Intraplate strike-slip deformation belts

F. STORTI[1], R. E. HOLDSWORTH[2] & F. SALVINI[1]

[1]*Dipartimento di Scienze Geologiche, Università "Roma Tre", Largo S. L. Murialdo 1, I-00146 Roma, Italy.*

[2]*Reactivation Research Group, Department of Geological Sciences, Durham University, Durham DH1 3LE, UK*

Abstract: Intraplate strike-slip deformation belts are typically steeply-dipping structures that develop in both oceanic and continental lithosphere where they form some of the largest and most spectacular discontinuities found on Earth. In both modern and ancient continental settings, intraplate strike slip deformation belts are of major importance in accommodating horizontal displacements where they additionally form very persistent zones of weakness that substantially influence the rheological behaviour of the lithosphere over very long time periods (up to 1 Ga or more). These deformation zones provide a fundamental geometric, kinematic and dynamic link between the more rigid plate-dominated tectonics of the oceans and the non-rigid, complex behaviour of the continents. During convergence, they help to transfer major displacements deep into the plate interiors. During divergence, they act as transfer zones that segment rifts, passive continental margins and, ultimately, oceanic spreading ridges. Such belts are also of great economic importance, controlling the location of many destructive earthquakes, offshore and onshore hydrocarbon deposits and metalliferous ore deposits. In the oceans, intraplate strike-slip movements are relatively minor along transform-related fracture zones, but there are an increasing number of documented examples that may reflect spatial and temporal variations in spreading rate along individual active ridge segments.

Strike-slip deformation belts are regions in which tectonic displacements occur predominantly parallel to the strike of the zone (for a general review, see Woodcock & Schubert 1994). The recognition of strike-slip-dominated plate boundaries or *transform faults* (Wilson 1965), together with their geometric linkage to, and kinematic interaction with, constructive and destructive plate margins were central to the emergence of plate tectonics (e.g. McKenzie & Parker 1967; Morgan 1968; Cox 1973). Plate tectonic theory assumes that the lithosphere in the plate interiors is, to a first approximation, rigid and that most deformation related to plate interactions will be concentrated into narrow belts along the plate margins. This model works reasonably well in regions underlain by oceanic lithosphere, with the result that much of the seismicity along transform faults occurs only where they form active plate boundaries. As they pass into the plate interior, transform faults become relatively tectonically quiescent features known as oceanic fracture zones which form some of the largest topographic structures on the Earths' surface (e.g. White & Williams 1986 and references therein). Many of these fracture zones segment or even bound major sedimentary basins where they impinge upon the continental margin. Examples include the southern Atlantic margins (Francheteau & Le Pichon 1972) and the western margin of Australia (Song *et al.* 2001).

In continental regions, intraplate structures and their relationship to plate tectonics are complicated by the non-rigid behaviour of substantial regions of continental lithosphere (e.g. Molnar 1988). This behaviour is well-illustrated by the broad, diffuse zones of seismicity observed in many continental regions – notably Central Asia from Tibet northwards – extending deep into the plate interiors. Geologically, these continental deformation zones may comprise interlinked systems of fault- and shear zone-bounded blocks and flakes that partition strains and other geological processes into complex regions of displacement, internal distortion and rotation on various scales (e.g. Dewey *et al.* 1986; Foster & Gleadow 1992; Park & Jaroszewski 1994; Tommasi *et al.* 1995; Butler *et al.* 1997; Salvini *et al.* 1997; Marshak *et al.* 2000). It may be useful to view such regions of non-rigid behaviour as broad, diffuse plate boundaries (e.g. see Gordon 1998 and references therein), but in the present paper we shall refer to all regions located away from the major plate boundaries as 'intraplate'.

The non-rigid behaviour of continental lithosphere probably arises from the presence of a weak

From: STORTI, F., HOLDSWORTH, R. E. & SALVINI, F. (eds) *Intraplate Strike-Slip Deformation Belts.* Geological Society, London, Special Publications, **210**, 1–14, 0305-8719/03/$15

quartzofeldspathic crustal layer and from pre-existing mechanical anisotropies. Such anisotropies, primarily old faults and shear zones, may undergo reactivation in preference to the formation of new tectonic structures during regional deformation episodes (e.g. Thatcher 1995; Holdsworth *et al.* 1997, 2001b). The buoyancy of continental crust means that it and its underlying lithospheric mantle are not normally subducted. As a result, zones of pre-existing weakness are effectively 'locked-in' to the continents and can potentially be reactivated many times during successive phases of continental deformation and accretion. This long-lived architecture of inheritance is not generally found in oceanic lithosphere (e.g. Sutton & Watson 1986). Crustal-scale reactivated fault systems in the upper crust broaden with depth into ductile shear zones of regional extent, in which substantial volumes of lower crustal and upper mantle rocks experience episodes of reworking (Holdsworth *et al.* 2001a).

In intraplate regions of the continents, deep-seated faults or shear zones often manifest themselves at the surface by the development of linear zones of geological, geophysical or topographic features known as lineaments (e.g. Sutton & Watson 1986 and references therein). Significantly, a majority of these lineaments seem to coincide with large, steeply-inclined strike-slip deformation zones (e.g. O'Driscoll 1986; Daly *et al.* 1989). Seismological and geodetic studies of neotectonic intraplate deformation (e.g. Molnar & Tapponnier 1975) suggest that horizontal movements are predominantly accommodated by strike-slip faulting. Thus intraplate strike-slip deformation belts are particularly important in determining the deformation response of continental lithosphere in plate interiors over long time scales. Interestingly, the longevity of intraplate continental strike-slip deformation zones and their deeply penetrating nature is central to many of the global tectonic hypotheses that have been proposed prior to plate tectonics (e.g. rhegmatic tectonics; Vening Meinesz 1947) or as alternative paradigms. It is probably significant that the most prominent of these – the Soviet endogenous regime model (Beloussov 1978; Pavlenkova 1995) – was developed by scientists based in an intraplate region almost entirely underlain by continental lithosphere.

Long-lived strike-slip deformation zones – or structures that reactivate them – are of considerable economic significance and represent important geological hazards. Although subordinate to the great concentrations of earthquakes around the plate margins, there are many regions of frequent and sometimes highly destructive seismicity focused along strike-slip deformation zones in intraplate regions. Examples include the N and E Anatolian faults in Turkey (e.g. Jackson & McKenzie 1988) and the New Madrid Seismic Zone in the USA (e.g. Johnston & Shedlock 1992; Marshak *et al. this volume*). Intraplate strike-slip faults often have a profound influence on the location, architecture and subsidence history of associated sedimentary basins, many of which are rich in hydrocarbons, forming important tectonic and palaeogeographic boundaries. Good examples are the many long-lived intraplate strike-slip faults that have repeatedly influenced the evolution of the rich hydrocarbon-bearing basins of SE Asia (e.g. Morley 2002 and references therein). Finally, there are numerous intraplate strike-slip deformation belts that have acted as channels for the flow of magma and hydrothermal fluids, leading to the accumulation of economically significant ore deposits (e.g. the Eastern Goldfields Province in the Yilgarn Block, western Australia; Cox 1999).

