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Strain partitioning in transpression zones

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Abstract—Transpressional strain acting upon structurally anisotropic rocks can be partitioned into separate deformational domains of pure shear and simple shear. This contrasts with homogeneous transpression in which both the pure shear and the simple shear strain components are uniformly distributed across the zone of deformation. Structural weaknesses capable of partially or fully accommodating one component of deformation include lithological contacts, rheological heterogeneities, and faults or shear zones situated within the deformation zone or lying along its boundaries. Partitioning of transpressional strain can occur when stress is applied oblique to pre-existing structural weaknesses, or can occur during later stages of progressive strain, when the early deformation of isotropic rocks imparts sufficient anisotropy to allow subsequent strain to be partitioned. Partitioning of transpressional strain into domains lying parallel to the deformation zone boundaries can be distinguished from 'fault-stepped' transpression, in which strain is partitioned along the length of a segmented fault zone.

Mesofracture analysis of rocks affected by mid-Devonian deformation on both sides of the Highland Boundary Fault Zone (HBFZ) in central Scotland shows that strain was not homogeneous. The mesofracture data suggest that regional north–south compression, orientated oblique to the pre-existing NE–SW-trending HBFZ, was partitioned into separate deformational domains. The HBFZ accommodated most of the simple shear component, whilst the rocks flanking the zone were deformed predominantly by pure shear. A contemporary example is the San Andreas Fault in central California, where analyses of neo-tectonic stresses show that the direction of principal compression is perpendicular to the fault zone. Many examples of the partitioning of transpressional strain have also been recognised at destructive plate margins where the direction of plate motion is oblique to the edge of the over-riding plate.

INTRODUCTION

The terms transpression and transtension have been adopted to describe oblique relative motions of lithospheric plates (Harland 1971). Sanderson & Marchini (1984) described the resulting deformation in terms of a transcurrent shear acting simultaneously with horizontal shortening and vertical lengthening of the deformation zone. Thus transpression can be conveniently described mathematically by factorizing the overall strain into pure shear and simple shear components of deformation (Sanderson & Marchini 1984).

The concept of transpression (including transtension) described by Sanderson & Marchini (1984) has subsequently been widened, and its applicability to observed deformation structures has been increased. Fossen & Tikoff (1993) presented a more generalized deformation matrix which includes the effects of volume change, and allows for simple shear in non-vertical planes. Robin & Cruden (1994) used a continuum mechanics model to describe heterogeneous transpression, which also describes oblique transpression in which the relative motion of the shear zone boundaries contains a component of dip-slip. Their solution removes a mechanical limitation of the Sanderson & Marchini (1984) model, which required the boundaries of the transpression zone to possess the unlikely property of allowing frictionless slip in the vertical direction but no slip in the horizontal direction.

In this paper we consider a further modification to the model of Sanderson & Marchini (1984), and present a model for transpression in which the pure shear and simple shear strain components are partially or wholly partitioned (or 'compartmentalized' or 'decoupled') into separate deformational domains within the transpression zone.

THEORY

Partitioning of transpressional strain

Sanderson & Marchini (1984) described a deformation (D) across a vertical shear zone (experiencing no volume change) in terms of simple shear and pure shear strain components:

$$D = \underbrace{\begin{pmatrix} 1 & \gamma & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix}}_{\text{simple shear}} \underbrace{\begin{pmatrix} 1 & 0 & 0 \\ 0 & \alpha^{-1} & 0 \\ 0 & 0 & \alpha \end{pmatrix}}_{\text{pure shear}} = \underbrace{\begin{pmatrix} 1 & \alpha^{-1}\gamma & 0 \\ 0 & \alpha^{-1} & 0 \\ 0 & 0 & \alpha \end{pmatrix}}_{\text{transpression}}, \quad (1)$$

where α^{-1} is the ratio of the deformed to original width of the zone, and γ is the bulk shear strain parallel to the zone. Under conditions of homogeneous transpression described by Sanderson & Marchini (1984), both components of deformation are uniformly distributed across

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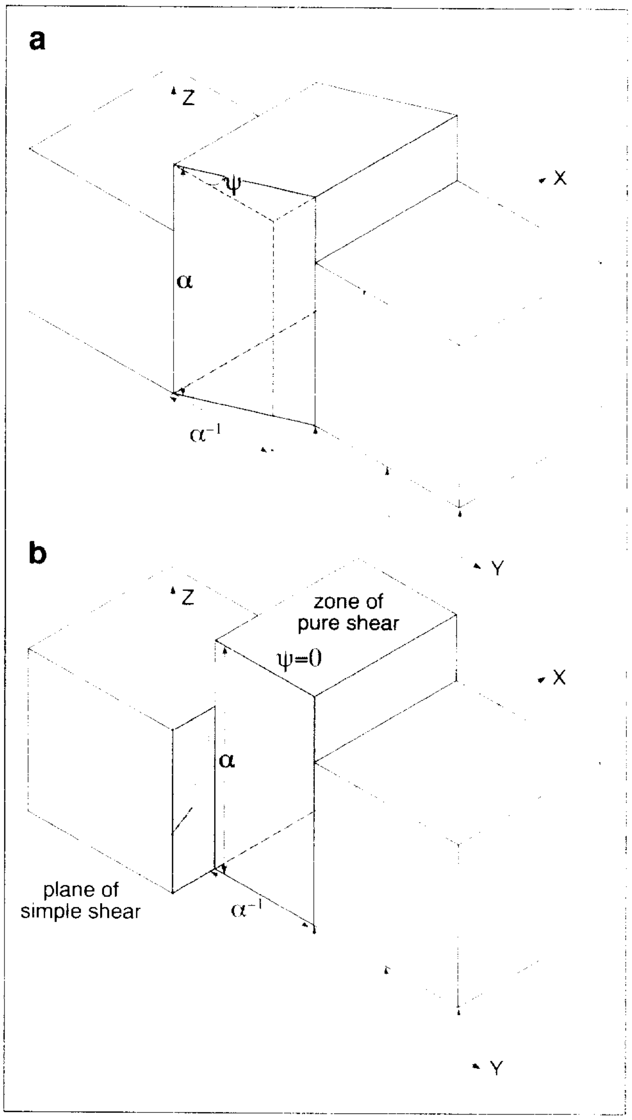

