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Scale dependence, strain compatibility and heterogeneity of three-dimensional deformation during mountain building: a discussion

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Abstract

Plate motion at convergent margins causes crustal shortening and orogenic thickening. Relative motion that is oblique to the plate margins is an inevitable consequence of plate kinematics on a sphere and results in non-coaxial three-dimensional deformation that cannot be approximated to simple shear. Models of mountain building that include a smoothly varying component of pure shear shortening allow strain compatibility to be maintained by deforming the upper free surface of the Earth without disrupting the material continuum. However such models do not reflect accurately the nature of deformation in many areas of high strain in the upper crust, which are characterized by interconnected arrays of kinematically linked faults that can be active on several scales of magnitude simultaneously. As brittle deformation increases, the coherence of the material continuum is highly reduced. In such situations, strain compatibility is maintained by partitioning the deformation amongst structures of varying kinematic significance over a wide range of scales and not by smooth variations in strain magnitude acting on a single scale across a material continuum. There is a marked tendency for such partitioned domains to be oriented parallel or sub-parallel to the orogenic grain. Alignment of domains in this way represents a strong structural anisotropy, which acts as a highly significant boundary condition that controls deformation at subordinate scales. Finite strain observed within an individual domain at a given scale need not therefore display the same magnitude or orientation as bulk finite strain at the plate scale and consequently data must be collected from as large an area as possible to relate outcrop-scale structures to global-scale tectonics.

Keywords: Heterogeneous strain; Scaling of deformation; Strain partitioning; Discontinuum mechanics

1. Introduction

“We have been somewhat alarmed at the separation we see between the conclusions of those investigating global tectonics with those looking at the smaller-scale structural features observed at outcrop or map scale. We hope to see in the future a better integration of the actual geometrical features observed in the rocks themselves with the larger-scale predictions of plate tectonics. These geometrical appreciations should be scale independent and we have to make more position linkages across the scale divide” (Ramsay and Lisle, 2000, preface).

From deformation mechanisms at the scale of the crystal lattice to plate motion on a lithospheric scale, the processes studied by structural geologists and tectonicists span at least 14 orders of magnitude (10^{-6}–10^{+7} m). Even restricting our analysis to macroscopic structures by taking the outcrop-scale as a lower limit, the scale divide highlighted by Ramsay and Lisle spans magnitudes from millimetres to tens of thousands of kilometres.

In this paper we discuss the three-dimensional geometric and kinematic predictions inherent in plate tectonic theory of orogenic processes and compare these with outcrop- and map-scale structures using examples from the Southern Uplands of Scotland that developed along a destructive plate

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margin of Iapetus during Palaeozoic times. By relating detailed field data at outcrop and map scales with larger scale information about regional deformation, our aim is to investigate the way in which upper-crustal strain compatibility is achieved across the scale divide.

2. Plate kinematics at convergent margins

Since the advent of plate tectonic theory, it has been well understood that there is an intimate relationship between mountain building and convergent plate margins. Dewey (1975) and Woodcock (1986) have demonstrated that the progressive movement of plates on the surface of a sphere will usually give rise to relative motions that are oblique (i.e. neither parallel nor orthogonal) to the margins of the plates. Oblique relative motion at convergent plate margins typically causes triaxial deformation (Dewey et al., 1998). Such deformation is non-coaxial and non-plane strain and cannot be approximated as a plane strain pure shear or simple shear (Fig. 1a–d). Irrespective of how this deformation is accommodated at lower orders of scale magnitude (e.g. in map- and outcrop-scale structures), there is a kinematic requirement at the uppermost scale (i.e. the scale of the plate margin) that the overall bulk strain is non-coaxial and non-plane strain. This gives rise to the typical characteristics of convergent plate boundaries, with shortening across the margin, orogenic thickening, oceanward overthrusting and orogen parallel strike-slip. In terms of bulk strain symmetry (see Paterson and Weiss, 1961), deformation is most likely to be triclinic (Fig. 1d).

This paper addresses the inherent difficulties regarding strain compatibility with respect to non-coaxial non-plane strains (Ramsay and Huber, 1987; Hudleston, 1999). Before we present field data from a zone of non-coaxial non-plane strain, we discuss theoretical ways in which strain compatibility can be achieved in areas of complex three-dimensional deformation.

2.1. Strain compatibility and strain heterogeneity

The cartoon depictions of Fig. 1 represent a simplistic abstraction of bulk (plate-scale) deformation assuming homogeneous strain. Clearly, if one considers the boundaries between the zones of deformation and the undeformed plate interiors depicted in Fig. 1a, c and d, it appears that these representations disregard the standard rules of strain compatibility (Ramsay, 1967, 1976; Ramsay and Graham, 1970; Ramsay and Huber, 1983, 1987; Ramsay and Lisle, 2000). As compatibility cannot be maintained between units of undeformed rock and adjacent units where the strain has a component of *homogeneous* pure shear, it is inevitable that there will appear to be compatibility problems with this kind of depiction.

There are a number of ways, described in the following sections, in which strain compatibility can be re-established. These depictions apply equally well to both orthogonal relative plate motion and the geometrically more complex case of non-orthogonal motion, without compromising the essential kinematic boundary condition that requires bulk deformation to be non-coaxial and non-plane strain when relative motion is oblique. These compatible solutions all involve heterogeneous strain, but differ in the way in which heterogeneity is distributed.

2.2. Strain compatibility and smoothly varying heterogeneity

The most straightforward way to satisfy compatibility conditions involves a deformation in which the magnitudes

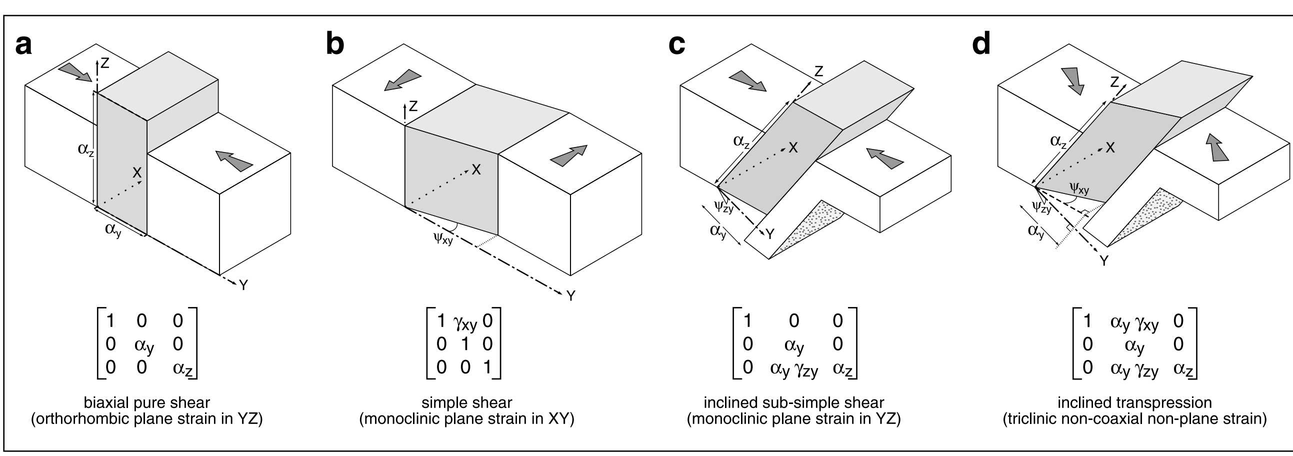


Fig. 1. Symbolic sketches of deformation at plate margins, with strain depicted homogeneously, in order to emphasize the overall symmetry of bulk, plate-scale deformation required by the relative plate motions shown by large arrows. (a) Reference deformation (coaxial plane strain) with orthorhombic strain symmetry. (b) Transcurrent margin (non-coaxial plane strain), monoclinic symmetry. (c) Convergent plate margin with orthogonal relative motion (non-coaxial plane strain), monoclinic symmetry. (d) Convergent plate margin with oblique relative motion (non-coaxial non-plane strain), triclinic symmetry. *X*, *Y* and *Z* are Cartesian reference coordinates with *X* horizontal and *XZ* parallel to the deformation zone boundary. α_y and α_z are ratios of deformed to original width of zone (including any volume change) parallel to *Y* and *Z*-axes, respectively. ψ_{XY}, ψ_{ZY} are angular shear strains where $\psi = \tan \gamma$.

of individual strain components are expressed as smoothly varying functions across the deformation zone, with zero strain at the zone margins and maximum strains in the centre of the zone (Fig. 2). Suitable functions in which strain is smoothly varying might include parabolic terms (fig. 26.25A of Ramsay and Huber, 1987; Robin and Cruden, 1994), sinusoidal terms (fig. 26.25B of Ramsay and Huber, 1987) or some other smoothly varying function of *y*.

