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RECOGNITION OF SYN-CONVERGENCE EXTENSION IN ACCRETIONARY WEDGES WITH EXAMPLES FROM THE CALABRIAN ARC AND THE EASTERN ALPS

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ABSTRACT. The treatment of accretionary complexes on convergent tectonic margins by continuum mechanics as weak wedge-shaped bodies above a decollement surface allows two testable predictions: (1) deformation within such wedges may include large amounts of horizontal extension as well as shortening, and (2) this extension may take place while plate convergence is continuing before continental collision. Extension may be recognized by the presence of abrupt downward increases in metamorphic grade and, at higher structural levels, the formation of extensional sedimentary basins above the orogenic wedge. Extension may also be recognized by using flow path modelling in broad zones of ductile shear. Flow-path modelling has the potential both to recognize and quantify extension but has until now received only little attention. Two former active margins, the Calabrian Arc and the Eastern Alps, are characterized by high-strain regional deformation that caused substantial subvertical thinning prior to continental collision. This deformation is consistent with the predictions of the model and can explain the exhumation of high P/T metamorphic material in both areas.

INTRODUCTION

Both theory and observation suggest that gravity exerts an important control on the nature of deformation in convergent tectonic settings (Houseman and England, 1986; England and Houseman, 1986), and that it may result in extension during convergence (Molnar and Tapponier, 1978; Royden and Burchfiel, 1987; Platt, 1986; England and Houseman, 1988). In zones of deformation hundreds of thousands of km^2 in extent such as Tibet or the Andes, thickening may affect the entire lithosphere, and the primary control on the type of deformation is the contrast in surface elevation between the zone and its surroundings (Suarez, Molnar, and Burchfiel, 1983; England and Houseman, 1988). In active thrust belts and accretionary wedges that adopt wedge-shaped profiles, on the other hand, extension may be caused by internal processes that change the angle of taper or the shear stress on the basal décollement (Dahlen, 1984). Platt (1986) has suggested that if an orogenic wedge is thick enough for prograde metamorphic reactions to occur, it is unlikely to exhibit a long-term yield strength. In this case, it will approach a configuration in which the traction exerted by the motion on its basal decollement (which tends to shorten the wedge) is equal to the

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gravitational forces (which tend to cause the wedge to extend). Accretion of material to the wedge will, in general, change this geometry, and the wedge will deform in response. If addition of material is by underplating at depth to the rear of the wedge, this will increase the taper, and the wedge will start to extend. Significant extension in the hanging wall of a subduction zone may result in the exhumation of deeply buried material, and this may be a principal mechanism by which regionally coherent high-pressure metamorphic terrains are emplaced at the Earth's surface (Platt, 1986).

The treatment of convergent tectonic margins as weak wedge-shaped continua that may extend as well as contract is a radical departure from the traditional view of mountain belts as the result of a relatively simple process of crustal shortening by folding and thrust imbrication. Two testable predictions can be made from this hypothesis for a convergent margin that undergoes accretion of material mainly by underplating.

1. Deformation within the upper rear of the wedge involves substantial horizontal extension associated with subvertical shortening.

2. Extension is a consequence of accretion by underplating and may therefore take place while convergence is still active. The timing of extension clearly distinguishes this process of localized extension above a decollement zone from whole-lithosphere extension that may occur after post-collisional convective removal of dense mantle (England and Houseman, 1989; Platt and Vissers, 1989). In this paper we use data from the Eastern Alps and the Calabrian Arc to test these two predictions. We begin with a review of the general problem of recognizing extension and its timing in orogenic belts.

RECOGNITION OF EXTENSION

Introduction

The recognition of structures formed by horizontal extension in the highly deformed and metamorphosed rocks characteristic of active convergent margins presents considerable difficulties. Strata are likely to have been folded, imbricated, or tilted into a steep orientation, and subsequent horizontal extension may then form folds or other contractional structures in the bedding. Similarly fault planes may be rotated at any stage during or after their development, and the attitude of a tectonic contact (with respect to the Earth's gravitational field) and the associated sense of movement are insufficient to determine whether it was originally associated with horizontal extension or shortening. Thrusts may be rotated by backfolding until they are technically normal faults at the front of a thrust culmination (Price, 1981, fig. 2), and normal faults may be domed by isostatic response to unloading until displacement is up-dip along parts of their trajectories (Lister and Davis, 1989). There are also difficulties associated with the scale of observation: local extensional phenomena such as footwall plucking have been recognized in association with thrust tectonics (Platt and Leggett, 1986) for example, and an extensional allochthon in an orogenic wedge may transfer displacement onto a thrust at the front of the system (for example, Platt, 1986, fig. 5d). During continuing deformation a complex sequence of cross-cutting faults may develop, which presents a further potential difficulty. For instance, an out-of-sequence thrust that cuts a previously formed imbricate stack will extend the individual thrust slices, but the large-scale effect is to cause horizontal contraction and crustal thickening (Boyer and Elliot, 1982, fig. 20).

In the following we discuss three possible ways to recognize extension in accretionary wedges: sudden downward increases in metamorphic pressure, formation of extensional sedimentary basins on the surface of the wedge, and flow path modelling in broad regions of ductile shear. Flow path modelling is a potentially powerful tool in tectonic reconstructions but has received little attention in the geological literature. We therefore give this section a fuller treatment than the other two.

1. Jumps in metamorphic grade.—Perhaps the most reliable indicator of horizontal extension in an active margin setting is a rapid or abrupt downward increase in the pressure of metamorphism. Because pressure is essentially depth dependent, this can only be explained by the extension or excision of part of the original thickness of rock that produced the pressure gradient. This type of evidence has been used to support extension in active margin settings by, for example, Platt (1986), Jayko, Blake, and Harms (1987), Sedlock (1988), Carmignani and Kligfield (1990), Jolivet and others (1990), Philippot (1990), and Schermer (1990). In principle it might also be possible to use fossil isotherms in the same way, but geothermal gradients vary greatly, and in regions of active tectonism temperature may vary horizontally as well as vertically, in which case the isotherms may have some complex predeformational geometry.

2. Formation of extensional sedimentary basins on the contemporary surface of a convergent margin.—Another clear indicator of extension is the formation of extensional sedimentary basins on the contemporary surface of an active margin. It should be borne in mind, however, that not all such basins will form as the consequence of extension. They may also form as a response to flexural loading and the formation of surface relief. Basins that have been suggested as evidence of extension on active convergent margins include the Gosau basins of the Eastern Alps (Wallis ms; Ratschbacher and others, 1989), the Stilo-Capo d'Orlando basins of the Calabro-Peloritani arc (Knott, ms; Weltje, 1992) and the actively extending forearc basin above the Hikurangi active margin in New Zealand (Walcott 1987; Cashman, 1990).

FLOW-PATH MODELLING

Throughout large parts of former convergent margins deformation is commonly distributed in broad shear zones of high ductile strain. Kinematic reconstruction within such zones can determine not only whether individual shear zones have undergone a change in thickness during deformation but also places some important geometric constraints on the nature of deformation in the surrounding material. Flow-path modelling in shear zones is a potentially powerful but as yet relatively untried technique for both detecting and quantifying vertical thinning in a mountain belt. In the following section we give the basic mechanical background to this approach.

The kinematics of steady-state deformation can be defined by knowledge of three parameters of finite deformation: finite strain, volume change, and the degree of non-coaxiality. Techniques for determining finite strain have been comprehensively reviewed by many workers (for example, Ramsay and Huber, 1983) and will not be discussed. Determining volume change in deformed rocks has similarly received considerable attention in the literature and will only be briefly discussed. It may be possible to estimate the amount of volume change by studying changes in the relative proportions of immobile minerals (Gratier, 1983). It may also be possible to quantify volume change if suitable strain markers are present. Wright and Platt (1982) use deformed graptolites to show the Martinsburg shale underwent 50 percent volume loss parallel to cleavage. Passchier (1990) has also shown that knowledge of the orientation of sets of material lines with different stretch histories, for example, deformed veins, is sufficient to determine all the parameters of finite deformation including volume change. Wallis (1992) uses this method in conjunction with an independent estimate of finite strain to constrain the amount of volume change in a metachert during deformation.

It is the determination of the third quantity, the degree of non-coaxiality, that has received the least attention in the literature. The degree of non-coaxiality of deformation gives a measure of the relative contributions of rotation to stretching during deformation (it can also be loosely thought of as the ratio of pure shear to simple shear 'components'). For instantaneous deformation this quantity is most commonly expressed in terms of the kinematic vorticity number, Wk (Truesdell, 1954; Malvern, 1969; Means and others, 1980; Lister and Williams, 1983; Passchier, 1986), although other measures are also possible (Ghosh and Ramberg, 1976; Ghosh, 1987). In the analysis of finite deformation the degree of non-coaxiality can be represented by a mean kinematic vorticity number, Wm (Passchier, 1988a,b). For steady-state deformation Wk = Wm. As an example, for zero volume change and $0 \leq \mathrm{Wm} \leq 1$ three types of steady-state plane deformation can be defined,

1. Wm = 0 pure shear,
2. $0 < \mathrm{Wm} < 1$ general non-coaxial deformation, and
3. Wm = 1 simple shear.

Several possible methods for determining the degree of non-coaxiality (Wm) have been suggested (Ghosh and Ramberg, 1976; Ghosh, 1987; Passchier, 1988a,b). Qualitative studies of the degree of non-coaxiality have been made using quartz c-axis fabrics (Platt and Behrmann, 1986; Law, Knipe, and Dayan, 1984; Law, Casey, and Knipe 1986; Schmid and Casey, 1986), the orientation of sets of deformed veins

(Hutton, 1982; Passchier, 1986), and the geometry of shear bands (Platt and Vissers, 1980). Quantitative estimates have also been made using rotated rigid objects (Passchier, 1987; Wallis, 1988; Vissers, 1989; Cowan, 1990), deformed vein sets (Passchier, 1988a; Wallis, 1992), and the stretch and rotation of several material lines (Passchier and Urai, 1988). If the finite strain is known then the geometry of crystallographic fabrics may also be used to quantify Wm (Vissers, 1989; Wallis, 1992). A full review of these techniques is beyond the scope of this paper, but an example of their use is given in the section on the Eastern Alps (see later).