Fig. 1. End-member models of transpression showing the transformation of a unit cube, assuming constant volume. (a) Homogeneous transpression: the pure shear and simple shear deformational components are both uniformly distributed across the transpression zone (based on fig. 1 of Sanderson & Marchini 1984). (b) Partitioning of transpressional strain: the pure shear and simple shear components are completely compartmentalized into two separate deformational domains (a zone of pure shear, and a plane of simple shear), giving rise to regionally heterogeneous deformation. The zones of deformation are shaded, the underformed rocks on either side of the zones are unshaded. α is the vertical stretch, α^{-1} is the ratio of the deformed to original zone width, and $\gamma = \tan \psi$ is the simple shear strain

the deformation zone (Fig. 1a), and the deformation at any point in the zone is described by the transpression strain matrix product of equation (1). In contrast, Fig. 1(b) shows the same resultant transpressional deformation, but this is now distributed heterogeneously by the complete partitioning of the overall strain into two separate deformational domains. One domain is deforming by homogeneous simple shear, and the deformation in this domain is fully described by the simple shear matrix of equation (1). The second domain, which is fully described by the pure shear matrix of equation (1), is schematically shown to be deforming by homogeneous pure shear. Thus, for the model shown in Fig. 1(b), the factorization of strain into two components in equation (1) is not a mathematical convenience to aid understanding, but rather reflects the actual partitioning of transpressional deformation.

Partitioning of transpressional strain requires the pre-existence of one or more surfaces or zones of weakness, along which a component of simple shear can be preferentially accommodated. Such weaknesses might include faults, shear zones, lithological boundaries and/or rheological anisotropies, and these may actually define the transpressional zone *boundaries* or be situated entirely *within* the zone. Because (as stated by Sanderson & Marchini 1984, p. 449), "many zones of deformation in the crust are bounded by steep, parallel planes, often representing strain discontinuities manifested as faults or shear zones", we expect some degree of strain partitioning to be a common phenomenon in transpression zones. It is important to emphasize that the term 'strain partitioning' is not synonymous with 'transpression'. Transpression is a description of deformation in relation to a specified reference frame and with respect to well-defined deformation zone boundaries or intra-zone anisotropies, whereas strain partitioning is a process that can occur under general strain conditions, including simple shear, pure shear, dilation, transpression or transtension.

The two models shown in Fig. 1 can be considered to be end-member descriptions of transpression. The transpression model shown in Fig. 1(a) (see also fig. 1 of Sanderson & Marchini 1984, and fig. 1 of Robin & Cruden 1994) is most relevant when the deformation zone is structurally isotropic and the margins of the zone are not planes of structural weakness. In contrast, the 'strain partitioning' model (Fig. 1b) is more applicable when the deformation zone has marked structural anisotropy, or when either (or both) of the zone margins is a pre-existing fault or shear zone. For the complete partitioning of strain, such planes must be cohesionless and frictionless. Many geologically realistic situations probably involve some intermediate strain state, in which the transpressional strain is only partially partitioned; that is, not all of the simple shear strain is accommodated along pre-existing discontinuities.

In the end-member model for transpression shown in Fig. 1(b) the component of simple shear is fully partitioned onto one deformation zone boundary. Kinematically, however, this zone margin is not a straightforward strike-slip fault boundary, because the vertical thickening that accompanies transpressional shortening imparts a component of dip-slip displacement (Fig. 2). This suggests that faults in transpression zones (particularly in zones in which the compression to transcurrence ratio is high) can display dominantly strike-slip displacements at depth but oblique-slip (with potentially large components of dip-slip) near the surface.

Simple Transpression (as defined by Harland 1971)

In order to predict the strain patterns that might develop during progressive deformation, we can analyze

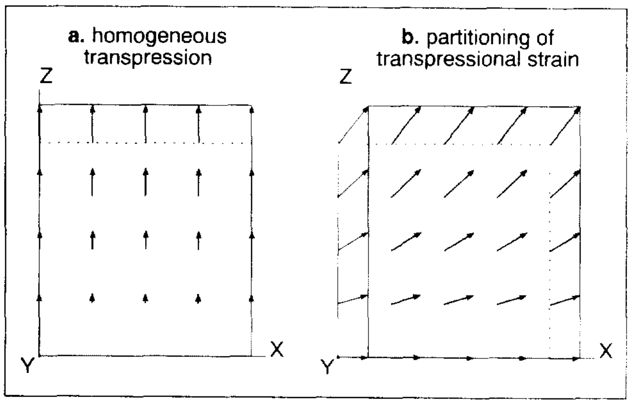

Fig. 2. Displacement vectors for points lying along the deformation zone boundary plane (i.e. the XZ plane in Fig. 1). (a) Homogeneous transpression. (b) Transpression in which strain is completely partitioned.

theoretical transpression models in which well-defined boundary conditions have been specified. Simple Transpression, resulting from non-orthogonal plate collision, has been defined by Harland (1971) in terms of a set of boundary conditions in which two rigid plate margins converge obliquely, with a constant direction of regional shortening. Harland schematically depicted the deformation in Simple Transpression to be accommodated by the progressive rotation and tightening of folds. The analysis of Simple Transpression is used as a basis for the interpretation of field data presented later in this paper.