Size and mechanical significance of intraplate strike-slip deformation belts

In all plate tectonic settings, strike-slip deformation belts at the surface are characterised by steeply-dipping anastomosing arrays of faults, often with many bends and offsets in individual fault strands (Fig. 1a; e.g. Sylvester 1988; Woodcock & Schubert 1994). Regionally significant fault zones are typically a few tens of kilometres wide and several hundred kilometres long. By definition, plate-boundary transform faults cut through the whole lithosphere in all settings. The deep geometry of intraplate faults is less straightforward, particularly as it is difficult to image steep structures at depth. However, our improved understanding of the relationships between fault dimensions and displacement (e.g. Walsh & Watterson 1988, Cowie & Scholz 1992) suggests that sub-vertical strike-slip fault zones with strike-lengths and offsets greater than 300 km and 30 km, respectively, are very likely to be of a size sufficient to cut much, if not all, of the lithosphere (Fig. 1a). A growing body of geological, geochemical and geophysical observations suggest a direct link between the mantle at depth and regional-scale strike-slip faults and shear zones in the crust. In summary, this evidence includes the following (see Vauchez & Tommasi, *this volume*, and references therein):

i) Geological observations in ancient exhumed mid- and lower crustal rocks preserve many examples of vertically cross-cutting strike-slip shear zones, with little evidence of detachment along sub-horizontal surfaces, even in partially molten rocks (e.g. Borborema Province, Brazil: Vauchez *et al.* 1995).
ii) Chemical and stable isotope studies of

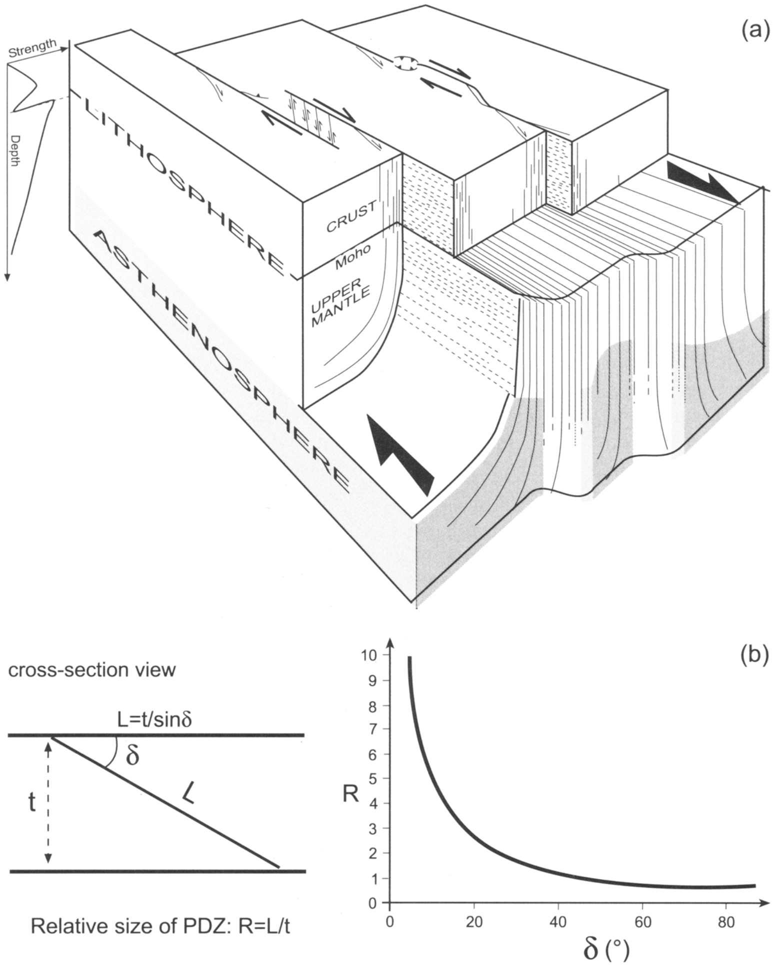

Fig. 1. (**a**) Cartoon showing how major continental intraplate strike-slip deformation belts may ultimately root into the asthenosphere (after Teyssier & Tikoff 1998 and Vauchez *et al.* 1998). Strike slip faults in the upper crust pass down into increasingly broad shear zones (fabric traces shown schematically) in the lower crust and lithospheric mantle. Dashed lines in exposed fault & fabric surface nearer to viewer are transport-parallel lineations. Schematic strength vs depth profile for continental lithosphere shown to the left. (**b**) On the left, sketch cross-sectional view of a load-bearing layer (such as the crust or lithosphere) of thickness t cut by a principal displacement zone (PDZ) of length L dipping at an angle $\delta°$. The relative size of the PDZ is given by L/t. Graph on the right shows how the relative size of a PDZ rapidly decreases as the dip increases.

magmas and hydrothermal fluids channelled along large strike-slip faults and shear zones suggest CO_2-rich, mantle origins (e.g. Madagascar: Pili *et al.* 1997).

iii) Many continental-scale shear zones and reactivated faults overlie large-scale anisotropies such as low-velocity zones in the upper mantle (e.g. Houseman & Molnar 2001) and/or positive gravity anomalies associated with localised uplift of the crust-mantle boundary (e.g. Pili *et al.* 1997).

iv) Shear wave splitting measurements (reviewed in Silver 1996) collected along several intraplate strike-slip faults and shear zones suggest that fabrics related to these structures are developed at all levels in the lithosphere, including the upper mantle (e.g. Fig. 1; Tommasi *et al.* 1996 and references therein). These results are broadly supported by magnetotelluric and electrical anisotropy measurements of deep mantle fabrics (e.g. Pous *et al.* 1995; Senechal *et al.* 1996).

v) Deep seismic reflection profiling studies have imaged numerous examples of regional-scale faults that cut the entire crust and that they penetrate deep into the mantle (e.g. Great Glen Fault: McGeary 1989).

The large-scale mechanical behaviour of the lithosphere has been investigated widely through the application of experimentally derived strength vs depth profiles (e.g. Goetze & Evans 1979; Brace & Kohlstedt 1980; Kirby 1983). Typically, these assume a simple horizontally layered lithosphere, with each layer having a uniform composition, a limited number of competing deformation mechanisms (usually brittle failure and dislocation creep) and specified environmental (P, T, strain rate etc) and lithological (composition, grain size, crustal thickness) conditions. These diagrams are gross simplifications of the likely rheological behaviour (see for example Paterson 1978; Schmid & Handy 1991) but as first order approximations they provide useful information concerning the vertical distribution of strength in the lithosphere. It is generally agreed that the mechanical properties of the strongest layer(s) will determine the overall behaviour of the lithosphere (e.g. England 1983). In most continental settings, the extrapolations of experimental data suggest that the main load-bearing region in the lithosphere should lie in the upper mantle, with a secondary strong region in the mid-crust (Fig. 1a; Molnar 1992 and references therein). This view has been questioned recently by Maggi *et al.* (2000a, b) who argue that the distribution of earthquake focal depths suggests that the main load-bearing region lies in the crust and that the aseismic upper mantle is weak. This conclusion is based on the premise that the presence of seismicity is indicative of strength, but it remains a distinct possibility that the upper mantle may be both aseismic and strong, even over long time scales.

Mechanically weak tectonic discontinuities will be most significant when they cut through the load-bearing regions of the lithosphere and, from the foregoing discussion, it is clear that this is particularly likely for steeply-inclined to sub-vertical, regional-scale strike-slip faults and shear zones (e.g. Fig 1a). From a simple geometric viewpoint, the steeper a fault, the smaller it has to be (in terms of length, area or displacement) in order to cut through a horizontal load-bearing layer of thickness t (Fig. 1b). This may be one reason why strike-slip faults and shear zones are particularly prone to reactivation in continental settings.

Field studies of regional-scale reactivated faults suggest that profound weakening can occur following textural and retrograde metamorphic modification of fault rocks under mid-crustal and upper mantle conditions (e.g. Vissers *et al.* 1995; Stewart *et al.* 2000; Imber *et al.* 2001; Holdsworth *et al.* 2001b). These processes seem to be particularly effective in regions where a large influx of H_2O- or CO_2-rich fluid has occurred during shearing. Sub-vertical strike-slip belts will focus the weakening effects of fault-related processes particularly strongly as all fault strands are vertically aligned. Thus, their persistence over long time scales and apparent importance in intraplate deformation regimes is perhaps not surprising.

Origins of intraplate strike-slip deformation belts

The generation of intraplate strike-slip belts is particularly favoured when one or more of the following occurs: (i) collision of irregularly shaped continental margins and indentors, a process that often leads to lateral escape (e.g. India-Eurasia collision; Tapponnier *et al.* 1986; Arabia – Eurasia collision generating the Anatolian fault block; McKenzie 1972; Dewey 1977); (ii) deformation of lithosphere in which marked lateral variations in rheological strength occur due to rift-related changes in crustal thickness or geothermal gradient (e.g. Borborema Province, Brazil; Tommasi & Vauchez 1997; Vauchez *et al.* 1998); (iii) convergence continues after initial continental collision (e.g. India-Eurasia collision; Molnar & Tapponnier 1975); (iv) relative motions among adjacent plates are governed by different Eulerian poles (e.g. Australia-East Antarctica-New Zealand; Stock & Molnar 1982); (v) differential rotations occur within a major plate (e.g. the Cenozoic motion between East and West Antarctica; Cande *et al.* 2000); (vi) kinematic strain partitioning of a regional intraplate transpressional or transtensional deformation (e.g. The Main Recent Fault, NW Iran; Talebian & Jackson 2002; see also Jackson 1992).

