Transpression and transtension zones

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Abstract: Transpression and transtension are strike-slip deformations that deviate from simple shear because of a component of, respectively, shortening or extension orthogonal to the deformation zone. These three-dimensional non-coaxial strains develop principally in response to obliquely convergent or divergent relative motions across plate boundary and other crustal deformation zones at various scales. The basic constant-volume strain model with a vertical stretch can be modified to allow for volume change, lateral stretch, an oblique simple shear component, heterogeneous strain and steady-state transpression and transtension. The more sophisticated triclinic models may be more realistic but their mathematical complexity may limit their general application when interpreting geological examples. Most transpression zones generate flattening ($k < 1$) and transtension zones constrictional ($k > 1$) finite strains, although exceptions can occur in certain situations. Relative plate motion vectors, instantaneous strain (or stress) axes and finite strain axes are all oblique to one another in transpression and transtension zones. Kinematic partitioning of non-coaxial strike-slip and coaxial strains appears to be a characteristic feature of many such zones, especially where the far-field (plate) displacement direction is markedly oblique (<20°) to the plate or deformation zone boundary. Complex foliation, lineation and other structural patterns are also expected in such settings, resulting from switching or progressive rotation of finite strain axes. The variation in style and kinematic linkage of transpressional and transtensional structures at different crustal depths is poorly understood at present but may be of central importance to understanding the relationship between deformation in the lithospheric mantle and crust. Existing analyses of obliquely convergent and divergent zones highlight the importance of kinematic boundary conditions and imply that stress may be of secondary importance in controlling the dynamics of deformation in the crust and lithosphere.

Transpression (TP) and transtension (TT) (Harland 1971) occur on a wide variety of scales during deformation of the Earth's lithosphere. On the largest scale, this is an inevitable consequence of relative plate motion on a spherical surface: plate convergence and divergence slip vectors are not commonly precisely orthogonal to plate boundaries and other deformation zones. Plate boundary zones will, therefore, experience oblique relative motions at some time during their history along some part of their length (Dewey 1975). Within a plate boundary zone, strain is focused generally into displacement zones that bound units of less-deformed material on several scales. This is particularly evident in continental orogens, where broad deformation belts develop in which fault- and shear-zone-bounded blocks partition strains into a series of complex displacements, internal strains and rotations in response to far-field plate tectonic stresses and large-scale body forces (Dewey *et al.* 1986). Here again, block convergence and divergence slip vectors are not commonly precisely orthogonal to or parallel to plate margins or to smaller-scale deformation zone boundaries. In many cases, this arises because block margins are inherited features that act as zones of weakness, repeatedly reactivated during successive crustal strains, often in preference to the formation of new zones of displacement (Holdsworth *et al.* 1997). Any displacement zone margin that is significantly curvilinear or irregular is bound to exhibit oblique convergence and/or divergence unless it follows exactly a small circle of rotation. In addition to collisional orogenic belts, TP and TT occur widely in a large range of other tectonic settings: oblique subduction margins in the forearc (TP), arc (TP and TT) and back-arc (TT) regions; 'restraining' (TP) and 'releasing' (TT) bends of transform and other strike-slip displacement zones; continental rift zones (TT), especially during the early stages of continental break-up and formation of new oceanic lithosphere; during late orogenic extension (TT) and in slate belts (TP), where deformation may be accompanied by large-scale volume loss. This paper sets out to introduce some of the basic features of transpressional and

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transtensional deformation zones and to highlight briefly some important problems that may be encountered in such regions and their implications for geologists. Our review is not exhaustive and is intended to set the scene for the papers in this volume.

Properties of transpression and transtension zones

Basic definitions

Relative plate, or block, motion vectors and the orientation of the plate or block margins are the principal boundary conditions during lithospheric deformation. Therefore, in contrast to the original definitions discussed by Harland (1971), we suggest that the terms *oblique convergence* and *oblique divergence* should be used to indicate the relative motion of boundary plates or blocks. These are measured using the angle α between the horizontal far-field (plate) displacement vector and the boundary of the deformation zone (Fig. 1; Tikoff & Teyssier 1994). We suggest that the terms *transpression* and *transtension* be restricted to the resulting combinations of non-coaxial and coaxial strains. Thus transpression and transtension can be defined as *strike-slip deformations that deviate from simple shear because of a component of, respectively, shortening or extension orthogonal to the deformation zone* (Fig. 1). The deformation zone is commonly steeply dipping or sub-vertical.

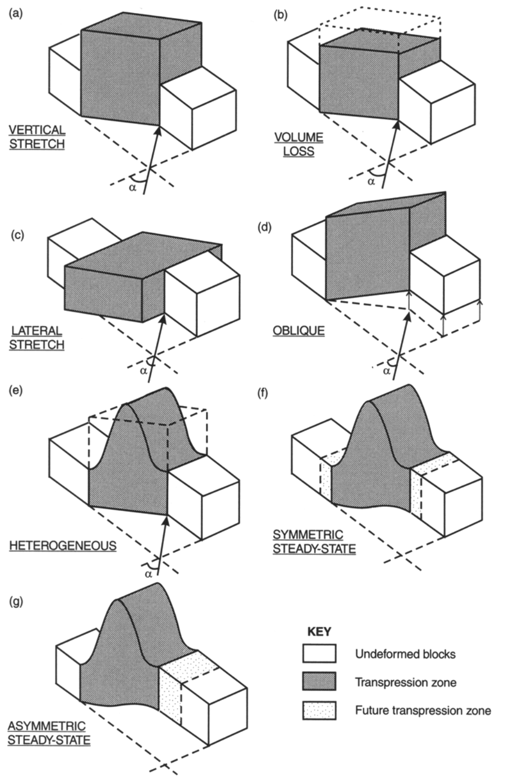

Fig. 1. Some examples of transpressional strain models. (Note that the arrows and angle α are omitted in (**f**) and (**g**) for simplicity.)