A quantitative mathematical description of Simple Transpression is given by the following equations from Sanderson & Marchini (1984):

pure shear contraction, $\alpha^{-1} = (1 - S)$ (2)

simple shear strain, $\gamma = \tan \psi = S(1 - S)^{-1} \cot \beta$, (3)

where S is the amount of shortening, β is the obliquity of the zone margins to the direction of shortening, and ψ is the angular shear (Fig. 3). Equations (2) and (3) describe both homogeneous Simple Transpression (Fig. 3b), and Simple Transpression in which strain is partitioned into separate deformational domains (Fig. 3c). In the latter case, equation (2) fully describes the strain in the domain of pure shear, whilst equation (3) fully describes the strain in the domain of simple shear.

For any given zone geometry (β) and amount of shortening (S) there are very many valid models of transpressional deformation. These models differ principally in the way that the simple shear component is distributed across the deformation zone. Simple shear strain that is completely partitioned might be distributed onto one or both of the zone boundaries (Figs. 3c and 4a, respectively) and/or onto faults within the deformation zone (Fig. 4b). Conversely, simple shear strain might not be completely partitioned (Fig. 4c). The amount of strain partitioning, and the way in which partitioned strain is accommodated, are dependent upon the nature and three-dimensional geometry of the zone boundaries and/or intra-zone weaknesses, the mechanical and geological properties of the material comprising the deformation zone, the orientation of any anisotropy with respect to the zone boundaries, and the extent of anisotropy within the deformation zone. Complex structural styles can therefore occur in deformational models that have well defined and relatively uncomplicated boundary conditions.

Strain partitioning during progressive transpressional deformation

Partitioning of transpressional strain may occur not only during episodes of fault reactivation (when a stress is applied with an orientation oblique to a pre-existing fault), but may also develop during *progressive* transpressional deformation. Minor faults or shear zones that initiate during the early stages of homogeneous transpression, can subsequently develop to form a major strain discontinuity that is capable of preferentially accommodating partitioned simple shear strain. In such a situation it is not clear whether strain partitioning would initiate suddenly as a single event, or would develop progressively over a period of time. In either case, the orientation of the major axis of the infinitesimal stress ellipsoid will change significantly, and can become orientated parallel to the zone boundaries after only relatively small amounts of shortening. This contrasts with *homogeneous* transpression, in which 100% shortening (infinite shear strain) is necessary to rotate the long axis of the finite strain ellipse into parallelism with the zone margins (Fig. 5, and Sanderson & Marchini 1984, fig. 3). This implies that during progressive deformation the onset of strain partitioning can result in en échelon structures (that formed early in the deformation sequence oblique to the zone boundaries), being subsequently overprinted by later structures orientated parallel or sub-parallel to the zone.

Strain partitioning along linked fault systems

The above examples demonstrate how pure shear and simple shear can be partitioned into sub-domains that lie broadly *parallel* to the margins of transpression zones. Harland (1971) also documented examples of strain partitioning in which a zone of deformation is segmented *along its length* into separate deformational sub-domains; Harland labelled this 'Fault-Stepped' transtension/transpression. The partitioning of strain between linked contractional, extensional and strike-slip faults is a common feature of deformation at several scales of magnitude. In strike-slip zones, localized partitioning of strain occurs in pop-ups and pull-aparts (e.g. Biddle & Christie-Blick 1985, Woodcock 1986, Woodcock & Fischer 1986). Compressional deformation in orogenic belts is often segmented between thrust sheets and strike-slip transfer zones (e.g. Cobbold *et al.* 1991). The width of the deformation zone in Fault-Stepped movement can be restricted. In the ridge transform geometry of the Mid Atlantic Ridge, the plate motion