In the special case of zero volume change, a value of Wk < 1 in a shear zone implies a rate of shortening or extension in the plane of the shear zone and a corresponding rate of thickening or thinning of the zone. Subject to certain assumptions, a component of shortening or extension in the plane of a shear zone requires the same component of deformation to be present in the surrounding rocks, and it is for this reason that flow path analysis of shear zones can help determine the nature of the bulk deformation in a mountain belt. This requirement arises from consideration of strain compatibility. This is an essential tenet of continuum mechanics, which states that the deformation and the rates of deformation can only vary spatially in ways that do not cause gaps or overlaps to develop between adjacent volumes of material. This establishes relations among the strain gradients, most simply expressed by the equation

$$\nabla \times \mathrm{E} \times \nabla = 0, \tag{1}$$

where E is the finite strain tensor (Malvern, 1969, p. 186).

If applied to an ideal shear zone, that is parallel-sided and effectively of infinite extent in its own plane, strain compatibility requires that for constant volume deformation, the deformation in the shear zone be the sum of any amount of simple shear plus whatever deformation occurs in the wall rocks outside it (Cobbold, 1977; Ramsay, 1980). The converse is also true, hence if there is any component of stretch or rate of stretching within the shear zone parallel to its boundaries, this must also be present in the wall rocks. If this were not the case material would be squeezed out, or sucked into, the shear zone, and for a shear zone of infinite extent the resulting flow velocities, rates of shear strain, and hence deviatoric stress would become infinite. Real shear zones, however, are not of infinite extent. How long do they have to be, compared with their thickness, for strain compatibility to be a requirement? This depends on the strength of the materials: once the deviatoric stresses arising from the viscous resistance to flow in or out of the shear zone exceed the strength of the wall-rocks, the latter will start to deform. Shear zones are necessarily weaker than their wall-rocks, but if the shear zone has formed by deformation of the surrounding rock, the flow stress within it is in any event likely to be close to the plastic limit of the wall rocks. Hence the latter are unlikely to be able to withstand any significant increase in deviatoric stress caused by a failure to obey flow compatibility requirements. For example, a value of Wm = 0.9 implies a ratio of zone-parallel stretching (or shortening) to simple shear of about 0.24:1. If the wall-rocks remain rigid, this implies that in a shear zone with an aspect ratio of, say, 10:1, and a γ of 1, that material would be extruded (or inhaled) at each end for a distance roughly equal to the thickness of the shear zone. The rates of shear strain associated with this extrusion would be about twice those associated with the simple shear itself, and there would be a corresponding increase in deviatoric stress. It is probably safe to say, therefore, that for parallel-sided shear zones with an aspect ratio of two or more orders of magnitude, strain compatibility constraints will apply fairly closely, and hence any shortening or extrusion in the plane of the shear zone also affects the surrounding rocks.

A value of Wm $<$ 1 may imply either shortening or extension in the plane of the shear zone and hence either thickening or thinning of the zone. These two possibilities can be distinguished by the orientations of the deformational fabrics at the margins of the zone. If deformation in the zone causes thickening, then the resulting foliation and stretching lineation will be inclined at more than 45° to the shear zone boundary in the margin of the zone and in the adjacent wall-rocks (fig. 1); if it causes thinning, this angle will always be significantly less than 45° (fig. 1).

Subject to some assumptions, therefore, we can use the kinematic vorticity number in large-scale shear zones to constrain the rates of the bulk deformation in a mountain belt. These assumptions are (1) constant volume deformation, and (2) that the shear zone is parallel-sided and has an aspect ratio of two or more orders of magnitude. We must also know the approximate orientation of the shear zone at the time of deformation. If the shear zone was vertical then thinning across the boundaries implies horizontal shortening and regional thickening. If, on the other hand, the shear zone was gently inclined, then shortening across the boundaries implies vertical shortening and regional thinning. The orientation of a shear zone may be determined if it can be related to the paleo-horizontal, for example, an erosional or isobaric surface.

A potential difficulty with this analysis is that if the shear zone boundaries rotate with respect to the regional deformation field, the flow will be non steady-state and cannot be analyzed simply in terms of the finite end state. For regional deformation it can be argued that the Earth's surface will be a surface of zero finite rotation and, therefore, that a band of high-strain deformation parallel to this surface will be non-rotating and undergo steady-state deformation. This implies that steady-state deformation may be inferred from the original orientation of the shear zone. Secondly, it may be possible to show that deformation in a shear zone was approximately steady-state by using a variety of different elements of the tectonic fabric to determine Wm. The different fabric-elements are likely to develop at different stages in the deformation history. Consistent results using a variety of methods therefore imply steady-state deformation (for example, Wallis, 1992).

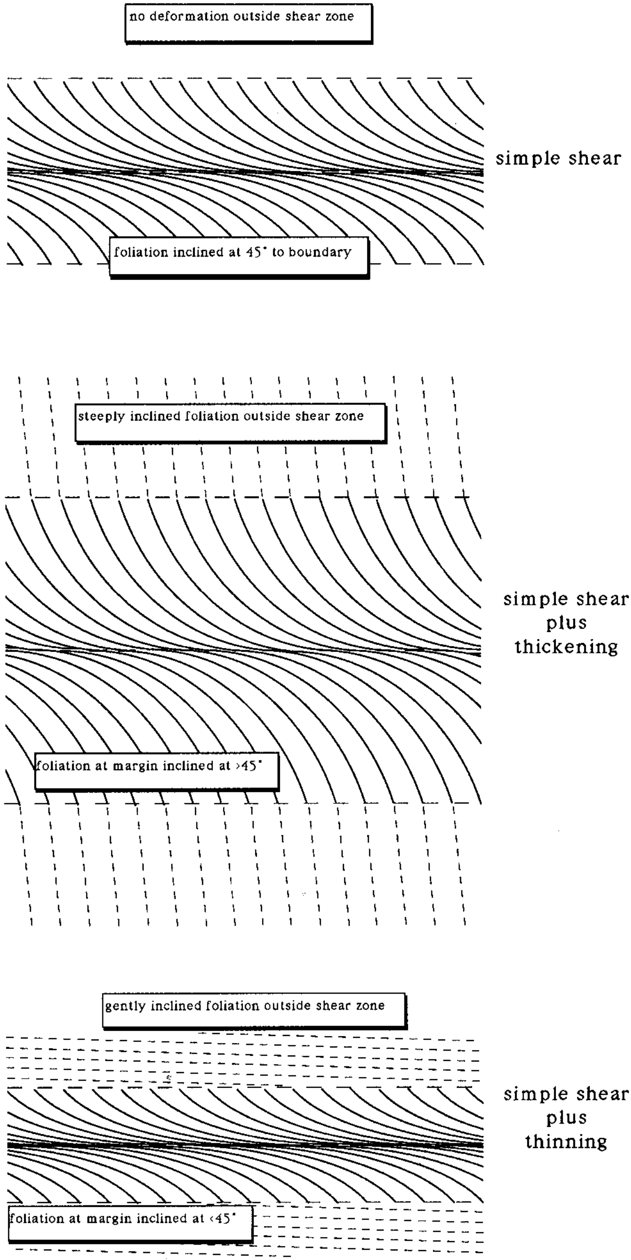


Fig. 1. Orientation of the foliation within three types of shear zones developed by plane steady-state deformation. In the case of simple shear (Wm = 1), there is no change in thickness across the shear zone after deformation. In the other two cases the degree of non-coaxiality is less than simple shear (Wm < 1), but in one case deformation causes a thickening of the shear zone and in the other thinning. Both these types of deformation can be described in terms of a simple shear deformation within the shear zone upon which a general deformation is imposed that affects both the shear zone material and the wall rocks (Cobbold, 1977). The predicted orientation of the foliation with respect to the shear zone allows the two possibilities to be distinguished.

DETERMINATION OF THE TIMING OF EXTENSION

Even if extension can be demonstrated within a convergent system, there may be some difficulty in establishing whether it occurred during the process of plate convergence. Many collisional orogens are known to have experienced extension after plate convergence ceased, for a variety of reasons, and this can confuse the issue. We can identify four main criteria for recognizing extension during convergence.

1. The most direct is the observation of present-day extension in active margin settings. The best documented example of this is the Hikurangi margin of New Zealand, where Walcott (1987) has shown that extension is taking place in the rear of the accretionary wedge and has shown by mass-balance calculations that it is probably related to underplating.

