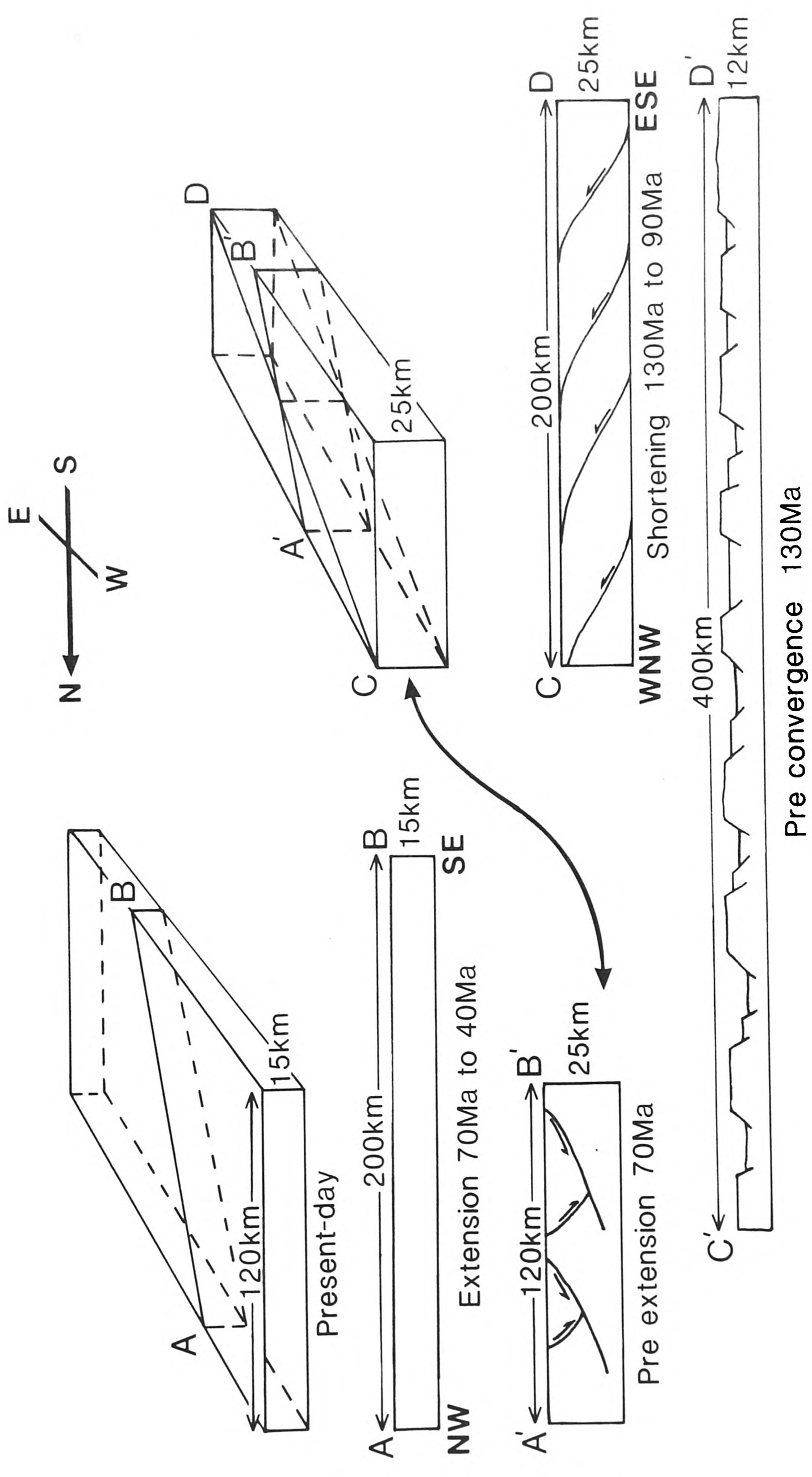
Around the Tauern Window the Austroalpine units have been thinned by about 10km during Ds, from an initial thickness of ≈ 25 km. Assuming

this is a representative value for the whole Austroalpine domain, implies a bulk extension of 66%. Kinematic indicators suggest that this extension took place in a NW direction (see chapters 3, 4). At present the Austroalpine sheet has an average thickness of 15km and a N-S dimension of $\Sigma 120$ km. Assuming these dimensions have not been significantly altered by erosion or late-stage deformation, the relative movement direction during deformation and the estimates of crustal thickness at various stages can be used to calculate the amounts of extension and shortening.

During Ds the Austroalpine domain was extended in a NW direction. To restore its present-day configuration to the original pre-Ds thickness of $\Sigma 25$ km, the Austroalpine domain has to be shortened by 80km in a NW direction. The pre-Ds NW dimension was, therefore, around 120km (fig.5.6). In the early stages of convergence the Austroalpine domain underwent strong shortening which produced a crustal thickening from around 12 to 25km (chapter 5), and kinematic indicators suggest this took place in a WNW direction. From the estimate of crustal thickness before and after shortening and using the pre-Ds dimensions of the Austroalpine domain given above, the original pre shortening WNW dimension can be calculated. This suggests that at the time of greatest extension the Austroalpine domain had a WNW dimension of 400km, and implies a shortening during convergence of 200km (fig.5.6). These calculations can only be regarded as a rough guide to the amounts of shortening and extension that have taken place within the Austroalpine domain. Once the detailed structure has been better documented, techniques of balancing cross-sections can be used to derive more precise estimates.

I have identified a number of low-angle extensional faults, which in contrast to thrusts restore to a shorter original section. If balanced cross-sections are to be drawn through the Eastern Alps they

Fig.5.6. Schematic block diagrams illustrating the evolution of the Austroalpine domain including the estimates for the amounts of shortening and extension.



must take into account both the nature of the tectonic contacts and changes in the tectonic transport direction. The existence of low-angle extensional faults and a dominant movement direction to the NW quadrant are both at variance with the conclusions of earlier workers in the Eastern Alps, and account for the large amounts of N-S shortening that have been suggested by some (e.g. Tollmann 1977a).

5.6 CONVERGENT TECTONICS AND PLATE MOVEMENTS IN THE EASTERN ALPS.

5.6.1 Tectonic synthesis: the Eastern Alpine orogenic wedge

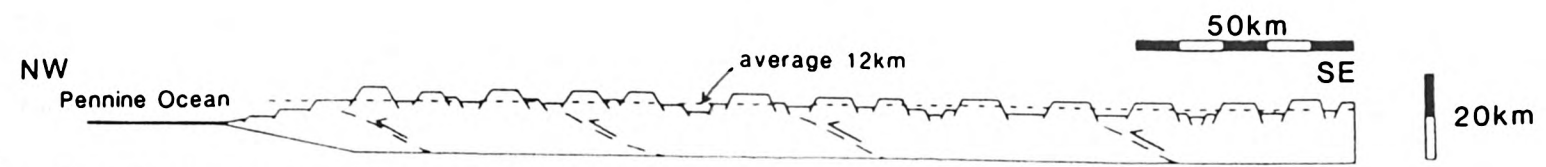
The dynamics of deforming thrust belts and accretionary prisms have been analysed using wedge-shaped continua with various specified rheologies (Chapple 1978, Elliot 1976, Davis et al 1983, Platt 1986). For each of these models there is a critical taper at which gravitational forces acting on the wedge are just able to overcome the frictional resistance to sliding along the base. In convergent tectonic settings, the slope necessary for forward motion of the thrust wedge is maintained by the compressive forces due to plate motion. Addition of material to the wedge will, in general, cause a change in geometry and to regain the critical taper the wedge has to undergo internal deformation. The change in geometry of the wedge is largely determined by the mode of accretion (Platt 1986). Frontal accretion will cause the wedge to lengthen and therefore reduce the overall taper. The wedge has to shorten internally in order to regain its stable geometry. Addition of material by underplating at depth (e.g. Byrne 1986) will thicken the wedge and increase the taper. To regain its stable taper, the wedge has to extend and thin. The regional structural development of the Austroalpine domain can be described in terms of such an orogenic system. This development is summarized in fig.5.7.

In the Eastern Alps convergence and subduction began in the Early Cretaceous (fig.5.7a.) and formed a broad shortened and thickened zone, which affected both the Austroalpine and Pennine domains (fig.5.7b.). The consequent burial metamorphism is characterized by Barrovian facies in the Austroalpine domain and by high P/T metamorphism in the Pennine domain. Kinematic indicators and estimates of the crustal thickness suggest that this was associated with 200km shortening in the Austroalpine domain in a WNW direction (fig.5.6). It was during this phase of shortening that the Austroalpine units were first thrust onto the Pennine domain and the Alpine orogenic wedge began to develop (fig.5.7b.).

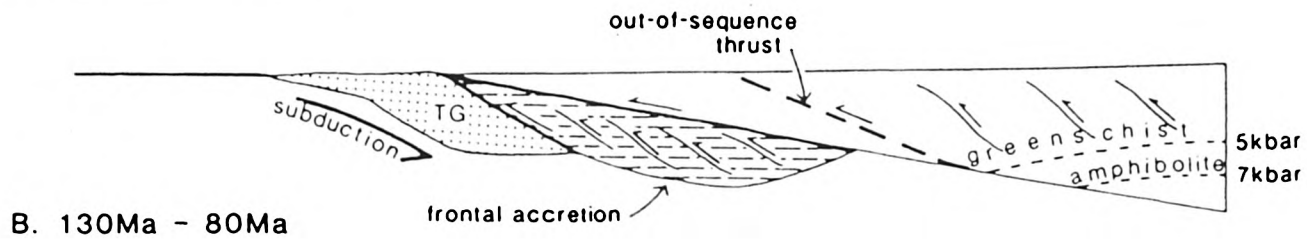
Radiometric dating suggests that the Austroalpine domain cooled at around 80Ma over a large area of the Altkristallin (chapter 3). To the south of the Tauern Window and in the Oetztal Alps there is evidence that this cooling took place relatively rapidly (Satir 1975, chapter 3). Along the southern margin of the Tauern Window the effects of the Cretaceous metamorphism are overprinted by an out of sequence thrust, which juxtaposes relatively high-grade metamorphic rocks of the Altkristallin against the low-grade Matrei Zone and Pennine units. The major discrepancy between the metamorphic grade of the Altkristallin and the underlying units, and the demonstrable existence of this structure over at least 30km in the study area suggests that this out of sequence thrust was associated with large displacements.