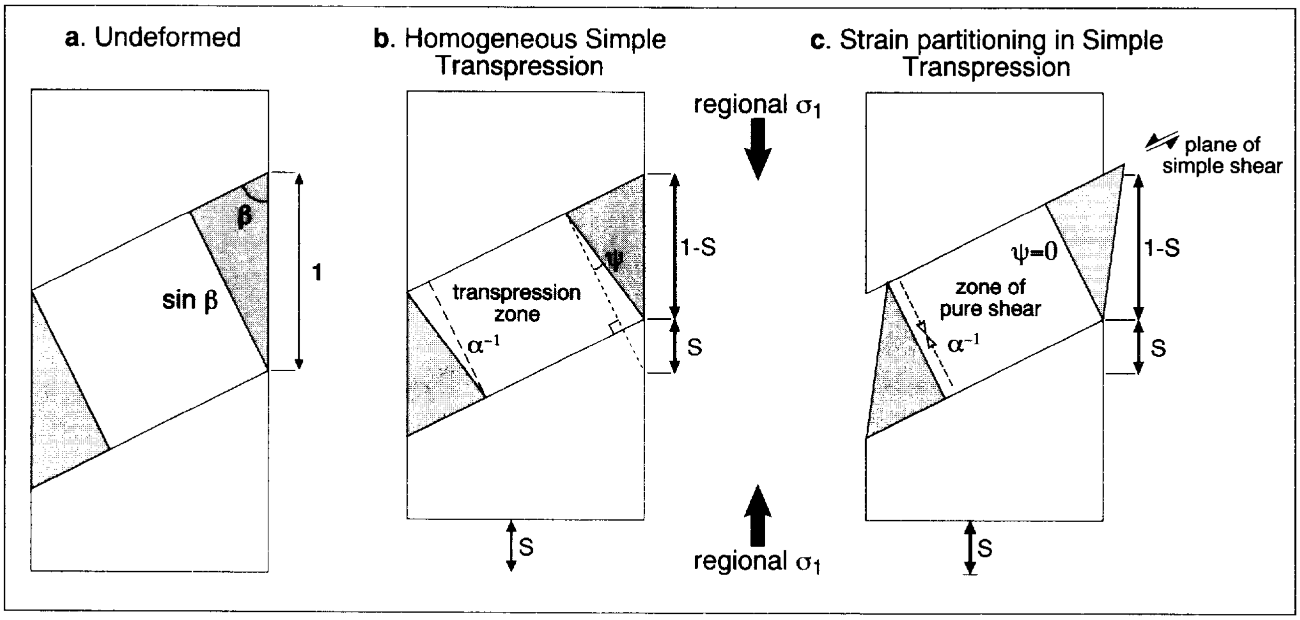

Fig. 3. End-member models of Simple Transpression (as defined by Harland 1971), in plan view. (a) Undeformed zone geometry. (b) Homogeneous Simple Transpression (based on fig. 6 of Sanderson & Marchini 1984). (c) Simple Transpression in which strain is partitioned between a wide zone of pure shear ($\psi = 0$), and a discrete plane of simple shear ($\psi = \tan^{-1}\gamma$) along one zone margin. The models show the transformation of a cube, assuming constant volume. The zone boundaries are vertical, planar and parallel, and the direction of shortening, S, remains constant during deformation. Within the deformed zones [shaded in (b) and (c)] there is an area decrease in the plane of the diagram, representing vertical thickening.

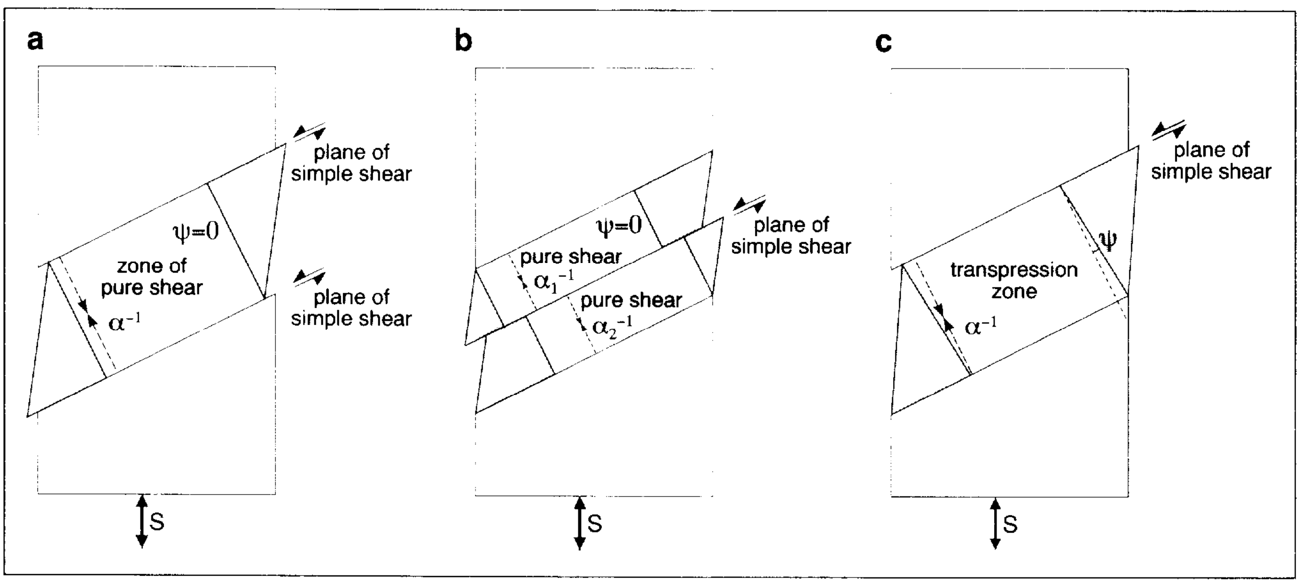

Fig. 4. Variants of the 'strain partitioning' model of Simple Transpression (plan view). (a) Partitioned simple shear is accommodated along both zone boundaries (e.g. oblique closure of a basin, or deformation within a wide fault zone, as depicted schematically in fig. 21 of Sylvester & Smith 1976). (b) Simple shear is wholly partitioned onto a weakness within the deformation zone (e.g. oblique reactivation of a pre-existing fault). Deformation within the zone is dominated by the intra-zone weakness, and the zone boundaries might not necessarily be well defined. (c) Simple shear is only partially partitioned onto one zone boundary.

vectors are oblique to the regionally transtensional constructive plate margins (e.g. Sykes 1967). The main active deformation zone (ignoring the effects of post-accretion vertical subsidence) is restricted to the ridge segments and the active parts of the transform faults, and kinematically, the deformation zone is a single stepped vertical plane.

Sheared Transpression (as discussed by Harland 1971)

Harland (1971, p. 30) discussed a third type of transpression, which he labelled 'Sheared Transpression'. Whereas Simple and Fault-Stepped Transpression are both expressed in terms of zone margin geometry and the direction of zone boundary displacement, Sheared

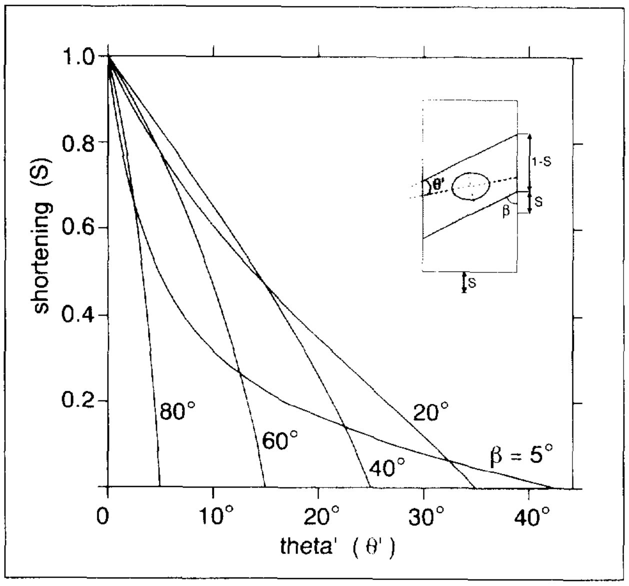

Fig. 5. Graph showing the rotation of the long axis of the finite strain ellipse in the horizontal plane during *homogeneous* Simple Transpression (see Fig. 3b). The angle, θ', between the zone boundary and the long axis of the finite strain ellipse, is derived from the strain matrix of Sanderson & Marchini (1984): $\tan 2\theta' = 2\gamma/(\alpha^2 + \gamma^2 - 1)$. By substituting equations (2) and (3) into this relationship, it is possible to express θ' in terms of shortening, S. Five curves are shown, representing different values of β. During homogeneous transpression the long axis of the finite strain ellipse achieves parallelism with the zone boundary only after 100% shortening ($S = 1$).