2. Stratigraphic evidence for deposition in extensional basins above active thrust wedges is ideal evidence, but such basins are only sporadically preserved, and it may be difficult to demonstrate that the basins were formed by extension. Possible examples include the Gosau beds of the Eastern Alps and the Stilo Capo d'Orlando Formation of Calabria, both of which are discussed later.

3. Radiometric dating of various stages on the P-T-time path of metamorphic complexes can determine the timing and rate of exhumation. If exhumation is known to have resulted from extension, this constrains the timing of extension. An example is the dated exhumation history of the eclogitic Sesia Zone of the Western Alps (Rubie, 1984).

4. Direct dating of extensional structures within orogenic wedges is difficult, but they may be dated in relation to earlier and later structures or to metamorphic events. Examples of this are given later in the sections on the Eastern Alps and the Calabrian Arc.

THE TWO CASE STUDIES: THE EASTERN ALPS AND THE CALABRIAN ARC

Introduction

The Calabrian Arc and the Eastern Alps formed within the overall convergent tectonic setting between Africa and Europe (Biju-Duval, Dercourt, and Le Pichon, 1977; Smith, 1971; Channel, D'Argenio, and Hovarth, 1979; Dewey and others, 1973, 1989), and both had a prolonged history as active subduction complexes from Cretaceous to Tertiary times culminating in continental collision. Both orogens contain regionally coherent terrains metamorphosed under high P/T conditions, which have been emplaced at relatively high levels in the Earth's crust.

In the following we present the basic data to show that both these orogens have undergone major extension while convergence was still active. The two case studies emphasize the use of different techniques to recognize extension. We concentrate, however, on the Eastern Alps because this section includes the example of flow path modelling in tectonic studies.

CALABRIAN ARC

Geological setting.—The Calabrian Arc is the curved segment of the southern Italian mountain belt that joins the north-south trending Apennine mountains of Italy with the east-west trending Maghrebide range of Sicily and North Africa. The present shape and structure of the Calabrian Arc (fig. 2) are due to a combination of subduction, intra orogenic extension, and back-arc extension, mainly during Cenozoic time (Knott, ms). In the north, the arc can be divided into three vertically stacked tectonic complexes (Amodio-Morelli and others, 1976; fig. 3). The structurally lowest complex is composed of mainly carbonate rocks of Mesozoic age that were originally deposited on the continental margins of Africa and Adria. These sediments were stripped from their basement during the Early Miocene collision of Calabria with Africa and Adria and presently form part of the Africa-verging Apennine-Maghrebide fold-thrust belt (Dewey and others, 1973; Scandone, Giunta, Liguori, 1974; Ogniben, Parotto, and Praturlon, 1975). The middle complex (Ligurides) is composed of two nappes (Lower Ophiolitic Nappe, Upper Ophiolitic Nappe) of Mesozoic to Cenozoic metasedimentary and ophiolitic rocks, which can be interpreted as the remains of an ancient accretionary wedge (Knott, 1987; Hill and Hayward, 1988). The uppermost complex (Calabrides) is composed of Palaeozoic thrust sheets of igneous and metamorphic rocks and Mesozoic to Cenozoic sediments considered to be the basement and cover respectively of the former European Margin of Neotethys (Ogniben, 1969; Bouillin, 1984; Bouillin, Durand-Delga, and Oliver, 1986; Knott, 1987; Dewey and others, 1989).

Extensional tectonics in the Calabrian Arc.—Some of the best evidence for regional extension in the Calabrian Arc is provided by the contrast in metamorphic grade between the unmetamorphosed or locally greenschist facies Upper Ophiolitic Nappe (UON) and the blueschist-facies LON. An exposure of the tectonic contact between these two units near Spezzano Albanese (fig. 4) is described by Bouillin (1984). At this locality the UON tectonically overlies the LON, and the two nappes are separated by a tectonic slice of Eocene sedimentary rocks. Both the UON (pillow basalt, radiolarite, and Calpionella limestone) and the Eocene sedimentary rocks are unmetamorphosed (Bouillin, 1984). The LON contains high-pressure low-temperature metamorphic assemblages including jadeite plus quartz, glaucophane, crossite, lawsonite, aragonite, and phengite (Spadea, Tortorici, and Lanzafame, 1976). The abrupt increase in metamorphic pressure going downward across the contact between the Eocene sedimentary rocks and the blueschist-facies LON suggests that the boundary is an extensional fault that has excised roughly 15 km of rock.

An analogous situation exists within the crystalline thrust sheets in southwestern Calabria. These have locally been affected by high-pressure greenschist facies metamorphism, probably related to the blueschist facies metamorphism of the LON, followed by a lower pressure upper greenschist to amphibolite facies metamorphism in Late Oligocene time

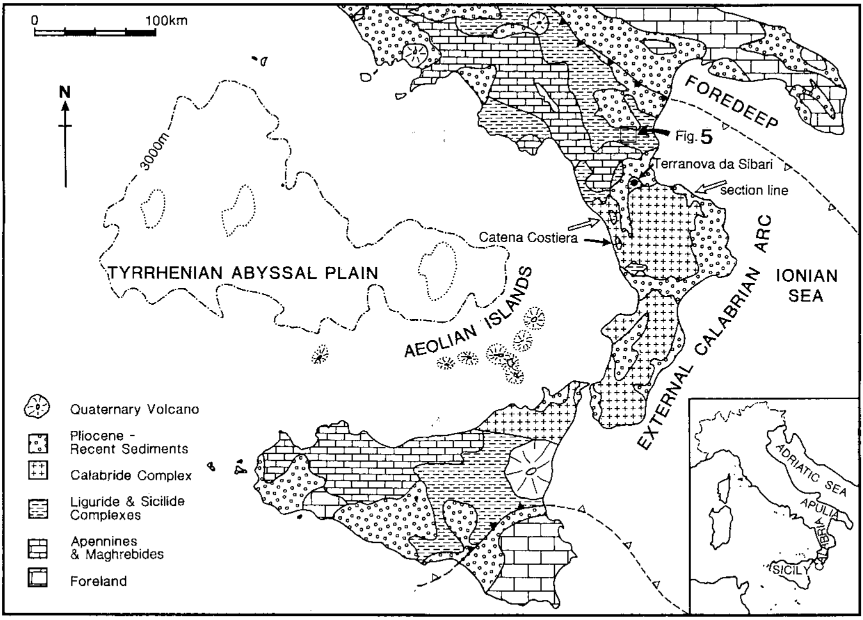


Fig. 2. Geological map of the Southern Italian mountain belt. Single barbed line represents the Apennine thrust front (internal thrusts are not shown for clarity).

(Bonardi and others, 1984, 1987; Platt and Compagnoni, 1990). The region affected by these metamorphic events is separated from rocks lacking evidence for Alpine metamorphism by a subhorizontal extensional shear zone about 1 km thick (Platt and Compagnoni, 1990).

Structures within the LON also indicate regional extension. Numerous ductile or semiductile shear zones are developed within the LON, many of which are extensional with respect to stratigraphic section. These zones formed within a well-defined stage in the deformation sequence, that is they truncate early ductile fold-thrust structures but pre-date late brittle out-of-sequence thrusts and are folded in their hanging walls (figs. 5, 6). This structural sequence is recognizable within the LON over the entire internal zone of the southern Apennines. The regional development of these structures shows that their extensional nature cannot be explained solely as a local effect.

A third line of evidence for extension comes from the deposition of conglomerates and turbidites of the Early Miocene Stilo-Capo d'Orlando Formation (Bonardi and others, 1980) in extensional basins above the crystalline basement rocks (Knott, 1988; Weltje, 1992). The basins are now exposed onshore in northeastern Sicily and southern Calabria. The Stilo-Capo d'Orlando Formation is up to 2 km thick and initiates in

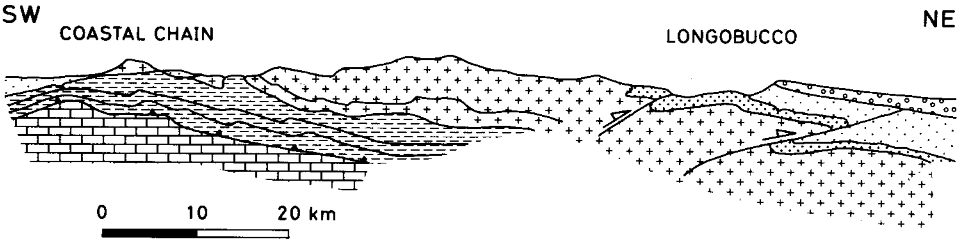


Fig. 3. Cross section through the Calabrian Arc, see figure 2 for location. Bricks-Apennine Platform (Panormide Complex); dashes—Liguride Complex; crosses—Calabrian basement nappes (Calabrian Complex); heavy dots—Longobucco Group (Lias) and Paludi Formation (Eocene to ?Oligocene); light dots—Miocene deposits; open circles—Pliocene to Recent deposits. The deformation sequence is: (A) Jurassic(?) Neotethys extension, (B) contraction, (C) ductile extension.

non-marine facies passing rapidly upward into deep marine turbidites and debris flow deposits (Bonardi and others, 1980).

In the following section we show that these three expressions of regional extension all developed during plate convergence and before continental collision.