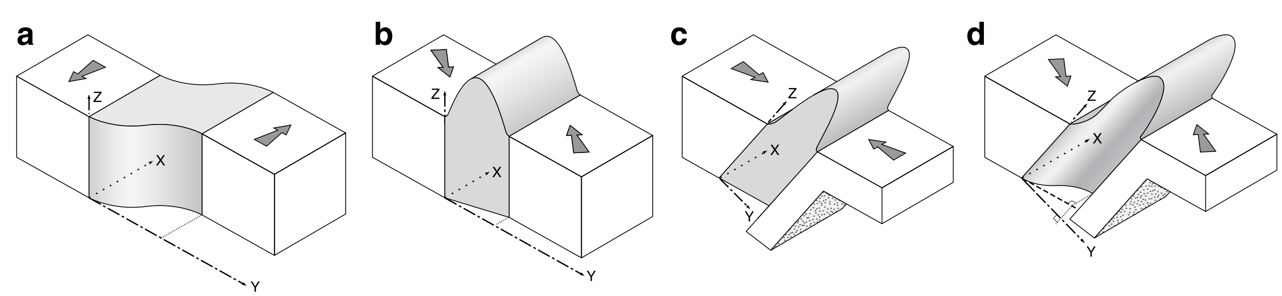
2.3. Strain compatibility and anastomosing shear zones

Some shear zones that develop in mid to lower crustal levels form anastomosing arrays in which strain is concentrated into narrow zones between lozenge-shaped blocks of undeformed or less deformed rocks (Ramsay and Allison, 1979; Ramsay, 1980; Bell, 1981; Ramsay and Huber, 1987, pp. 612–617; Hudleston, 1999; Carreras, 2001; Burg, 2002). In two dimensions these arrays commonly appear as conjugate sets (Fig. 3a and b), although they can also sometimes have a quadrimodal geometry in three-dimensions (Hudleston, 2002), a probable indication of bulk non-plane strain (Fig. 3c). Comparable quadrimodal geometries in brittle fault systems have been described by Reches (1983) and Krantz (1988). The compatibility issues of anastomosing shear zone arrays have been extensively discussed by Bell (1981) and Hudleston (1999), who have shown how displacements along anastomosing sets of shear zones can produce a bulk strain that is non-coaxial and non-plane strain without generating compatibility problems.

When the deformation across a broad area is accommodated by anastomosing arrays of shear zones, variations in magnitude of individual coaxial and non-coaxial components are no longer smoothly varying mathematical functions and it is not feasible to define a strain matrix that accurately describes deformation at every point across the broad zone using standard linear equations (Fig. 4). Although it may be possible to describe the deformation accurately within specific parts of localized individual shear zones using the types of strain matrix shown in Fig. 2, the use of such matrices to describe the bulk deformation across the whole area requires a degree of abstraction. The need to describe bulk deformation in terms of the aggregation of the strain seen in several smaller sub-areas is an extremely common feature of heterogeneous deformation patterns in crustal rocks and epitomises the challenges that we face in scaling data between outcrop measurements and plate scale interpretations. Detailed measurements made at one scale of observation often have to be synthesised and 'homogenised' to encapsulate their *bulk* deformational properties before they can be depicted with clarity at higher scales. This approach allows us to replace the detailed depiction of anastomosing shear zones in Fig. 3 with the more abstract depiction of smoothly heterogeneous strain in Fig. 2b (or an even cruder depiction using homogeneous strain; c.f. Fig. 1), depending on the level of accuracy required (i.e., the detail that is possible to show at the chosen scale of analysis without loss of clarity), and the level of mathematical complexity which we are able to model using the strain matrix.

2.4. Strain compatibility and strain partitioning

The partitioning of strain is a fundamental characteristic of deformation throughout most if not all of the Earth's lithosphere. The concept of strain partitioning is used in connection with grain-scale deformation mechanisms (e.g. Ramsay and Huber, 1983, pp. 113–124), tectonic processes operating at plate margins (e.g. Fitch, 1972; McCaffrey,



$$\begin{bmatrix} 1 & f_1(\gamma_{xy}, y) & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix} \quad \begin{bmatrix} 1 & \alpha_y \gamma_{xy} & 0 \\ 0 & \alpha_y & 0 \\ 0 & 0 & f_2(\alpha_z, y) \end{bmatrix} \quad \begin{bmatrix} 1 & 0 & 0 \\ 0 & \alpha_y & 0 \\ 0 & \alpha_y \gamma_{zy} & f_3(\alpha_z, y) \end{bmatrix} \quad \begin{bmatrix} 1 & f_4(\alpha_y \gamma_{xy}, y) & 0 \\ 0 & \alpha_y & 0 \\ 0 & f_5(\alpha_y \gamma_{zy}, y) & f_6(\alpha_z, y) \end{bmatrix}$$

Fig. 2. Heterogeneous non-coaxial deformations in which strain compatibility is achieved by progressively varying the magnitude of strain components from zero at the zone margins to a maximum in the centre of the zone. f_1–f_6 are unspecified functions in which the magnitude of the strain component is dependent upon position along the *Y*-axis. (a) Heterogeneous simple shear, with *y*-dependant non-coaxial strain component (after Ramsay and Graham, 1970). (b) Zone bounded by vertical margins, with *y*-dependant coaxial thickening and *y*-constant non-coaxial component (after Robin and Cruden, 1994; Dutton, 1997). (c) Inclined zone, with *y*-dependant coaxial shortening component, and non-coaxial component parallel to *x*-axis = 0 (after Dutton, 1997). (d) Inclined zone (c.f. Jones et al., 2004a), with *y*-dependant coaxial shortening component, and *y*-dependant non-coaxial component. Zones shown in (b)–(d) involve a component of pure shear deformation and achieve compatibility by deforming the upper free surface.