Strain patterns

Existing studies have modelled transpression and transtension zones using either finite and incremental strain (e.g. Sanderson & Marchini 1984; Fossen & Tikoff 1993; Tikoff & Teyssier 1994) or strain rate (e.g. Ramberg 1975). Models based on strain have been shown to be effective starting points in the analysis of three-dimensional deformation zones and examples are shown in Fig. 1a–g (for transpression zones), ordered approximately in terms of increasing realism and complexity. The basic model of Sanderson & Marchini (1984) (e.g. Fig. 1a) involves a constant-volume, homogeneous strain in a vertical zone in which horizontal shortening (or extension) across the zone was accommodated by vertical extension (or shortening). It is possible to modify this basic model by changing each of the boundary conditions to allow for volume change (e.g. Fig. 1b; Fossen & Tikoff 1993), lateral stretch (Fig. 1c; Dias & Ribeiro 1994; Jones *et al.* 1997) or an oblique simple shear component in which the bounding blocks are displaced vertically and laterally (Fig. 1d; Robin & Cruden 1994; Jones & Holdsworth this volume). All these homogeneous strain models are idealized compared with the strain patterns in naturally occurring shear zones. In particular, the deformation zone boundaries are unlikely to be unconstrained, as they cannot simultaneously allow free slip in all directions while transmitting the shear stress imparted by the component of simple shear (Schwerdtner 1989; Robin & Cruden 1994). A more realistic model would display a strain gradient developed from zero slip at the boundary walls to maximum vertical extension in the centre of the zone, yielding a heterogeneous strain (Fig. 1e; Robin & Cruden 1994). Unfortunately, such models are extremely complex and do not easily allow generalizations to be made concerning finite strains; therefore, they may be of limited use in the analysis of most naturally deformed rocks. In all these models, the deformation zone has a fixed width, so the strain rates will exponentially increase (transpression) or decrease (transtension). Further variants are steady-state transpressional or transtensional models, either symmetrical (Fig. 1f) or asymmetrical (Fig. 1g), that yield a constant strain rate (Dutton 1997).

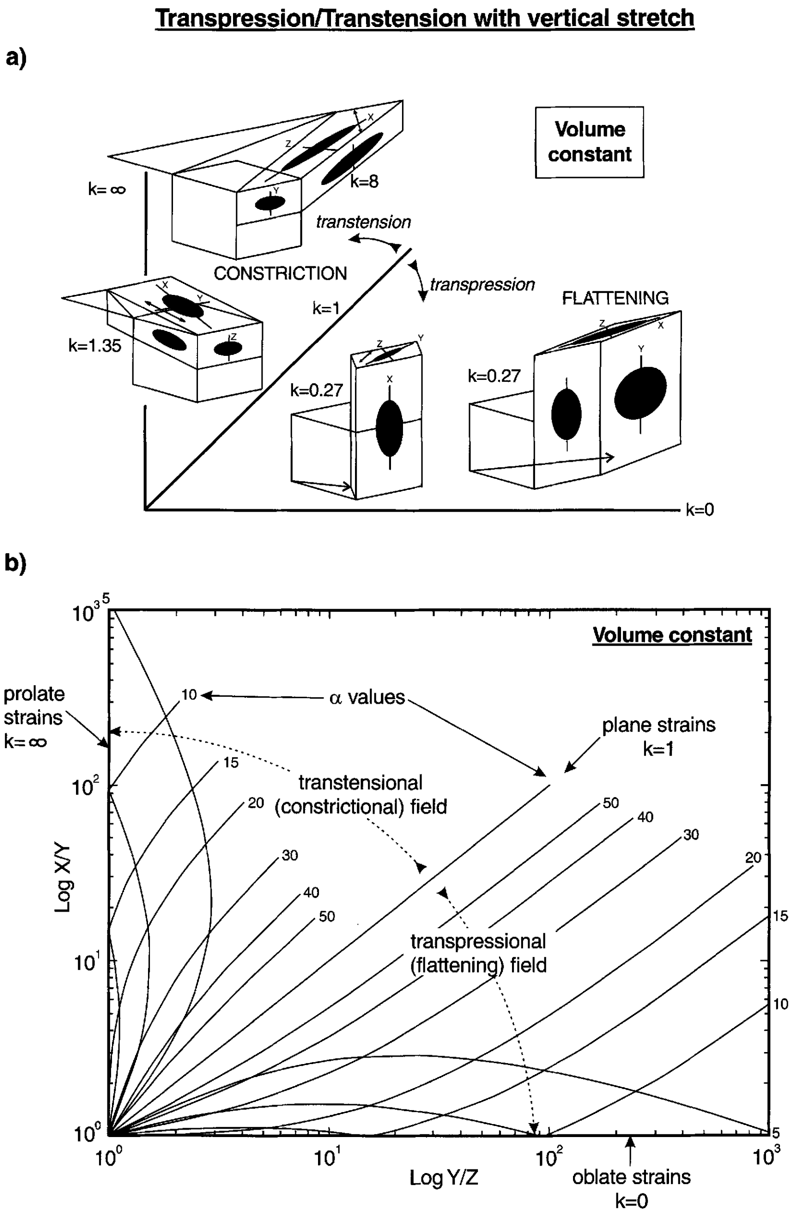

Fig. 2. (**a**) Flinn plot to illustrate examples of transpressional flattening and transtensional constrictional deformations. The double-headed arrows are incremental stretching directions. (**b**) Logarithmic Flinn plot to illustrate some transpressional (flattening) and transtensional (constrictional) strain paths.

This is important because the correlation of flow parameters, such as vorticity, with finite strains will be valid only if the deformation is steady state (Jiang & White 1995).

In the more straightforward strain models for transpression and transtension (e.g. Fig. 1a–c), one of the principal axes of finite strain remains fixed and vertical as deformation progresses, and the other two axes rotate in the horizontal plane because of the non-coaxial wrench simple shear component. This produces strains that, in common with simple shear, have a monoclinic symmetry. In contrast, more complex models, requiring an oblique simple shear component (e.g. Fig. 1d–g) generally have a triclinic symmetry in which all three axes of finite strain rotate relative to a fixed external reference frame.