Timing of extension.—The Liguride accretionary complex formed as a result of the subduction of a part of the Neotethys oceanic crust beneath the Adriatic microcontinent. A small continental block was attached to the northwest of this piece of oceanic crust. Palaeomagnetic data and radiometric dating show that this block first began to rotate in the Early Oligocene which marks the onset of convergence (Montigny, Edel, and Thuziat, 1981; Dewey and others, 1989). At this time the block consisted of Calabria and Sardinia. From late Oligocene, Corsica also began to rotate as part of a composite block consisting of Calabria, Sardinia, and Corsica (CSC block) (Dewey and others, 1989). Rotation of the CSC block and hence plate convergence were largely complete by Burdigalian time (about 18 Ma; Montigny, Edel, and Thuziat, 1981). Convergence throughout this time is also demonstrated by the presence of coeval thrusting and synorogenic turbidite deposition in the external zone (lower part of the Albidona Formation, Early Oligocene) in the flexural foredeep to the east of Calabria (Knott, ms). Convergence in the Calabrian Arc therefore took place from early Oligocene ($\approx$ 35 Ma) to 18 Ma.

The extensional structures in the Liguride complex can be dated by their relationships with metamorphism and sedimentation. Both the extension recognized along the boundary between the LON and UON and also that which is developed within the metamorphic LON must have occurred after the high P/T metamorphic peak, which affects Late Eocene sedimentary rocks. A younger age limit for extension in the LON is given by the presence of unmetamorphosed late Oligocene to Burdigalian flysch (upper part of the Albidona Formation), which was deposited in perched basins and unconformably overlies the HP/LT metamor-

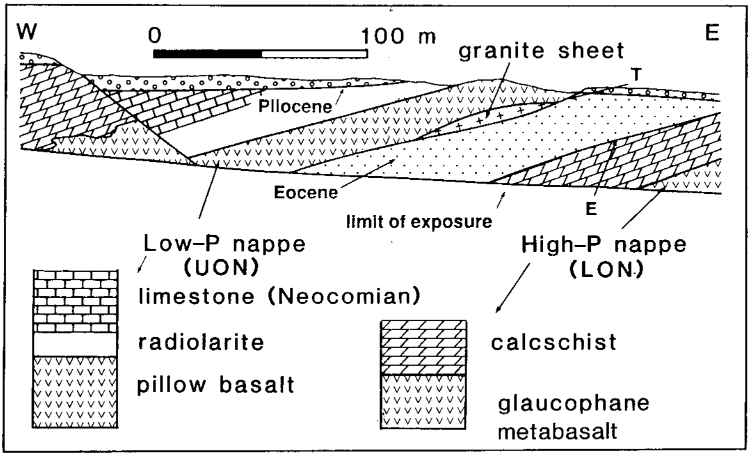


Fig. 4. Cross section through the lower (high P) and upper (low P) ophiolitic nappes near Terranova da Sibari (see fig. 2 for location). The contact between Eocene sedimentary rocks and high-P nappe is an extension fault omitting roughly 15 km of rock. The contact between the low-P nappe and the Eocene sediment is a thrust (Late Jurassic rocks over Eocene rocks). The granite sheet is probably a raft of upper crust which was stranded in the oceanic material during Jurassic rifting/spreading. There is a general repetition of stratigraphy by thrusting, but thrust contacts have been overprinted by layer-parallel semi-ductile extension faults during crustal thinning. Cross section reinterpreted from Bouillin (1984). T = thrust, E = extension fault. Vertical and horizontal scales equal.

phic rocks in the internal zones of northern Calabria and Lucania (Bonardi, Clampo, and Perrone, 1985). In the internal zones of southern Calabria and Sicily extensional structures are postdated by the deposition of the Stilo-Capo d'Orlando Formation. The Silo-Capo d'Orlando Formation formed in extensional basins above the crystalline and flysch nappes of the internal zones of southern Calabria and Sicily. The lower part of this formation is dated between post-late Eocene to pre-late Oligocene (Bonardi, Clampo, and Perrone, 1985), which is synchronous with plate convergence and overlaps with the timing of other extensional structures discussed previously.

Locally the sedimentary basins are affected by more recent extension (for example, Knott and Turco, 1991). However, the above sedimentary ages (late Eocene to late Oligocene) bracket at least the onset of extension within the Liguride Complex. This shows that extension in the internal part of the system was active at the same time as convergence. In figure 7 we summarize the convergent tectonics of this region.

EASTERN ALPS

Geological setting.—The Eastern Alps formed as a consequence of convergence between the continent of Europe and the Adriatic microplate (Smith, 1971; Dewey and others, 1973, 1989; Biju-Duval, Dercourt,

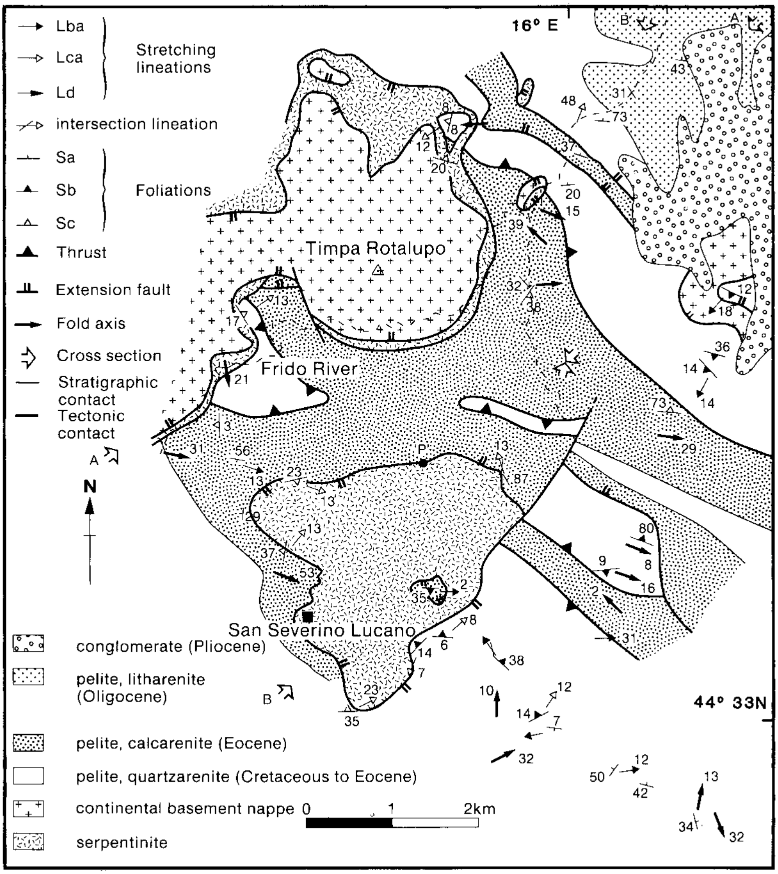


Fig. 5. Geological map of part of the southern Apennines (based on mapping by S.D.K. at 1: 10 000 scale and reinterpretation of published geological maps from the 1:25 000 series, Carta della Calabria). See figure 2 for locations. Sa = bedding, Sb = first cleavage (formed during deformation Db), Sc = crenulation cleavage, Lba = stretching lineation formed during deformation Db, Lca = stretching lineation formed during deformation Dc, Ld = stretching lineation formed during ductile extension event, E = extension fault.

and Lepichon, 1977; Laubscher and Bernouilli, 1982; Platt and others, 1989). The intervening Pennine Ocean was subducted to the south or southeast beneath Adria, forming an accreted packet of thickened crust (Dietrich, 1976; Tollmann, 1977; Frisch, 1979). The northern fringe of Adria is represented in the present day Eastern Alps by the largely continental Austroalpine domain (fig. 8). During the early stages of convergence both the Pennine and Austroalpine domains underwent substantial crustal thickening. In the Austroalpine domain this can be

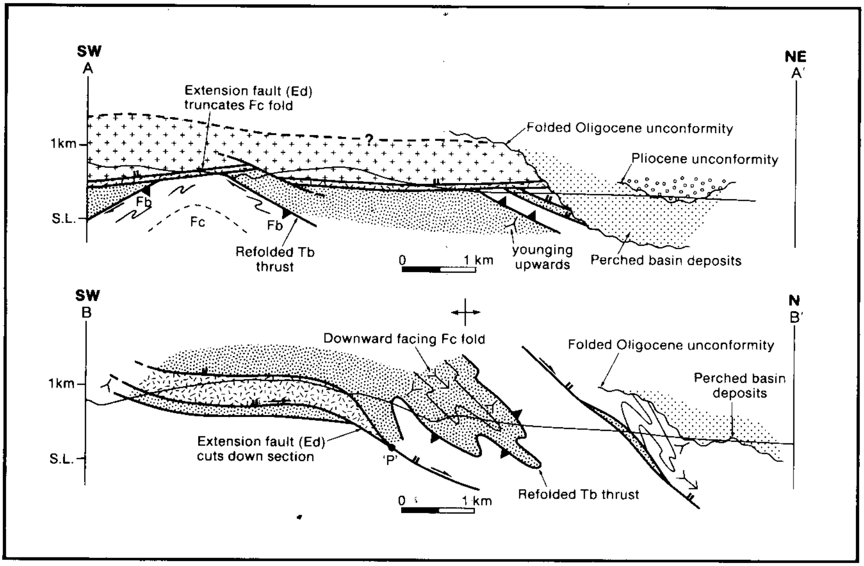


Fig. 6. Cross sections through part of the internal zone of the southern Apennines around San Severino Lucano. See figure 5 for location of cross sections (A-A', B-B') and explanation of symbols. The folded uncomformable contact between the Ligurian nappes and the Oligocene basin strata is present in the northeast corner of the map. Extension is seen to predate the Oligocene deposits and thrusting by a number of features: (1) Oligocene strata unconformably overlie extension faults that have been folded in the hangingwalls of thrusts (see sec. A-A'). (2) A large number of minor ductile extension faults, too small scale to be shown on the map, are also unconformably overlain by Oligocene sediments. (3) The number of phases and style of deformation is different above and below the unconformity: above there is evidence for only one phase of folding and associated thrusting; below there were firstly two phases of folding/thrusting (Db, Dc) followed by an extension phase (Dd) and then a final phase of folding/thrusting (De).