An out of sequence thrust may develop as one response to the mechanical requirement for an over-extended wedge to regain its stable taper (Platt 1988, Platt 1986). The out of sequence thrust seen at the base of the Altkristallin could, therefore, be interpreted as indicating that the dominant mode of accretion at that time was by frontal imbrication, which caused the orogenic wedge to become overextended and necessitated further shortening at the rear (fig.5.7c.). The

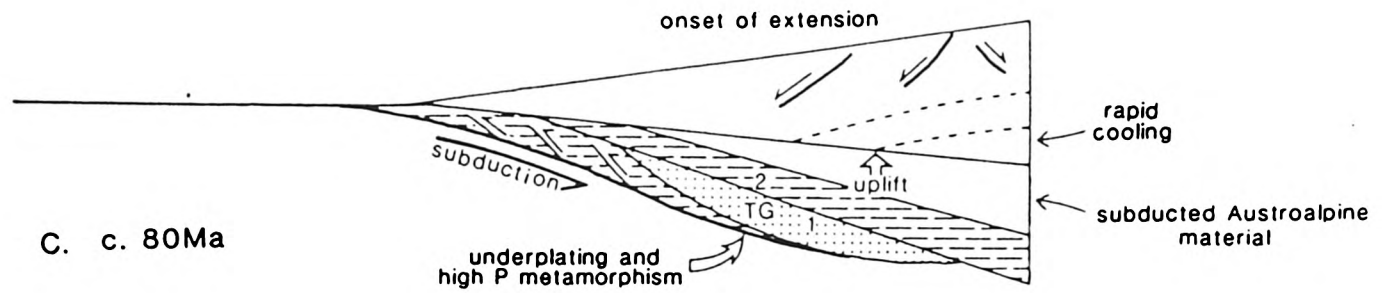
Fig.5.7. Illustrative cross-sections showing the formation and suggested evolution of the Eastern Alpine orogenic wedge, based on the estimates of shortening and extension given in the text. The sections have been drawn parallel to the dominant NW movement direction in the Austroalpine domain, and are area balanced. However, kinematic indicators suggest that substantial movement of material in the Pennine domain has taken place out of the plane of section. '1' marks the initial position of the 10kbar metamorphism recorded by Selverstone (1985) in the lower Pennine sheet, and '2' the position of 8kbar metamorphism (Holland & Ray 1985) in the lower Pennine sheet. The two tectonic units were juxtaposed during extension. The main expanse of amphibolite grade rocks at the rear of the wedge may be represented by the Koralpe region. Some spreading continues after collision. N.C.A. = Northern Calcareous Alps, T.G. = Tauern Gneiss.



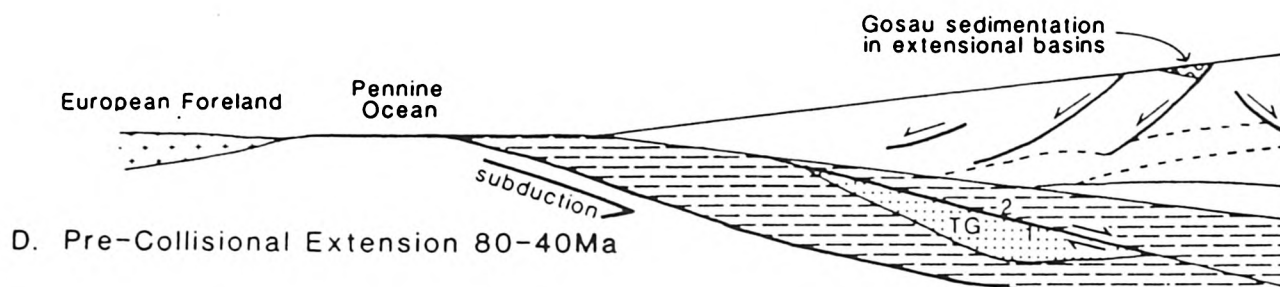
A. Pre-Convergence 130Ma



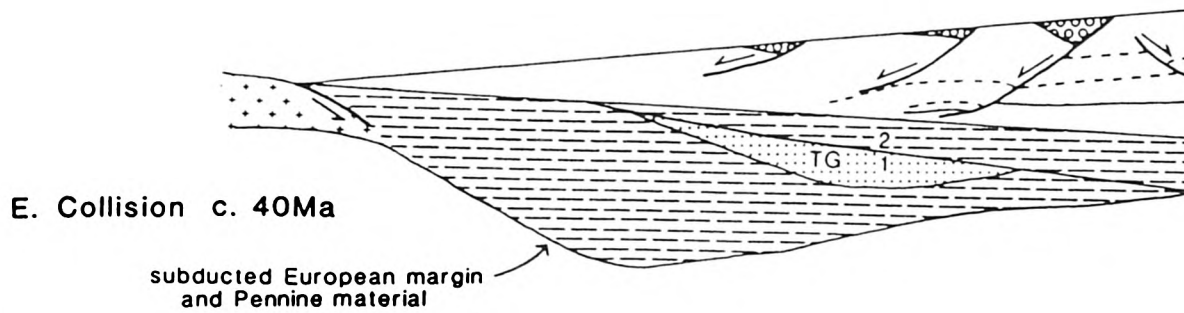
B. 130Ma - 80Ma



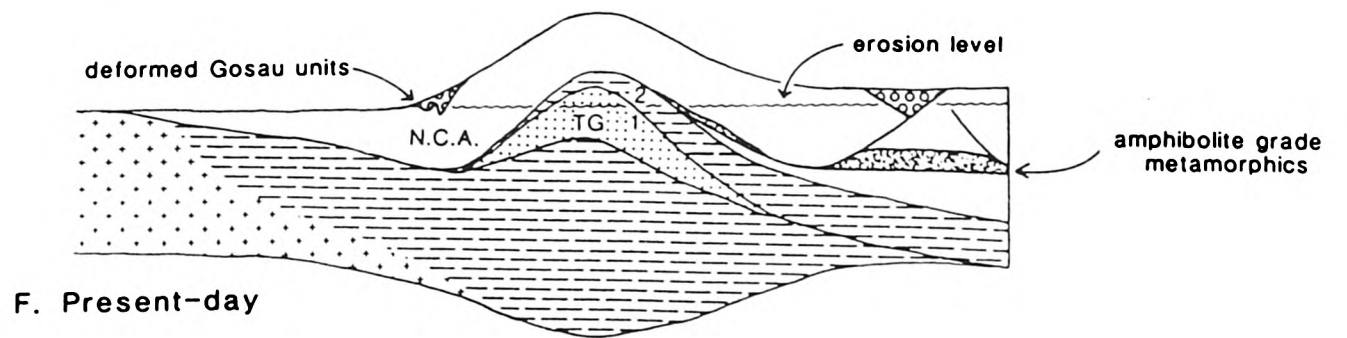
C. c. 80Ma



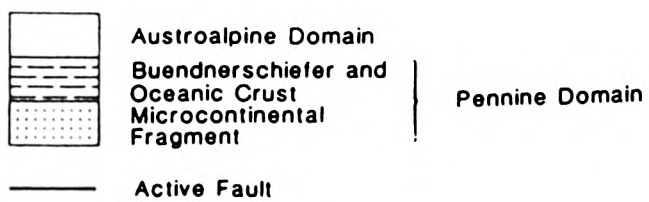
D. Pre-Collisional Extension 80-40Ma



E. Collision c. 40Ma



F. Present-day



development of an out of sequence thrust may also be triggered by collision of the destructive margin with a region of thick crust. The Tauern Gneisses exposed in the Tauern Window probably represent major fragments of continental crust which became incorporated within the Pennine ocean basin during rifting (Frisch 1979). If the involvement of these bodies in the subduction zone is related to the formation of the out of sequence thrust, their high P/T metamorphism should post date the formation of the thrust at around 80Ma (see above). There are insufficient radiometric age determinations to test this prediction.

A direct consequence of the development of the out of sequence thrust is that part of the Austroalpine domain must have been displaced and carried down the subduction zone. The phenomenon of subducted Austroalpine material is already well-documented from the Western Alps, although this probably occurred at an early stage in the convergent history (Hunziker 1974). In the Eastern Alps a similar situation may be represented in the Koralpe region, where Alpine amphibolite grade basement overlies a Variscan mica schist series largely unaffected by Alpine events. The contact between the two shows a reverse metamorphic gradient, which Frank (1983) interprets as a zone of contact metamorphism. There have been as yet no reports of high P/T Alpine metamorphism from this unit. However, Frank et al (1983) suggest that some of the eclogites within the Koralpe may be of Alpine age. The rapid cooling seen to the south of the Tauern Window (see 3.9.2) and in the Oetzal Alps (Satir 1975) may be related to the subduction of relatively cool low-grade material beneath the main Austroalpine edifice at this time.

Immediately post dating the thrusting along the southern margin of the Tauern Window, the Austroalpine domain begins to undergo extensional deformation, which caused a thinning of at least 10km (fig.5.7c,d,e.). Extension of a similar age is seen throughout the Austroalpine domain

(see above) and is represented by discrete low-angle extensional faults as well as broad zones of distributed shear. At higher structural levels extensional sedimentary basins began to develop, which are represented by the Gosau sedimentary units (fig.5.7d,e.). The mechanical analysis of orogenic wedges by Platt (1986) directly predicts the existence large-scale extension in a dominantly convergent tectonic setting, assuming the material has negligible long term yield strength and that underplating is a major mode of accreting new material to the wedge. Platt (1986) and Selverstone (1985) have both recognized the metamorphic evidence for post metamorphic thinning in the Eastern Alps, and Platt (1986) proposes that this can be explained by extension in a weak orogenic wedge, which is being continually thickened at depth. One of the main objections to this hypothesis is the long-held view that the Austroalpine domain can be treated as a largely rigid and undeformable plate, which overrode the much weaker sedimentary sequences of the Pennine units (Bickle & Hawkesworth 1978, Oxburgh 1972, Clar 1965, Roeder 1977). However, metamorphic and structural evidence clearly shows that the Austroalpine domain has undergone major internal deformation during convergence. The evidence presented in this thesis for regional extension taking place during convergence (see above) strongly suggests that gravity spreading was a major driving force for deformation in the Eastern Alps.

To maintain extension, the orogenic wedge has to be continually thickened at depth by underplating. One of the major controls on the mode of accretion is likely to be the ease of detachment. If subducted material is only weakly attached to the down-going slab this is likely to be accreted to the overlying plate before being subducted very far, e.g. the Bündnerschiefer overlying oceanic crust. If, however, there is strong mechanical attachment to the down-going slab (e.g the Tauern Gneiss bodies, which probably formed part of the ocean floor) the

material will be carried down to great depths before buoyancy forces are sufficient to cause detachment from the greater density oceanic slab, and underplating takes place. In the Eastern Alps regional high P/T metamorphism locally reached $P_{\max} \sim 10 \text{ kbar}$ $\Sigma 35 \text{ km}$ (Selverstone 1985) which represents burial to depths equivalent to normal crustal thickness and is not compatible with frontal imbrication. The timing of this underplating is uncertain since the only radiometric dating on high P minerals in the Eastern Alps is the K-Ar dating of glaucophanes, which gives ages of $78 \pm 12 \text{ Ma}$ (Raith et al 1978).