In continental regions, many intraplate strike-slip deformation belts are reactivated structures that formed initially at continental plate boundaries as transform faults, or, as in the case of trench-linked and indent-linked strike-slip faults, due to the operation of plate-boundary processes (e.g. Woodcock 1986). Others have initiated as major dip-slip discontinuities such as oceanic sutures, thrusts or rift-bounding faults. Reactivated oceanic transforms are restricted to ophiolites in ancient settings and seem to be relatively uncommon. In all other cases, the discontinuities have become intraplate features following continental collision and may have undergone steepening into a sub-vertical attitude that is particularly favourable to reactivation. Some intraplate strike-slip faults may form as new structures if no favourably oriented zones of pre-existing weakness are present.

Once present in the continental lithosphere, strike-slip deformation zones clearly influence the segmentation of rifts and the resulting location of salient-re-entrant features in passive continental margins during break-up (e.g. Daly *et al.* 1989). As first recognised by Wilson (1965), the resulting irregularities in the continental margin significantly determine the location and development of transform faults in the evolving spreading ridge and their associated intraplate fracture zones. Significantly, many regions of enhanced seismicity occur along pre-existing strike-slip deformation belts adjacent to and continuous with the terminations of transform-related fracture zones in passive continental margins (e.g. Sykes 1978). These observations suggest a direct link between intraplate faulting in continental and oceanic lithosphere and illustrate that the structural inheritance locked-up in the continents ultimately plays an important role in controlling the geometric and kinematic evolution of oceanic plates.

Termination zones

Two main classes of intraplate strike-slip deformation zones are recognised based on the nature of their terminations (Fig. 2). *Transfer* intraplate strike-slip faults occur when displacement is accommodated at a plate boundary, either by the extrusion of a single, rigid block (rigid escape), or by extruding a number of linked blocks with a rotational component (rotational escape). *Confined* intraplate strike-slip faulting occurs when the displacement decreases and is fully accommodated by strain within the plate interior. Deformation patterns at these fault terminations fall into four end-member types: *extensional, contractional, strike-slip* or *rotational* (Fig. 2). In some cases, more than

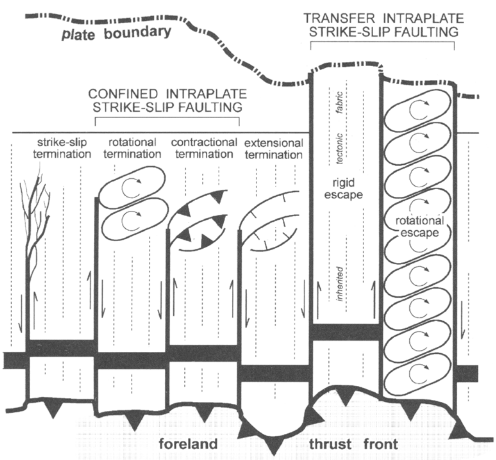

Fig. 2. Highly conceptual sketch showing the two main classes of intraplate strike-slip deformation belts and their different modes of termination.

one type may occur associated with individual terminations (see below). These second-order accommodation structures form at an angle to the master strike-slip fault. The dominant type(s) formed will probably depend on the interaction of the strain fields related to fault motion and shape (local bends, offsets, tips) with the mechanical properties of the adjacent host rocks, particularly the orientation of pre-existing anisotropies in the basement (e.g. see Sylvester 1988; Woodcock & Schubert 1994 and references therein). In case of block rotation, the angle between the block-boundary faults and the master strike-slip belt changes markedly through time (Scotti *et al.* 1991).

Once formed, secondary accommodation structures provide weaknesses into which part of the residual strike-slip displacement can be transferred from the master strike-slip belt (e.g. Storti *et al.* 2001). Repeated propagation of the master strike-slip fault system into the plate interior can produce a sequence of accommodation structures becoming younger towards the fault tip (as seen in small-scale faults: Willemse & Pollard 1998). Thus, complex and superimposed structures can develop in the termination region of intraplate strike-slip deformation belts.

Examples of termination structures in the tip region of intraplate strike-slip deformation belts include the strike-slip faults in the northern Aegean Sea, which end in a region of normal faulting in central Greece (Taymaz *et al.* 1991), and the Priestley Fault, a Cenozoic intraplate right-lateral fault system in north Victoria Land, Antarctica which terminates on its southern side into a series of extensional and transtensional faults including the Terror Rift (Fig. 3, Salvini *et al.* 1997, 1998; Storti *et al.* 2001). The northern side of the fault termination is characterised by ESE-WNW striking strike-slip and transpressional splay faults illustrating that both contractional and extensional structures can form on opposite sides of a single termination. The geometric arrangement of these termination structures is a major clue to the sense of fault movement (e.g. see Fig. 3 inset). As yet there are no palaeomagnetic data to constrain the amount of block rotation about vertical axes that may have occurred, but there is no independent geological evidence to suggest that this is significant (Storti *et al.* 2001).

A rotational and contractional structural architecture is developed at the termination of the right-lateral San Gregorio-Sur-San Simeon-Hosgri fault system, in Southern California (Sorlien *et al.* 1999). The southern Hosgri Fault comprises two main strands bounding compressional folds that lie at high angle to the faults (Fig. 4). Right-lateral shear across the southern Hosgri Fault is absorbed mainly by clockwise vertical-axis rotation of the

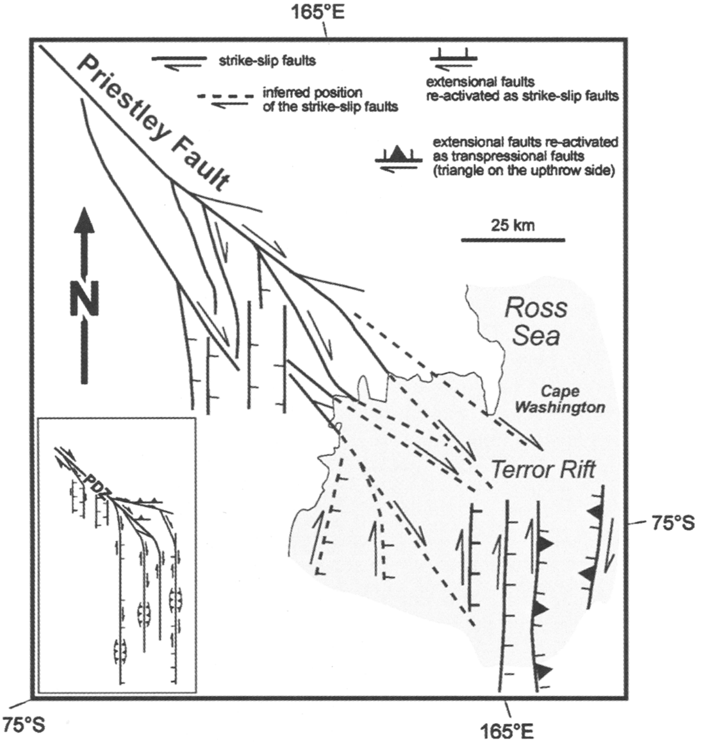

Fig. 3. Structural architecture at the termination of the Priestley Fault, north Victoria Land, Antarctica (see Fig. 9 for location). The inset shows how extensional, contractional, and strike-slip deformation accommodates the residual horizontal displacements at the tip of the fault system (after Storti *et al.* 2001).

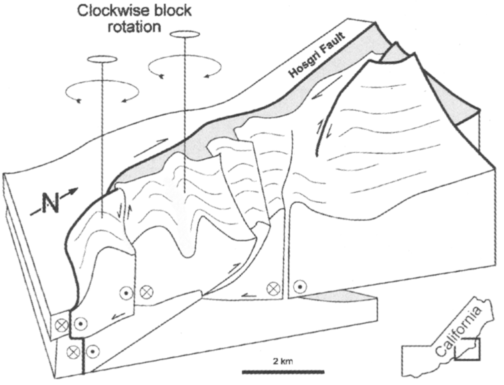

Fig. 4. Cartoon showing the tip of the Hosgri Fault, California, where contractional and strike-slip deformation, together with block rotation about vertical axes accommodate horizontal displacements (after Sorlien *et al.* 1999).

elongated blocks between the fault strands, as well as by folding and thrusting (Sorlien *et al.* 1999). Other examples of rotational terminations of intraplate strike-slip belts include the Marlborough fault system of New Zealand (Little & Roberts 1997), the San Jacinto fault system of southeastern California (Armbruster *et al.* 1998) and the Whittier fault system in the Los Angeles Basin (Wright 1991).