related to a widespread Barrovian facies metamorphism (Brewer, 1969; Hawkesworth, 1976; Thöni, 1986; Krohe, 1986; Frank and others, 1987). In the Pennine domain thickening also caused regional metamorphism but under high P/T conditions (Raith and others, 1978; Holland, 1979; Franz and Spear, 1983; Selverstone and others, 1984; Selverstone, 1985; Holland and Ray, 1985). Estimates of the peak metamorphic pressures indicate Pmax > 10 kb in the Pennine domain implying parts of this domain were buried to depths of > 35 km. However, the full structural thickness of the overburden to these high pressure rocks, including the upper Pennine units and the Austroalpine nappes, is less than 20 km (Clar, 1965; Oxburgh and Turcotte, 1974; Tollmann, 1977). This leaves a shortfall of around 15 km required to account for the metamorphic pressures. Both Selverstone (1985) and Platt (1986) have suggested that this may be explained by extension of the higher structural units within the Eastern Alps. Two of the main objections to this proposal are the traditional perception of the Austroalpine domain as a rigid lid largely unaffected by the Alpine orogeny (Bickle and Hawkesworth, 1978; Laubscher and Bernoulli, 1982) and the lack of convincing structural evidence for extension in the Austroalpine domain (Hsü, 1991).

Extensional tectonics in the Austroalpine domain.—To investigate the possibility that the Austroalpine domain underwent regional extension during plate convergence, several areas straddling the Austroalpine-Pennine boundary were studied in the southeastern Tauern region. This major tectonic boundary is the contact between the continental Austroalpine units and the dominantly oceanic Pennine units below (Tollmann, 1977; Frisch, 1979; Oxburgh, 1968) and is well-exposed along the southern boundary of a large tectonic window called the Tauern Window (figs. 8, 9A). In this region the Austroalpine-Pennine boundary is marked by the development of a highly deformed zone, the Matrei Zone, which contains elements of both Austroalpine and Pennine affinities (Cornelius and Clar, 1939; Bickle and Hawkesworth, 1978; Frisch and others, 1987; Wallis, ms). Along this contact zone, early thrust imbricates have been strongly overprinted by a penetrative ductile deformation (Behrmann and Wallis, 1987). We refer to this fabric as Ds. Ds deformation defines a zone of high-strain shear in the base of the Austroalpine domain that can be traced around the southeastern margin of the Tauern Window for a distance of about 80 km (Waters, ms; Hawkesworth, 1976; Hoke, ms; Wallis, ms). This zone is of the order of 1 to 2 km thick and is associated with a fabric that dips at moderate angles to the south (fig. 9B). These dimensions suggest that the zone of Ds deformation within this region is unlikely to depart significantly from the constraints imposed by strain compatibility conditions. Within the zone, Ds is associated with a strong platy foliation and stretching lineation, and the opening angle of Ds folds is consistently $<10°$ all of which suggest relatively high strain. The associated schistosity, Ss, is the dominant fabric of the area and can be traced continuously from within the Austroalpine basement into the Pennine units beneath (fig. 10). Ds is also present at higher structural levels in the Austroalpine domain (fig. 10), but here it is lower strain and not associated with the same penetrative platy fabric. There is no major change in the orientation of the Ss foliation between the zone of strong Ds deformation and the region of weaker Ds deformation. These observations suggest that the shear zone was not shortening during deformation (fig. 1).

One of the principal lines of evidence for recognizing extension in this region is flow-path modelling within the zone of distributed Ds shear. However, to relate shortening across this zone to the larger scale movements and possible vertical shortening, it is necessary to know the approximate orientation of the zone of Ds deformation during its formation. Cross sections through the eastern Alps show that the Austroalpine domain has a wedge-shaped profile with an angle of taper of a few

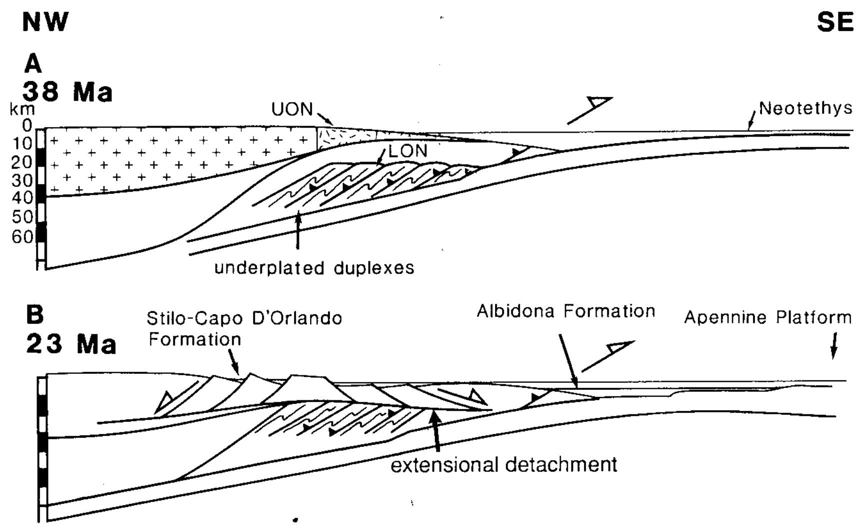


Fig. 7. Interpretation of the structural history of the Cenozoic convergent margin near Calabria from Late Eocene to Early Miocene time. Calabria lies on the southeast margin of the European Plate (left of diagram).

(A) Eocene time (38 Ma). Northwest-directed subduction of Neotethys beneath Calabria and underplating of the Neotethys oceanic/transitional crust. The future upper ophiolitic nappe (UON) forms the ophiolitic forearc region of the upper plate and the lower ophiolitic nappe (LON) is composed of a series of major underplated duplexes derived from the lower plate. (B) Aquitanian time (23 Ma). Following and during underplating extension occurs by normal faulting in the upper crust and ductile extensional flow at lower structural levels. Underplating is favored by rapid subduction (3 cm/yr based on the reconstructions of Dewey and others, 1989) and the relatively thick (> 2 km) carapace to the Neotethyan Ocean (see section on the controls on extension). Progressive thinning at the back of the wedge causes subsidence and deposition in the forearc region (Stilo-Capo d'Orlando Formation). The lower ophiolitic nappe is progressively exhumed by continued underplating and lies in close proximity to the upper ophiolitic nappe. The detailed relationships are, however, complex, and the LON is not specifically identified on the diagram. The present day Calabride Complex and Liguride complex in Calabria represent the most internal and the deepest part of the wedge exposed. The Liguride Complex in the southern Apennines represents a more shallow and external part of the wedge. The two regions have been displaced left laterally by several tens (or even hundreds) of kilometers along the southern Apennines lateral shear zone (Knott, ms). The cross sections are area balanced, and lengths are constrained by a recent plate reconstruction of the Western Mediterranean (Dewey and others, 1989).

degrees (Tollmann, 1977; Clar, 1965; Bickle and Hawkesworth, 1978). This is in agreement with observations of present day accretionary complexes which generally have an angle of taper of $< 10^0$ (for example, Davis, Suppe, and Dahlen, 1983). The upper surface of the Austroalpine nappes was a site of active deposition for most of the convergent history (for example, Tollmann, 1977) suggesting that there was only local deviation from sealevel. This implies that during the same period the lower boundary of the Austroalpine domain and therefore also the zone