The metamorphic history gives one possible constraint on the age of the high P metamorphism in the Pennine domain of the Eastern Alps. Before 80 Ma the Barrovian facies metamorphism of the Austroalpine domain is incompatible with the high P/T metamorphism of the Pennine units, implying that the Austroalpine was not underlain by an extensive accretionary wedge of subducted Pennine material until a later stage. Fig. 5.7c. shows major underplating and high P metamorphism beginning at 80 Ma with the development of the out of sequence thrust. This hypothesis is compatible with the evidence that major extensional flow in the Eastern Alps did not begin until $\sim 80 \text{ Ma}$ (see 5.3.4) and makes the falsifiable prediction that the high P metamorphism in the Tauern Gneisses is younger than 80 Ma.

The stable taper of the viscous wedge proposed by Platt (1986) is maintained by the frictional resistance to sliding along the base. If subduction and underplating stopped, the wedge should continue to deform until it attained a zero relief. Collision and the end of subduction is marked by the onset of deformation in European foreland sediments, which occurred around $\Sigma 40 \text{ Ma}$ in Switzerland (Pfiffner & Milnes 1977). The evidence for continued westward extension after this time (Behrmann 1987c, Selverstone 1988, fig. 5.7f.) can be interpreted as reflecting

continued gravity spreading in an orogenic wedge with a very low long-term yield strength.

5.6.2 Constraints on the Adria - Europe convergence vector

From the Early Cretaceous the tectonic development of the Alpine chain can be described in terms of the convergence between two continental masses, Europe and Adria, with a subduction zone dipping beneath Adria (Hawkesworth et al 1978, Platt 1986, Frisch 1979, Gillet et al 1986). Across the Austroalpine - Pennine boundary in the Eastern Alps the pre-collisional convergent history can be divided into two major tectonic phases:

(i) crustal thickening, due to shortening within the Austroalpine and Pennine domains; and

(ii) crustal thinning, which caused substantial extension within both the Austroalpine and the upper parts of the Pennine domain.

Throughout the Alpine chain the dominant pre-collisional stretching direction is NW-SE. Some of these stretching lineations are associated with crustal shortening and are likely to be directly related to the movement of Adria over the Pennine domain. However, I suggest the majority are related to post metamorphic crustal extension caused by gravity spreading (see earlier). The extension direction in a body deforming under the action of gravity will be determined by the line of maximum surface slope (Chappell 1978). The geometry of a destructive margin is partially dependent on the subduction direction, but will also be affected by lateral variations in the volume and nature of the subducting material and irregularities in the margins of the convergent tectonic domains. The consistency of the pre-collisional stretching direction in the Austroalpine domain is strong evidence that the movement of the Austroalpine nappes and hence of the Adriatic platform was to the NW quadrant with respect to Europe throughout convergence. A

further constraint on the convergence vector between Europe and Adria is given by the concave to the south east curvature of the western Alpine arc. For compressional tectonics to have been active all around this sector of the arc requires convergence to have been to the NW quadrant. Radiometric dating of high P minerals suggests that subduction was taking place in this area from 100Ma onwards (e.g. Frey et al 1974).

There have been several attempts to correlate the movement directions in the Alps with the motions of Africa and Europe derived from palaeomagnetic and regional studies. The latest work by Hellman (in progress, Dewey et al 1988)) shows that the path of Africa with respect to Europe is dominantly N-S from the Early Cretaceous. There is no prolonged period of NW movement as predicted by some other reconstructions (Dewey et al 1973, Smith 1971). The kinematic data from the Alps, both E and W, indicates movement of the Austroalpine domain and hence Adria to the NW quadrant with respect to Europe throughout the Cretaceous. This movement direction is incompatible with the N motion of Africa with respect to Europe. In agreement with e.g Frisch (1979), I, therefore, consider Adria to have moved as an independent microplate from the Lower Cretaceous onwards.

The present configuration of the Alpine mountain belt is E-W, and both the distribution of sedimentary facies belts during the Mesozoic and the orientation of Cretaceous palaeocurrents (Gwinner 1978), shows that the European margin has remained approximately E-W since rifting in the Permian. If the Adriatic platform moved to the NW quadrant towards an E-W trending European margin, it would impinge on the European margin earlier in the E than in the W. The subduction of buoyant continental material in the East of the collisional zone would produce rapid preferential uplift in this region. Once the uplift has produced a sufficient surface slope, which is dependent on the bulk rheology of the orogenic wedge, extension into the orogenic foredeep will take place.

For the main part of the Alpine chain, this extension is constrained to the NW quadrant. The contemporaneous NE movement in the Koralpe may represent localized extension into the Carpathian flysch basin (fig.5.8), which was still a site of active sedimentation at this time (Savostin et al 1986).

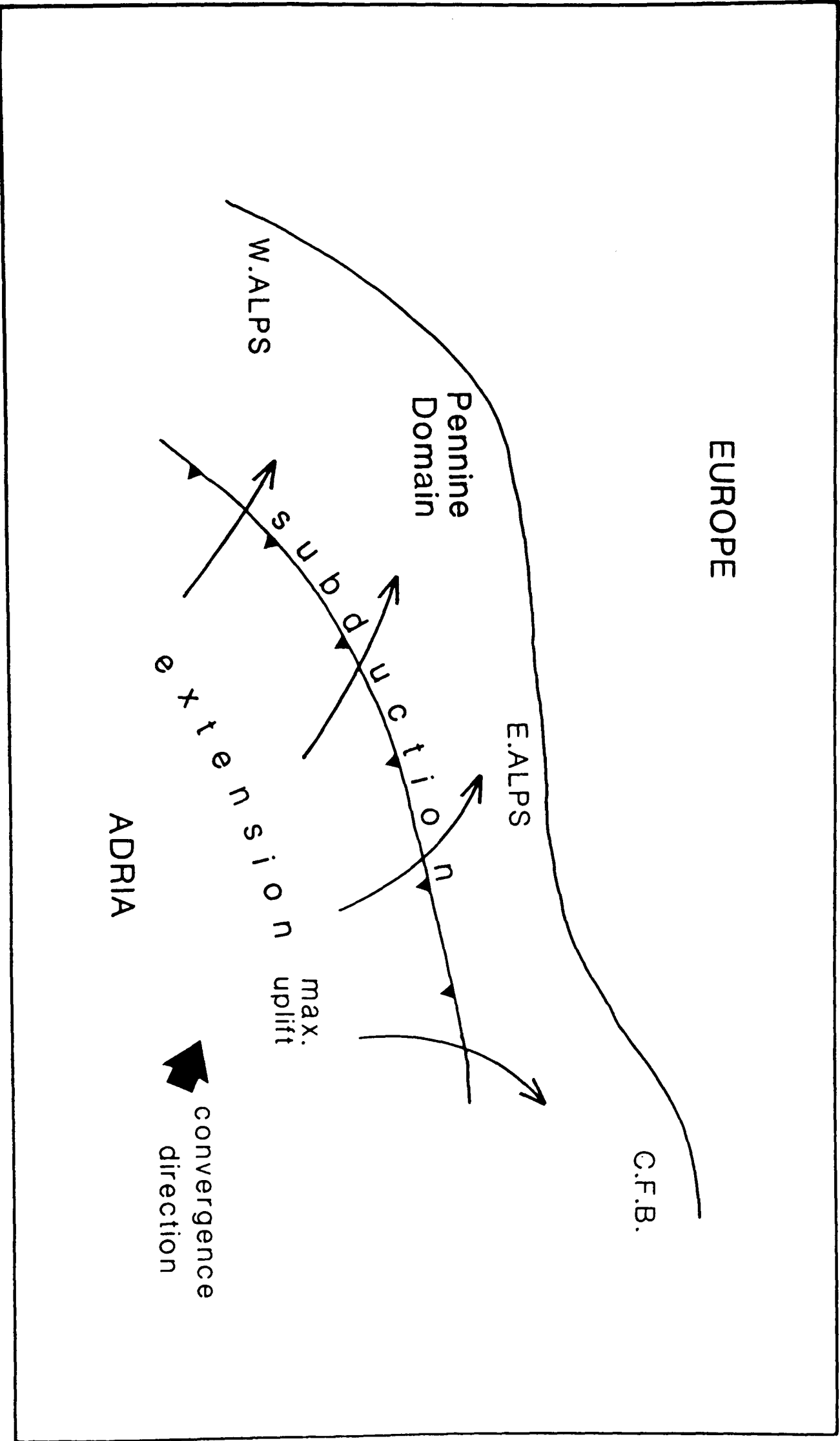


Fig.5.8. Suggested control on the extension direction during gravity spreading by the European continental margin. The region of maximum uplift corresponds to the Koralpe of the Eastern Alps (fig.5.1, 5.3). C.F.B. = Carpathian Flysch Basin.

CHAPTER 6 CONCLUSIONS

6.1 INTRODUCTION

The Austroalpine - Pennine boundary is a major tectonic contact in the Eastern Alps, which separates a dominantly continental Austroalpine domain from the Pennine oceanic units. In this thesis I have documented the structural and kinematic development of this boundary and attempted to set this in the wider context of major tectonic events in the Eastern Alps. The conclusions are divided into several thematic sections.

6.2 DISRUPTION IN THE MATREI ZONE

In the S.E. Tauern the Austroalpine - Pennine boundary is marked by a strongly disrupted zone known as the Matrei Zone, which shows lithological affinities to both the overlying and underlying domains. Transition analysis and the study of field relationships showed that most of the disruption within the Matrei Zone can be explained as a result of high-strain inhomogenous deformation. However, the presence of olistoliths does make a minor contribution to the discontinuous nature of the units within this zone. These sedimentary blocks can be identified by their fractured margins, commonly associated with neptunian dykes and veins, and the presence of finer-grained clastic material within the surrounding matrix. The general age of the matrix and the olistoliths themselves, suggests that their emplacement took place during the Jurassic extensional block faulting and not in a Cretaceous subduction zone trench.

Some postulated tectonic boundaries in the Pennine domain are recognized by their association with dolomite blocks (Tollmann 1977a).

The criteria established in the Matrei Zone can be used to determine whether these blocks have an olistolithic or tectonic origin.

6.3 PALAEOGEOGRAPHY

Structural sections through the area show that the Matrei Zone can be divided into two distinct units (fig.3.40, 3.42). The higher unit is dominantly composed of quartz phyllite associated with thin slivers of metamorphic basement. Bündnerschiefer lithologies are very rare. I interpret this unit as the stacked-up continental margin of the Austroalpine domain and assign it to the Lower Austroalpine tectonic sheet.

The base of the quartz phyllite sheet is a major tectonic discontinuity, which is locally decorated by serpentinite lenses. The structurally lower part of the Matrei Zone consists of a complex zone dominated by Bündnerschiefer lithologies. Slivers of serpentinite occur locally, which are associated with pelagic deposits, suggesting that this part of the Matrei Zone was originally floored by a serpentinitic oceanic crust. The good evidence for the oceanic affinities of some of the rocks in this unit suggests that this dominantly Bündnerschiefer unit represents part of the Pennine ocean basin. However, at deeper structural levels within the same tectonic unit, there are slivers of tectonically emplaced continental basement and associated cover sequence, which are overlain by Bündnerschiefer lithologies. The thinned fragments of continental material included at lower structural levels within this zone indicate that the floor of the Pennine ocean consisted of a heterogenous mixture of serpentinite and thinned fragments of the original continental crust.