Many intraplate strike-slip belts end in areas of distributed thrusting, like those in the eastern and northern Tibet (Molnar & Lyon-Caen 1989; Meyer *et al.* 1998). Bayasgalan *et al.* (1999) described field examples of contractional termination of intraplate strike-slip belts in Mongolia. At both the eastern end of the Artz Bogd fault system (Fig. 5a) and of faults in the Toromhon region (Fig. 5b), thrust faults developed at high angles to the strike-slip fault systems. Thrust displacement decreases progressively away from the strike-slip faults, suggesting the relative rotation of the thrust footwall and hangingwall blocks about vertical axes (Fig. 5c; Bayasgalan *et al.* 1999).

Bends and stepovers

The fault systems associated with strike-slip deformation zones are rarely perfectly straight as the host rocks are invariably mechanically anisotropic

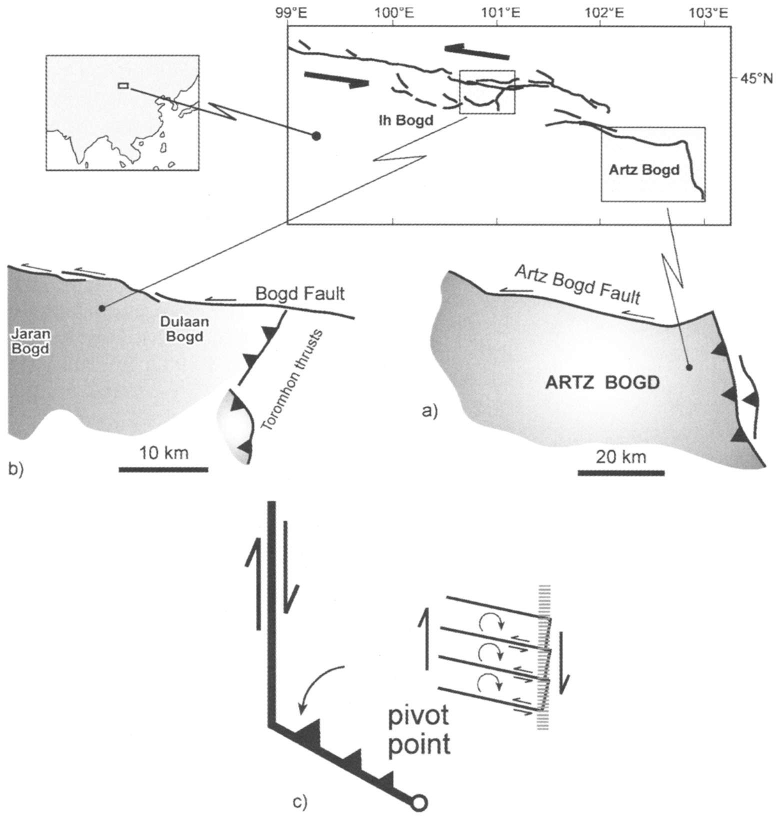

Fig. 5. Simplified sketch map of the structural architecture at the termination of the Artz Bogd Fault (**a**) and Bogd Fault (**b**), Mongolia, showing the dominant role of contractional deformations that accommodate residual horizontal displacement at fault tips. The progressive decrease of thrust displacement away from the master strike-slip fault suggests the occurrence of block rotation about vertical axes (**c**) (after Bayasgalan *et al.* 1999).

and the faults typically grow by the linkage of second-order non-parallel fault segments (e.g. Wilcox *et al.* 1973). Irregularities can be subdivided into two end-member types: *bends* where the fault trace is continuous and *stepovers* or *jogs* where a discontinuity occurs in the fault trace (Fig 6a; Sylvester 1988; Woodcock & Schubert 1994 and references therein). In many cases, stepover zones developed in sedimentary cover sequences near to the surface may pass downwards with depth into bends in the fault where it cuts the basement. Bends and stepovers form local zones of transpressional (restraining) or transtensional (releasing) deformation depending on the sense of overstep or bending relative to the overall sense of movement along the principal displacement zone (PDZ). At the surface, restraining bends or offsets produce localised regions of uplift referred to as push-ups whilst releasing bends or offsets are associated with the development of pull-apart basins. In cross-sections derived from seismic reflection data across many strike-slip deformation belts, upward diverging fault patterns are commonly imaged originating from a single sub-vertical discontinuity at depth (e.g. Harding 1985). These are known as flower structures and they are particularly common in the region of fault bends and stopovers.

A good example of the effects of fault bends and offsets – and what happens when the sense of shear is reversed during successive reactivation episodes – is provided by the Late Archaean to the Late Proterozoic Carajás-Cinzento strike-slip fault systems in the Amazonian Craton of Brazil (Fig. 6b; Pinheiro & Holdsworth 1997a & b; Holdsworth & Pinheiro 2000). Late Archaean brittle dextral movements along E-W trending, sub-vertical fault zones reactivated pre-existing basement fabrics in the underlying Itacaiúnas shear zone, down-faulting cover sequences of low grade and unmetamorphosed rocks into a series of releasing bends and offsets (Fig. 6b). Later fault reactivation and partial inversion of the cover sequences

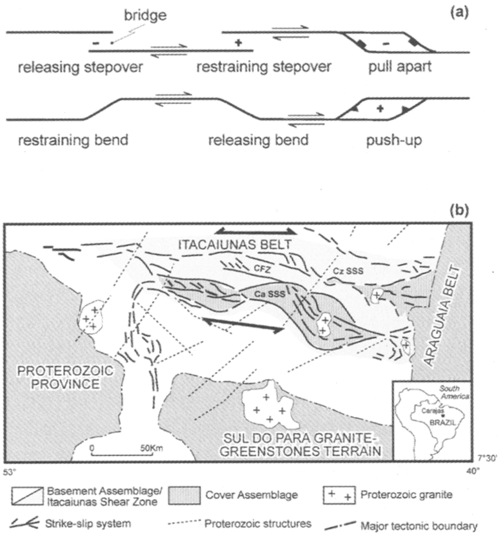

Fig. 6. (**a**) Map views of stepover and bend geometries found along strike-slip faults (after Woodcock & Schubert 1994). (**b**) Simplified map of the structural architecture along the Carajás-Cinzento strike-slip fault systems, in the Amazon Craton of Brazil (after Holdsworth & Pinheiro 2000). Shading highlights the location of the Archean cover rocks and the Itacaiúnas shear zone in the older, underlying basement rocks. Late Archaean faults include the Carajás Fault Zone (CFZ), the Carajás strike-slip system (CaSSS) and the Cinzento strike-slip system (CzSSS).

occurred during a subsequent sinistral shearing episode. This led to the formation of complex assemblages of folds, thrusts, oblique slip and strike-slip faults which were preferentially developed in the cover rocks close to the pre-existing fault traces in bends and offsets that had become restraining features due to the reversal in the sense of shear (Pinheiro & Holdsworth 1997a; Holdsworth & Pinheiro 2000). The adjacent basement gneisses remained comparatively undeformed during these later episodes, undergoing regional uplift and exhumation that stripped away the cover sequences everywhere except where they were initially downfaulted in bends and offsets during dextral movements.

Intraplate strike-slip belts and plate convergence

Intraplate strike-slip belts have been extensively studied in convergent settings (e.g. Vauchez *et al.* 1998). These belts of localised intracontinental deformation are typically several tens of km wide and many hundreds of km long (e.g. Molnar & Tapponnier 1975; Pili *et al.* 1997; Ludman 1998). Contractional-, extensional-, and strike-slip-related structures characteristically alternate along these impressively long deformation belts, in which the internal architecture is thought to be influenced by inherited crustal fabrics (e.g. Vauchez *et al.* 1995; Rossetti *et al.* 2002).