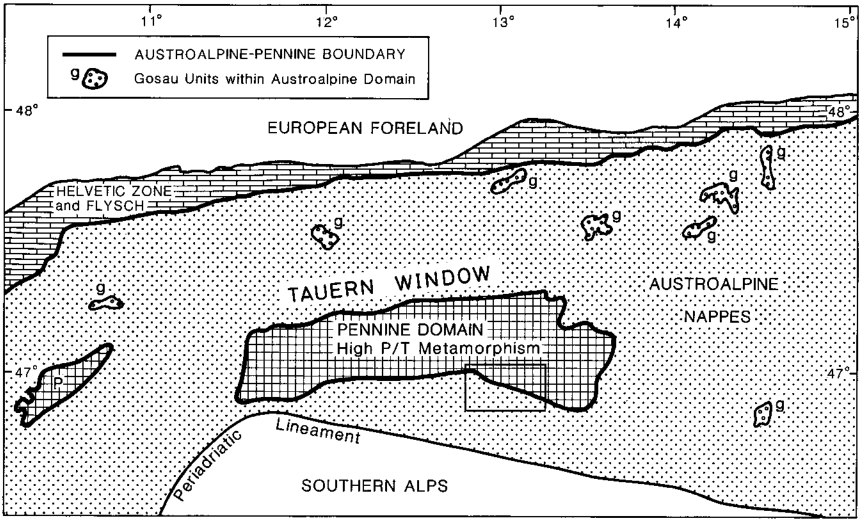
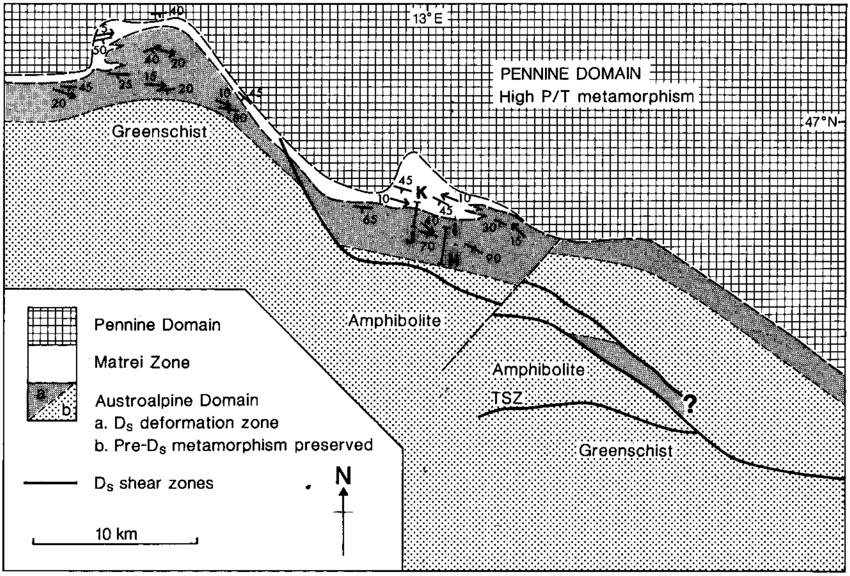


Fig. 8. The main tectonic units of the Eastern Alps. The location of figure 9A, a more detailed map across the Austroalpine-Pennine boundary, is shown by the box. The underlying Pennine units are exposed in two tectonic windows through the Austroalpine nappes. The largest of these tectonic windows is the Tauern Window.

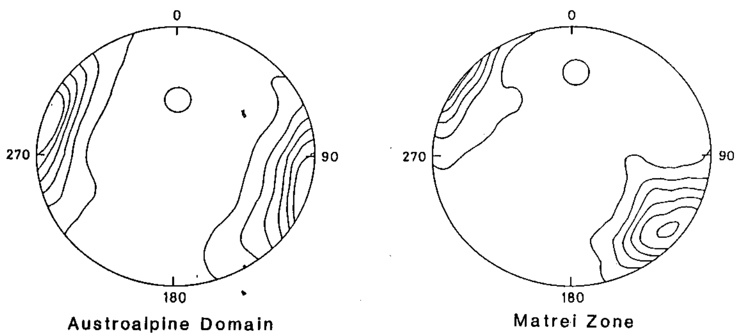
of distributed Ds shear dipped a few degrees to the south, that is, the zone of high-strain Ds deformation formed in a subhorizontal orientation. Therefore, assuming strain compatibility requirements are adhered to, any thinning across the zone of Ds deformation also caused vertical shortening of the Austroalpine domain. The present orientation is a result of post-collisional deformation.

The key parameter that needs to be determined for the flow-path modelling is the associated degree of non-coaxiality, Wm. A significant change in thickness across the shear zone should be reflected in a deviation away from simple shear, that is, $Wm < 1$. The lineation and foliation both inside and outside the zone of Ds shear were originally gently inclined, therefore a value of $Wm < 1$ implies thinning across the zone (fig. 1). Three features of the Ds tectonic fabric were used to investigate the degree of non-coaxiality: the orientation of shear bands, quartz c-axis fabrics, and the rotation of albite porphyroclasts. The study was restricted to the approx 2 km thick zone of high-strain penetrative deformation in the base of the Austroalpine units.

Shear bands are a common feature of high-strain micaceous tectonites, and the formation of these micro-shear bands is strongly controlled by the kinematics of deformation (Platt and Vissers, 1980; Bobyarchick, 1986). The kinematic analysis of Platt and Vissers (1980) suggests that coaxial deformation produces conjugate sets while simple shear is associ-



A.



B.

Fig. 9(A) Metamorphic and structural division of the southeastern Tauern region of the Eastern Alps. Location of shear zones after Hoke (ms), Waters (ms), and Wallis (ms). Ds deformation produced the dominant fabric throughout the Matrei Zone and Pennine domain of this area. Ds also forms a broad zone of distributed shear in the base of the Austroalpine Nappes. The mesoscopic orientation data for this zone and the underlying Pennine domain are also shown. At higher structural levels above this zone of distributed shear, Ds is associated with the development of a series of discrete shear zones. One of these shear zones clearly offsets the base of the Austroalpine nappes in the form of a large-scale shear band. TSZ = Teuchl Shear Zone. H-I J-K show the location of sections shown in figure 10. These sections show the orientation of the Ss foliation and the location of samples used for quartz c-axis work.

(B) Ds stretching lineation data from the base of the Austroalpine nappes and the immediately underlying Matrei Zone. The large circle shows the average orientation of the pole to the Ss foliation. Stretching lineation data: Austroalpine domain N = 136, Matrei Zone N = 238. Contoured using an algorithm following the suggestions of Fisher, Lewis, and Embleton (1987).

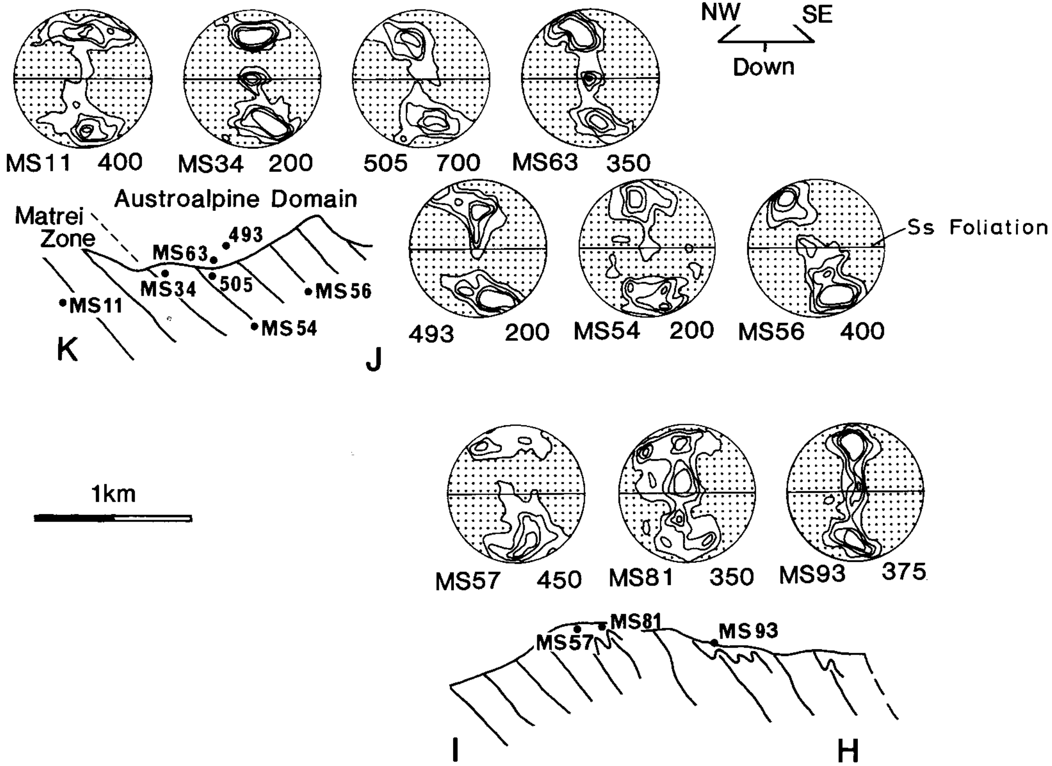


Fig. 10. Cross sections across the southern boundary of the Tauern Window showing the orientation of the Ss fabric and the location of samples used for quartz c-axis work. See figure 8A for location of sections. The position of some fabrics has been projected along strike and parallel to the local stretching lineation. The fabrics are all drawn looking to the northeast and with respect to the Ss foliation. Sample numbers are given beneath the fabric on the left-hand side. The numbers on the right hand side of each fabric refer to the number of c-axes measured. Contours at 1 percent per 1 percent area with < 1 percent shaded. There is a consistent asymmetry in point density indicating a non-coaxial deformation with a top to the northwest sense of shear. Many of the fabrics have a kinked or roughly orthorhombic topology suggesting deformation with $0 < Wm < 1$.

ated with a single set developed synthetically with the bulk sense of shear. This also implies that a general non-coaxial deformation ($0 < Wm < 1$) would be represented by unequal development of conjugate sets, the dominant set indicating the bulk sense of shear.