The transitional material between the stacked-up continental margin and the oceanic Pennine domain is not preserved and has probably been subducted away.

6.4. NATURE OF THE AUSTRALPINE - PENNINE BOUNDARY

Within the Mafrei Zone the Austroalpine - Pennine boundary can be placed along two different contacts according to definition. One possible site for this boundary is the major tectonic discontinuity along the base of the quartz phyllite sheet in the Mafrei Zone, which separates units of continental Austroalpine affinity above from those of oceanic Pennine affinity below. However, the upper contact of the quartz phyllite sheets is another major tectonic boundary which separates amphibolite to upper greenschist Austroalpine basement from low-grade quartz phyllite. The quartz phyllite may have undergone low-grade high P/T metamorphism and its metamorphic history is more akin to the Pennine domain than to the overlying Austroalpine basement units. The quartz phyllite sheet locally thins out where the Austroalpine basement directly overlies the Pennine sediments and only one boundary can be defined.

The palaeogeographic boundary lies along the base of the quartz phyllite sheet but tectonically the upper contact of this unit is probably the most important since it coincides with a major discordance in the metamorphic history. This upper boundary represents a major out of sequence thrust which overprints the Cretaceous metamorphism in the Altkristallin and implies that the leading edge of the Austroalpine domain has been subducted away in this area. Radiometric dating suggests that the Altkristallin cooled rapidly at around 80Ma in this region, possibly as a result of the juxtaposition of the relatively cool material of the Pennine domain and the Mafrei Zone with the hotter

Altkristallin. The geometry of the imbricate stack requires a two phase development, the second of which may be associated with a top to the NE sense of shear. The geometry of the imbricate stack is complicated by being overprinted by a later phase of extensional flow, Ds.

The Austroalpine - Pennine boundary, therefore, lies within the Matrei Zone and it can be defined either on the basis of lithology or metamorphic history. Along the southern margin of the Tauern Window the Austroalpine - Pennine boundary is the result of a combination of brittle tectonics representing out of sequence thrusting and a subsequent phase of extensional deformation.

6.5 OROGENIC WEDGE MODEL FOR THE EASTERN ALPS

The insufficient overburden in the Eastern Alps to account for the high P low T metamorphism in the Pennine units has been attributed to extensional tectonics in the rear of a thickened and gravitationally unstable orogenic wedge (Platt 1986). In the Eastern Alps the major objection to this model has been the lack of evidence for substantial syn-convergence extension in the Austroalpine domain. To the south of the Tauern Window a kinematic analysis of deformation within the Austroalpine units indicates a reduction in thickness of $\approx 10\text{km}$ during ductile deformation (Ds). Radiometric dating suggests this deformation began around 80Ma and continued until collision at around 40Ma. This major tectonic phase can be correlated over a large part of the Eastern Alps, implying that the Eastern Alpine destructive margin underwent substantial thinning whilst convergence was still active. This is strong evidence in support of gravity spreading being a major driving force for deformation in the Eastern Alps and suggests that the tectonic development of the Eastern Alps can be described in terms of a weak

orogenic wedge that became gravitationally unstable at around 80Ma and deformed under its own weight.

The evidence for continued extension after collision suggests that the surface slope to the orogenic wedge was sufficient to continue to drive deformation after the push from the rear, due to plate motion, had ceased. This implies a low long-term yield strength for the orogenic wedge.

6.6 CONSTRAINTS ON PLATE MOTIONS

The regional kinematic data collated in this thesis can make a contribution to answering the vexed question of ancient plate motion in the Tethyan realm. The dominant sense of shear throughout convergence was to the NW quadrant. This can be divided into two distinct phases, an early episode of shortening in both the Austroalpine and Pennine domains and a later stage when the movement direction represents a combination of continued thickening by subduction of Pennine material and concomitant extension of the overlying material. The movement direction during crustal thickening is likely to be controlled by the plate convergence vector. However, the extension direction associated with gravity spreading will be determined by the line of maximum slope on the surface of the orogenic system. Although not directly related to plate convergence, gravity spreading is unlikely to take place at an angle greater than 90° to the convergence direction, suggesting that plate motion of the Adriatic microcontinent continued to the NW quadrant.

The proposed plate motion of Adria with respect to Europe is incompatible with the latest Africa - Europe motions in the Cretaceous, and suggests that Adria was an independent microplate throughout this time.

6.7. THE IMPLICATIONS FOR PLATE TECTONICS

The Austroalpine - Pennine boundary separates two very different palaeogeographic domains and represents the site of subduction at some time in the history of the Eastern Alps. The paradigm of plate tectonics suggests this could be thought of as a boundary between a rigid upper plate and more strongly deformed metasediments forming the cover to the down-going plate. In this thesis I have shown that the Austroalpine domain was first shortened by about 200km in a WNW direction, forming a region of thickened crust. This region then became gravitationally unstable and with continued thickening at depth by underplating, progressively extended under its own weight causing a regional 10km thinning of the Austroalpine domain. The leitmotiv is one of deformation not stability and this cannot, therefore, be described as a rigid plate.

Plate tectonic theory describes continent-scale displacements on the Earth's surface, but the kinematics need not be those of destructive margins. The primary cause of orogenesis is plate motion, however, in these regions of thickened crust body forces can also be a major driving force for deformation. The kinematic development of the Austroalpine domain in the Eastern Alps represents the interplay between compressive forces due to plate convergence and concomitant gravity-driven extensional flow.

APPENDIX ONE : STATISTICAL ANALYSIS

A1.1 TRANSITION ANALYSISA1.1.1 Theory

In complexly deformed areas, the sequence of transitions between different units can be tested for non-randomness. If the units are randomly disposed throughout the area, the frequency with which a particular transition occurs depends only on the number of times the relevant units are recorded. In an area divided into n different elements the number of transitions between the different elements can be expressed by the $n \times n$ transition matrix n_{ij} . The variable

$$\sum_{j=1}^n (n_{ij} - e_{ij})^2 / e_{ij}$$

has an approximate χ^2 distribution, with $v = (n-1)^2$ degrees of freedom (Hoel 1971). Where e_{ij} = the expected frequency of the transition from the i th to j th element; n_{ij} = the number of transitions recorded from the i th to j th elements. It is commonly not possible to distinguish transitions between two occurrences of the same element. If these transitions are ignored, then e_{ij} has to be replaced by $e'_{ij} = e_{ij} \sum n_{ij} / (\sum e_{ij} - \sum e_{ii})$ and the number of degrees of freedom is reduced by n to

$$v = (n-1)^2 - n \quad (\text{Naylor \& Woodcock 1977}).$$

The probability of a particular transition occurring is given by the probability matrix P , where

$$P = n_{ij} / \sum_{j=1}^{j=n} n_{ij} \quad (\text{Naylor \& Woodcock ibid})$$

A1.1.2 Application

The lithologies within the Matrei Zone of area 1 and 3 were grouped into several different lithological associations. The upper contact with the main Austroalpine basement edifice was not included in the analysis. The transitions were counted along N-S traverses at 100m intervals across the map areas, and the results compiled in the form of transition matrices.

The χ^2 value derived from these matrixes can be used to test the null hypothesis of randomness in the order of the units. The likelihood of particular transitions occurring was determined by constructing the probability matrix, P .

A1.1.3 Results - Area 3

Four main units were distinguished: Basement, quartz phyllite, Triassic carbonates (dolomite and marble), Bündnerschiefer. The most variable lithological group is the Bündnerschiefer. The reasons for grouping these diverse lithologies together is given in chapter 2. The very small patches of serpentinite were ignored. The expected frequencies of all the transitions are greater than 5.

Basement = A, quartz phyllite = B, Triassic carbonates = C,
Bündnerschiefer = D (fig.A1.1).

H_0 : the transitions from one unit to another are independent and their likelihood depends only on the number of times the elements are recorded.

H_1 : the transitions are not independent.

Under H_0 the test statistic χ^2 can be calculated:

$$\begin{aligned} \chi^2_0 = & (20 - 24.9)^2/24.9 + (52 - 40.2)^2/40.2 + (1 - 9.7)^2/9.7 \\ & + (20 - 36.8)^2/36.8 + (73 - 51.3)^2/51.3 + (0 - 12.4)^2/12.4 \\ & + (67 - 68.9)^2/68.9 + (72 - 59.2)^2/59.2 + (35 - 23.2)^2/23.2 \\ & + (20 - 17.5)^2/17.5 + (0 - 14.9)^2/14.9 + (24 - 24.3)^2/24.3 \end{aligned}$$

$$= \underline{65.6}$$

For $\nu=5$ there is a 0.01 probability that $\chi^2 > 15.09$. This transition matrix is, therefore, significant at the 1% level, i.e. reject H_0 .

A1.1.4 Results - Area 1

A similar analysis was carried out in area 1. In this area, however, serpentinite occurs as an important rock type and cannot be ignored. The expected frequencies of both the serpentinite and basement transitions were not great enough for the assumption that

$\sum_{j=1}^n (n_{ij} - e_{ij})^2 / e_{ij}$ has a χ^2 distribution. Throughout this area serpentinite occurs associated with fragments of continental crust, which I interpret as representing a heterogeneous ocean floor consisting of both serpentinite and continental fragments (chapter 2). This interpretation suggests that the serpentinite and basement elements can be grouped together. Unless this grouping is carried out, a χ^2 distribution of the test statistic cannot be assumed.

H_0 : the transitions from one unit to another are independent and their likelihood depends only on the number of times the elements are recorded.

H_1 : the transitions are not independent.

Under H_0 the test statistic χ_0^2 can be calculated:

$$\begin{aligned} \chi_0^2 = & (10 - 12.4)^2/12.4 + (5 - 12.1)^2/12.1 + (26 - 16.3)^2/16.3 \\ & + (8 - 13.7)^2/13.7 + (15 - 11.4)^2/11.4 + (16 - 15.6)^2/15.6 \\ & + (14 - 10.2)^2/10.2 + (8 - 8.8)^2/8.8 + (7 - 11.6)^2/11.6 \\ & + (21 - 19.7)^2/19.7 + (19 - 17)^2/19 + (16 - 16.4)^2/16.4 \\ & = \underline{17.54} \end{aligned}$$

For $\nu=5$ there is only a 0.01 probability that $\chi^2 > 15.09$. This

AREA 3

n_{ij}	A	B	C	D
A		35	0	1
B	24		73	52
C	0	72		20
D	20	67	20	
	44	174	93	73
				384

e'_{ij}

	23.2	12.4	9.7
24.3		51.3	40.2
14.9	59.2		24.9
17.5	68.9	36.8	

P_{ij}	A	B	C	D
A		0.97	0.0	0.03
B	0.16		0.49	0.35
C	0.0	0.78		0.22
D	0.19	0.63	0.19	

AREA 1

n_{ij}	A	B	C	D
A		10	5	26
B	8		15	16
C	14	8		7
D	21	19	16	
	43	37	36	49
				165

e'_{ij}

	12.4	12.1	16.3
13.7		11.4	15.6
10.2	8.8		11.6
19.7	17.0	16.4	

P_{ij}	A	B	C	D
A		0.24	0.12	0.63
B	0.21		0.39	0.41
C	0.48	0.28		0.24
D	0.38	0.34	0.29	

Matrices for the transition analysis. n_{ij} = measured transitions; e_{ij} = expected number of transitions if the sequence was random; P_{ij} = probability matrix representing the likelihood of a given transition occurring. A = basement, B = quartz phyllite and quartz schist, C = Triassic carbonate, D = Bündnerschiefer.