There is an ongoing debate concerning the relative importance and roles of crustal thickening, extensional collapse and strike-slip faulting in bringing about lateral extrusion in the Tibetan Plateau and Asian regions to the N and the relationship between these processes and the collision and indentation of India (e.g. Tapponnier *et al.* 1982; Davy & Cobbold 1988; England & Molnar 1990; Shen *et al.* 2001). Irrespective of the relative merits of the various competing models, there is clear evidence in much of Asia that significant deformation and displacement have occurred along a series of very large intraplate strike-slip deformation belts during the Cenozoic.

The simplified tectonic sketch map of Asia published by Jolivet *et al.* (1999) (Fig. 7) illustrates the tectonic architecture of major intraplate deformation belts, their impressive length and complexity. In this interpretation, the Pamir-Baikal-Okhotsk shear zone comprises interlinked extensional rifts, such as the Baikal basin, and strike-slip fault segments. The deformation belt appears to extend from the collision zone to the Bering Strait, separating the stable Eurasian block from the

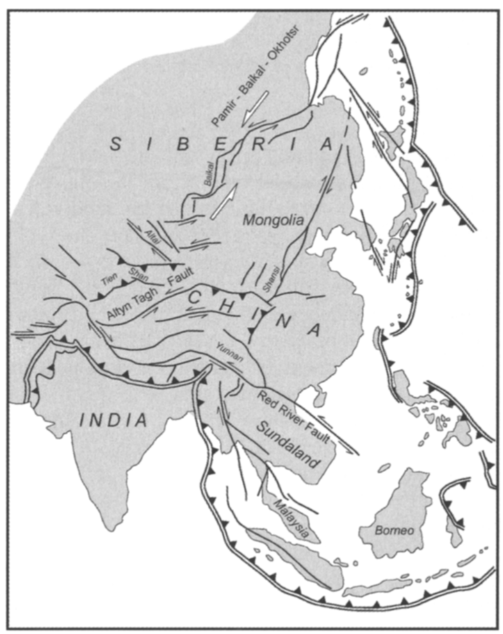

Fig. 7. Highly simplified tectonic sketch map of Asia based on the interpretation of Jolivet *et al.* (1999).

deformed part of the Asian plate (Davy & Cobbold 1988). Thus, the Pamir-Baikal-Okhotsk shear zone represents a possible example of a transfer intraplate strike-slip deformation belt, since it connects the northwest corner of the Indian indenter, the western Himalayan syntaxis, to the boundary region of the Pacific Plate. Another example of a transfer intraplate strike-slip deformation belt may be provided by the roughly N-S envelope of right-lateral strike-slip fault systems and extensional basins (both pull-apart and back-arc) that developed along the eastern border of Asia (Fig. 7). This right-lateral intraplate deformation belt connects the northeast corner of the Tibetan Plateau with the Pacific plate boundary region where it abuts a complex array of left-lateral strike-slip fault systems (e.g. Jolivet *et al.* 1999).

Examples of confined intraplate strike-slip faults include the Red River Fault and the Altyn Tagh Fault (Molnar & Tapponnier 1975; Leloup *et al.* 2001) (Fig. 7). The Red River Fault is a left-lateral intraplate strike-slip deformation belt which bounds the Indonesian block to the north and terminates in the extensional domain of the South China Sea (e.g. Morley 2002). Despite its internal complexity, the Red River Fault can be broadly described as having an extensional termination. The Altyn Tagh Fault is a ENE to E-W striking left-lateral strike-slip deformation belt bounding the Tibetan Plateau to the north. At the point where the fault trajectory starts bending clockwise, it shows a compressional component (Fig. 7). The Altyn Tagh Fault terminates against the NNE-SSW thrust system that bounds the Tibetan Plateau to the E and can thus be described as having a contractional termination.

Intraplate strike-slip belts and plate divergence

The occurrence of strike-slip belts that are significantly active in intraplate regions past or present is uncommon in divergent plate boundaries that are more generally dominated by sea floor spreading and passive margin development. Substantial strike-slip movements do not occur along oceanic fracture zones once they pass outboard of their associated ridge segments and away from the plate boundary (Fig. 8a).

A good example of intraplate strike-slip faulting in a divergent setting comes from the Cenozoic tectonic evolution at the eastern edge of the Antarctic Plate, which includes the Southern Ocean east of 139°E, north Victoria Land, and the Ross Sea (Fig. 9). Interpretation of seismic reflection profiles in the Ross Sea and correlation of the offshore tectonic fabric with the onshore major structural lineaments allows reconstruction of a tectonic architecture dominated by NW-SE-striking right-lateral strike-slip fault systems in north Victoria Land, which to transfer their horizontal displacement in to the N-S trending basins of the Ross Sea (Salvini *et al.* 1997). The continuity of the NW-SE striking right-lateral strike-slip deformation belts connecting the Ross Sea into the impressive, co-linear fracture zones of the Southern Ocean is demonstrated by the development of prominent recent positive flower structures in reflection seismic profiles recorded across the seismically active Balleny Fracture Zone adjacent to the continental shelf (Spezie *et al.* 1993). This evidence suggests that major fracture zones in the Southern Ocean, east of 139°E, are tectonically active and that right-lateral partitioned transtension in the western Ross Sea (Wilson 1995; Rossetti *et al.* 2000) accommodates transform shear in the Southern Ocean (Salvini *et al.* 1997). Such an excess shear appear to be transmitted from the oceanic ridges to the Ross Sea through a network of long, intraplate strike-slip deformation belts cutting across both oceanic and continental lithosphere (Fig. 9). Similar processes might also explain why some of the largest intra-

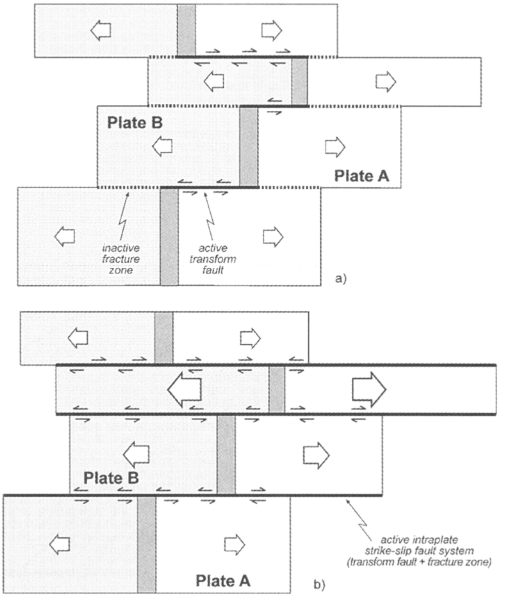

Fig. 8. Conceptual cartoon showing the possible relationships between transform faulting and spreading rates at mid oceanic ridges. (**a**) "conventional" geodynamic framework with constant spreading rate and transform faulting confined between ridge segments. Out-of-ridge transform segments are inactive (fracture zones). (**b**) Differential spreading rates at the plate boundary cause plate segmentation by active intraplate strike-slip fault systems that include both "classical" transform faults and their associated fracture zones along strike.