Transpression is discussed in relation to a number of different possible types of deformation path. One deformation path involves two distinct stages of deformation, an orthogonal compression followed by a separate transcurrent displacement, implying that a major change in the orientation of the regional stress field must occur between the separate deformational stages. A second deformation path involves oblique movement (homogeneous transpression?) followed by transcurrent shear (partitioned strain?), and might be included in our description of strain partitioning during progressive transpressional deformation. A third deformation path involves *synchronous* strike-slip faulting together with normal or oblique compression, and consequently this type of deformation is included in our general model of strain partitioning in transpression zones. Although Harland's discussion served to emphasize that identical bulk strains can result from different deformation histories, to avoid ambiguity we have not adopted the term 'Sheared Transpression'.

EXAMPLES OF STRAIN PARTITIONING IN TRANSPRESSION ZONES

Mid-Devonian upper crustal deformation in central Scotland

The Highland Boundary Fault Zone (HBFZ) in central Scotland is a major crustal fracture with a long and complex structural history, and separates Neoproterozoic Dalradian metasediments from late Silurian–Devonian Old Red Sandstone (ORS). The fault zone contains rocks of the Highland Border Complex, a disparate suite of igneous and sedimentary fault-bounded slivers that have experienced variable amounts of metamorphism, metasomatism and tectonism. Some authors consider that the HBFZ represents a major terrane boundary along which Lower Palaeozoic terrane accretion occurred (e.g. Soper & Hutton 1984, Hutton 1987, Haughton 1988, Bluck 1990). In mid-Devonian times, central Scotland experienced a deformation event that caused the reactivation of the pre-existing steeply-dipping HBFZ, involving the thrusting of Dalradian rocks southwards over the Highland Border Complex and Lower ORS. The deformation also gave rise to the long wavelength, low amplitude, laterally continuous Strathmore Syncline/Sidlaw Anticline fold pair in the Midland Valley (Fig. 6), and probably caused reactivation of regionally significant faults such as the Tyndrum/Killin/Loch Tay fault array in the Southern Highlands (Watson 1984, Treagus 1991).

Figure 6 presents the interpreted results of a mesofracture analysis from central Scotland that utilized the 'brittle microtectonics' methodology outlined by Hancock (1985). The rationale of this approach is to interpret the direction of principal shortening within individual sampling stations, and then to use such data to infer the principal stress orientations on a regional scale. In this study the geometric and kinematic data describing 2600 mesofractures (faults and joints) from 40 stations were used to interpret the dynamics of brittle deformation (Jones 1990). In addition, Fig. 6 incorporates the results of a two-dimensional study of fractured pebbles in Lower ORS conglomerates adjacent to the HBFZ (Ramsay 1964). Because most fractures in these conglomerates are steep to vertical, the two-dimensional results correlate closely with the more detailed three-dimensional analysis, and the two-dimensional data have been further validated by an unpublished comparative three-dimensional study of fractured clasts.

The principal limitation of this type of study lies in the difficulty of ascertaining the age of the fractures at each sampling station. Wherever possible, therefore, adjacent sampling sites were chosen that allowed a direct comparison between the fracture patterns in Lower ORS (or older) rocks that have been affected by the mid-Devonian deformation, and nearby Upper ORS (or younger) rocks which only show the imprint of Carboniferous or later deformation events.

The majority of mesofracture stations situated close to the HBFZ display a direction of principal shortening that tends to be orthogonal to the fault zone. In contrast, stations sited more than a few kilometres from the HBFZ show a direction of principal shortening that is generally orientated approximately north–south. Most of the mesofracture stations contain four sets of fractures, and fracture geometry and kinematic indicators show that the strain is triaxial (Reches 1978, 1983, Reches & Dietrich 1983). Within such stations, mesofracture geometry and kinematics suggest that the inter-