Fig.A1.1.

transition matrix is, therefore, significant at the 1% level, i.e. reject H_0 .

A1.1.5 Conclusions

In both areas the transition matrices are significantly non-random at the 1% level. The greater disruption of area 1 is reflected in the lower value for χ_0 and the lack of any well-defined preferred sequence. The statistical significance of the matrices from both areas indicates that parts of the original stratigraphic sequence are still preserved.

The probability matrix P (fig.A1.1) shows the likelihood of a particular transition occurring. In area 3 there are two clearly defined preferred sequences, which correspond to incompletely preserved stratigraphic sequences (see chapter 2 for further discussion). In both areas there is a relatively high probability of transitions within the proposed sequence as well as transitions to Bündnerschiefer. Together with the statistical significance of the transition matrices, this suggests the Matrei Zone in these two areas is not simply an array of discontinuous units embedded in a Bündnerschiefer matrix.

A1.2. K-Ar RADIOMETRIC AGE DATES

A1.2.1 Grouping the data

The area to the north of the Kreuzeck, Sadnig and Schober Groups was divided into two zones. One to the north, where Ds deformation was indicated by greenschist retrogression of the Cretaceous mineral assemblages, and another to the south where undisturbed Mr assemblages were still preserved. Only white mica K-Ar dates were used for this

comparison and any which were associated with excess argon were ignored (Brewer 1970).

DATA USED

Brewer (1970)

Ds Deformation		Mr Metamorphism	
L49	73±1	Z124	91±1
L11	65±1	Z125	88±1
G61	81±1	Z126	88±1
		Z127	85±2
		Z129	85±1
		Z132	83±1
		Z133	82±1
		Z101	84±2
		Z134	73±1
		L134	73±1

Oxburgh et al (1966)

PA2	86±2	W67	84±1
C613	91±3	PG	78±2
C614	66±2	C589	80±2
		C552	84±3
		PA1	86±2
		C578	80±2

Waters (1976)

SP203	66±2	0201	83±4
SP89	76±2	0204	94±4

Lambert (1970)

3045	78±2	3046	86±2
3043	76±2	3049	79±2
3042	80±2		

n=11

$$\bar{x} = 76.2$$

$$s^2 = 8.36$$

$$\tilde{\sigma}^2 = 7.98$$

n=20

$$\bar{x} = 83.2$$

$$s^2 = 5.23$$

$$\tilde{\sigma}^2 = 5.09$$

n= sample number, $\bar{x}$ = sample mean, s=sample variance, $\tilde{\sigma}$ =unbiased estimate for population variance.

Al.2.2 Significance testing

The two sets of data grouped in this way have different means and $\bar{x}$ from the zone dominated by Ds is lower than $\bar{x}$ of the rocks dominated Mr. The statistical significance of this difference can be tested (Student t distribution method for testing the significance of the difference of two means, see Hoel 1971 p.264).

For two samples X and Y with means $\bar{x}$, $\bar{y}$ and variances s_x , s_y which are taken from populations with means μ_x , μ_y ,

$$\text{the variable } \tau = \frac{(\bar{x} - \bar{y}) - (\mu_x - \mu_y)}{\sqrt{(\tilde{\sigma}_x^2/n_x + \tilde{\sigma}_y^2/n_y)}}$$

has an approximate Student t distribution with the degrees of freedom, v , given by:

$$v = \frac{(\tilde{\sigma}_x^2/n_x + \tilde{\sigma}_y^2/n_y)^2}{\frac{(\tilde{\sigma}_x^2/n_x)^2}{n_x + 1} + \frac{(\tilde{\sigma}_y^2/n_y)^2}{n_y + 1}}$$

H_0 : the population means of the two areas are the same i.e. $\mu_x = \mu_y$

H_1 : the ages associated with the zone of Mr are greater than those associated with the zone of Ds i.e. $\mu_x > \mu_y$.

If $\mu_x = \mu_y$ then for the samples given above,

$$\begin{aligned} \tau &= 7 / \sqrt{(0.73 + 0.25)} \\ &= \underline{7.14} \end{aligned}$$

$$v = \frac{(0.73 + 0.25)^2}{0.73^2/12 + 0.25^2/21} - 2$$

$$= \underline{18.27}$$

For $v=18$. there is only a 0.5% chance that τ exceeds 2.878, therefore reject H_0 . There is a greater than 99.5% probability that the mean radiometric age of the area dominated by Ds is less than the mean age from the area dominated by Mr.

MS88 is a sample of chloritic schist of the Bündnerschiefer.

Major elements (Wt. %)		R.E.E. (p.p.m)			
SiO ₂	30.8	Rb	19	La	51
TiO ₂	1.08	Cs	0.76	Ce	94
Al ₂ O ₃	22.8	Sr	17	Nd	45
Fe ₂ O ₃	7.81	Ba	207	Sm	5.4
MnO	0.03	Sc	10.9	Eu	0.56
MgO	25.2	Zr	16.0	Tb	0.81
CaO	0.42	Hf	5.5	Yb	1.00
Na ₂ O	0.56	Th	15.1	Lu	0.18
K ₂ O	0.53	U	7.7		
P ₂ O ₅	0.1	Ta	2.6		
L.O.I.	11.4				
	100.63				

Major elements measured using X.R.F, R.E.E measured using neutron diffraction, Technische Universität, München.



APPENDIX THREE: SAMPLE LOCALITIES MENTIONED IN TEXT

Grid references refer to individual map only.

Map Area 1:

B127 277232
 B125 095233
 B117 110210
 B114 095140
 B113 095216
 B101 065194
 B69 159219
 B61 178062
 B41 193117
 K3 128159
 K2 131161
 K1 128161

For the locations of samples B135, B129, B124, B110, B109, B90, B83, B69, B41, B35, B8, B3 and K5 see fig.4.9.

Mapping to the east of line 06 west, after Bickle (1973) and Behrmann & Wallis (1987).

Map Area 2:

SW16 052072
 SW13 069021
 SW11 055105

Map Area 3:

MS111 039127
 MS101 055045
 MS100 055058
 MS99 058182
 MS95 075113
 MS90 072191
 MS88 071183
 MS78 109130
 MS68 095185
 MS58 085107
 MS43 064191
 MS40 098151
 281 080184

D17 2100m on path from Fraganter Hütte to Duisburger Hütte.
 D11 (ctd. schist) 1½km SSE of the Hagener Hütte. See map by Droop (1978) for location of coarse ctd. schist band.
 D2 2350m on the path 1km to the SSW of Hagener Hütte.

For the locations of samples MS97, MS93, MS83, MS82, MS81, MS63, MS57, MS56, MS54, MS53, MS49, MS40, MS38, MS34, MS29, MS24, MS10, 512, 508, 483, 463, 433, 412, 228 and 173 see fig.4.11.

Mapping within 000150 - 040150, 040150 - 040200 after Gommeringer (1985). NE of 130190 reconnaissance mapping only.

REFERENCES CITED

(For an exhaustive listing of German and Austrian literature on the Eastern Alps, see Tollmann 1977a, 1986.)

Ampferer, O. 1906. Ueber das Bewegungsbild von Faltengebirgen. Jb. Geol. R.-A. 56, 539-622.

Angel, F. 1928. Gesteinskundliche und geologische Beiträge zur Kenntnis der Schobergruppe in Osttirol. Verh. Geol. B.-A. 153-182.

Angel, F. 1930. Ueber Plagioklasfüllungen und ihre genetische Bedeutung. Mitt. Nat-wiss. Ver. f. Steiermark 67, 36.

Baird, A.W. & Dewey, J.F. 1986. Structural evolution in thrust belts and relative plate motion: the upper Penninic Piedmont zone of the internal Alps, southwest Switzerland and northwest Italy. Tectonics 5, 375-387.

Barber, A.J., Tjokrosapoetro, S. & Charlton, T.R. 1986. Mud volcanoes, shale diapirism, wrench faults and mélanges in accretionary complexes, Eastern Indonesia. A.A.P.G. Bull. 70, 1729-1741.

Bechstädt, T. 1978. Faziesanalyse permotriadischer Sedimente des Drauzuges als Hinweis auf eine großräumige Lateralverschiebung des Ostalpins. Jahrb. Geol. B.-A. 121, 1-121.

Becke, F. 1909. Ueber Diaphthorite. Tschm. Min. Pet. Mitt. 28, 369-375.

Behrmann, J.H. 1986. Abstracts from the Tübinger meeting. Symposium für Strukturgeologie, Kristallingeologie und Tektonik, Tübingen.

Behrmann, J.H. 1987a. Abstracts to the Tauern Tagung. Uniakuell - Zeitschrift der Universität Salzburg. 10, 9.

Behrmann, J.H. 1987b. A precautionary note on shear bands as kinematic indicators. J.S.G. 6, 659-666.

Behrmann, J.H. 1987c. Large-scale subhorizontal extension in a convergent orogen: The Sterzing-Steinach mylonite zone in the Eastern Alps. Abstracts to the E.U.G. meeting, Strasbourg.