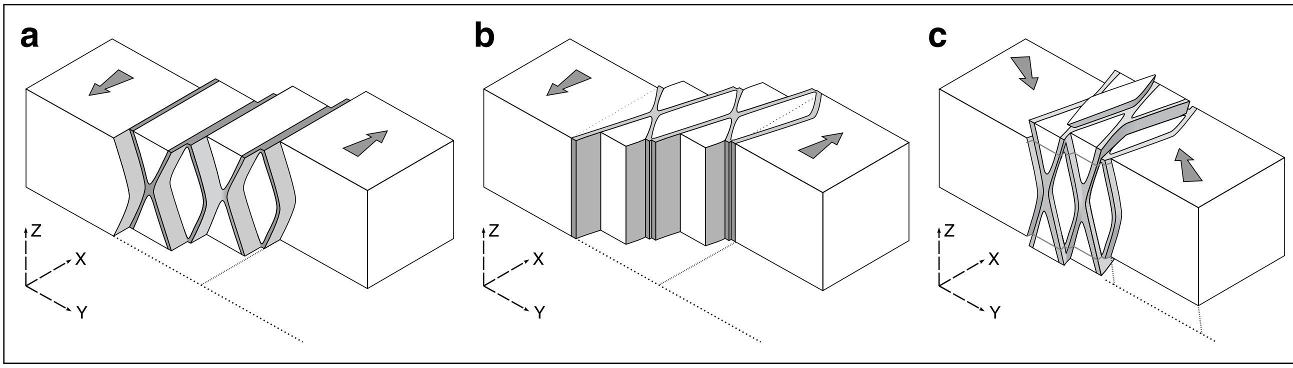


Fig. 3. Accommodation of non-coaxial strains by arrays of anastomosing shear zones. (a) Accommodation of simple shear along dipping sets of strike-parallel shear zones (c.f. figure 6b of Hudleston, 1999). The anastomosing zones intersect parallel to the horizontal *X*-axis. (b) Accommodation of simple shear along vertical sets of shear zones (c.f. figure 7 of Bell, 1981). Anastomosing zones intersect parallel to the vertical *Z*-axis. The lozenge blocks between the shear zones must deform and/or rotate in order to maintain compatibility during progressive deformation. (c) Accommodation of non-coaxial non-plane strain by quadrimodal arrays of shear zones. The quadrimodal array is asymmetric in plan view due to the component of strike-parallel non-coaxial strain. The shear zone array accommodates bulk shortening and thickening of the whole deformation zone (c.f. figure 8 of Bell, 1981 and figure 6c of Hudleston, 1999).

1992; Molnar, 1992; Platt, 1993), and rock deformation in general (e.g. Lister and Williams, 1983). The term is most commonly used to describe the spatial segregation (compartmentalisation) of a bulk strain or displacement into separate structural domains with contrasting strain characteristics, although it is sometimes also used in relation to the *conceptual* factorisation of homogeneous strains into separate strain components as an aid to understanding (e.g. Lister and Williams, 1983, p. 6).

Kinematic strain partitioning at convergent plate margins is usually depicted in terms of the separation of regional bulk strain into a broad zone of predominantly orogen-normal shortening in the frontal parts of the over-riding plate, flanked by a major strike-slip fault or zone of strike-slip further from the plate margin. Structural observations and seismic studies of active convergent margins suggest that strain partitioning is so ubiquitous (e.g. references cited in Tikoff and Teyssier (1994), Fossen et al. (1994), Jones and Tanner (1995) and Holdsworth et al. (1998)) that it must be considered as the normal way for non-coaxial non-plane strains to be accommodated. With plate-scale strain partitioning of this type, the standard compatibility rules need not be violated, as long as the boundaries between the deformational domains are mutually compatible (Fig. 5).

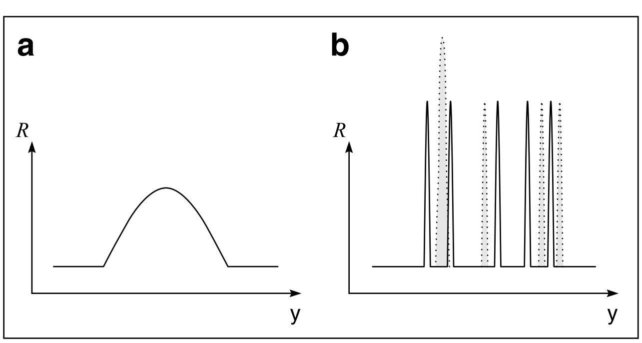


Fig. 4. Graphs showing variation in strain intensity, *R*, across the width of the shear zone, *y*. (a) Heterogeneous strain intensity varies smoothly across the shear zone (e.g. Fig. 2b), so strain can be described by a regular mathematical function, such as $R=f(\sin y)$. (b) Deformation is concentrated into narrow zones of high strain intensity and the bulk strain distribution cannot be described by standard linear equations. In any anastomosing system of shear zones, adjacent strain profiles will inevitably vary (c.f. solid and dotted curves) along different vertical (Fig. 3a and c) and/or horizontal transects through the deformation zone (Fig. 3b and c).

2.5. Distributed strain partitioning at multiple scales

Examples of the compartmentalization of strain components into distinct deformational domains are shown simplistically in Fig. 5 with partitioning occurring only at the plate scale. However, in our experience, deformation zones in the upper crust that have been subjected to bulk non-coaxial non-plane strain are commonly characterized by a high degree of strain heterogeneity. In such cases, mutually cross-cutting and inter-linking sets of structures show that both strain intensity and kinematic partitioning have occurred at several scales simultaneously. This is illustrated in the following section using outcrop, map and regional-scale structures from the Southern Uplands Terrane in Scotland.

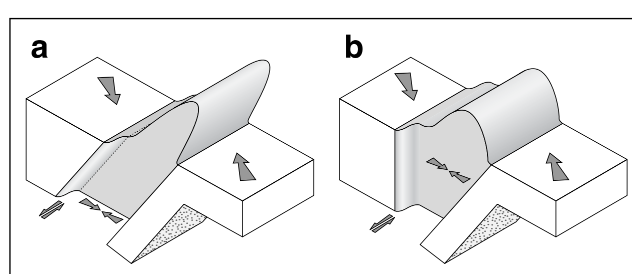


Fig. 5. Strain partitioning of bulk non-coaxial non-plane strain (large arrows) into a broad zone of heterogeneous coaxial deformation (crustal shortening and thickening) and a narrower zone of heterogeneous non-coaxial (strike-slip) deformation. (a) Inclined deformation zone with parallel zone margins. (b) Zone with wedge geometry.

3. Scaling, heterogeneity and partitioning in the Southern Uplands Terrane, Scotland

The Southern Uplands Terrane is a broad belt of rocks consisting largely of turbiditic deposits of Ordovician and Silurian age that were deformed during the oblique closure of the Iapetus Ocean in Palaeozoic times (Soper and Hutton, 1984). The terrane consists of a series of NW-dipping fault-bounded tectonostratigraphic units (Fig. 6). Within each unit, the beds mostly young to the NW but adjacent collective units young successively to the SE on a regional scale. The terrane is variously interpreted as an accretionary prism (Leggett et al., 1979; Needham and Knipe, 1986) or other type of subduction-related basin (Hutton and Murphy, 1987; Stone et al., 1987). Rocks currently exposed at the surface have experienced low grade metamorphism (diagenetic to prehnite–pumpellyite facies), suggesting that deformation occurred at shallow crustal levels (Oliver and Leggett, 1980).

The structural character and evolution of the Southern Uplands have been studied extensively over many decades: recent overviews are given by Woodcock and Strachan (2000) and Holdsworth et al. (2002a,b) and references therein. In this paper, we focus on deformation at a range of scales and describe the way in which strain is heterogeneously distributed from the plate-scale down to small-scale structures in specific outcrops.

3.1. Supra-kilometric regional-scale structures ($>10^3$ m)

The overall structure of the Southern Uplands was first described in detail by Peach and Horne (1899), who delineated a series of NE/SW tracts that lie parallel to orogenic strike and highlighted the contrast between the extensive continuity of the tracts along strike, compared with the variation in stratigraphy and structure perpendicular to the tract boundaries (Fig. 6). Several tracts can be traced along the entire strike length of the Southern Uplands into equivalent rocks of the Longford Down Terrane in Ireland, a distance of 400 km or more. Peach and Horne's original division of the terrane into Northern, Central and Southern belts has been subsequently further sub-divided into narrower fault bounded tracts with typical widths of 10^3–10^4 m (Leggett et al., 1979; Murphy et al., 1991). Thus the along-strike dimension of the tracts is typically at least one or two orders of magnitude larger than their across-strike width.