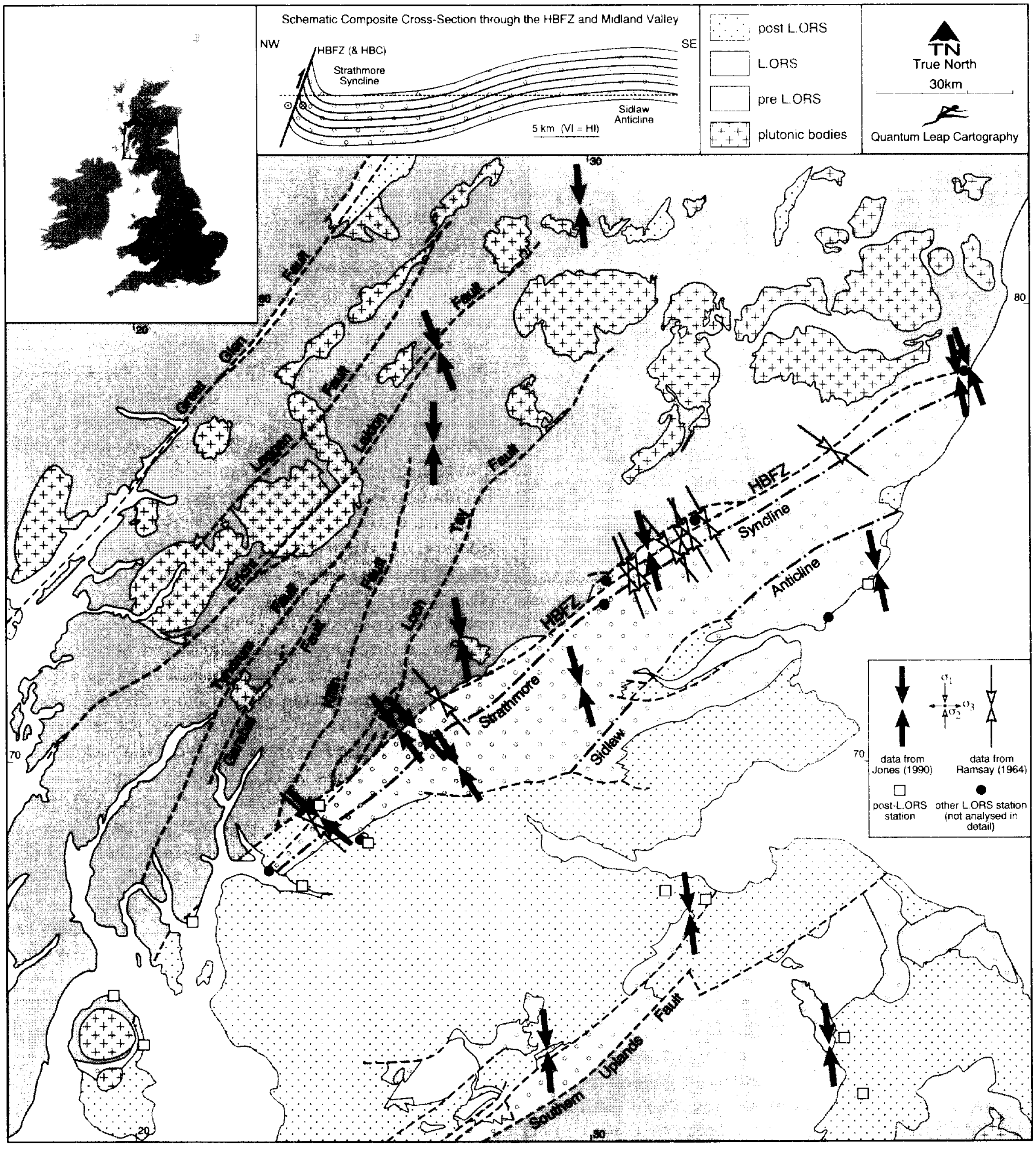

Fig. 6. Regional interpretation of mid-Devonian deformation in central Scotland based on the analysis of results from Jones (1990) and Ramsay (1964). Interpretive solid geology and schematic cross-section compiled from BGS data. Strathmore and Sidlaw fold traces from Armstrong & Paterson (1970) and Mykura (1991). Fault traces collated from Johnson & Frost (1977) and Treagus (1991). HBFZ, Highland Boundary Fault Zone; HBC, Highland Boundary Complex.

mediate stress axis (σ_2) was vertical and that the effective intermediate stress was tensile, i.e. strains are oblate. The orientation of σ_2 is also interpreted to have been vertical in stations showing plane strain, corroborating the results of Ramsay (1964). Some stations sited within a few hundred metres of the Highland Border Complex show an intense brittle deformation, and the abundance of sinistral mesofaults suggests that the finite brittle strain is (at least partially) a result of progressive sinistral simple shear. This contrasts with stations sited farther away from the HBFZ, in which the strain is sufficiently low to suggest that the finite strain ellipse has not been noticeably rotated by non-coaxial strain.

The large-scale Strathmore/Sidlaw folds in the Midland Valley are very open structures which trend NE–SW (i.e. parallel to the Highland Boundary Fault Zone), and which appear not to have any associated cleavage development. The brittle deformation which affects the Lower ORS in central Scotland, though widespread, is generally very low in magnitude in rocks situated more than a few hundred metres from the HBFZ, showing that large amounts of shortening did not occur during mid-Devonian deformation. The results of the mesofracture analysis (Fig. 6), and the parallelism of the Strathmore/Sidlaw fold pair with the HBFZ, are therefore inconsistent with a *homogeneous* model of transpression. The presence of the pre-existing, structurally weak Highland Border Complex, and the close proximity of the HBFZ to sampling sites that show evidence of non-coaxial deformation, suggests that a heterogeneous model of transpression is more relevant. Schematically, this involves the partitioning of strain into separate deformational domains in a model broadly comparable with Fig. 4(b).

We use the direction of principal compression observed in the mesofracture stations sited farthest from the HBFZ to infer an approximate north–south direction of regional shortening, equivalent to the far field maximum principal stress (σ_1). Because such stations show no evidence of rotational strain, we apply the boundary conditions of Simple Transpression, in which the obliquity of the HBFZ to the direction of regional shortening (i.e β) is interpreted to have been approximately 60°. Most of the simple shear component of the deformation occurring at this time seems to have been partitioned onto the HBFZ, whilst the pure shear was mostly accommodated by open folding in the Midland Valley (reconstruction of balanced cross-sections suggests that the Strathmore/Sidlaw fold pair represents a contraction of only a few kilometres, i.e. less than 10% shortening), followed by low-magnitude pervasive brittle fracturing throughout cental Scotland.

In detail, the actual distribution of mid-Devonian strain in central Scotland is complex, and several deformational sub-domains can be recognized within the first-order model shown in Fig. 4(b). For example, a significant component of simple shear appears to have been partitioned onto the synthetic faults of the Tyndrum/Killin/Loch Tay array, and the overthrusting of Dalradian rocks onto the northern margin of the Midland Valley represents a component of pure shear within the HBFZ (Fig. 7). In addition, the boundaries to the deformational zone are poorly defined (the HBFZ in the centre of the zone was the dominant anisotropy controlling the partitioning of strain), and the distribution of strain, which decreases progressively away from the HBFZ, is truly heterogeneous.