- Behrmann, J.H. & Platt, J.P. 1982. Sense of nappe emplacement from quartz c-axis fabrics; an example from the Betic Cordilleras (Spain). *E.P.S.L.* 59, 208-215.
- Behrmann, J.H. & Wallis, S.R. 1987. Hangendverschuppung des Tauernfenster-Südrandes bei Kals (Osttirol) als Zeuge von eo-alpinem Underplating. *Jb. Geol. B.-A.* 130, 133-138.
- Bell, T.H. 1978. Progressive deformation and the reorientation of fold axes in a ductile mylonite zone: the Woodroffe thrust. *Tectonophysics* 44, 285-321.
- Bernoulli, D. & Jenkyns, H.C. 1974. Alpine, Mediterranean and Central Atlantic Mesozoic facies in relation to the early evolution of the Tethys. In: *Modern and ancient geosynclinal sedimentation* (edited by Dott, D.H. & Shaver, R.). *SEPM Special Publication* 19, 129-160.
- Bickle, M.J. 1973. Studies across the Mafrei Zone. Unpublished D.Phil. Thesis, Oxford.
- Bickle, M.J. & Hawkesworth, C. 1978. Deformation phases and the tectonic history of the Eastern Alps. *Geol. Soc. Am. Bull.* 89, 293-306.
- Bickle, M.J. & Nisbet, E. 1972. The ocean affinities of some Alpine mafic rocks based on their Ti-Zr-Y contents. *Geol. Soc. Lond. Jour.* 128, 267-271.
- Bickle, M.J. & Powell R. 1977. Calcite-dolomite geothermometry for iron-bearing carbonates. *Contrib. Min. Pet.* 59, 281-292.
- Bobyarchick, A.R. 1986. The eigenvalues of steady flow in Mohr space. *Tectonophysics*, 122, 35-51.
- Bouchez, J.L. & Pecher, A. 1981. The Himalayan main central thrust pile and its quartz-rich tectonites in central Nepal. *Tectonophysics*, 78, 23-50.
- Braumüller, E. 1939. Der Nordrand des Tauernfensters zwischen dem Fuscher und dem Rauristal. *Mitt. Geol. Gesell.* 30, 37-150.

- Brewer, M.S. 1969. Excess argon in metamorphic micas from the Eastern Alps, Austria. *E.P.S.L.* 6, 321-331.
- Brewer, M.S. 1970. Geochronological studies in the Altkristallin; Goldeck- and Kreuzeck Groups. Unpublished D.Phil thesis. Oxford.
- Brunel, M. & Geyssant, J. 1978. Mise en évidence d'une déformation rotationnelle est-ouest dans la fenêtre des Tauern. *Revue Géogr. Phys. Géol. Dyn.* 20, 335-346.
- Butler, R.W.H. 1982. The terminology of structures in thrust belts. *J.S.G.* 4, 239-245.
- Byrne, T. 1986. Eocene underplating along the Kodiak shelf Alaska: implications and regional correlation. *Tectonics*, 5, 403-421.
- Carreras, J., Estrada, A. & White, S.H. 1977. The effects of folding on the c-axis fabrics of a quartz mylonite. *Tectonophysics*, 39, 3-24.
- Casa, J.M. 1986. Shear bands and related extensional structures in a mylonitized quartz dyke. *J.S.G.* 8, 693-699.
- Channel, J.E.T., D'Agénio, B. & Hovarth, F. 1979. Adria the African promontory. In: *Mesozoic Mediterranean palaeogeography*. *Earth Sci. Rev.* 15, 213-292.
- Chapple, W.M. 1978. Mechanics of thin-skinned fold and thrust belts. *Geol. Soc. Am. Bull.* 89, 1189-1198.
- Clar, E. 1940. Von der Tⁿrtaler Breccie (Lizum). *Sitz. Ber. Akad. Wiss. in Wien, Mathem.-naturw. Klasse, Abteilung I*, 149, 71-84.
- Clar, E. 1954. Ein zweikreisiger Geologen- und Bergmannskompaß zur Messung von Flächen und Linearen. *Verh. Geol. B.-A. Hft.4*, 201-215.
- Clar, E. 1965. Zum Bewegungsbild des Gebirgsbaues der Ostalpen. *Verh. Geol. B.-A. Sonderheft G*, 11-35.
- Cliff, R.A., Droop, G.T.R. & Rex, D.C. 1985. Alpine metamorphism in the south east Tauern Window, Austria: rates of heating, cooling and uplift. *J. Met. Geol.* 3, 403-415.

- Cliff, R.A., Norris, R., Oxburgh, E.R. & Wright, R.C. 1971. Structural, metamorphic and geochronological studies in the Reifbeck and Ankogel Groups. *Jahrb. Geol. B-A.* 114, 121-272.
- Cloos, M. 1982. Flow mélanges: numerical modelling and geologic constraints on their origin in the Franciscan subduction complex, California. *Bull. Geol. Soc. Am.* 93, 330-345.
- Cobbold, P.R. 1977. Description and origin of banded deformation structures - I Regional strain, local perturbation and deformation bands. *Can. J. Earth Sci.* 14, 1721-1731.
- Cornelius, H.P. & Clar, E. 1939. Geologie des Großglocknergebietes (I Teil). *Abh. Zweigst. Wien der Reichsanst. für Bodenforschung*, 25, 1-305.
- Davis, D., Suppe, J. & Dahlen, F.A. 1983. Mechanics of fold and thrust belts and accretionary wedges. *J. Geoph. Research*, 88, 1153-1172.
- Deutsch, A. 1983. Datierungen an Alkaliamphibolen und Stilpnomelan aus der südlichen Platta-Decke (Graubünden). *Ecl. Geol. Helv.* 76, 295-308.
- Dewey, J.F., Pitman, W.C., Ryan, W.B.F. & Bonnin, J. 1973. Plate tectonics and the evolution of the Alpine system. *Geol. Soc. Am. Bull.* 84, 3137-3180.
- Dewey, J.F., Hellman, M.L., Turco, E., Hutton, D.H.W, Knott, S.D. 1988. Kinematics of the western Mediterranean. *Geol. Soc. Lond Spec. Pub. Alpine tectonics* (in press).
- Dietrich, V.J. & Franz, U. 1976. Alpidische Gebirgsbildung in den Ostalpen: ein plattentektonisches Modell (Kurzfassung). *Geol. Rndsch.* 65, 361-374.
- Dodson, M.H. 1973. Closure temperature in cooling geochronological and petrological systems. *Contrib. Min. Pet.* 40, 259-274.

- Dodson, M.H. 1979. Theory of cooling ages. In: lectures in isotope geology (edited by Jäger, E. & Hunziker, J.C.). Springer-Verlag, Berlin, 207-214.
- Droop, G.T.R. 1978. Metamorphic studies in the S.E. Tauern. Unpublished D.Phil. Thesis, Oxford.
- Droop, G.T.R. 1981. Alpine metamorphism of pelitic schists in the south-east Tauern Window, Austria. Schweiz. Min. Pet. Mitt. 61, 237-273.
- Droop, G.T.R. & Cliff, R.A. 1985. Alpine metamorphism in the S.E. Tauern Window, Austria. 1. P-T variations in space and time. J. Met. Geol. 3, 371-402.
- Elliot, D. 1976. The motion of thrust sheets. J. Geoph. Res. 81, 949-963.
- England, P.C. 1987. Diffuse continental deformation: length scales, rates and metamorphic evolution. Phil. Trans. Roy. Soc. Lond. A321, 3-22.
- England, P.C. & Holland, T.J.B. 1979. Archimedes and the Tauern eclogites: the role of buoyancy forces in the preservation of exotic tectonic blocks. E.P.S.L. 44, 287-294.
- Ernst, W.G. 1973. Interpretative synthesis of metamorphism in the Alps. Bull. Geol. Soc. Am. 84, 2053-2078.
- Ernst, W.G. 1984. Californian blueschists, subduction and the significance of tectonostratigraphic terranes. Geology 12, 436-441.
- Exner, C. 1962. Sonnblicklamelle und Mölltallinie. Jb. Geol. B-A. 105, 273-286.
- Exner, C. & Prey, S. 1964. Erläuterung zur geologischen Karte der Sonnblickgruppe 1:50,000. Geol. B-A., pp.170.

- Faupl, P. & Sauer, R. 1978. Zur Genese roter Pelite in Turbiditen der Flyschgosau in den Ostalpen (Oberkreide - Alttertiär). Neues Jahrb. Geol. Pal. Monatsh. 65-86.
- Ferguson, C.C. 1979. Rotation of elongate rigid particles in slow non-Newtonian flows. Tectonophysics, 60, 247-262.
- Flügel, H. 1963. Das Paläozoikum in Oesterreich. Mitt. Geol. Gesell. 56, 401-443.
- Forsyth, D. & Uyeda, S. 1975. On the relative importance of the driving forces of plate motion. Geophysics. J. 43, 163-200.
- Frank, W. 1983. Argumente für ein neues Entwicklungsmodell des Ostalpins. Jber. Hochschulschwerpunkt, 4, 249-262.
- Frank, W., Esterlus, M., Jung, G., Krohe, A. & Weber, J. 1983. Die Entwicklungsgeschichte von Stub- und Koralpen-Kristallin, und die Beziehung zum Grazer Paläozoikum. Jahresbericht. 1982. Hochschulschwerpunkt 15, 263-293.
- Franz, G. & Spear, F. 1983. High pressure metamorphism of siliceous dolomites from the central Tauern Window, Austria. Am. J. Sci. 283, 396-413.
- Frasl, G. 1953. Zum Auftreten von Füllungsmineralien in Kalifelspaten und ihre Verbreitung in Oesterreichischen Graniten und Granitgneisen. In: Skizze zum Antlitz der Erde, (edited by Hollinek), 255-272.
- Frasl, G. 1958. Zur Seriengliederung in den mittleren Hohen Tauern. Jb. Geol. B.-A. 101, 323-472.
- Frasl, G. & Frank, W. 1964. Exkursion 1/2, Mittlere Hohe Tauern. Mitt. Geol. Gesell. Wien. 57, p.17-31.
- Frey, I. 1984. Gefügeuntersuchungen und petrographische Untersuchungen im Stainzer Plattengneiss. Unpublished Ph.D. Thesis, Vienna.

- Frey, M., Hunziker, J.C., Frank, W., Bocquet, J., Dal Piaz, D., Jäger, E. & Niggli, E. 1974. Alpine metamorphism of the Alps: a review. Schweiz. Min. Pet. Mitt. 54, 247-289.
- Frisch, W. 1979. Tectonic progradation and plate tectonic evolution of the Alps. Tectonophysics, 60, 121-139.
- Frisch, W. 1980. Tectonics of the western Tauern Window. Mitt. Oesterr. Geol. Gesell. 71/72, 65-71.
- Frisch, W. 1983. Plattendrift, Orogenese und Mantelkonvektion. Deutsche Forschungsgemeinschaft, Rundgespräch ' Plattentektonik und Schwerefeldstruktur', 1-28 (als Ms. vervielfältigt).
- Frisch, W. 1984. Sedimentological response to late Mesozoic subduction in the Penninic windows of the E. Alps. Geol. Rndsch, 73, 33-45.
- Frisch, W., Gommeringer, K., Kelm, U. & Popp, F. 1987. The upper Bündnerschiefer of the Tauern Window - a key to understanding Eo-alpine orogenic processes in the Eastern Alps. In: geodynamics of the Eastern Alps (edited by Flügel, H.W. & Faupl, P.), 55-69.
- Garcia-Celma, A. 1983. C-axis and shape fabrics in quartz mylonites of Cap de Creus (Spain); their properties and development. Unpublished Ph.D. thesis, Utrecht.
- Ghosh, S.K. 1987. Measure of non-coaxiality. J.S.G. 9, 111-113.
- Ghosh, S.K. & Ramberg, H. 1976. Reorientation of inclusions by combination of pure and simple shear. Tectonophysics, 34, 1-70.
- Ghosh, S.K. & Sengupta, S. 1973. Compression and simple shear of models with rigid and deformable inclusions. Tectonophysics, 17, 133-175.
- Gillet, P., Choukroune, P., Ballèvre, M. & Davy, P. 1986. Thickening history of the Western Alps. E.P.S.L. 78, 44-52.
- Gommeringer, K. 1985. Die Matreier Zone zwischen Mohar, Makernisitze und Ofenspitze in der Sadniggruppe (Kärnten). Unpublished Diplomarbeit der Universität Tübingen, pp.143.