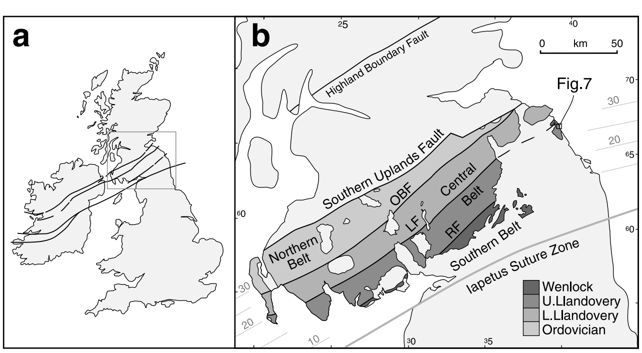


Fig. 6. Regional tectonostratigraphy of the Southern Uplands Terrane, Scotland. (a) Location of the Southern Uplands in Scotland, and its along-strike equivalent, the Longford Down Massif in Ireland. (b) Delineation of the Southern Uplands Terrane into Northern, Central, and Southern Belts. Tract-bounding faults are: OBF, Orlock Bridge Fault; LF, Laurieston Fault; RF, Riccarton Fault. Dotted contours in the North and Irish Seas show depths in kilometres to postulated Iapetus seismic reflector (after Soper et al., 1992).

Throughout the Southern Uplands Terrane there is widespread evidence of extensive, long-lasting shortening recorded by top-to-SE thrusting on tract-bounding faults (Needham and Knipe, 1986). In addition, there is also a significant component of sinistral shear which is of regional importance from late Llandovery onwards (Anderson and Oliver, 1986; Anderson, 1987). Strike-slip deformation is typically concentrated within narrow fault zones, some of which in the north of the region were probably active earlier as thrust faults. Contractional deformation is regionally diachronous, with the main period of shortening becoming younger to the SE, as progressively younger sediments became incorporated into the thrust sequence (Barnes et al., 1989). Consequently, in the Northern Belt and northern part of the Central Belt (e.g. Anderson and Oliver, 1986; Holdsworth et al., 2002b), the evidence for sinistral shear clearly overprints earlier shortening structures, whereas deformation within the tracts further south in late Llandovery times is characterized by synchronous shortening and strike-slip (e.g. Anderson, 1987; Barnes et al., 1989; Holdsworth et al., 2002a). Later sections of this paper focus on progressively smaller-scale structures from rocks that were deformed in this transpressional regime.

Although most of the Southern Uplands typically displays top-to-SE vergence, there are zones in which back-thrusting and top-to-NW verging structures are much more prevalent. These zones are commonly localized within the Central Belt in the southern part of the tract that lies to the north of the Laurieston Fault and appear to separate areas of the Southern Uplands Terrane in which shortening precedes sinistral strike-slip from those in which shortening and strike-slip are synchronous (e.g. fig. 6 of Holdsworth et al., 2002b).

In summary, at the uppermost scale of observation, the Southern Uplands is a distinct tectonostratigraphic terrane with a width that is of the same scale as the thickness of the crust and a length along strike that has (at least) the same order of magnitude as the thickness of the lithosphere. Taking the Southern Uplands as a whole, the terrane represents structural anisotropy on a crustal scale that helps define the NE/SW structural grain of the British/Irish Caledonides. Although deformation of Caledonide blocks adjacent to the Southern Uplands has also been the focus of much geological research over many decades, the critical point in relation to this paper is that there is no direct mechanical continuity of rock units between the Southern

Uplands Terrane and adjacent regions across strike. At the present day surface the terrane as a whole is entirely bounded by mechanical discontinuities and it is therefore not possible to determine by direct observation whether or not strain compatibility is maintained across strike on a regional scale.

Anisotropy of deformation is also reflected at subordinate scales of observation *within* the Southern Uplands Terrane. On a regional scale, individual tracts and tract-bounding faults are oriented parallel to the tectonic grain. Adjacent tracts can display highly contrasting styles of deformation (e.g. Barnes et al., 1989; Holdsworth et al., 2002b), which raise issues of strain compatibility between neighbouring tracts. However, because the tracts appear to be entirely bounded by brittle faults that either extend along the whole length of the terrane or are mechanically linked to other faults of similar or larger magnitude, it is not possible to determine how strain compatibility is maintained (or whether it is maintained at all).

3.2. Decametric–kilometric map-scale structures (10^1–10^3 m)

As with regional deformation described at larger scales, strain on the scale of tens to thousands of metres is characterized by variations in strain intensity and the kinematic partitioning of strike-slip components of deformation into individual domains. Fig. 7 shows an example of map-scale strain anisotropy from the well-exposed coastal section south of Eyemouth, SE Scotland, described in more detail by Holdsworth et al. (2002a), Clegg (2002), Jones et al. (2004a) and Tavarnelli et al. (2004). Domain boundaries are aligned parallel to the larger scale tract-bounding faults described earlier. The domains differ in the ratio of strike-slip and shortening deformation, and the way in which each of these strain components is accommodated by folding and faulting. The different types of domain are shown schematically as block diagrams within Fig. 7. The domains, which are described in detail in Holdsworth et al. (2002a), can be summarized as follows:

(i) *Domain type 1*: a bulk inclined transpression characterized by NW-dipping and NW-younging homoclinal beds and moderately to steeply SW-plunging and-facing folds.
(ii) *Domain type 2*: shortening dominated, with subsidiary top-to-SE shear, represented by upright or slightly inclined, upwards-facing whaleback folds with low angles of plunge.
(iii) *Domain type 3*: strike-slip dominated, characterized by sinistral faults and sinistrally verging, highly curvilinear, upwards and downwards facing folds, in which the bisector of the facing direction is approximately parallel to strike and plunges shallowly to the SW.
(iv) *Domain type 4*: a bulk inclined transpression with a significant strike-slip component. This type of domain is characterized by highly heterogeneous deformation in which strain is further partitioned at subsidiary scales (10^{-1}–10^1 m) into separate sub-domains of interlinked strike-slip, dip-slip and oblique-slip detachments and contractional fold structures with variable facing directions (Fig. 8). This type of domain is well exposed at Dulse Craig at the northern end of the Eyemouth section shown in Fig. 7, and is described in detail by Tavarnelli et al. (2004).

Because strain is kinematically partitioned into map-scale domains, the bulk strain within a single domain at the map-scale does not necessarily reflect the overall bulk strain of domains at larger scales or the bulk strain of the Southern Uplands Terrane as a whole. Because of this, we obviously need to address issues of strain compatibility between adjacent domains. Whereas regional-scale domains ($>10^3$ m) are delineated by large tract-bounding faults, the map-scale domains at Eyemouth (10^1–10^3 m) are generally not bounded by single discrete discontinuities that are easily recognized in the field. Instead, the domains on this scale are separated by numerous bedding-parallel detachments, and by highly interlinked arrays of small-scale structures (typically 10^{-1}–10^1 m) that are described in subsequent sections. Consequently, compatibility between map-scale domains is dependant upon the combined compatibility of myriads of individual outcrop-scale structures that are linked together to form pervasive fracture arrays.

3.3. Millimetric–decametric outcrop-scale structures (10^{-3}–10^1 m)

Throughout the Eyemouth section, there is a wide variety of macroscopic structures visible at the outcrop scale. These are described in detail by Holdsworth et al. (2002a), Clegg (2002) and Tavarnelli et al. (2004). Widespread shortening across the section is shown by folding of homoclinal bedding, predominantly through flexural-slip mechanisms. The rocks are also pervasively fractured and faults generally form anastomosing arrays of fractures at all observed scales. There is a large amount of vein material lining fracture surfaces and filling tension gashes and small-scale intra-bed pull-aparts. Evidence for non-coaxial strain is shown by the predominance of sinistral detachment faults and fault duplexes (Riedel and P shears), steeply plunging folds with 'S' vergence, transected fold cleavage, flexural-slip lineations that are systematically oblique to fold hinges and the asymmetry of pull-aparts and other structures.