San Andreas Fault Zone, central California

Present-day strain partitioning has been recognized along the San Andreas Fault in California. Mount & Suppe (1987) have documented neo-tectonic stresses orientated approximately perpendicular to the San Andreas Fault, based on borehole elongations and breakouts, in rocks flanking a 150 km section of the fault in central California along which demonstrable strike-slip movement is taking place (Fig. 8). They have also shown that many of the fold traces in rocks immediately adjacent to the San Andreas Fault lie parallel or sub-

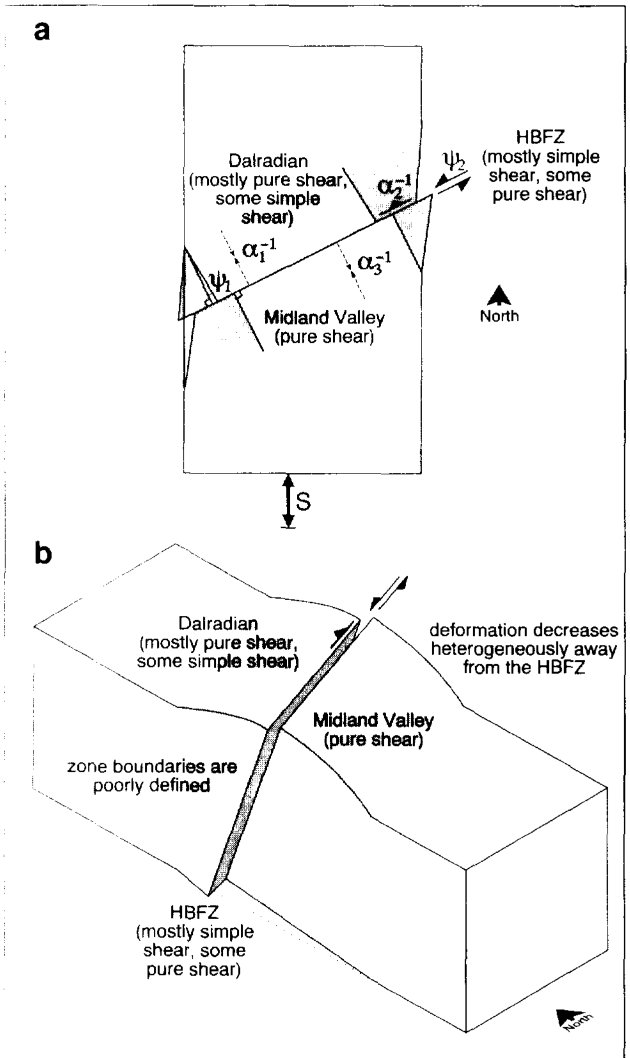

Fig. 7. Schematic interpretation of strain-partitioning of mid-Devonian transpression in central Scotland, in plan view (a), and three-dimensional block diagram (b). The amount of regional pure shear, α^{-1}, is equal to the sum of α_1^{-1}, α_2^{-1} and α_3^{-1}, the separate components of pure shear, and the amount of regional simple shear, ψ, is equal to the sum of ψ_1 and ψ_2, the separate components of simple shear in the Dalradian block and the fault zone. For further discussion see text.

parallel to the fault zone. Zoback *et al.* (1987), Namson & Davis (1988) and Lettis & Hanson (1991) have provided further evidence for the partitioning of strain between the San Andreas and associated fault strands, and immediately adjacent regions of folded and horizontally shortened rocks.

Harding (1988) and Wickham (1988) have discussed and clarified the conclusions of Mount & Suppe (1987) concerning the implications that such observations have for transpressive wrench tectonics in general. The San Andreas Fault in central California may demonstrate how transpressional strain partitioning can vary temporally. The early stages of the deformational sequence in the sedimentary cover rocks are marked by the development of wrench-related en échelon folds and fractures at much lower angles than expected from the simple-shear clay-box experiments of Wilcox *et al.* (1973). Sanderson & Marchini (1984, p. 453) show that such a discrepancy is consistent with their model of

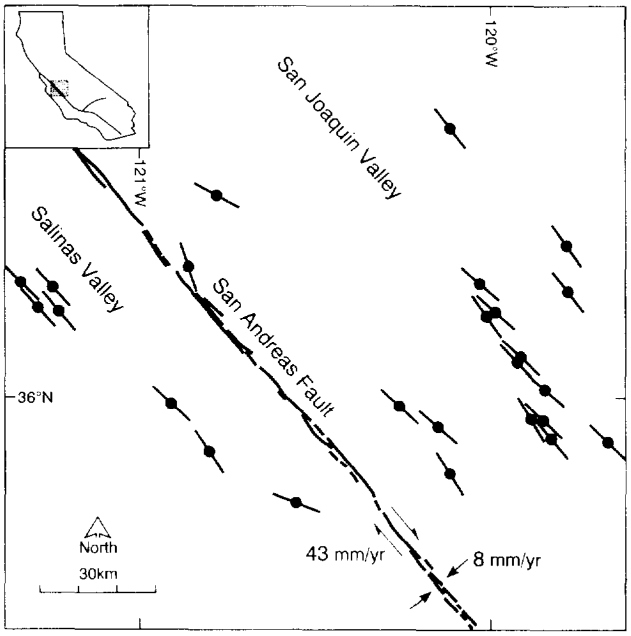

Fig. 8. Present-day movement rates and stress orientation data from the San Andreas fault zone, central California, after Mount & Suppe (1987). Bars with circles indicate the direction of borehole elongation (minimum horizontal stress).