Figure 2a and b shows examples of finite

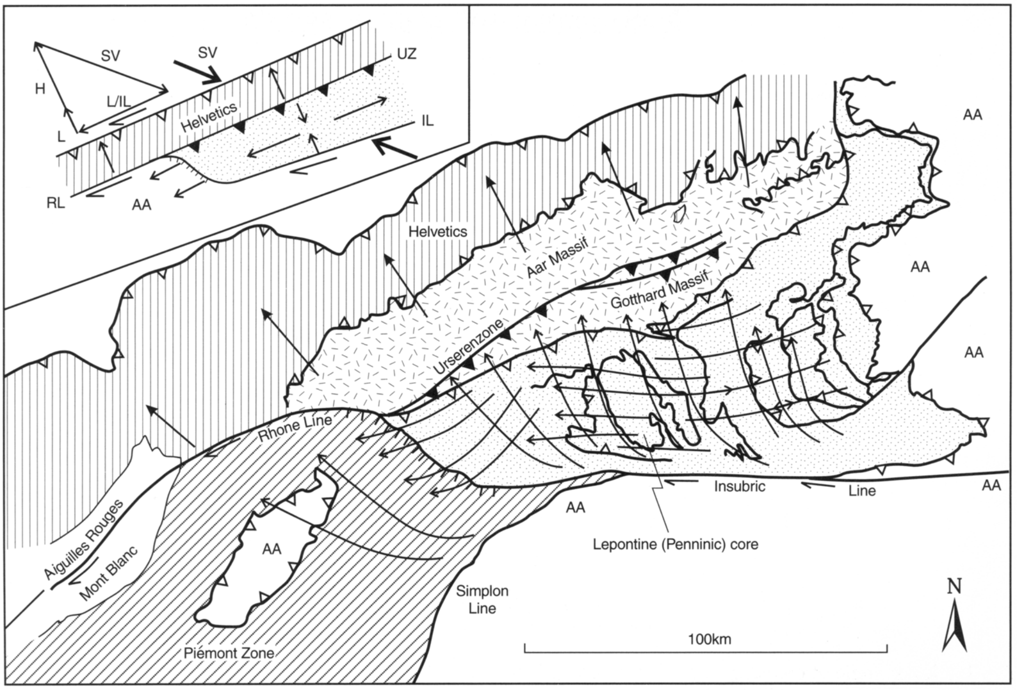

Fig. 3. Simplified tectonic map of the Central Alps. Lines with open arrowheads are Eocene–Oligocene lineations; lines with filled arrowheads are Miocene lineations; AA, Austro-Alpine zones. Inset map shows a schematic illustration of the Miocene trans-Alps slip vector partitioning (H, Helvetics; L, Lepontine; IL, Insubric Line; RL, Rhône Line; UZ, Urserenzone; SV, slip vector).

strains and strain paths using the basic constant-volume, vertical stretch model of Sanderson & Marchini (1984) (e.g. Fig. 1a). These show that transpression generates flattening ($k < 1$) and transtension constrictional ($k > 1$) bulk strains. This is also generally the case in constant-volume deformation zones where lateral extrusion (Fig. 1c) or oblique simple shear (Fig. 1d) or heterogeneous transpression (Fig. 1e) occur, although the range of k values is reduced, clustering increasingly towards plane strains as the lateral stretch or vertical simple shear components of finite strain are increased, respectively (Robin & Cruden 1994; Jones *et al.* 1997; Jones & Holdsworth this volume). However, it is possible to develop constrictional or prolate strains in transpression zones and flattening or oblate strains in transtension zones in special circumstances. For example, constrictional strains will develop in transpression zones where there is a component of lateral stretch and vertical shortening (or volume loss) (e.g. Dias & Ribeiro 1994; Fossen & Tikoff this volume). A good example of this developed during Miocene oblique-dextral convergence in the Swiss Alps (Fig. 3), where partitioning developed between roughly zone-orthogonal Helvetic thrusting and zone-parallel dextral slip on the Insubric Line, which passed northwestwards into the extensional Simplon Line and then westward into the dextral Rhône Line. During dextral slip, the Lepontine Alps were extensionally unroofed and vertically shortened simultaneously with north–south horizontal shortening and east–west extension, thus leading to a bulk constrictional deformation in a plate-boundary zone of oblique convergence.

In constant-volume transpression or transtension zones with a vertical stretch, finite strain paths at low angles of convergence ($\alpha < 20°$; 'wrench-dominated' models of Tikoff & Teyssier 1994) are strongly non-linear and complex with 'bouncing' off $k = 0$ and $k = \infty$ axes (Fig. 2b). This coincides with a switching (or 'swapping') in the orientation of the finite strain axes (e.g. Fig. 4; Sanderson & Marchini 1984; Tikoff & Teyssier 1994; Tikoff & Greene 1997). In transpression zones, an initially horizontal x-axis swaps orientation with the vertical y-axis with increasing finite strain, whereas in transtension zones, the y-axis swaps with an initially vertical z-axis.

Far-field plate slip vectors will be parallel to the axes of instantaneous or finite strain in a

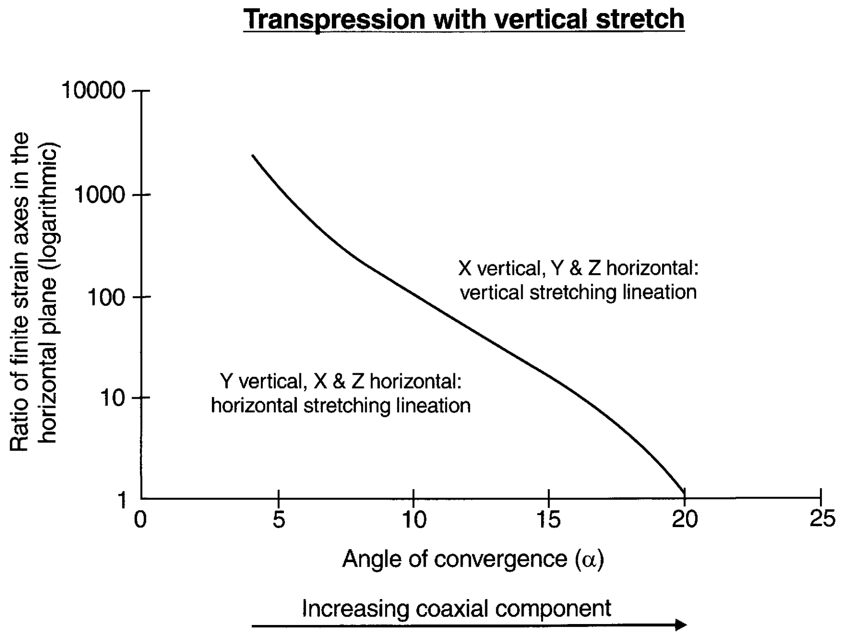

Fig. 4. Plot of angle of convergence (α) v. horizontal finite strain ellipse ratio for homogeneous transpression (after Tikoff & Greene 1997, fig. 5). A line of oblate strain separates the field in which the long axis of finite strain is horizontal from the field in which it is vertical.

deformation zone only when they are oriented orthogonally to the zone boundaries (α = 90°). Thus, in all zones of transpression and transtension, plate motions, instantaneous strain (or stress) axes and finite strain axes are oblique to one another (e.g. see top of blocks in Fig. 2a). However, the horizontal plate motion will always correspond to one of three flow apophyses that define the maximum, intermediate and minimum rates of particle movement in the zone (Fossen *et al.* 1994). Such flow apophyses critically control the passive rotation of marker structures in the deformation zone even though these structures may have formed initially in response to instantaneous or finite strains. This emphasizes that the relationships between plate motions and resulting strain patterns are complex.