The orientation of shear bands was measured throughout the zone of distributed high strain (fig. 11). Samples were collected throughout the zone of Ds shear and are representative of the zone of deformation as far as it is exposed. The shear bands form in conjugate sets with their poles at a high angle to the extension direction, suggesting approximately plane-strain deformation with a low rotational component. The asymmetry in the density of the data suggests a top to the northwest sense of shear.

Quartz-rich tectonites commonly develop crystallographic preferred orientation patterns by slip along particular crystallographic planes. Well-developed c-axis fabrics show geometries that can be used to characterize the kinematics of the flow (Lister and Hobbs, 1980; Schmid and Casey, 1986). Theory predicts that deformation by pure shear will produce orthorhombic quartz c-axis patterns (Lister and Hobbs, 1980). This is supported by the observations of Law, Schmid, and Wheeler (1984) in samples where there is independent evidence of a very low degree of non-coaxiality. In contrast, in many natural examples simple shear deformation is associated with the formation of single girdles inclined to the foliation in the direction of shear (Schmid and Casey, 1986; Law, Schmid, and Wheeler, 1990). Fabrics with a geometry intermediate between these two types, commonly with a kinked outline, can be interpreted as representing deformation intermediate between simple and pure shear (Law, Knipe, and Dayan, 1984; Law, Casey, and Knipe; Platt and Behrmann, 1986; Schmid and Casey, 1986; Vissers, 1989, Wallis, 1992) (fig. 12).

The preferred orientation patterns of the quartz c-axes were measured in a variety of quartz-rich tectonites throughout the zone of distributed Ds shear. A selection of the data is shown in figure 10. These fabrics generally show some degree of asymmetry in both topology and density distribution, indicating a non-coaxial flow within the rock with a top to the northwest sense of shear. Four of the samples show a clear asymmetric kinked outline. However, in no case is there an inclined single girdle fabric indicative of simple shear. These fabrics, therefore, suggest bulk deformation with $0 < Wm < 1$.

The above two methods of investigation allow qualitative study of the degree of non-coaxiality. In order to quantify the degree of non-coaxiality, the equations governing the rotation of rigid objects in a flowing viscous medium were used (Jeffery, 1922; Ghosh and Ramberg, 1976; Passchier, 1987). For any flow regime with $Wk < 1$, not all rigid particles are free to rotate continuously. Particles with an aspect ratio above a certain critical value will rotate until they reach a stable orientation. For aspect ratios less than the critical value, rotation is unrestricted (fig. 13). The value of the aspect ratio, R_c, that divides freely rotating

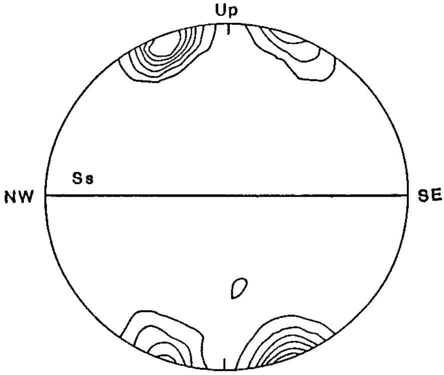


Fig. 11. Orientation of poles to shear bands with respect to the Ss foliation. Shear bands form as conjugate sets suggesting a lower degree of non-coaxiality than simple shear, but there is an asymmetry in the density of the data indicating a top to the northwest sense of shear; that is, $0 > Wm > 1$. $N = 75$. Each data point represents the average of 4 to 5 readings. Contoured using an algorithm following the suggestions of Fisher, Lewis, and Embleton (1987).

objects from those that have reached a stable orientation is a function of the degree of non-coaxiality only:

$$Wm = \frac{R_c^2 - 1}{R_c^2 + 1}$$

(Passchier, 1987).

Samples of augen gneiss containing abundant albite porphyroclasts were selected, and the rotation angles of the porphyroclasts measured with respect to the mesoscopic foliation. The results for one of the samples (MS63) are shown in figure 13B. There is a clear distinction between the group of measurements that scatter across a wide range of orientations and those at higher aspect ratios, which have a much more restricted range of possible orientations (fig. 13B). The change in behavior takes place at around $R = 2.4$. We interpret this as the critical value for R separating those porphyroclasts that have rotated without restriction from those that have attained a stable orientation. A value of $R_c = 2.4$ is equivalent to $Wm = 0.7$ and represents a significant departure from simple shear ($Wm = 1$) within the zone of distributed Ds shear.

Quartz c-axis fabrics, shear band orientation data, and the orientations of rotated porphyroclasts all suggest that Ds was associated with a degree of non-coaxiality between simple and pure shear. The consistency of these kinematic indicators throughout the zone of Ds deformation suggests that deformation was approximately time constant. With estimates of volume change and finite strain it is therefore possible to

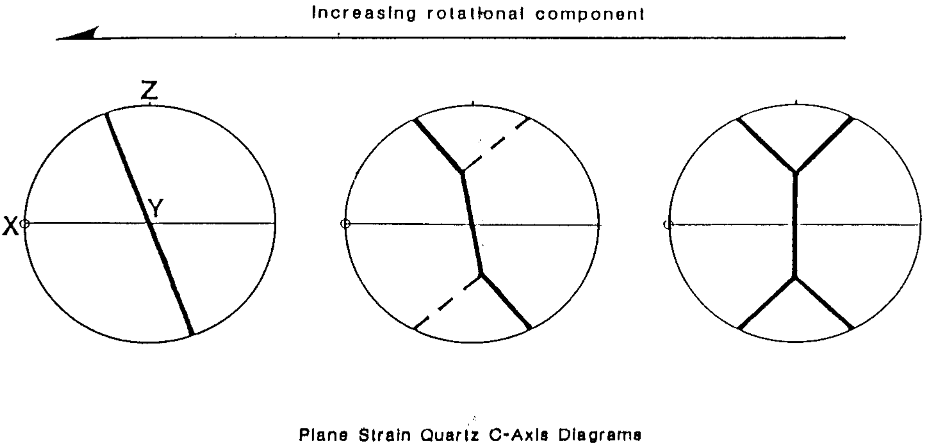
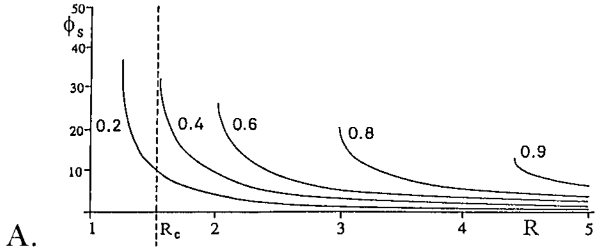


Fig. 12. Schematic quartz c-axis diagrams for plane strain deformation (after Schmid and Casey, 1986). With an increase in Wm there is a progressive change in the topology of the associated c-axis fabric.

quantify the amount of shortening across the zone of distributed Ds shear.

Finite strain was estimated from the grain shape of deformed quartz grains. These are recrystallized grains and only indicate a minimum estimate. The results suggest $X/Z > 4$ (fig. 14). Another indication of finite strain is given by the orientation of fold axes with respect to the stretching lineation. In regions of low finite Ds strain structurally above the zone of penetrative Ds shear, the Ds stretching lineation and fold axis locally show an angular separation of as much as 45°. In contrast, throughout the region of Ds distributed shear this angle does not exceed 10°. The change in angle indicates a minimum strain increase of $Rf = 5.5$. Volume change during deformation is difficult to quantify. However, the lack of prominent veining and pressure solution seams and the microstructural evidence for intracrystalline plasticity being the dominant deformation mechanism imply that the volume change was small. Using these estimates the amount of shortening across the zone of Ds deformation can be calculated. These calculations can be most easily performed using a Mohr circle construction (Passchier, 1988a, b; Wallis, 1992). An analytical solution can also be derived that gives the relationship between the values of Rf, Wm, and the stretch perpendicular to the shear zone (S) as:

$$S = \left[\frac{1}{2}(1 - Wm^2)\left[\left(Rf + Rf^{-1} + \frac{2(1 + Wm^2)}{(1 - Wm^2)}\right)^{1/2} + (Rf + Rf^{-1} - 2)^{1/2}\right]\right]^{-1} \quad (2)$$



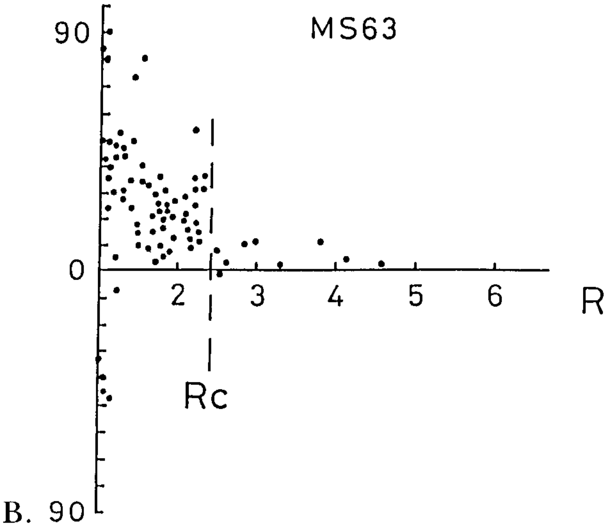


Fig. 13(A) Curves showing stable orientation of rigid particles in a viscous flowing matrix for different degrees of non-coaxiality (Wm = 0.2, 0.4, 0.6, 0.8, 0.9). R = aspect ratio of particles, ϕ_s = calculated stable orientation. The boundary between aspect ratios of particles that rotate continuously and those that attain a stable orientation is shown by a dashed line for Wm = 0.4. This is the critical aspect ratio, Rc, which is a function of Wm only. The relationship is given by:

$$\mathrm{Wm} = \frac{R_c^2 - 1}{R_c^2 + 1}$$

(Passchier, 1987).