- Gwinner, M.P. 1978. Geologie der Alpen. 2te Auflage. Schweizerbart, Stuttgart, pp.480.
- Hawkesworth, C.J. 1974. Geochemical studies in and around S.E. corner of the Tauernfenster. Unpublished D.Phil. thesis, Oxford.
- Hawkesworth, C.J. 1976. Rb/Sr geochronology in the Eastern Alps. Contrib. Min. Pet. 54, 225-244.
- Hawkesworth, C.J., Waters, D.J. & Bickle, M.J. 1975. Plate tectonics in the Eastern Alps. E.P.S.L. 24, 405-413.
- Höck, V. 1974. Lawsonit Pseudomorphosen in den Knotenschiefer der Glockner Gruppe (Salzburg - Kärnten, Oesterreich). Der Karinthian, 71, 110-119.
- Höck, V. & Miller, Ch. 1980. Chemistry of Mesozoic metabasⁱtes in the middle and eastern part of the Hohe Tauern. Mitt. Geol. Gesell. 71/72, 99-118.
- Hoel, P.G. 1971. Introduction to mathematical statistics (4th edition). Wiley Int., pp.409.
- Hofmann, K.-H., Kleinschmidt, R., Lippert, R., Mager, D. & Stöckhert, B. 1983. Geologische Karte des Altkristallins südlich des Tauernfensters zwischen Pfunderer Tal und Tauferer Tal (Südtirol). Der Schlern (Monatszeitschrift für Südtiroler Landeskunde), 57, 572-590.
- Hoinkes, G. 1973. Die Anatexis des Winnebachgranits (Oetztaler Alpen, Oesterreich) am Beispiel eines Aufschlusses. Tschm. Min. Pet. Mitt. 20, 225-239.
- Hoke, L. 1987. Geology of part of the Altkristallin sheet in the Eastern Alps. Unpublished Ph.D. thesis, Cambridge.
- Holland, T.J.B. 1979. High water activities in the generation of high pressure kyanite eclogites of the Tauern Window, Austria. J. Geol. 87, 1-27.

- Holland, T.J.B. & Powell, R. 1985. An internally consistent thermodynamic dataset with correlated uncertainties 2: data and results. *J. Met. Geol.* 3, 343-370.
- Holland, T.J.B. & Ray, N.J. 1985. Glaucophane and pyroxene breakdown reactions in the Pennine units of the Eastern Alps. *J. Met. Geol.* 3, 417-438.
- Hunziker, J.C. 1974. Rb-Sr and K-Ar determinations and the Alpine tectonic history of the Western Alps. *Mem. Inst. Geol. Min. Padova*, 31, 1-54.
- Janoschek, W.R. & Matura, A. 1980. Outline of the geology of Austria. *Abh. Geol. B.-A.* 34, 7-98.
- Jeffery, G.B. 1922. The motion of ellipsoidal particles immersed in a viscous fluid. *Proc. Roy. Soc. Lond.* A102, 161-179.
- Karig, D.E. 1971. Origin and development of marginal basins in the W. Pacific. *J. Geophys. Res.* 76, 2542-2561.
- Karig, D.E. 1980. Material transport within accretionary prisms and the knocker problem. *Journ. Geol.* 88, 27-39.
- Kober, L. 1955. *Bau und Entstehung der Alpen*, 2te Auflage, Deuticke, Wien, pp.379.
- Krohe, A. 1986. Kinematics of Cretaceous nappe tectonics in the Austroalpine basement of the Koralpe region (eastern Austria). *Tectonophysics*, 136, 171-196.
- Lambert, R.St.J. 1970. A potassium argon study of the margin of the Tauernfenster at Döll^alch, Austria. *Eclog. Geol. Helv.* 63, 197-205.
- Law, R. 1987. Heterogeneous deformation and quartz crystallographic fabric transitions: natural examples from the Moine Thrust zone at the Stack of Clencoul, northern Assynt. *J.S.G.* 7, 819-833.
- Law, R., Knipe, R.J. & Dayan, H. 1984. Strain path partitioning within thrust sheets: microstructural and petrofabric evidence from the Moine Thrust, Loch Eriboll. *J.S.G.* 6, 477-497.

- Leg 107 Shipboard scientific party. 1986. A microcosm^u_Λ of ocean basin evolution in the Mediterranean, *Nature*, 321, 383-384.
- Lemoine, M., Tricart, P. & Boillot, G. 1987. Ultramafic and gabbroic ocean floor of the Ligurean Tethys (Alps, Corsica, Apennines). In search of a genetic model. *Geology*, 15, 622-625.
- Lisle, R. J. 1985. The effects of composition and strain on quartz fabric intensity in pebbles from a deformed conglomerate. *Geol. Rndsch.* 74, 657-663.
- Lister, G.S., & Hobbs, B.E. 1980. Influence of deformation history in the simulation of fabric development. *J.S.G.* 2, 355-370.
- Lister, G.S. & Williams, P.F. 1979. Fabric development in shear zones. *J.S.G.* 1, 283-297.
- Lister, G.S. & Williams, P. 1983. The partitioning of deformation in flowing rock masses. *Tectonophysics*, 92, 1-33.
- Lloyd, G.E., Law, R.D. & Schmid, S.M. 1987. A spherical electron channelling pattern map for use in quartz petrofabric analysis: correction and verification. *J.S.G.* 2, 251-253.
- Lowrie, W. 1986. Palaeomagnetism and the Adriatic promontory. *Tectonics*, 5, 797-807.
- Malvern, L.E. 1969. Introduction to the mechanics of a continuous medium. Prentice Hall, Engle Wood Cliffs, pp.713.
- Means, W.D. 1983. Application of the Mohr circle construction to problems of inhomogenous strain. *J.S.G.* 5, 279-286.
- Means, W.D., Hobbs, B.E., Lister, G.S. & Williams, P.F. 1980. Vorticity and non-coaxiality in progressive deformations. *J.S.G.* 2, 371-378.
- Milnes, A.G. 1978. Structural collision zones and continental collision central Alps. *Tectonophysics*, 47, 369-392.

- Milnes, A.G. & Pfiffner, O.A. 1977. Structural development of the
infrahelvetic complex, Eastern Switzerland. *Ecl. Geol. Hel.* 70,
83-95.
- Miyashiro, A. 1973. Paired and unpaired metamorphic belts.
Tectonophysics 17, 241-254.
- Moore, G.F. & Karig, D.E. 1980. Structural history of Nias Island,
Indonesia. Implications for subduction zone tectonics. *Am. J. Sci.*
280, 193-223.
- Naylor, M.A. & Woodcock, N.H. 1977. Transition analysis of structural
sequences. *Bull. Geol. Soc. Am.* 88, 1488-1492.
- Neugebauer, J. & Hammerschmidt, K. 1986. Alpidische Krustenausdünnung
im Ostalpin - Neuinterpretation des Metamorphose- und
Deformationsablaufs an der Grenze Mittel / Oberostalpin in der
Sausalpe (Ostalpen). Symposium für Strukturgeologie,
Kristallingeologie und Tektonik, Tübingen.
- Nowy, W. 1977. Die Geologie des Gleinalm Autobahntunnels (I Teil:
Tektonik). *Mitt. Ges. Geol. Bergbaustud.* 54, 59-123.
- Oberhauser, R. 1968. Beiträge zur Kenntnis der Tektonik und der
Paläogeographie während der Oberkreide und Paläogen im
Ostalpenraum. *Jahrb. Geol. B.-A.* 111, 115-145.
- Oxburgh, E.R. 1966. Superimposed fold systems in the Altkristallin on
the S.E margin of the Tauern Window. *Verh. Geol. B-A.* 33-46.
- Oxburgh, E.R. 1968. An outline of the geology of the central Eastern
Alps. *Proc. Geol. Ass.* 79, 1-46.
- Oxburgh, E.R. 1972. Flake tectonics and continental collision. *Nature*,
239, 202-204.
- Oxburgh, E.R., Lambert, R., Baadsgaard, H., & Simons, J. 1966.
Potassium-argon studies across the S.E. margin of the Tauern
Window, E. Alps. *Verh. Geol. B-A.* 17-35.

- Oxburgh, E.R. & Turcotte, D.L. 1970. Thermal structure of island arcs. Bull. Geol. Soc. Am. 81, 1665-1688.
- Oxburgh, E.R. & Turcotte, D.L. 1974. Thermal gradients and regional metamorphism in overthrust terrains with special reference to the Eastern Alps. Schweiz. Min. Pet. Mitt. 54, 641-662.
- Passchier, C.W. 1986. Flow in natural shear zones - the consequence of spinning flow regimes. E.P.S.L. 77, 70-80.
- Passchier, C.W. 1987a. Efficient use of the velocity gradients tensor in flow modelling. Tectonophysics, 136, 159-163.
- Passchier, C.W. 1987b. Stable positions of rigid objects in non-coaxial flow - a study in vorticity analysis. J.S.G. 9, 679-690.
- Passchier, C.W. 1988. Flow path analysis in shear zones. Geol. Rndsch. (in press). *old, 1988 - 1989*
- Passchier, C.W. & Simpson, C. 1986. Porphyroclast systems as kinematic indicators. J.S.G. 8, 831-843.
- Peer, H. und Zimmer, W. 1980. Geologie der Nordrahmenzone der Hohen Tauern (Klammkalkzone). Jahrb. Geol. B-A. 123, 411-466.
- Pfiffner, O.A. & Ramsay, J.G. 1982. Constraints on geological strain rates, arguments from finite strain states of naturally deformed rocks. J. Geophys. Res. 87, 311-321.
- Platt, J.P. 1984. Secondary cleavages in ductile shear zones. J.S.G. 6, 439-442.
- Platt, J.P. 1986. Dynamics of orogenic wedges and the uplift of high-pressure metamorphic rocks. Bull. Geol. Soc. Am. 97, 1037-1053.
- Platt, J.P. 1987. The uplift of high pressure - low temperature rocks. Phil. Trans. Roy. Soc. Lond. A321, 87-103.
- Platt, J.P. 1988. The mechanics of frontal imbrication: a first order analysis. Geol. Rndsch. (in press).