Strain partitioning

Strains may be kinematically non-partitioned or partitioned in transpression and transtension zones (Fig. 5a and b). There are many ways in which partitioning may occur (e.g. Fig. 3 and 6a–f; see also Oldow *et al.* 1990). Fitch (1972) showed how oblique plate convergence in the Indonesian arc from Java to the Andaman Islands is partitioned between orthogonal subduction and intra-arc (or volcanic axis) strike-slip, with the strike-slip component becoming more important northwestwards. In the Andes, a fairly constant 100 mm/a ENE slip vector between the Nazca and South American Plates is accommodated and partitioned in various ways along the mountain belt (Dewey & Lamb 1992). Depending upon the sense (sinistral or dextral) of obliquity, left-lateral or right-lateral intra-arc strike-slip faulting, mainly along the weaker arc volcanic zone (where present), is combined with orthogonal Benioff Zone slip and/or back-arc thrusting.

Molnar (1992) has provided the best rationale yet for plate boundary-scale partitioning. Molnar's argument is that in a strong, yet ductile, continental upper mantle, a viscous continuum should generate principal stresses and strain rates parallel to or perpendicular to the Earth's surface. Because upper-crustal block rotation about vertical axes is coupled, via a viscous lower crust, to the vorticity field of the upper mantle, oblique-slip faults should not be stable during bulk finite strain. Strain partitioning may also be facilitated by the reactivation of pre-existing structural weaknesses that are in suitable orientations to minimize work done (e.g.

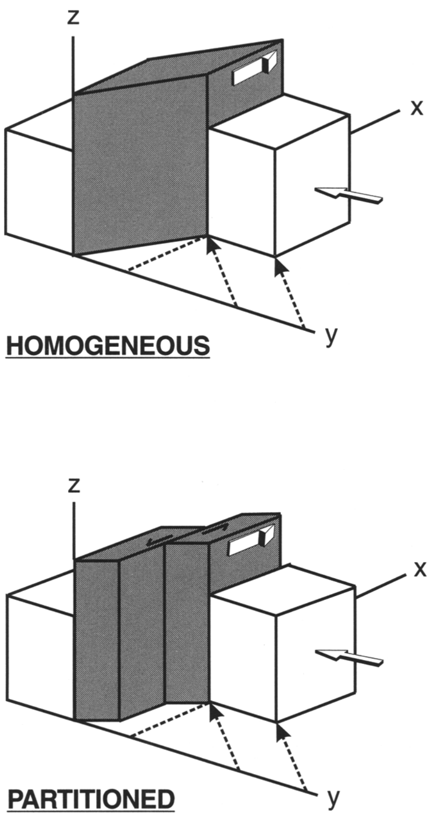

Fig. 5. (**a**) Non-partitioned transpression. (**b**) Partitioned transpression in which a significant component of the wrench component is accommodated by a discrete strike-slip fault.

Jones & Tanner 1995) or at angles at which the coefficient of sliding friction is less than the coefficient of internal friction for new faults. Michael (1990) demonstrated that the rate of energy release on oblique-slip faults is greater than that on pairs of strike-parallel dip-slip and strike-slip faults, by invoking Hamilton's Principle that systems choose configurations that minimize work done. This could apply, particularly, to simple bimodal partitioning at island arcs where a very weak zone along the magmatic axis of the arc represents a suitable surface to accommodate the strike-slip component. Platt (1993) showed, by modelling a variety of accretionary-prism rheologies, that a linear viscous model generates a continuously partitioned strain from orthogonal near the Benioff Zone with a rapidly increasing strike-slip component near the back-stop. However, a plastic rheological model demonstrated strong dependence upon the angle of convergence; at high angles, there is little or no partitioning, whereas at low angles, there is strong partitioning into thrust and strike-slip components. This is well illustrated by the low degree of partitioning in the South Island of New Zealand and the strong partitioning along the San Andreas zone, where obliquity is, respectively, high and low (Teyssier *et al.* 1995). Tikoff & Teyssier (1994) further suggested that partitioning is especially likely where the angles of convergence or divergence are low (<20°) because of non-parallelism of incremental and finite strain axes.

Fabric patterns and structural style

The angle of obliquity (α), intensity of finite strain and degree of kinematic partitioning principally control fabric orientation in both transpression and transtension zones (e.g. Fig. 6a–f; McCoss 1986; Fossen & Tikoff 1993; Fossen *et al.* 1994; Tikoff & Teyssier 1994; Tikoff & Greene 1997). In monoclinic transpression zones where the bulk deformation follows the vertical stretch model of Sanderson & Marchini (1984), the strikes of the principal flattening surface (cleavage, schistosity, gneissosity) will vary with the non-coaxial component of the strain (Fig. 6a–f) but the dip is always vertical. Characteristically, associated stretching lineations can switch from vertical to horizontal (Fig. 4) where α <20° ('wrench-dominated transpression'). If no kinematic partitioning occurs, stretching lineations will be initially horizontal but, as the finite strain increases, they will switch to vertical; the exact strain threshold required depends on the value of α (Fig. 4). If, however, the wrench non-coaxial component is kinematically partitioned into boundary-parallel high-strain zones, vertical lineations may switch back into sub-horizontal orientations. Possible field examples of such switching behaviour have been described by Hudleston *et al.* (1988), Holdsworth (1989) and Tikoff & Greene (1997). Conversely, in transtension, the stretching direction is always horizontal, but the shortening direction switches from vertical to horizontal where $\alpha < 20°$ with increasing finite strain (McCoss 1986). Thus, if no kinematic partitioning occurs, any planar fabric switches from vertical to horizontal with increasing strain or vice versa if boundary-parallel wrench simple shear-dominated high-strain zones form as a result of partitioning. Switching behaviour is suppressed in transpression and transtension zones where there is a significant component of lateral stretch and lineations are mainly horizontal (e.g. Jones *et al.* 1997).