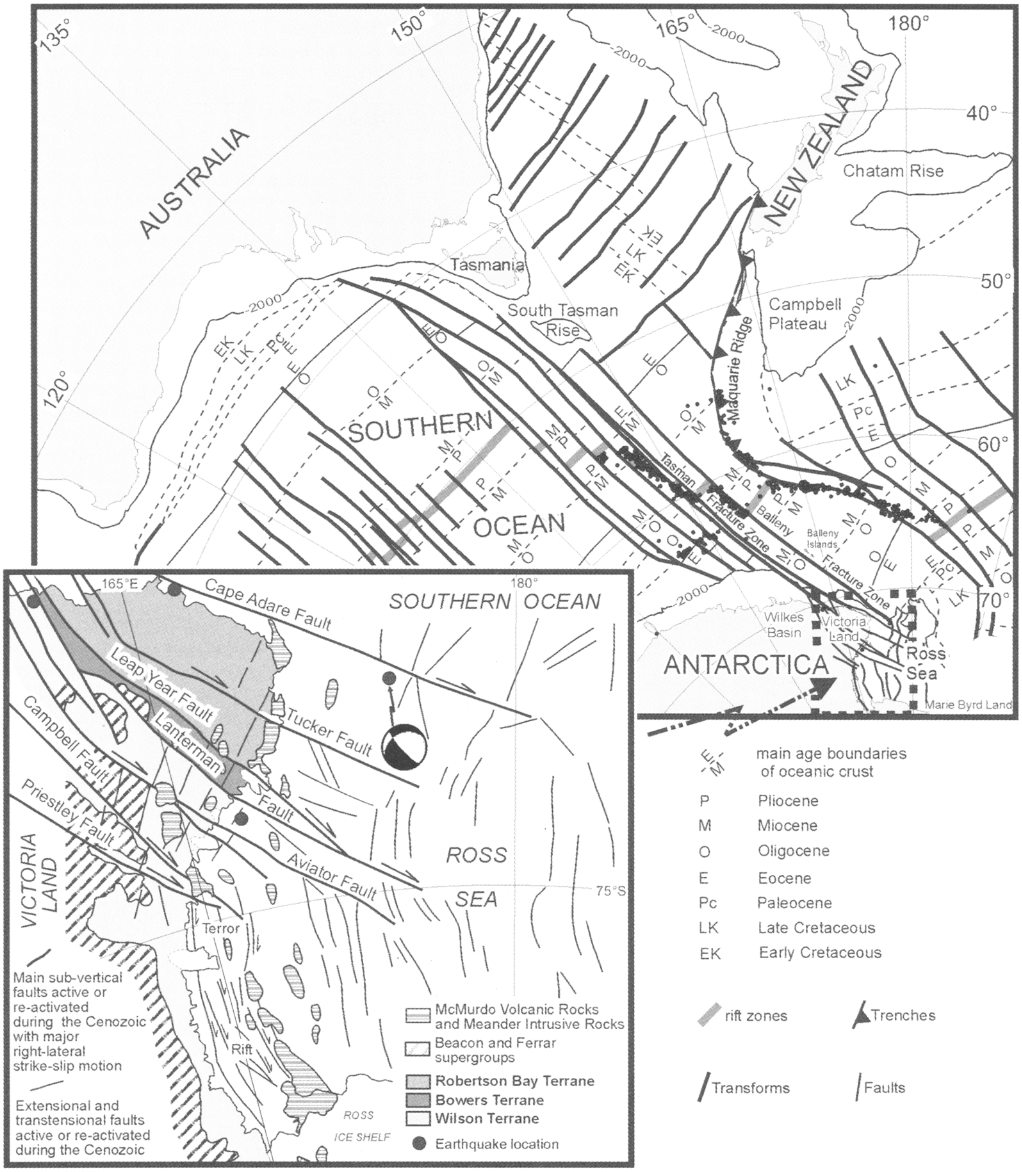

Fig. 9. Cenozoic geodynamic framework at the northeastern edge of the Antarctic Plate showing the intraplate termination of transform shear by transtensional faulting the western Ross Sea. (after Salvini *et al.* 1997). Earthquake location is from the Harvard CMT catalog.

plate shocks in the continents are located along pre-existing faults located inland from the end of oceanic transform fault fracture zones (e.g. see Sykes 1978).

One possible explanation for the occurrence of intraplate strike-slip deformation belts in passive margin settings may relate to changes in the spreading rate along mid-oceanic ridges. Plate tectonic theory generally assumes rigidity so that the rate of spreading is constant and is proportional to the distance from the Eulerian pole. If the rigidity constraint is relaxed (Gordon 1998), however, intraplate strike-slip movements along transform fracture zones and at their terminations are possible. In particular, differences in the spreading rate at the mid-oceanic ridge in adjacent transform fault-bounded compartments could lead to strike-slip shear along the intraplate fracture zones (Fig. 8b). The sense of shear in the intraplate segments is towards the ridge in the low-spreading plate sectors and away from the ridge in the fast spreading sectors. The excess shear along the intraplate strike-slip belts can terminate in the oceanic plate interior or in the continental passive margin following one or more of the termination mechanisms described earlier.

Conclusions

Intraplate strike-slip deformation belts form some of the most prominent tectonic and topographic features on both the Earth and, possibly, other planets (e.g. Crumpler *et al.* 1986). A majority of these structures appear to originate in plate boundary deformation zones and in the continents where the lithosphere is not subducted, they become incorporated into the plate interior by the processes of collision and accretion. Once established they actively transfer displacements from plate margins into the interior regions, fundamentally influencing the location and evolution of a broad range of geological features, including sedimentary basins, orogenic belts, active sesimicity, hydrothermal activity and magmatism. In the continents especially, they form major persistent zones of apparent weakness whose influence may be felt over many hundreds or even thousands of million years. It therefore seems likely that intraplate strike-slip deformation belts form one of the most significant sources of long-term mechanical anisotropy in the lithosphere.

Financial support for this work was provided by the Italian Programma Nazionale di Ricerche in Antartide (PNRA; grants to F. Salvini). Nigel Woodcock, Mark Allen and Jonathan Turner are thanked for detailed and thoughtful reviews.

References

ARMBRUSTER, J. G., SEEBER, L., SORLIEN, C. C. & STECKLER, M. S. 1998. Rotation vs. Rifting in an extensional jog: Salton trough, California. *Eos* (Transactions American Geophysical Union), **79**, F565.

BAYASGALAN, A., JACKSON, J., RITZ, J. -F. & CARRETIER, S. 1999. Field examples of strike-slip fault terminations in Mongolia and their tectonic significance. *Tectonics,* **18**, 394–411.

BELOUSSOV, V. V. 1978. *Endogenic Regimes of Continents.* Nedra, Moscov, pp. 232 (in Russian).

BRACE, F. W. & KOHLSTEDT, D. L. 1980. Limits on lithospheric stress imposed by laboratory experiments. *Journal of Geophysical Research,* **50**, 6248–6252.

BUTLER, R. W. H., HOLDSWORTH, R. E. & LLOYD, G. E. 1997. The role of basement reactivation in continental deformation. *Journal of the Geological Society, London,* **154**, 69–71.

CANDE, S. C., STOCK, J. M., MÜLLER, R. D. & ISHIHARA, T. 2000. Cenozoic motion between East and West Antarctica. *Nature,* **404**, 145–150.

COWIE, P. & SCHOLZ, C. H. 1992. Growth of faults by accumulation of seismic slip. *Journal of Geophysical Research,* **97**, 11,085–11,095.

COX, A. (ed), 1973. *Plate Tectonics and Geomagnetic Reversals.* Freeman, San Francisco, pp. 702.

COX, S. F. 1999. Deformational controls on the dynamics of fluid flow in mesothermal gold system. *In*: MCCAFFREY, K. J. W., LONERGAN, L. & WILKINSON, J. J. (eds) *Fractures, fluid flow and mineralization.* Geological Society, London, Special Publications **155**, 123–140.

CRUMPLER, L. S., HEAD, J. W. & CAMPBELL, D. B. 1986. Orogenic belts on Venus. *Geology,* **14,** 1031–1034

DALY, M. C., CHOROWICZ, J. & FAIRHEAD, J. D. 1989. Rift basin evolution in Africa: the influence of reactivated steep basement shear zones. *In*: COOPER, M. A. & WILLIAMS, G. D. (eds) *Inversion Tectonics.* Geological Society, London, Special Publications **44**, 309–344.

DAVY, P. & COBBOLD, P. R. 1988. Indentation tectonics in nature and experiments. Experiments scaled for gravity. *Bulletin of the Geological Institutions of Uppsala,* **14**, 129–141.

DEWEY, J. F. 1977. Suture zone complexities: a review. *Tectonophysics,* **40**, 53–67.

DEWEY, J. F., HEMPTON, M. R., KIDD, W. S. F., SAROGLU, F. & SENGÖR, A. M. C. 1986. Shortening of continental lithosphere in neotectonics of Eastern Anatolia – a young collision zone. *In*: COWARD, M. P. & RIES, A. C. (eds) *Collision tectonics.* Geological Society, London, Special Publications, **19**, 3–36.

ENGLAND, P. 1983. Constraints on extension of continental lithosphere. *Journal of Geophysical Research,* **88**, 1145–1152.

ENGLAND, P. & MOLNAR, P. 1990. Right-lateral shear and rotation as the explanation for strike-slip faulting in eastern Tibet. *Nature,* **344**, 140–142.

FOSTER, D. A. & GLEADOW, J. W. 1992. Reactivated tectonic boundaries and implications for reconstruction of southeastern Australia and northern Victoria Land, Antarctica. *Geology,* **20**, 267–270.