Individual outcrop-scale structures have widely varying kinematic significance with respect to larger-scale domains and regional deformation (Jones et al., 2004a). Both folds and faults display strike-slip, dip-slip, oblique-slip or shortening, or combinations of these end-member strain components (Fig. 9). Thus, deformation at the outcrop scale is highly heterogeneous and is characterized by widespread

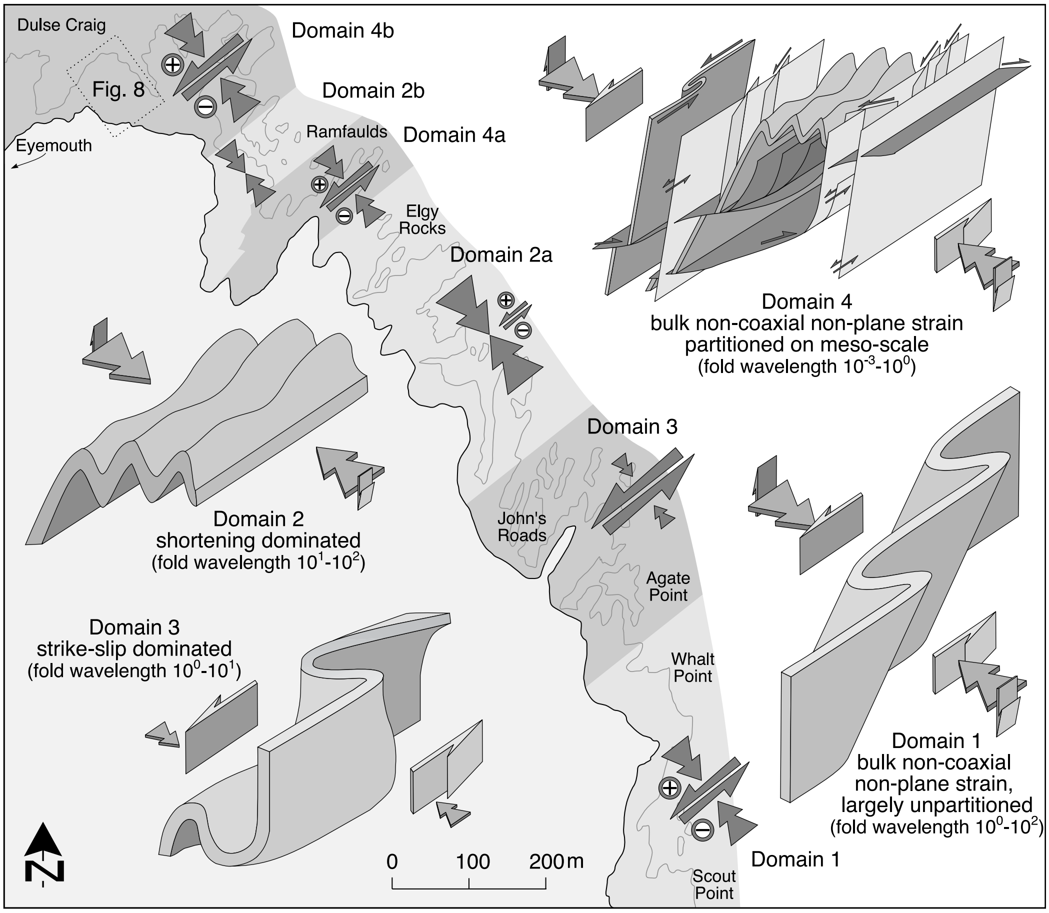


Fig. 7. Map scale domains at Eyemouth. The type of bulk strain for each domain is shown schematically in the 3D sketch diagrams. The boundaries between map scale domains are not defined by individual map-scale faults; domains are defined by a predominance of specific types of smaller scale structures (see Figs. 8 and 9) and are separated by arrays of interlinked detachments.

kinematic partitioning and local variations in strain intensity.

One consequence of the relatively high level of upper crustal strain in the Eyemouth section is that the compatibility of outcrop-scale structures can be difficult to evaluate. Some vein-filled structures such as tension gashes and pull-aparts clearly represent a breakdown in the mechanical continuum on a small-scale (c.f. fig. 33.1G of Ramsay and Lisle, 2000). These structures are typically encompassed by largely coherent rock material and do not seem to represent any significant compatibility issue beyond the small scale of the actual localized structure itself. In contrast, the vast majority of fractures in the section are not isolated discontinuities embedded within a rock continuum, but rather, are geometrically linked with larger (and smaller) scale fractures to form a continuous network of mechanical discontinuities that completely disrupts the coherence of the rock continuum. Individual segments of faults can have highly variable geometries and often curve in 3D space to give changes in strike and/or switches in dip-direction over short distances. Slickenlines and slickenfibres commonly show curved displacement trajectories and/or overprinting histories in which the displacement vector changes dramatically between successive movements. Displacement along such fracture arrays must have caused significant strain compatibility problems within the intervening blocks and it is rarely clear from field observations how compatibility was achieved.

3.4. Scale attributes of strain heterogeneity at Eyemouth

Clearly, the distribution of strain at Eyemouth associated with bedding, cleavage, folding, veins and especially discontinuities (i.e. fractures) is highly heterogeneous. Heterogeneity is not smoothly distributed, but occurs at several punctuated scales from 10^{-3}–10^{5} m. At some intervening scales, the deformation can be approximated as homogeneous, but any area of broadly homogeneous strain is bounded by heterogeneity at both smaller and larger scales. In particular, heterogeneity is most pronounced at scales in which there is inherent anisotropy prior to deformation, or mechanical anisotropy that develops during deformation. Examples of the former include the scale of typical bed thickness (10^{-3}–10^{0} m), and multilayer

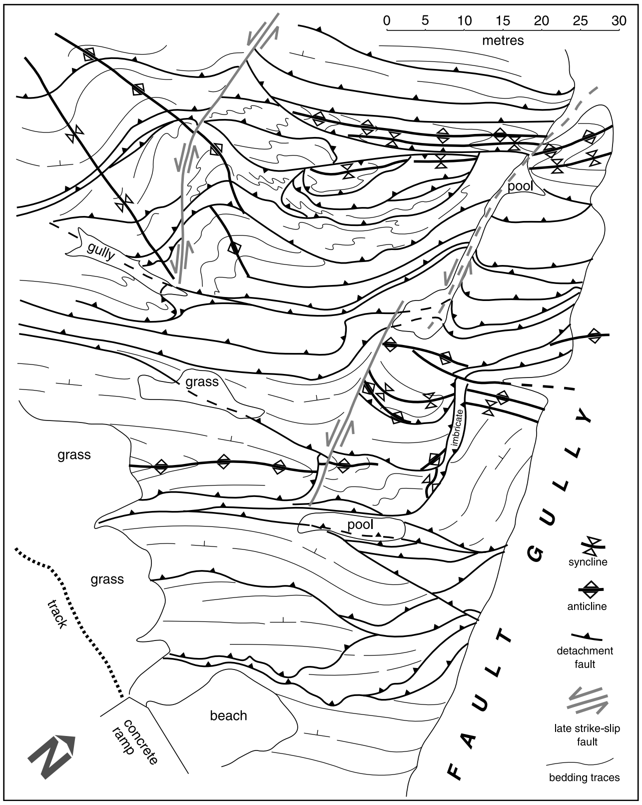


Fig. 8. Generalised map of Dulse Craig showing kinematic linking of arrays of strike-slip, oblique-slip and dip-slip detachments (simplified from Tavarnelli et al., 2004). Individual detachment surfaces have complicated 3D geometry and often show variable kinematics along their length. Dip of bedding varies considerably but is typically 40–60° NW.