- Platt, J.P. & Behrmann, J.H. 1986. Structures and fabrics in a crustal-scale shear zone. Betic Cordillera, S.E. Spain. *J.S.G.* 8, 15-33.
- Platt, J.P. & Leggett, J.K. 1986. Stratal extension in thrust footwalls, Makran accretionary prism: implications for thrust tectonics. *A.A.P.G. Bull.* 70, 191-203.
- Platt, J.P., van den Eeckhout, B., Janzen, E., Konert, G., Simon, O.J. & Weijermars, R. 1983. The structure and tectonic evolution of the Aguilon fold-nappe, Sierra Alhamilla, Betic Cordilleras, S.E. Spain. *J.S.G.* 5, 519-538.
- Platt, J.P. & Vissers, R.L.M. 1980. Extensional structures in anisotropic rocks. *J.S.G.* 2, 397-410.
- Poirier, J-P. 1985. Creep of crystals. Cambridge university press, pp.259.
- Poirier, J.P. 1980. Shear localization and shear instability in minerals in the ductile field. *J.S.G.* 2, 63-68.
- Prey, S. 1961. Der ehemalige Großfragant^{er} Kupfer- und Schwefelkiesbergbau. *Mitt. Geol. Gesell.* 54, 163-200.
- Price, N.J. 1975. Rates of deformation. *J. Geol. Soc. Lond.* 131, 553-575.
- Raith, M., Raase, P., Kreuzer, H. & Müller, P. 1978. Alpidic metamorphism in the western Tauern Window; radiometric dating. In: *Alps, Apennines, Hellenides*. (edited by Closs, H., Roeder, F. & Schmidt, K.), 140-148.
- Ramberg, H. 1975. Particle paths, displacement and progressive strain applicable to rocks. *Tectonophysics*, 28, 1-37.
- Ramsay, J.G. 1967. Folding and fracturing of rocks. Mc.Graw Hill, New York, pp.568.
- Ramsay, J.G., Casey, M. & Kligfeld, R. 1983. Role of shear in the development of the Helvetic fold-thrust belt of Switzerland. *Geology* 11, 439-442.

- Ratschbacher, L. 1987. Strain, rotation and translation of Austroalpine nappes. In: Geodynamics of the Eastern Alps, Deuticke (edited by Flügel, H.W. & Faupl, P.), 237-243.
- Ratschbacher, L. & Oertel, G. 1987. Superposed deformations in the Eastern Alps: strain analysis and microfabrics. J.S.G. 9, 263-276.
- Roeder, D. 1977. Continental convergence in the Alps. Tectonophysics, 40, 339-350.
- Roeder, D. & Bögel, H. 1978. Geodynamic interpretation of the Eastern Alps. In: Alps, Apeninnes, Hellenides (edited by Closs, H. & Roeder, D.), 191-212.
- Sander, B. 1950. Einführung in die Gefügekunde der geologischen Körper. Springer, Wien, pp.409.
- Satir, M. 1975. Die Entwicklungsgeschichte der westlichen Hohen Tauern und der südlichen Oetztalmasse aufgrund radiometrischer Altersbestimmungen. Mem. Inst. Geol. Min. Univ. Padova. 30, 1-84.
- Savostin, L.A., Sibuet, J-C., Zonenshain, L-P., Le Pichon, X. & Roulet, M-J. 1986. Kinematic evolution of the Tethyan belt from the Atlantic Ocean to the Parais since the Triassic. Tectonophysics, 123, 1-35.
- Schmid, S.M. & Casey, M. 1986. Complete fabric analysis of some commonly observed quartz c-axis patterns. A.G.U. Mon. 36 - the Paterson volume (edited by Hobbs, B.E. & Heard, H.C.), 263 -286.
- Schmid, S.M., Casey, M. & Starkey, J. 1981. The microfabric of calcite tectonites from the Helvetic nappes. Geol. Soc. Lond. Spec. Pub. No.9. (edited by Mc.Clay, K.R. & Price, N.J.), 151-158.
- Schmid, S.M. & Haas, R. 1987. The transition from near surface thrusting to intra basement décollement along the Schlieren Thrust. Terra Cognita 7, 68.

- Schmidt, W.J. 1950-1952. Die Matreier Zone in Oesterreich (drei Teile).
 Sitz. bericht. der Oesterr. Akad. der Wiss. Mathem.- naturw. kl.
 Abt. I Teil. 1950, 159, 291-332.
 II Teil. 1951, 160, 711-760.
 III Teil. 1952, 161, 343-371.
- Schonefeld, C. 1977. A study of some typical inclusion patterns in
 strongly paracrystalline - rotated garnets. Tectonophysics, 39,
 453- 471.
- Silverstone, J. 1985. Petrologic constraints on imbrication metamorphism
 and uplift in the S.W. Tauern Window, Eastern Alps. Tectonics, 4,
 687-704.
- Silverstone, J. 1988. Evidence for east-west crustal extension in the
 Eastern Alps: implications for the unroofing history of the Tauern
 Window. Tectonics, 7, 87-105.
- Silverstone, J. & Spear, F. 1985. Metamorphic P-T paths from pelitic
 schists and greenstones in the S.W. Tauern Window, Eastern Alps. J.
 Met. Geol. 3, 439-465.
- Silverstone, J., Spear, F., Franz, G. & Morteani, G. 1984. High Pressure
 metamorphism in the S.W Tauern Window, Austria: P-T paths from
 hornblende - kyanite - staurolite schists. J. Petrol. 25, 501-531.
- Senarclens-Grancy, 1964. Zur Grundgebirgs- und Quartärgeologie der
 Def^eregger Alpen und ihrer Umgebung. Zt. Dtsch. Geol. Gesell. 116,
 502-511.
- Shakleton, R.M. & Ries, A.C. 1984. The relationship between regionally
 consistent stretching lineations and plate motions. J.S.G. 6,
 111-117.
- Simpson, C. & Schmid, S. 1983. An evaluation of criteria to deduce the
 sense of movement in sheared rocks. Bull. Geol. Soc. Am. 94,
 1281-1288.

- Smith, A.G. 1971. Alpine deformation and the oceanic areas of the Tethys, Mediterranean and Atlantic. Bull. Geol. Soc. Am. 82, 2039-2070.
- Stöckhert, B. 1984. K-Ar determinations on muscovites and phengites from deformed pegmatites, and minimum age of the Old Alpine deformation in the Austridic basement to the south of the western Tauern Window (Ahrn valley, southern Tirol, E. Alps). Neues Jahrb. Min. Abh. 150, 103-120.
- Thöni, M. 1980. Zur Westbewegung der Oetztaler Masse. Mitt. Geol. Bergbaustud. Oesterreich. 26, 247-275.
- Thöni, M. 1986. The Rb-Sr thin slab isochron method - an unreliable method for dating geologic events in polymetamorphic terrains? Mem. Inst. Geol. Min. Uni. Padova. 38, 283-352.
- Tollmann, A. 1961. Die Rolle des Ost-West Schubes im Ostalpenbau. Mitt. Geol. Gesell. 54, 229-247.
- Tollmann, A. 1977a. Geologie von Oesterreich Bd.1: die Zentralalpen. Deuticke, Wien. pp.766.
- Tollmann, A. 1977b. Die Bruchtektonik Oesterreichs im Satellitenbild. N. Jb. Geol. Paläon. Abh. 153, 1-27.
- Tollmann, A. 1986. Geologie von Oesterreich Bd.3: Gesamtübersicht. Deuticke, Wien. pp.718
- Troll, G., Forst, R. & Söllner, F. 1976. Ueber Bau und Metamorphose des Altkristallins der Schobergruppe, Osttirol. Geol. Rndsch. 65, 483-511.
- Truesdell, C. 1954. The kinematics of vorticity. Indiana university Press, Bloomington.
- Trümpy, R. 1980. Geology of Switzerland. Part A. Schweizerische geologische Kommission, pp.104.

- Tullis, J., Christie, J.M. & Griggs, D.T. 1973. Microstructures and preferred orientations of experimentally deformed quartzites. Bull. Geol. Soc. Am. 84, 297-314.
- Urai, J.L., Means, W.D. & Lister, G.S. 1986. Dynamic recrystallization of minerals. In: mineral and rock deformation: laboratory studies - the Paterson volume. A.G.U. Mon. 36, 161-199.
- Vernon, R.H. Metamorphic processes. Allen & Unwin, London, pp.247.
- Visser, R.L.M. 1987. The effect of foliation orientation on the inferred rotation axes and rotation angles of porphyroblasts. Tectonophysics, 139, 275-283.
- Visser, R.L.M. 1988. On quartz c-axis fabrics in non-coaxial deformation using rotated skeletal garnets and their included ghost fabrics. J.S.G. (in press).
- v. Gosen, W. 1982. Geologie und Tektonik am Nordostrand der Gurktalerdecke (Steiermark/Kärnten - Oesterreich). Mitt. Geol. Pal. Inst. Uni. Hamburg, 53, 33-149.
- v. Gosen, W., Haiges, K.H., Neugebauer, F., Pistotnik, J. & Thiedig, F. 1985. Die tektonischen Baueinheiten am Nord- und Westrand der Gurktaler Decke (Oesterreich). Jahrb. Geol. B.-A. 127, 693-699.
- Waters, D.J. 1976. Structural, metamorphic and geochronological studies in the S.E. Tauern: Kreuzeckgroup. Unpublished D.Phil. thesis, Oxford.
- Weber, J. 1983. Petrologische Untersuchungen im Raum Wolfsberg/Koralpe. Unpublished Ph.D. thesis, Vienna.
- Wendt, J. 1971. Genese und Fauna submarinischer sedimentärer Spaltenfüllungen im mediterranen Jura. Palaeographica, 136, 105-116.
- White, S.H., Burrows, S.E., Carreras, J., Shaw, N.D. & Humphries, F.J. 1980. On mylonites in ductile shear zones. J.S.G. 2, 175-187.

- Wiedenmayer, F. 1963. Obere Trias bis mittleren Lias zwischen Saltrio und Tremona (Lombardische Alpen). Die Wechselbeziehungen zwischen Stratigraphie, Sedimentologie und syngentischer Tektonik. Ecl. Geol. Helv. 56, 529-640.
- Williams, G. 1978. Rotation of contemporary fold axes into the X-direction during overthrust processes in Lakesfjord, Finnmark. Tectonophysics, 48, 29-40.
- Winkler, W & Bernoulli, D. 1986. Detrital high-pressure low-temperature minerals in a late Turonian flysch sequence of the Eastern Alps (western Austria). Implications for early Alpine tectonics. Geology, 14, 598-601.
- Winterer, E.L. & Bosellini, A. 1981. Subsidence and sedimentation on Jurassic passive continental margin, Southern Alps, Italy. A.A.P.G. Bull. 65i, 394-421.
- Wright R.C. 1973. Studies on metamorphic rocks of the upper east alpine sheet, Kärnten. Unpublished D.Phil. thesis, Oxford.
- Zimmermann, R., Lucassen, F. & Franz, G. 1987. Die Metamorphoseentwicklung der unteren Schieferhülle im Frosnitztal / Südvenediger. Uniaktuell - Zeitschrift der Universität Salzburg. 10, 26.