The relationship between planar and linear fabrics in triclinic transpression and transtension

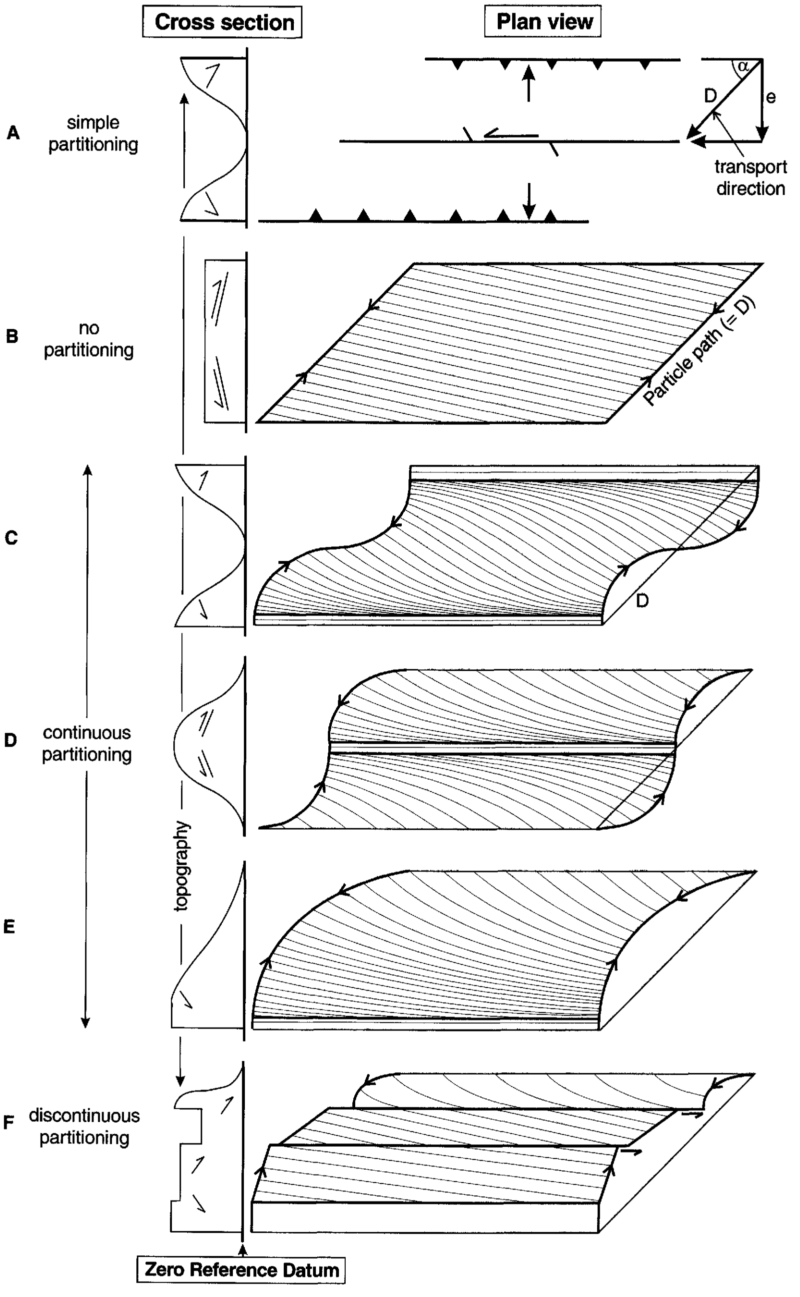

Fig. 6. Plan view of six transpressional zones showing various degrees and styles of partitioning, particle paths in the transpression zone (arrows on sides) and cleavage traces (fine lines). D, tectonic transport direction; α, angle between D and zone margin; e, zone-normal direction. Topographic profiles and possible gravity-driven thrust movements (half arrows) are indicated in the cross-sections. (**a**) Simple partitioning into marginal thrust belts and a central strike-slip zone. (**b**) A non-partitioned zone of oblique cleavage (particle paths parallel to D). (**c**) Continuous symmetrical partitioning of coaxial and non-coaxial components towards the edges and centre of the zone, respectively. (**d**) Continuous symmetrical partitioning of the coaxial and non-coaxial components towards the centre and edges of the zone, respectively. (**e**) Continuous asymmetrical partitioning of the coaxial and non-coaxial components to either side of the zone. (**f**) Discontinuous partitioning of the coaxial and non-coaxial components.

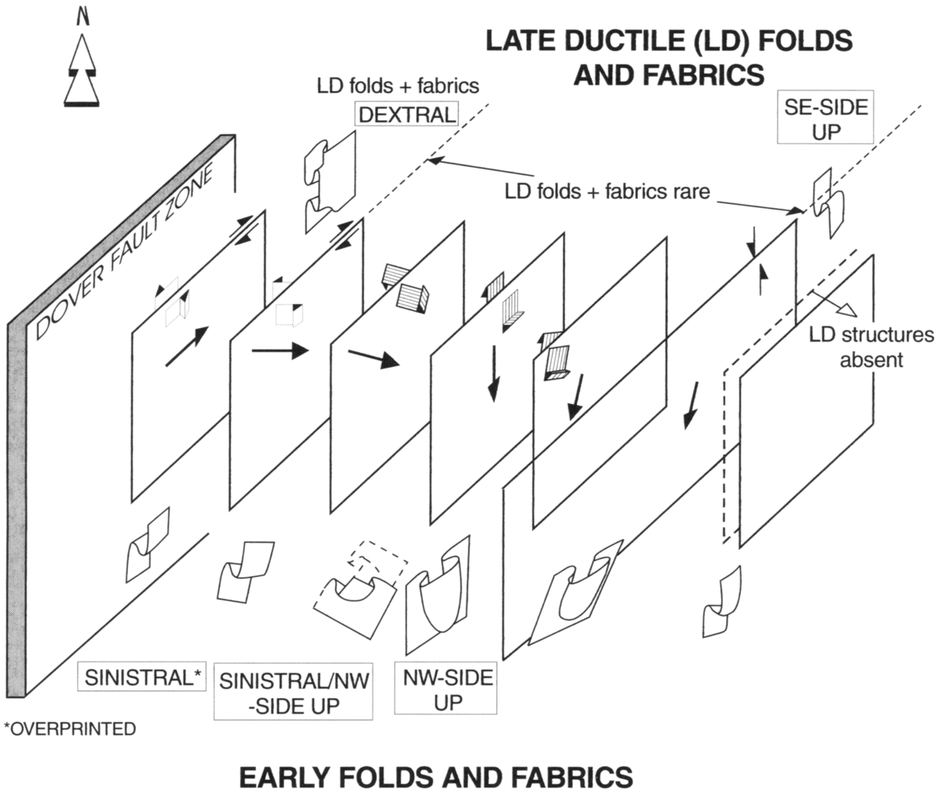

Fig. 7. Three-dimensional diagram showing foliation, lineation, shear sense and associated folding patterns from a transpressional terrane boundary shear zone developed in the Avalon Zone, NE Newfoundland Appalachians (redrawn after Holdsworth 1994). Two phases of ductile transpressional structures, 'early' and 'late ductile' (LD), are recognized, both of which appear to have an overall triclinic symmetry. (Note that in both cases, the wrench component is concentrated adjacent to the present-day terrane boundary, the Dover Fault.)