(B) Stable orientation analysis using rotated albite porphyroclasts in sample MS63. R_c is interpreted to be close to R = 2.4, which corresponds to a value of Wm = 0.7.

Using a value of Rf = 5 and Wm = 0.7 gives a shortening of around 45 percent within the zone of Ds deformation (Wallis, ms).

Evidence given above suggests that the zone of Ds deformation was subhorizontal at the time of deformation and that strain compatibility requirements are likely to hold. With these assumptions, shortening across the zone of Ds shear at the base of the Austroalpine nappes should be reflected in extension of the higher structural units. Two lines of evidence support this prediction: (1) A number of discrete Ds shear zones can be mapped in the Austroalpine basement immediately above the zone of distributed Ds shear. The base of the Austroalpine domain is offset and extended by at least one of these zones (fig. 9A). The same Ds shear zones cause retrogression of a Barrovian metamorphic fabric formed during pre Ds thickening in the Austroalpine domain. The Teuchl Shear Zone (TSZ—fig. 9A) separates greenschist facies material above from amphibolite facies below (Hoke, ms). Even at elevated geothermal gradients this indicates the removal of several kilometers of structural section from along the path of this zone. (2) Unconformably overlying a pile of previously deformed and thickened units there are a series of sedimentary units, the Gosau Units. Locally these basins can be shown to be bounded by extensional faults (Krohe, 1986; Ratschbacher and others, 1989). The formation of at least some of the Gosau basins can therefore be related to extension at the highest structural level of the evolving mountain belt (Wallis, ms; Ratschbacher and others, 1989).

The Ds fabric can be traced throughout a large part of the southern Tauern Window and has been interpreted by most workers as the result of imbrication and crustal thickening (for example, Bickle and Hawkesworth, 1978; Ratschbacher and others, 1989). The above results suggest, however, that Ds actually formed as a result of extensional tectonics.

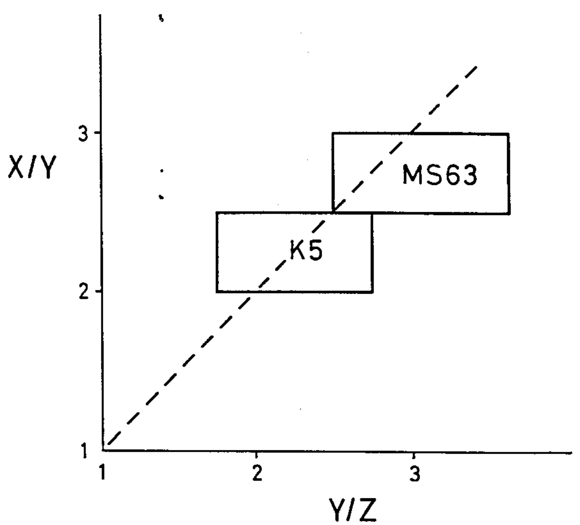


Fig. 14. Finite strain estimates from quartz grains of two samples MS63 and K5 used for vorticity analyses. This gives a minimum estimate of finite strain of X/Z greater than about 4. The results for MS63 are given in figure 13.

Extension controlled by the geometry of an accretionary wedge will develop in the line of maximum slope. However in the Eastern Alps the extension direction is dominantly northwest-southeast, oblique to the trend of the belt (Platt and others, 1989; Ratschbacher and others, 1989). It is intuitively likely that this is related to oblique convergence. However, the 3-dimensional modelling of the dynamics of orogenic wedges that is required to test this idea has yet to be fully developed. The sense of shear during extension is not tightly constrained. However, the dominant sense will be away from the buttress of the accretionary prism. During Ds deformation a top to the northwest sense of shear causes emplacement of the Austroalpine nappes over the Pennine domain. The above analysis shows that in contrast to the traditional view this nappe emplacement is likely to be related to regional thinning not thickening.

Pre-Ds fabrics related to prograde metamorphism are also preserved in the study region (Wallis, ms). Analogous fabrics have also been reported elsewhere in the Austroalpine domain (Krohe, 1986; Schmid and Haas, 1989; Ratschbacher and others, 1989). Where they are developed the associated kinematic indicators are very variable and no clear sense can be determined. At the leading edge of the accretionary complex many thrust relationships are preserved (for example, Tollmann, 1976). These are mainly younger features than Ds and may be related to crustal thickening during continental collision.

Timing of extension.—A critical aspect of the weak orogenic wedge model as proposed by Platt (1986) is that substantial extension can take place during convergence, before collision and major emergence of the mountain belt. Convergence in the Alpine system began in early to mid-Cretaceous (for example, Frisch, 1979; Dewey and others, 1989). Continental collision in the Alps can be dated by the onset of deformation in the European continental basement, which occurred at around 40 Ma in central Switzerland (Milnes and Pfiffner, 1977). At high structural levels in the Pennine domain of the Eastern Alps the peak of a largely static metamorphic event, the Tauern metamorphism, can also be dated at around 35 to 40 Ma (Lambert, 1970; Hawkesworth, 1976; Waters, ms). The growth of metamorphic minerals associated with this metamorphism overprints Ds structures. Ds therefore took place before 40 Ma and before collision. The timing implied by these overprinting relationships is confirmed by radiometric dating from within the zone of distributed Ds shear in the base of the Austroalpine units. The dates obtained from K-Ar and Rb-Sr methods range from 75 to 40 Ma (Oxburgh and others, 1966; Brewer, 1969; Hawkesworth, 1976; Waters, ms; Wallis and Bickle, unpublished), that is, pre-Tauern metamorphism and post Barrovian metamorphism in the Austroalpine nappes. The main sedimentation of the Gosau beds took place in the time interval 80 to 40 Ma (Tollmann, 1976, 1977; Leiß, 1990) and is therefore contemporaneous with the ductile extension during Ds at lower structural levels.

The evidence presented above shows that regional subvertical shortening of several kilometers took place in the Austroalpine nappes during a period of overall plate convergence.

CONTROLS ON EXTENSION

Not all destructive margins need be associated with regional extension. The Eastern Alps and the Calabrian Arc are examples of a more general dynamic situation representing the interplay between forces due to plate convergence and those generated by gravity within the region of thickened crust. In the Southern Uplands of Scotland, for example, there is no evidence for the regional exhumation of high P/T rocks, and this may be an example of a convergent margin that did not experience extension during convergence. What are the major controls that decide whether extension will occur or not?

The principal condition necessary for extension during convergence in an accretionary wedge is that accretion of new material occurs mainly by underplating. Significant frontal accretion will always maintain the wedge in a state of longitudinal compression and require shortening, not extension in the interior (Platt, 1988). Controls on underplating include the following.

1. The thickness of the sedimentary section on the lower plate entering the subduction zone;
2. the mechanical properties of the incoming sedimentary section;
3. the rate of subduction, which may influence the rate of dewatering of the sediments and their thermal history.

A thick sedimentary section on the underthrust plate is likely to be comparatively well-consolidated and lithified near the base and may contain significantly overpressured horizons. This allows decollement within the section rather than at its base and underthrusting and underplating of the lower part of the section. Seismic evidence for this process comes from a number of active accretionary wedges, such as the Makran (White and Louden, 1982), Barbados Ridge (Westbrook and others, 1982), Nankai Trough (Karig, 1986), and Middle America Trench (Sample and Moore, 1987).

Rapid subduction causes rapid loading of the sedimentary section, which again favors the development of overpressured horizons. It also decreases the thermal gradient in the subducted slab, thereby delaying the onset of dehydration reactions and temperature-sensitive ductile deformation mechanisms, both of which are likely to lead to disruption and accretion of the underthrust section.

Other factors that may affect the likelihood of extension are changes in the rate of subduction and erosion or deposition in the rear of the wedge. If the basal shear stress is generated by a purely frictional mechanism, it should be independent of the rate of subduction. It may well be a consequence of plastic deformation in overpressured clay along a decollement horizon; however, in this case, the basal shear stress will be a function of the relative velocity, and a decrease in the rate of subduction would favor extension.

CONCLUSIONS

Modelling of convergent tectonic margins as weak wedge-shaped continuous media has been successful in predicting that mountain belts may undergo a prolonged phase of extension during their convergent history. We suggest generally applicable methods for recognizing such extension and its timing and demonstrate the practical application of these methods in two mountain belts. Flow path modelling in particular has the potential to identify ductile extension in broad zones of ductile shear. The timing and amount of subvertical shortening within both the Eastern Alps and the Calabrian Arc suggest that the exhumation of high density high P/T metamorphic rocks in these regions can be adequately explained by such wedge mechanics.

When recognizing extension two important factors should be borne in mind: (1) If it is related to wedge mechanics, extension can occur before convergence ceases, and (2) extension may be a localized phenomenon. It is necessary to show that it occurred contemporaneously at a variety of different structural levels and that the maximum shortening direction was subvertical.

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