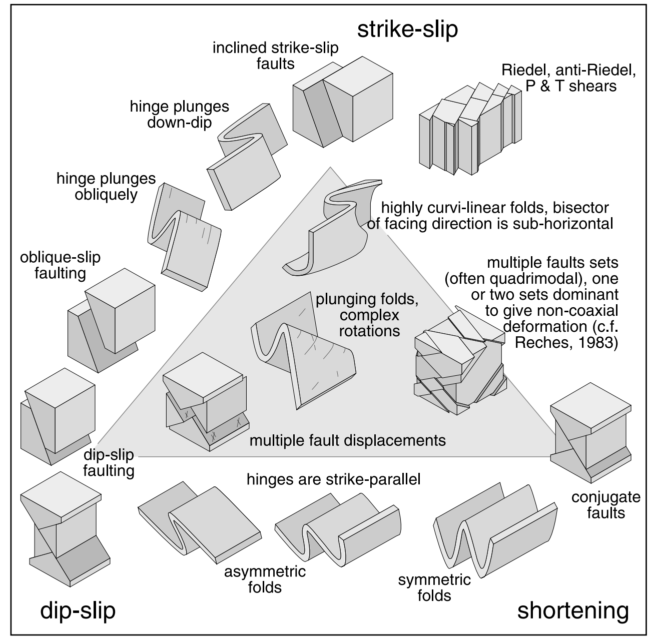


Fig. 9. Typical outcrop-scale structures seen at Eyemouth showing their inferred kinematic significance with respect to domain and deformation zone boundaries. The structures are plotted on a strain triangle in which the apices represent individual plane strain components and the inside of the triangle represents combinations of these components to give non-coaxial non-plane strains.

attributes (Ramsay and Huber, 1987), such as packing distances of competent layers (10^{-1}–10^{1} m) and the overall thickness of the whole turbidite cover sequence (10^{3}–10^{4} m). Heterogeneities that exist prior to the onset of deformation influence the growth of structures, which in turn add further anisotropy on different scales as deformation progresses. Examples of the latter include the dominant wavelength and amplitude of folds (10^{-2}–10^{2} m), and the typical width of regional tracts between tract-bounding faults (10^{3}–10^{4} m).

For some particular scales of observation, strain can be markedly heterogeneous in some areas but more homogeneous in others. For example, we might take a 100 m$\times$100 m sampling area divided into 10 m$\times$10 m grids to delineate structures with a magnitude of 10 m or greater (smaller structures are ignored). Sampling within many areas of the homoclinal sequence of domain 1 (see Fig. 7) would reveal a relatively homogeneous deformation at this scale, compared with equivalent sample areas in domain 4 at Dulse Craig in which strain is partitioned between a variety of different styles of sub-domain, representing high strain heterogeneity at the chosen scale of analysis (compare Figs. 7 and 8).

In summary, the structural style at Eyemouth shows that strain heterogeneity is highly scale dependant and that partitioning can occur at several scales simultaneously. Kinematics and geometry at a given scale principally reflect the deformation of that localized domain, which to maintain compatibility need not necessarily mirror the kinematics of larger domains in an immediately obvious way. When this is the case, a large amount of structural data must be collected from as wide an area as possible and inferences about kinematics cannot be reliably made at scales of magnitude greater than the largest sampling size for which data are available.

3.5. Future work at Eyemouth

One way to investigate the spatial variation across the wide range of scales observed in deformation zones such as Eyemouth, is to combine geostatistical sampling and analysis methods with recent technological developments in digital field data acquisition (e.g. Jones et al., 2004b; McCaffrey et al., in press). Geostatistical methods provide a powerful tool to characterise the spatial dependence of any rock property that varies continuously or discontinuously in space (Chilès and Delfiner, 1999). Indeed in geostatistics, 'structural analysis' refers to quantifying and modelling an observation's spatial variability, including regional trends, anisotropy and characteristic scales. Highly portable geographical information systems (GIS) systems linked to global positioning systems (GPS), provide automatic three-dimensional geospatial control on observations/ sample positions and allow multi-attribute mapping (Maerten et al., 2001). In addition to providing efficient field data capture facilities, the digital work-flow provides spatially referenced visualization of the field data and the subsequent geostatistical output. Field data collection requires a more rigorous approach to sampling, using hierarchical and nested methods, to determine sampling grids that optimise sampling efficiency, for heterogeneous deformation zones such as that exposed at Eyemouth. In this way, field observations may be interpolated throughout unexposed parts of the deformation zone and errors associated with those predictions are automatically determined. The new digital structural analysis is theoretically capable of providing a more rigorous approach to the characterization and determination of spatial patterns in heterogeneous deformation zones. In addition to increased spatial precision, digital structural mapping also provides a suitable platform for more precise mapping of temporal relationships of cross-cutting structures, using the quantitative approach of Potts and Reddy (1999, 2000).

4. Discussion

By emphasising the fundamental importance of the principle of strain compatibility more than three decades ago, John Ramsay laid the foundation for the quantitative description, measurement and interpretation of field structures that is in common use today. Application of such methodologies in orogenic regions continues to help structural geologists in bridging the scale divide between local outcrop and global tectonics referred to by Ramsay and Lisle (2000).

Bulk deformation (c.f. *mean strain* of Ramsay, 1976, p. 6) at convergent plate margins must be transpressional unless relative motion is orthogonal; this has lead to the wide application of transpression models in areas that have experienced oblique shortening (e.g. papers in Holdsworth et al., 1998). Symbolic models of homogeneous transpression (c.f. Fig. 1) can appear problematical when interpreted literally (e.g. Schwerdtner, 1989), based on considerations of strain compatibility that are easily resolved in terms of smoothly varying heterogeneous strain (c.f. Figs. 2–4). Paradoxically, the compatibility issues we discuss in this paper are not restricted to transpression, but are also applicable to homogeneous plane strains involving pure shear (Fig. 1a). Therefore, the following discussion about strain compatibility is relevant to deformation at convergent plate margins irrespective of whether relative plate motion is oblique or orthogonal.

4.1. Compatibility and ‘cream cake’ tectonics

Whereas strain in an ideal simple shear zone is inherently compatible with surrounding wall rock, there are potential compatibility issues associated with zones of coaxial deformation (pure shear). This has been referred to by Ramsay and Huber (1987, pp. 610–613) as the ‘cream cake effect’ and applies both to pure shear that is plane strain (as shown in fig. 26.24 of Ramsay and Huber, 1987) or non-plane strain. Compatibility of pure shear can be achieved at the plate scale because the upper surface of the Earth is a free surface. In terms of the mathematical compatibility criteria, the Earth’s surface represents a degree of freedom that allows crustal scale shortening to be accommodated by orogenic thickening, without causing upper crustal incompatibility (issues regarding lower crustal- and lithosphere-scale compatibility are discussed later). This applies whether deformation is a bulk pure shear (Fig. 1a) or a more complicated deformation that contains a pure shear component (such as fig. 26.22 of Ramsay and Huber, 1987; sub-simple shear of Simpson and DePaor 1993) and it applies both to the non-coaxial plane strains shown in Figs. 1c and 2c and the non-coaxial non-plane strains in Figs. 1d, 2b and d, and 3c.

In most orogenic belts there is abundant evidence for non-coaxial deformation associated with orogenic shortening and thickening. Large scale fold and thrust nappes typically transport hanging wall units many kilometres across orogenic strike, and the last two decades has seen a general consensus of opinion that mountains form primarily by thrust stacking and that this is essentially a simple shear process. However, it is invalid to automatically equate non-coaxiality with simple shear (Paterson and Weiss, 1961; Hossack, 1968) and the existence of large-scale thrusts does not necessarily imply that orogenic shortening is accomplished entirely by crustal-scale simple shear. The implications of a component of coaxial shear during thrusting are discussed by Sanderson (1982). Workers such as Law et al. (1984, 1986) and Holdsworth and Grant (1990) have shown the importance of coaxial deformation related to the Moine Thrust and overlying Moine Nappe in NW Scotland. Several recent studies in the Himalayas have indicated that bulk strain contains an important component of pure shear, leading to significant orogenic shortening and associated up-dip extrusion (Beaumont et al., 2001; Vannay and Grasemann, 2001; Grujic et al., 2002). Law et al. (2002) quantify sub-simple shear strains with a dominant pure shear component in the footwall to the South Tibetan Detachment System, an active top-to-the-north extensional fault on the north side of the Everest massif. In the Nanga Parbat syntaxis in the Pakistan Himalaya, ongoing shortening is causing extrusion of crustal rocks from approximately 25 km depth (Butler et al., 2002). Although there is evidence for highly non-coaxial strains on the margins of the Nanga Parbat Massif, deformation away from the margins is dominated by up-dip extrusion on a crustal scale. Ongoing stretch is aided by the arid climate, sparse vegetation and the enormous sediment-bearing capacity of the adjacent Indus river.