zones (e.g. Fig. 1d–g) is poorly understood. The preliminary transpressional models of Robin & Cruden (1994) suggest that complex and systematic variations in the orientation of both foliations and lineations will occur depending upon the intensity of finite strain, the obliquity of the simple shear component and the nature of any kinematic partitioning within the deformation zone. This seems to be consistent with the field observations made in steeply dipping transpression zones where there is evidence for a vertical simple shear component (e.g. Fig. 7 based on Holdsworth (1994); other examples have been given by Robin & Cruden (1994) and Goodwin & Williams (1996)). Figure 8a–e illustrates schematically in cross-section what may happen to foliation trajectories in vertically extruding, heterogeneous, triclinic transpression zones with a range of boundary wall morphologies. It should be noted that the broadening or narrowing of the zones will lead to significant changes in both the intensity of finite strain and the strain rate.

The depth variation in the style of transpressional and, especially, transtensional deformation structures and how such systems may be kinematically linked is very poorly understood. The shallower crustal levels are generally characterized by brittle deformation among complex domains of rotating crustal blocks. Exhumed examples of middle-crustal transpression zones display similar configurations except that the blocks may be internally deformed and the block-bounding structures are now ductile shear zones (e.g. Hudleston *et al.* 1988; D'Lemos *et al.* 1992). The nature of transpressional and transtensional structures in the weak ductile lower crust and much stronger ductile upper mantle is uncertain but is of some importance if the coupled model proposed by Molnar (1992) for zones of oblique convergence and divergence is correct. Transpressional slate belts also present

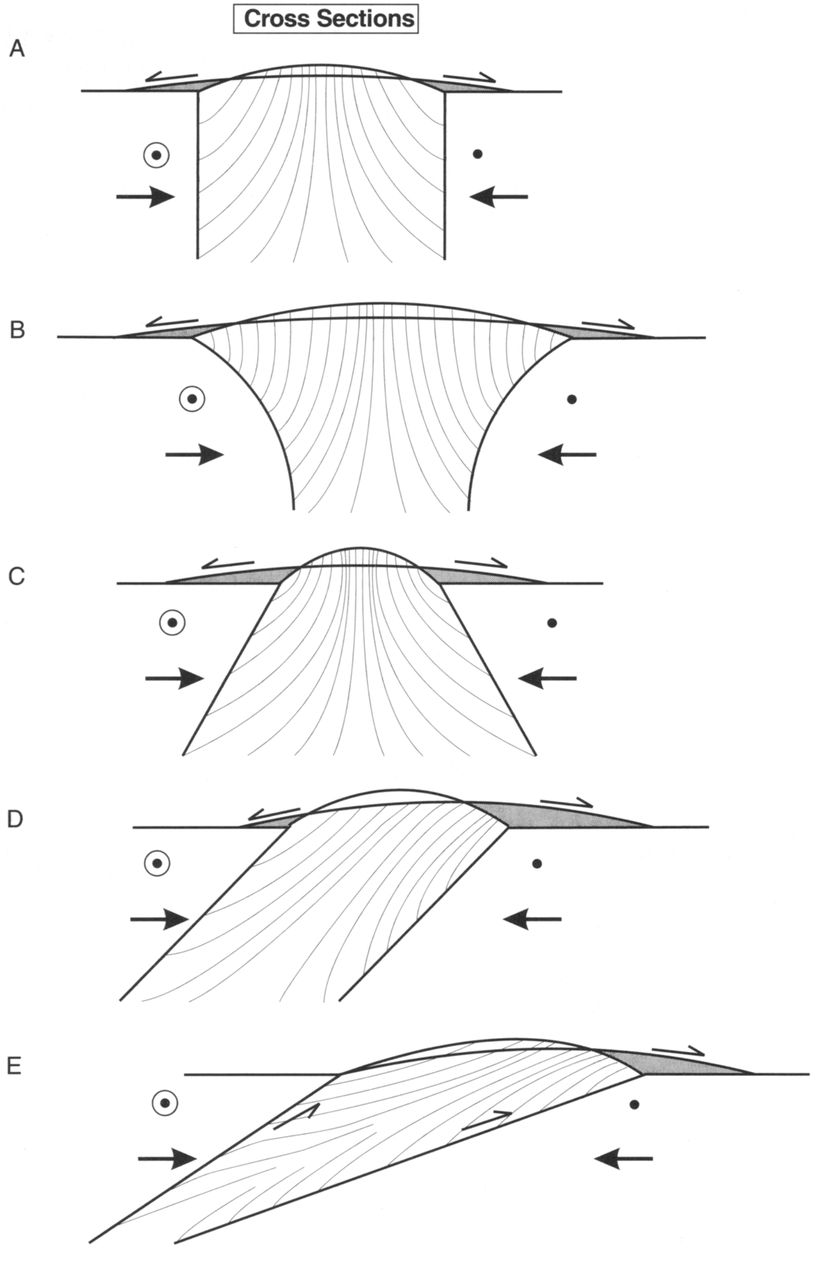

Fig. 8. Transpression with various boundary wall orientations showing possible cleavage trajectories in section. (**a**) Vertical walls with vertical escape. (**b**) Zone widening upwards between inward-dipping walls. (**c**) Zone narrowing upwards between outward-dipping walls. (**d**) Inclined walls dipping in same direction. (**e**) Inclined walls, more steeply dipping extensional detachment above more gently dipping basal thrust. Possible gravity-driven thrust movements are also indicated by the shaded zones and arrows.

a special problem. These are, structurally, high-level anchimetamorphic zones in which transecting cleavages and cleavage sequences are developed often with very large volume losses (up to 55%) because of fluid-assisted diffusive mass transfer mechanisms (e.g. Wood 1973; Cox & Etheridge 1989). It is not yet clear where modern slate belts develop and how they maintain compatibility with subjacent high-grade crustal levels.