FRANCHETEAU, J. & LE PICHON, X. 1972. Marginal fracture zones as structural framework of continental margins in South Atlantic Ocean. *American Association of Petroleum Geologists Bulletin,* **56**, 991–1007.

GOETZE, C. & EVANS, B. 1979. Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophysical Journal of the Royal Astronomical Society,* **59**, 463–478.

GORDON, R. G. 1998. The plate tectonic approximation: plate nonrigidity, diffuse plate boundaries and global plate reconstructions. *Annual Review of Earth and Planetary Science,* **26**, 615–642.

HARDING, T. P. 1985. Seismic characteristics and identification of negative flower structures, positive flower structures and positive structural inversion. *American Association of Petroleum Geologists Bulletin,* **69**, 582–600.

HOLDSWORTH R. E. & PINHEIRO, R. V. L. 2000. The anatomy of shallow-crustal transpressional structures: insights from the Archaean Carajas Fault Zone, Amazon, Brazil. *Journal of Structural Geology*, **22**, 1105–1123.

HOLDSWORTH, R. E., BUTLER, C. A. & ROBERTS, A. M. 1997. The recognition of reactivation during continental deformation. *Journal of the Geological Society*, **154**, 73–78.

HOLDSWORTH, R. E., HAND, M., MILLER, J. A. & BUICK, I. S. 2001a. Continental reactivation and reworking: an introduction. *In*: MILLER, J. A., HOLDSWORTH, R. E., BUICK, I. S. & HAND, M. (eds) *Continental reactivation and reworking*. Geological Society, London, Special Publications, **184**, 1–12.

HOLDSWORTH, R. E., STEWART, M., IMBER, J. & STRACHAN, R. A. 2001b. The structure and rheological evolution of reactivated continental fault zones. A review and case study. *In*: MILLER, J. A., HOLDSWORTH, R. E., BUICK, I. S. & HAND, M. (eds) *Continental reactivation and reworking*. Geological Society, London, Special Publications, **184**, 115–137.

HOUSEMAN, G. A. & MOLNAR, P. 2001. Mechanisms of lithosphere renewal associated with continental orogeny. *In*: MILLER, J. A., HOLDSWORTH, R. E., BUICK, I. S. & HAND, M. (eds) *Continental reactivation and reworking*. Geological Society, London, Special Publications, **184**, 13–37.

IMBER, J., HOLDSWORTH, R. E., BUTLER, C. A. & STRACHAN, R. A. 2001. A reappraisal of the Sibson-Scholz fault zone model: the nature of the frictional to viscous ("brittle-ductile") transition along a long-lived, crustal-scale fault, Outer Hebrides, Scotland. *Tectonics*, **20**, 601–624.

JACKSON, J. 1992. Partitioning of strike-slip and convergent motion between Eurasia and Arabia in eastern Turkey and the Caucasus. *Journal of Geophysical Research*, **97**, 12,471–12,479.

JACKSON, J. & MCKENZIE, D. P. 1988. The relationship between plate motions and seismic tremors, and the rates of active deformation in the Mediterranean and Middle East. *Geophysical Journal of the Royal Astronomical Society*, **93**, 45–73.

JOHNSTON, A. C. & SHEDLOCK, K. M. 1992. Overview of research in the New Madrid seismic zone. *Seismological Research Letters*, **63**, 193–208.

JOLIVET, L., FACCENNA, C., D'AGOSTINO, N., FOURNIER M. & WORRALL, D. 1999. The kinematics of back-arc basins, examples from the Tyrrhenian, Aegean and Japan Seas. *In*: MAC NIOCAILL, C. & RYAN, P. D. (eds) *Continental Tectonics*. Geological Society, London, Special Publication, **164**, 21–53.

KIRBY, S. H. 1983. Rheology of the lithosphere. *Review in Geophysics and Space Physics*, **21**, 1458–1487.

LELOUP, P. H., ARNAUD, N., LACASSIN, R., KIENAST, J. R., HARRISON, T. M., PHAN TRANG, T. T., REPLUMAZ, A. & TAPPONIER, P. 2001. New constraints on the structure, thermochronology and timing of the Ailao Shan-Red River shear zone, SE Asia. *Journal of Geophysical Research*, **106**, 6683–6732.

LITTLE, T. A. & ROBERTS, A. P. 1997. Distribution and mechanism of Neogene to present-day vertical axis rotation, Pacific-Australia plate boundary zone, South Island, New Zealand. *Journal of Geophysical Research*, **102**, 20447–20468.

LUDMAN, A. 1998. Evolution of a transcurrent fault system in shallow crustal metasedimentary rocks: the Norumbega fault zone, eastern Maine. *Journal of Structural Geology*, **20**, 93–107.

MAGGI, A., JACKSON, J. A., MCKENZIE, D. & PRIESTLY, K. 2000a. Earthquake focal depths, effective elastic thickness, and the strength of the continental lithosphere. *Geology*, **28**, 495–498.

MAGGI, A., JACKSON, J. A., PRIESTLY, K. & BAKER, C. 2000b. A re-assessment of focal depth distribution in southern Iran, the Tien Shan and northern India: do earthquakes really occur in the continental mantle? *Geophyscial Journal International*, **143**, 629–661.

MARSHAK, S., KARLSTROM, K. & TIMMONS, J. M. 2000. Inversion of Proterozoic extensional faults: An explanation for the pattern of Laramide and Ancestral Rockies intracratonic deformation, United States. *Geology*, **28**, 735–738.

MARSHAK, S., NELSON, W. J. & MCBRIDE, J. H. Phanerozoic strike-slip faulting in the continental interior (mid-continent cratonic platform) of the United States. *This volume*.

MCGEARY, S. 1989. Reflection seismic evidence of a Moho offset beneath the Walls boundary strike-slip fault. *Journal of the Geological Society, London*, **146**, 261–269.

MCKENZIE, D. P. 1972. Active tectonics of the Mediterranean region. *Geophysical Journal of the Royal Astronomical Society*, **30**, 109–185.

MCKENZIE, D. & PARKER, R. L. 1967. The North Pacific: an example of tectonics on a sphere. *Nature*, **216**, 1276–1280.

MEYER, B., TAPPONNIER, P., BOURJOT, L., METIVIER, F., GAUDEMER, Y., PELTZER, G., SHUNMIN, G. & ZHITAI, C. 1998. Crustal thickening in Gansu-Qinghai, lithospheric mantle subduction, and oblique, strike-slip controlled growth of the Tibetan Plateau. *Geophysical Journal International*, **135**, 1–47.

MOLNAR, P. 1988. Continental tectonics in the aftermath of plate tectonics. *Nature*, **335**, 131–137.

MOLNAR, P. 1992. Brace-Goetze strength profiles, the partitioning of strike-slip and thrust faulting at zones of oblique convergence and the stress-heat flow paradox of the San Andreas Fault. *In*: EVANS, B. & WONG, T.-F. (eds) *Fault mechanics and transport properties in rocks*. Academic Press, London, 359–435.

MOLNAR, P. & LYON-CAEN, H. 1989. Fault plane solutions of earthquakes and active tectonics of the Tibetan Plateau and its margins. *Geophysical Journal International*, **99**, 123–153.

MOLNAR, P. & TAPPONNIER P. 1975. Cenozoic tectonics of Asia: effects of a continental collision. *Science*, **189**, 419–426.

MORGAN, W. J. 1968. Rises, trenches, great faults and crustal blocks. *Journal of Geophysical Research*, **73**, 1959–1982.

MORLEY, C. K. 2002. A tectonic model for the Tertiary evolution of strike-slip faults and rift basins in SE Asia. *Tectonophysics*, **347**, 189–215.

O'DRISCOLL, E. S. T. 1986. Observations of the lineament-ore relation. *Philosophical Transactions of the Royal Society, London*, **A317**, 195–218.

PARK, R. G. & JAROSZEWSKI, W. 1994. Craton Tectonics, Stress and Seismicity *In*: HANCOCK, P. L. (ed) *Continental Tectonics*, Pergamon Press, Oxford, pp. 200–222.

PATERSON, M. S. 1978. *Experimental rock deformation-the brittle field*. Springer, Berlin.

PAVLENKOVA, N. I. 1995. Structural regularities in the lithosphere of continents and plate tectonics. *Tectonophysics,* **243,** 223–239.