Studies such as these lead us to believe that Ramsay and

Huber (1987) have over-emphasized the potential compatibility problems arising from large scale ‘cream cake’ tectonics and have underestimated the role of the Earth’s free surface in allowing compatibility to be achieved through extensive up-dip stretching arising from crustal scale deformation with a significant pure shear component. Compatibility can be achieved in this way for both plane strain as well as non-coaxial non-plane strain (i.e. transpressional) deformation. Furthermore, as the boundaries to real examples of such deformation zones are typically major crustal discontinuities, it appears that sufficient compatibility can be achieved in nature without needing smoothly varying strain heterogeneity and that therefore the representations shown in Fig. 1 might be as applicable to some convergent margins as those in Fig. 2 (as first-order approximations), however heretical this might seem at first glance.

4.2. Compatibility of lateral and basal zone boundaries

In the discussion so far, we have focused on the upper boundary surface and the strike-parallel front and rear deformation zone boundaries that define the orogenic grain. Equally important are the other boundaries to the zone, which must also maintain compatibility with surrounding rock. All margins have finite lateral extent and are bounded by other plates at triple junctions (Dewey, 1975; Dewey et al., 1998). Also, most convergent margins have along-strike strain heterogeneity caused by variations in the orientation of the zone boundary and by lateral variations in the extent of regional batholiths and other tectono-stratigraphic units. At the plate scale, lateral compatibility must be maintained by the mutual interaction of plates and micro-plates on a global scale.

Orogenic shortening in the upper crust must be matched in the lower crust and upper mantle to achieve lithospheric balance (e.g. Oldow et al., 1990). Although the extent to which brittle deformation in the upper crust is mechanically coupled to ductile deformation in the lower crust remains controversial, it is assumed that thin-skinned tectonic shortening must, in some way, be balanced at depth in order to preserve compatibility. Unfortunately, attempts at orogenic mass balancing on a crustal or lithospheric scale are not simple because crustal mass can be added at continental margins by obduction, accretion and by fractionation during melting of subducting oceanic lithosphere, leading to batholith emplacement and orogenic underplating of the overlying continental crust.

Deformation in the lower crust is generally believed to be more distributed and less localized than upper-crustal deformation, although studies of terranes that deformed at such depths clearly show that strain is commonly heterogeneous (e.g. Coward, 1984). Kinematic strain partitioning at lower crustal levels is common on a wide range of scales, and strain commonly localizes into ductile shear zones of high strain intensity separated by areas of lesser strain. The boundary between upper and lower crust is often envisaged as a sub-horizontal transition zone in which rheological variations between brittle and ductile deformation are accommodated (Richard and Cobbold 1989; Lettis and Hanson, 1991; Teyssier and Tikoff, 1998; Teyssier et al., 2002; Tikoff et al., 2002). Crustal strength profiles based on laboratory experiments (discussed in Molnar (1988, 1992) and England and Jackson (1989)) predict such a brittle–ductile transition zone at mid to lower crustal levels, as well as a comparable zone at the crust–mantle boundary (continental Moho). Field examples of possible mid-crustal transition zones are described by Holdsworth and Strachan (1991), Pavlis and Sisson (1995), Wynn (1995) and Garde et al. (2002).

In summary, lateral and basal compatibility are important issues in relation to deformation at plate margins. Although our current understanding of both lateral and basal compatibility issues is not complete, potential compatibility problems are equally relevant to models in which orogenesis is viewed in terms of plane strain orthogonal shortening, as to those which model non-coaxial non-plane strains arising from oblique motion. In particular, we agree with Teyssier and Tikoff (1998) that consideration of basal boundary conditions at convergent plate margins does not raise compatibility issues that invalidate the use of transpressional models.

4.3. Validity of the continuum approximation

At the core of the principle of strain compatibility lies the notion that rock undergoing deformation forms a mechanical continuum. Solutions for strain compatibility shown in Figs. 2–5 conform to this broad assumption. In structural geology, the concept of the continuum is used in subtly different ways. In the most general sense, the concept of a continuum is used to denote that when a body of rock is strained, it deforms without voids or gaps appearing within the body (Ramsay, 1976, p. 6; Ramsay and Huber, 1983, p. 33). This is essentially a geometrical description and we therefore refer to it here as the ‘geometrical continuum’. In a more rigorous sense, the concept of a continuum is also used to mean that every point in a deforming body remains physically connected to adjacent points in the material throughout the deformation process, such that material discontinuities do not develop. This is termed here the ‘material continuum’. It is this sense of the term ‘continuum’ that forms the basic assumption behind the compatibility equations in the form given by Ramsay and Graham (1970, p. 793), Ramsay and Huber (1983, p. 34), etc. and the use of continuum mechanics to study geological deformation (Ramsay and Lisle, 2000). Although the concept of a geometrical continuum can be applied to any deformation from sub-grain scales to the whole Earth, the applicability of a material continuum is inherently limited at upper and lower scales. In crustal regions, the lower scale limit at which the approximation of a material continuum is valid is represented by grain-scale processes (Ramsay, 1976, pp. 6–7; Ramsay and Huber, 1983, pp. 19–20), and the upper limit by the finite lateral and vertical extent of individual lithospheric plates.

When rocks in the upper crust are deformed beyond the point of brittle failure, they no longer represent a material continuum. In some situations the loss of material continuum occurs only on a small scale and the lack of a strict material continuum can be readily ignored by considering deformation on a larger scale of observation (Ramsay and Lisle, 2000, p. 813). In such cases, structures that represent material discontinuities should be heterogeneities of localized extent that are entirely enveloped by the continuum at a larger scale or that have negligible effect on the bulk strain at the larger scale. For example, individual small-scale structures seen at Eyemouth, such as tension gashes, pull-aparts, small-scale shear joints, boudinage and vein arrays, all represent localized departures from a strict mechanical continuum. But because these structures are usually of a similar magnitude (or smaller) than typical bedding thickness, on the outcrop scale their effect is generally minor and can usually be ignored when considering deformation on a larger scale. Similarly, folds at Eyemouth that have developed by flexural-slip contain discontinuities along bedding planes, but since the layer sequence as a whole forms recognizable continuous fold structures, bedding-parallel discontinuities can be ignored on the scale of the folds as a whole. Similar arguments can be applied to individual faults that are mechanically isolated and limited in extent, such as faults that pass into a region of elastic or ductile strain (see fig. 33.1F of Ramsay and Lisle, 2000). Examples of structures such as these reinforce the point emphasized by workers such as Paterson and Weiss (1961) and Lister and Williams (1983) that the concept of a continuum is highly scale-dependent.

The linked detachment fault arrays in zones such as Eyemouth do not conform to the kind of limited loss of material continuum described above. In our experience, upper crustal deformation zones in which strain is moderate or high are dominated by faults that are not mechanically isolated, but instead form highly interconnected arrays of fractures that span several orders of magnitude (e.g. Tchalenko, 1970; Dewey et al., 1986; Davison, 1994 and references therein). In such situations, the majority of faults do not terminate in areas of elastic or ductile deformation but are, instead, kinematically linked to faults of similar or larger orders of magnitude. In the case of Eyemouth, this is visible from the centimetre to kilometre scale and probably continues upwards to the plate-scale of the Southern Uplands as a whole. In such situations, it might be possible to view parts of the deformation as approximating a material continuum by considering only limited volumes of rock at one particular, arbitrary scale, but this scale will always be bounded by both larger and smaller scales in which the approximation of material continuum breaks down. In other words, when there is a high degree of fault interconnectivity, the general state is a lack of material continuum throughout the rock and a high degree of kinematic continuity amongst the block-bounding faults of different sizes. In such cases, the concept of a material continuum does not adequately capture the nature of deformation; the key to understanding areas of high brittle strain may be to increase focus on the heterogeneity of strain rather than trying to find homogeneity at arbitrary scales.