Lateral stretch, extrusion and escape

There is some semantic and kinematic confusion in the literature arising from failure to distinguish between lateral extrusion (or lateral escape) and lengthening of a transpression or transtension zone developed in other ways. There are three ways in which lateral stretches may develop in plate boundary or other deformation zones: lateral extrusion, non-coaxial stretching, and radial spreading.

Lateral extrusion. Lateral extrusion (or escape) *sensu stricto* involves a stretch in the horizontal direction that causes the deformation zone to lengthen relative to the undeformed or less-deformed rocks that form the zone margins (e.g. Fig. 1c; Jones *et al.* 1997). Homogeneous lateral extrusion necessarily involves slippage along zone walls with changing slip along strike, but heterogeneous extrusion may be boundary wall compatible where a zone-orthogonal strain gradient exists, although the amount of extrusion will be severely limited. Lateral extrusion, therefore, involves horizontal, along-strike mass movement of material towards the end (or ends) of the zone where compatibility problems of material extruding from the zone are solved by creating space at the end of the zone (Harland 1971; Ramsay & Huber 1987). In theory, this should be kinematically possible at certain types of triple junction or at a transform-bounded subduction zone. These are rare situations and escape is, therefore, extremely unlikely at a large plate-boundary zone scale. However, exceptions are known to exist in some transpressional and transtensional settings at smaller scales, where specific geometric and mechanical boundary conditions and internal rheologies in the deformation zone may favour lateral extrusion. It should be noted that, to apply such models, it is necessary to account for the lateral space problem at the terminations of the deformation zone.

(1) Transpression involving lateral extrusion affects deformed serpentinites intruding a complex suture zone in SW Cyprus (Bailey 1997; Jones *et al.* 1997). In this case, the serpentinites were intruded as irregular and possibly isolated bodies into a pre-existing fault system. Subsequent transpressional deformation then redistibuted the rheologically weak material laterally and, to a lesser extent, vertically within the fault zones. Similar deformation patterns may be common in plutonic bodies emplaced syn-tectonically along pre-existing faults and shear zones in active transpressional arcs.

(2) Small but significant components of deformation that may be attributed to a component of lateral extrusion have been documented in continental strike-slip duplex systems (Laney & Gates 1996) and broader regions of transpressional deformation, e.g. mid-Devonian, central Scotland (Jones *et al.* 1997). The complex and irregular nature of plate and 'block' boundaries and their mutual interaction means that small components of lateral extrusion may be possible in continental deformation zones.

Non-coaxial components of stretching. These can lead to horizontal zone elongation at low angles to the plate or deformation zone boundary in transpression and transtension zones. This stretch is boundary wall compatible and does not lead to extrusion. On a plate boundary scale, the arc-parallel along-strike extension in Sumatra (McCaffrey 1991) and the Aleutians (Ekström & Engdahl 1989), and the transpressional zone-parallel extension of northeast Venezuela that may be an important mechanism for the exhumation of high-pressure–low-temperature metamorphic rocks in accretionary wedges (Avé Lallemant & Guth 1990), are probable examples of this non-coaxial stretching.

Radial spreading. Radial spreading driven by body forces in curved transpressional arcs, accretionary prisms and thrust wedges may lead to zone-parallel plate-boundary scale extension. Good examples occur by radial thrusting in the Himalayas, which leads to orogen-parallel stretching, and the radial back-arc spreading of the Aegean, which is causing arc-parallel extension in the Cretan fore-arc. In such radial systems, body-force driven coaxial strains are combined with the coaxial and non-coaxial strains of transpression caused by relative plate motion across oblique portions of the plate boundary zone. Compatibility is not required along the free-slipping basal thrust or Benioff Zone but is maintained with the zone-orthogonal extending thrust sheet or upper plate.

Transpression, topography and exhumation

Large-scale steeply dipping or vertical transpression zones are likely to develop significant surface topographies both above and at the lateral terminations of the deformation zones. These topographies may be sufficient to generate regionally important episodes of gravity-driven deformation that may aid the exhumation of deep crustal rocks. Variations in strain intensity, kinematic partitioning and orientation of the deformation zone boundaries will all affect