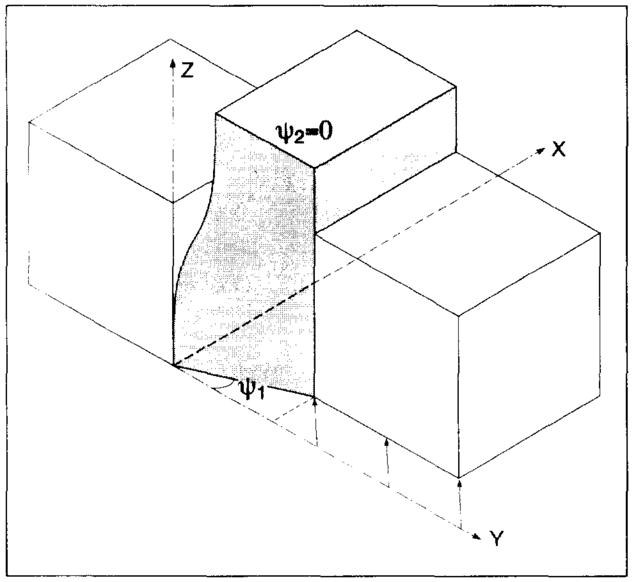

Fig. 9. Partitioned transpressional strain in the brittle upper crust balanced by homogeneous transpression in the ductile lower crust.

transpressional deformation. As deformation progressed, individual fractures propagated to form a through-going anastomosing fault system capable of preferentially accommodating the simple shear component. After this time, folds developed parallel or sub-parallel to the fault zones, and the system changed from one of homogeneous transpression to partitioned heterogeneous transpression.

Strain-partitioning at convergent plate margins

The oblique subduction of oceanic crust can cause regional-scale strain-partitioning in the fore-arc and back-arc regions of the over-riding plate (Fitch 1972, Walcott 1978, Hamilton 1979, Jarrard 1986), as reviewed by Oldow *et al.* (1990), and by Molnar (1990) who discussed possible causes of crustal-scale partitioning. Cashman *et al.* (1992) describe how strain is partitioned into paired domains of strike-slip and dip-slip deformation, in response to oblique subduction of the Pacific plate beneath New Zealand. Other examples include the Sunda arc, the Alaska arc and the Chilean trench (Lettis & Hanson 1991). A transpressional model has been proposed by Ratschbacher (1986) for Austro-Alpine nappe tectonics in the Eastern European Alps.

DISCUSSION

Partitioning of strain can occur on a range of scales, from microscopic grain-scale processes (e.g. Schmid 1982, Evans & Dunne 1991) to global-scale plate boundary movements (e.g. Jackson 1992), and has been demonstrated in laboratory modelling of oblique convergence (Richard & Cobbold 1989). The observations that many rocks are rheologically anisotropic, that many zones of deformation are bounded by faults or shear zones, and that such weaknesses are often reactivated by subsequent deformation, all suggest that strain partitioning may be a common feature in transpression zones.

Strain partitioning in transpression zones can occur at different levels in the lithosphere. Although several of the examples discussed in this paper demonstrate the partitioning of brittle strain in the upper crust, it is clear that ductile deformation can also be partitioned (e.g Lister & Williams 1983, Girard 1993). In large-scale transpression zones it is possible that brittle fault systems in the upper-crust (onto which a component of simple shear strain is preferentially partitioned), pass downwards into analogous ductile shear zones at lower crustal levels. Alternatively, heterogeneity may vary with depth, and rheological anisotropy in the upper crust need not necessarily imply significant structural weakness at depth. In such situations, partitioned strain in the anisotropic fractured upper crust might be accommodated without partitioning in the ductile lower crust (Fig. 9), as implied by Molnar (1990) and Cobbold *et al.* (1991). The transition between heterogeneous and homogeneous deformation may be gradual, or might occur within a region of rheological change (such as the base of the seismogenic crust), as envisaged by Lettis & Hanson (1991).

The recognition of strike-parallel faults and shear zones is essential in understanding strain in orogenic belts. When transpressive strain is partitioned, the relationship between regionally consistent stretching lineations and the direction of plate motion is likely to be complex. Kinematic reconstructions that assume homogeneous transpression can give erroneous estimates of shortening, and can lead to invalid assumptions about the orientation of σ_1 on a regional scale. Although two-dimensional line-balancing and area-balancing reconstruction techniques are invalid methods to use in homogeneously transpressive regimes, their use may be entirely justified within domains of pure shear in heterogeneous transpression. Similarly, homogeneous deformation should not be assumed when using three-dimensional balancing methods (e.g. the method described by McCoss 1986), or reconstructions based on quantifying the rotation of objects (including palaeomagnetic vectors). Consequently, the whole deformation zone and its boundaries must therefore be analyzed in detail to ascertain whether strain partitioning might have occurred. The measurement of the orientation of regional σ_1 (i.e. the direction of zone boundary displacement) is invaluable in the restoration of transpression zones. Mesofracture analysis of joint arrays beyond the zone of deformation can be a particularly useful technique.

It is important to consider transpression and strain-partitioning relative to the scale of geological analysis. Segmentation of fault systems can result from localized stress orientations and mechanical anisotropy, and need not necessarily reflect *regional* stress regimes. Strain-partitioning at one scale of observation (for example, at restraining or releasing bends of strike-slip faults, or side-wall ramps of thrust sheets) need not imply transpression or transtension on a more regionally significant scale. This underlines an important philosophical point regarding the scale of geological observation and the homogeneity and heterogeneity of strain. For example, the heterogeneity of mid-Devonian deformation in central Scotland varies according to the volume of rock considered: individual microfractures and mesofractures represent heterogeneities that are directly observable and measurable to a field geologist, whereas deformation across individual outcrops is much more homogeneous (indeed, this was an underlying assumption when interpreting the results of the mesofracture analysis). Macro-faults (often detected by aerial photography or displacements of stratigraphy, rather than by direct observation), impart further heterogeneity to the rocks on a map-scale, though deformation on a regional scale, within each limb of the Strathmore/Sidlaw folds for example, appears more homogeneous. Across central Scotland, the strain is transpressive *with respect* to the pre-existing Highland Boundary Fault Zone, and at this scale of observation the transpressional strain is heterogeneous.

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