During deformation, upper crustal rocks generally maintain *geometrical* continuity even when material continuity is diminished or absent. This is central to the rationale of section balancing techniques, which allow geologists to test geometrical compatibility in deformed rocks that are dissected by material discontinuities (i.e. fault duplexes). The validity of the technique relies on the ability to match stratigraphy across faults, rather than direct measurements of strain. In some regions of high brittle strain, even when material continuity is lost and fracture continuity is high, the kinematics of upper-crustal deformation nevertheless often appear to directly reflect the orientation of regional plate motion derived from geodetic data and geophysical measurements from the lower crust and lithospheric mantle. In such situations, it is difficult to use finite strain to test whether the assumption of material continuity is strictly valid: assuming an approximation of continuity is generally adequate to explain the finite strain observed. However, in some areas of upper-crustal deformation, particularly when plate convergence is oblique and bulk strain is non-coaxial and non-plane strain, the lack of material continuity leads to micro-plate rotations and large scale motion of crustal blocks along the orogenic margin.

Although the material continuum in the lower crust is maintained by deformation mechanisms that characterize ductile rheologies, the geometrical continuum in areas of high strain in the upper crust is maintained by gravity and the frictional properties of fractures. Brittle–ductile transition zones mark the level at which the material continuum can start to lose coherence with increasing strain and to be replaced upwards by interconnected arrays of faults acting on different scales simultaneously. Transition zones allow rocks in the upper crust to be *mechanically* linked to the lower crust, even when material continuity is lost at higher crustal levels. Mechanical coupling is provided both by material connectivity and by frictional force, the former dominant in ductile deformation and in brittle deformation when strain is low and the latter increasing in importance as strain increases in brittle regimes.

4.4. Temporal dependency and fault weakening

Once isolated fractures interact and propagate to form pervasive interconnecting fault arrays, the upper crust becomes markedly anisotropic. If internal fault weakening occurs along large-scale faults, the fracture network as a whole should represent a significant crustal weakness. Whilst the heterogeneity of the rock mass is dependant upon the spatial scale of observation, the localization of strain onto individual weak faults and transmission of the weakening within the network also depend on the *temporal* scale of analysis. Most fault networks are geometrically irregular, with numerous bends, branches, steps and jogs (e.g. Scholz, 2002), making it difficult to maintain a continuous interconnected weak layer from grain to crustal scales at all times during displacement. Irregular segments will inevitably experience cycles of periodic lock-up and breaching at all scales within the fault network. This geometrically controlled stick-slip behaviour means that the whole network may only be considered as weak on geological timescales, whilst its mechanical behaviour over shorter timescales—including those captured by most geophysical methods (e.g. in situ stress analysis)—is ambiguous (Holdsworth, 2004). This may explain why much of the evidence concerning the apparent weakness of major structures such as the San Andreas Fault has remained problematic and controversial (e.g. Scholz, 1996, 2000; Zoback, 2000).

5. Conclusions

Dominant factors controlling the deformation of the upper crust during mountain building include:

(a) the boundary conditions arising from relative plate motion orthogonal or oblique to the plate margin, including basal boundary conditions and the coupling between the more brittle upper crust and the ductile lower crust,
(b) the boundary conditions at the Earth's upper free surface, including gravitational instability and the capability of erosional systems to denude uplifted crust,
(c) the development of a kinematically interlinked discontinuity network (spanning several scales of magnitude) that cuts the Earth's surface, with the corresponding degradation of the material continuum,
(d) frictional properties of fault discontinuities and intervening regions of intact rock,
(e) rheological heterogeneity and material anisotropy of crustal rocks,
(f) longitudinal variations in (a)–(e) along the orogen and lateral boundary conditions at the ends of the mountain belt.

The structural style of deformation seen at Eyemouth (and similar zones of moderate to high strain in upper crustal rocks) helps us to make inferences about the way in which deformation can be accommodated during mountain building processes, and how upper crustal strain compatibility at the plate scale can be achieved by the partitioning of strain at several smaller scales simultaneously (Jones et al., 2004a). It is possible to define basic models of mountain building that reflect bulk orogenic shortening and thickening without violating the standard rules of mathematical strain compatibility (Fig. 2). Crustal thickening is represented by a component of heterogeneous pure shear that increases smoothly towards the centre of the deformation zone, allowing compatibility to be maintained with the zone boundaries. These models recognize the critical importance of the upper free surface of the Earth in maintaining bulk strain compatibility and in generating the global topography we see in basins and mountain belts. Compatibility is maintained whether relative plate motion is orthogonal or oblique to the plate margins. The models can be extended to show other types of strain heterogeneity, including deformation characterized by anastomosing shear zones (Fig. 3) or strain partitioning (Fig. 4). In all these simplistic models, strain compatibility is ensured by maintaining material continuity from the undeformed margins across the whole deformation zone.

In areas of ductile deformation that form during mountain building processes, the approximation of a continuum is largely applicable, at least within lower and upper limits (the grain-scale and plate-scale, respectively). In the brittle upper crust (the 'schizosphere' of Scholz, 2002), rocks in many areas may also approximate to a material continuum, particularly when the magnitude of brittle strain is low. However, with increasing brittle strain, fracture networks begin to propagate, the material continuity starts to diminish and the rock increasingly acts as a semi-continuum (c.f. Shubnikov, 1930).

Further deformation causes the coherence of the material continuum to become severely compromised, with the development of arrays of kinematically interlinked fault discontinuities acting on several orders of magnitude simultaneously. In such situations it might be more appropriate to describe the strain using transformation equations that contain highly non-linear (power-law) functions in order to capture the range of active scales of deformation. If the network of discontinuities becomes inter-connected along the whole length of an orogenic belt and through the main load-bearing regions of the lithosphere at depth, the material continuum is violated and there is increased likelihood that the finite strain observed within individual crustal blocks (of different scales) along the deforming plate margin does not directly reflect overall deformation caused by plate motion (e.g. figs. 1 and 2 of England and Jackson, 1989). In this situation, the overall kinematic and mechanical behaviour of the system will be determined respectively by the kinematic boundary conditions and the rheological properties of the fault rocks within the interlinked discontinuities (e.g. Rutter et al., 2001).

Much further work is needed to improve our understanding of the way in which deformation caused by plate motion is balanced between upper crust, lower crust and lithospheric mantle (Molnar, 1988; Jackson, 2002; Tikoff et al., 2002). Deformation in the lower and upper crust may be mechanically linked by flat-lying transition zones that span the brittle–ductile boundary. Although brittle deformation in the upper crust can cause a breakdown in the *material* continuum, deformed rocks retain *geometrical* continuity. Kinematic linkage of the fault networks means that fault-bounded blocks do not move randomly with respect to one another in response to deformation. However, the heterogeneity represented by the fault networks can cause complex patterns of three-dimensional fault movements in which it is not always clear how individual fault displacements relate to plate motion. Furthermore, the temporal relationship between the accommodation of displacement and mechanical behaviour of discontinuous fault networks is complicated by their evolving geometric irregularity. The deformation is not chaotic, but the way in which it is ordered from grain to plate scale is difficult to understand from a starting point that assumes material continuity. Strain compatibility in areas of high strain in the schizosphere (*sensu* Scholz, 2002) is maintained by partitioning strain amongst structures that have a wide variety of kinematic effects, over a wide range of scales simultaneously (at least on geological timescales) and not by smooth variations in strain magnitude acting on a single scale across a material continuum.

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