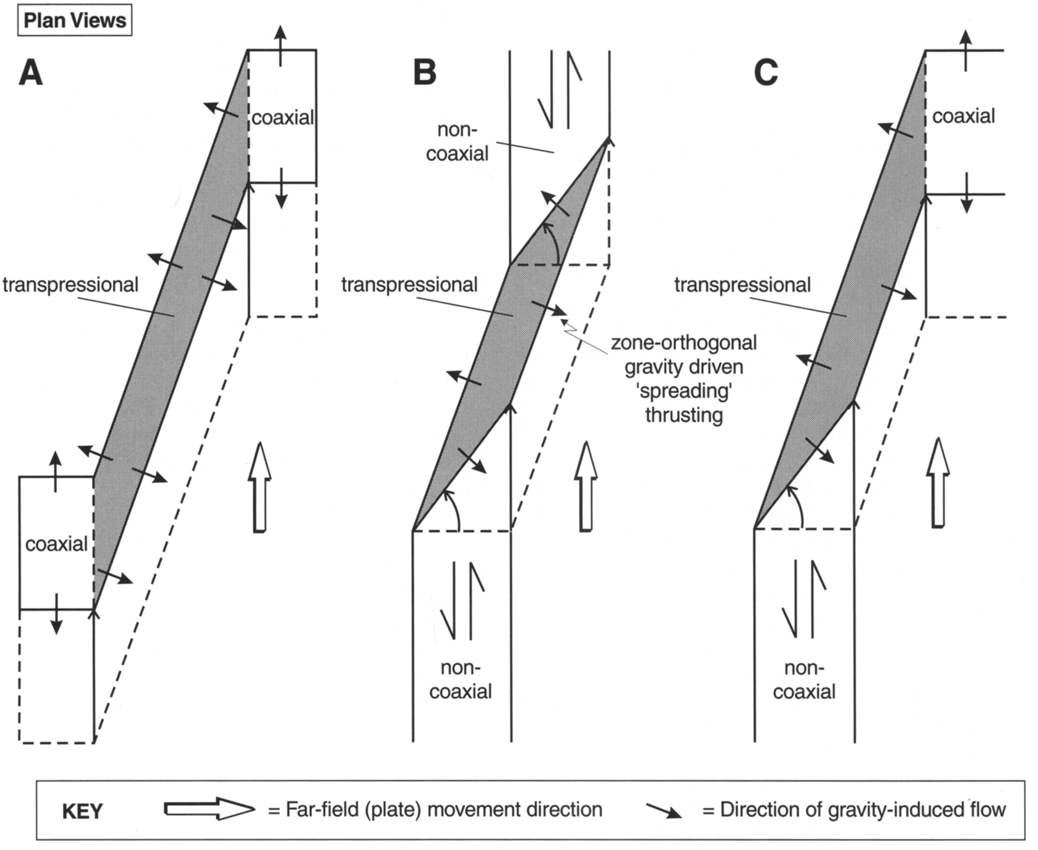

Fig. 9. Three possible modes of termination of transpressional zones. (**a**) Termination in coaxial zones where the zone termination boundaries do not rotate and where there is no compatibility problem between the surface topography of the coaxial and transpressional zones. The only compatibility problem exists where zone orthogonal, gravity-driven thrusting is in different directions at the coaxial–transpressional boundary. (**b**) Termination in non-coaxial zones where the termination boundaries rotate with non-coaxial zone vorticity and where a compatibility problem exists between the growing topography of the transpressional zone and the 'non-topography' of the non-coaxial zone. (**c**) One end terminated by a non-coaxial zone, the other by a coaxial zone. Gravity-driven lateral flow will be towards the non-coaxial zone.

the surface topography of transpression zones and resulting geometries and distributions of gravity-driven deformation (Fig. 6a–f and 8a–e).

The terminations of transpression (and transtension) zones are rarely considered. They may be terminated by coaxial or non-coaxial zones in one of three ways (Fig. 9a–c). Where coaxial zones terminate the transpressional zone (Fig. 9a), the transpressional boundary does not rotate and there are no compatibility problems between transpressional and coaxial zones. Where the transpressional zone is terminated by non-coaxial simple shear zones (Fig. 9b), not only do the non-coaxial–transpression zone boundaries rotate with the non-coaxial rotational sense but there is also a topographic mismatch between the vertically thickened transpression zone and the non-thickened non-coaxial zone that is likely to engender high-level gravity-driven flow. Third, the transpressional zone may be terminated by both non-coaxial and coaxial strain zones, allowing unidirectional gravity-driven flow towards the non-coaxial zone (Fig. 9c).

Implications

All transpression and transtension zones display three-dimensional, non-coaxial strains; existing and future attempts to model their development represent part of a broader and long overdue attempt to deal with geological deformation zones in a realistic manner. There are important practical difficulties. In particular, it is not clear whether strain models need to strive for realism, making them complex and unwieldy, or simplicity, meaning that they may only be applied qualitatively to finite bulk strain patterns.

On a broader note, the strain patterns deduced in transpression and transtension zones have several important implications for the broader geological community:

(1) *Balanced cross-section techniques.* Many quantitative studies of crustal deformation use cross-sections and explicitly or implicitly assume two-dimensional (i.e. $k = 1$ plane) strains. Examples include most palinspastic or balanced section methods in convergent and divergent settings (e.g. Buchanan & Neiuwland (1996) and references therein) and some forward-modelling techniques employed in the analysis of sedimentary basins (e.g. Kusznir *et al.* 1991). If the region being considered has suffered any component of wrenching, extension (transpression) or contraction (transtension) occurs normal to the line of section (Jamison 1991). Used in isolation, such two-dimensional cross-section-based methods may yield misleading results, and the development and use of three-dimensional techniques is required (e.g. Ma & Kuzsnir 1992, 1993).

(2) *Strain ellipses and faulting.* Wilcox *et al.* (1973) and Harding (1974) used a two-dimensional finite strain ellipse to account for the wide range of second-order contractional, extensional and strike-slip structures that formed in analogue models of strike-slip deformation zones viewed in plan. These workers demonstrated that such models may be used qualitatively to explain deformation patterns within strike-slip zones where the shear plane is well defined and the deformation approximates to a steeply dipping wrench simple shear. However, such ellipse models have also been very widely used in the interpretation of faulting patterns in offshore sedimentary basins and, in some cases, they are used to support strike-slip or oblique opening hypotheses (e.g. Gibbs 1986; Fossen 1989; Doré & Lundin 1996). The use of such two-dimensional, simple shear models in regions of three-dimensional transtensional strain is inappropriate and may lead to serious errors in the interpretation of crustal deformation patterns.

(3) *Structural complexity.* Complex and, as yet, poorly understood deformation patterns are likely within many transpression and transtension zones. Existing field studies (e.g. Fig. 7) show that structures that differ significantly in orientation may form simultaneously. Many of these features would previously have been and, in some areas, still are interpreted as the products of polyphase deformation events. A more fundamental difficulty exists in all transpression zones because there is not, in general, a simple relationship between stretching lineations and the direction of tectonic transport. Marked changes in lineation pattern may occur because of spatial variations in finite strain and/or kinematic partitioning (Tikoff & Greene 1997). This problem is particularly acute in triclinic transpression zones, where the precise relationship between the development of geological structures and deformation is very poorly understood.

(4) *Deduction of plate motions.* Zones of transpression and transtension present particular problems when trying to relate crustal deformation patterns to relative plate motions. If kinematic partitioning occurs, as is fairly common, the deformation seen in one region may not be representative of the system as a whole. This is particularly important in the recognition of major strike-parallel motions in crustal deformation zones because structures recording such movements very commonly occur in narrow zones of high strain of limited areal extent that may be poorly exposed (Goodwin & Williams 1996). In addition, attempts to relate regional lineation patterns to plate motions (e.g. Shackleton & Ries 1984; Ellis & Watkinson 1987) may also be unwise in transpression zones.

(5) *Boundary conditions versus stress.* The analysis of transpression and transtension zones has highlighted the importance of kinematic boundary conditions during deformation of the crust and lithosphere (e.g. Molnar 1992; Tikoff & Teyssier 1994). Many traditional approaches in deformation zones follow Anderson (1951) in considering stress to be the main deformation control, particularly in the brittle crust. If, however, the development of most geological structures is controlled by the strain imposed by the boundary conditions, as appears to be the case in transpression zones, then there may not be a simple or significant relationship between large-scale crustal deformation structures and stress.

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