PILI, E., RICARD, Y., LARDEAUX, J. -M & SHEPPARD, S. 1997. Lithospheric shear zones and mantle-crust connections. *Tectonophysics,* **280**, 15–29.

PINHEIRO, R. V. L. & HOLDSWORTH, R. E. 1997a. Reactivation of Archaean strike-slip fault systems, Amazon region, Brazil. *Journal of the Geological Society, London,* **154**, 99–104.

PINHEIRO, R. V. L. & HOLDSWORTH, R. E. 1997b. The structure of the Carajas N-4 Ironstone deposit and associated rocks: relationship to Archaean strike-slip tectonics and basement reactivation in the Amazon region, Brazil. *Journal of South American Earth Sciences,* **10,** 305–319

POUS, J., LEDO, J., MARCUELLO, A. & DAIGNIÈRES, M. 1995. Electrical resistivity model of the crust and upper mantle from magnetotelluric survey through the central Pyrenees. *Geophysical Journal International,* **121**, 750–762.

ROSSETTI, F., STORTI, F. & SALVINI, F. 2000. Cenozoic noncoaxial transtension along the western shoulder of the Ross Sea, Antarctica, and the emplacement of McMurdo dyke arrays. *Terra Nova,* **12**, 60–66.

ROSSETTI, F., STORTI, F. & LÄUFER, A. L. 2002. Brittle architecture of the Lanterman Fault and its impact on the final terrane assembly in north Victoria Land, Antarctica. *Journal of the Geological Society, London,* **159**, 159–173.

SALVINI, F., BRANCOLINI, G., BUSETTI, M., STORTI, F., MAZZARINI, F. & COREN, F. 1997 Cenozoic geodynamics of the Ross Sea Region, Antarctica: Crustal extension, intraplate strike-slip faulting and tectonic inheritance. *Journal of Geophysical Research,* **102,** 24669–24696.

SALVINI, F., STORTI, F., BRANCOLINI, G., BUSETTI, M. & DE CILLIA, C. 1998. Strike-slip induced Cenozoic basin inversion in the Ross Sea, Antarctica: Kinematic versus dynamic deformation. Terra Antartica, **5**, 209–215.

SCHMID, S. M. & HANDY, M. R. 1991. Towards a genetic classification of fault rocks. Geological usage and tectonophysical implications. *In*: MÜLLER, D. W., MCKENZIE, J. A. & WEISSERT, H. (eds) *Controversies in modern geology: Evolution of geological theories in sedimentology, earth history and tectonics*. Academic, Sand Diego, California, 339–361.

SCOTTI O., NUR A. & ESTEVEZ, R. 1991. Distributed deformation and block rotation in three dimensions. *Journal of Geophysical Research,* **96**, 12225–12243.

SENECHAL, G., RONDENAY, S., MARESHAL, M., GUILBERT, J. & POUPINET, G. 1996. Seismical and electrical anisotropies in the lithosphere across the Grenville Front, Canada. *Geophysical Research Letters,* **23**, 2255–2258.

SHEN, F., ROYDEN, L. H. & BURCHFIELD, B. C. 2001. Large-scale crustal deformation of the Tibetan Plateau. *Journal of Geophysical Research,* **106**, 6739–6816.

SILVER, P. G. 1996. Seismic anisotropy beneath beneath the continents: probing the depths of geology. *Annual Review of Earth and Planetary Science,* **24**, 385–432.

SONG, T., CAWOOD, P. A. & MIDDLETON, M. 2001. Transfer zones normal and oblique to rift trend: examples from the Perth Basin, Western Australia. *In*: WILSON, R. C. L., WHITMARSH, R. B., TAYLOR, B. & FROITZHEIM, N. (eds) *Non-volcanic rifting of continental margins: a comparison of evidence for land and sea*. Geological Society, London, Special Publications, **187**, 475–488.

SORLIEN, C. C., KAMERLING, M. J. & MAYERSON, D. 1999. Block rotation and termination of the Hosgri strike-slip fault, California, from three-dimensional map restoration. *Geology,* **27**, 1039–1042.

SPEZIE, G., TUCCI, S., BUDILLON, G., CORRADI, N., PICCAZZO, M., FANUCCI, F., FIRPO, M., MIRABILE, L. & FERRARI, M. 1993. Water masses characteristics, sediment dispersal and seismic stratigraphy in the continental margin of the Ross Sea (Antarctica). *Istituto Universitario Navale di Napoli, Annali della Facoltà di Scienze Nautiche*, vol. **LX**.

STEWART, M., HOLDSWORTH, R. E. & STRACHAN, R. A. 2000. Deformation processes and weakening mechanisms within the frictional-viscous transition zone of major crustal-scale faults: Insights from the Great Glen Fault Zone, Scotland. *Journal of Structural Geology,* **22**, 543–560.

STOCK, J. M. & MOLNAR, P. 1982. Uncertainties in the relative positions of the Australia, Antarctica, Lord Howe, and Pacific plates since the Late Cretaceous. *Journal of Geophysical Research,* **87**, 4697–4714.

STORTI, F., Rossetti, F. & SALVINI, F. 2001. Structural architecture and displacement accommodation mechanisms at the termination of the Priestley Fault, northern Victoria Land, Antarctica. *Tectonophysics,* **341**, 141–161.

SUTTON, J. & WATSON, J. V. 1986. Architecture of the continental lithosphere. *Philosophical Transactions of the Royal Society, London, Series A,* **317**, 5–12.

SYKES, L. R. 1978. Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism and other tectonism post-dating continental fragmentation. *Reviews of Geophysics and Space Physics,* **16**, 621–688.

SYLVESTER, A. G. 1988. Strike-slip faults. *Geological Society of America Bulletin,* **100**, 1666–1703.

TALEBIAN, M. & JACKSON, J. 2002. Offset on the Main Recent Fault of NW Iran and implications for the late Cenozoic tectonics of the Arabia-Eurasia collision zone. *Geopysical Journal International,* **150,** 422–439.

TAPPONIER, P., PELTZER, G., LE DAIN, A. Y., ARMIJO, R. & COBBOLD P. 1982. Propagating extrusion tectonics in Asia: new insights from simple experiments with plasticine. *Geology,* **10**, 611–616.

TEYSSIER, C. & TIKOFF, B. 1998. Strike-slip partitioned transpression of the San Andreas fault system: a lithosphere-scale approach. *In*: HOLDSWORTH, R. E., STRACHAN, R. A. & DEWEY, J. F. (eds) *Continental Transpressional and Transtensional Tectonics*, Geological Society, London, Special Publications, **135,** 143–158.

THATCHER, W. 1995. Microplate versus continuum descriptions of active tectonic deformation. *Journal of Geophysical Research,* **100**, 3885–3894.

TAYMAZ, T., JACKSON, J. & MCKENZIE, D. 1991. Active tectonics of the north and central Aegean Sea. *Geophysical Journal International,* **106**, 433–490.

TOMMASI, A., VAUCHEZ, A. & DAUDRÉ, B. 1995. Initiation and propagation of shear zones in a heterogeneous continental lithosphere. *Journal of Geophysical Research,* **100**, 22083–22101.

TOMMASI, A., VAUCHEZ, A. & RUSSO, R. 1996. Seismic anisotropy in ocean basins: resistive drag of the sublithospheric mantle? *Geophysical Research Letters,* **23**, 2991–2994.

VAUCHEZ, A. & TOMMASI, A. Wrench faults down to the asthenosphere :geological and geophysical evidence and thermo-mechanical effects. *This volume.*

VAUCHEZ, A., PACHECO-NEVES, S., CABY, R., CORSINI, M., EGYDIO-SILVA, M., ARTHAUD, M. & AMARO, V. 1995. The Borborema shear zone system. *Journal of South American Earth Sciences,* **8**, 247–266.

VAUCHEZ, A., TOMMASI, A. & BARRUOL, G. 1998. Rheological heterogeneity, mechanical anisotropy and deformation of the continental lithosphere. *Tectonophysics,* **296**, 61–86.

VENING MEINESZ, F. A. 1947. Shear patterns of the Earth's crust. *Transactions of the American Geophysical Union,* **28**, 1–61.

VISSERS, R. L. M., DRURY, M. R., HOOGERDUIJIN STRATING, E. H., SPIERS, C. J. & VAN DER WAL, D. 1995. Mantle shear zones and their effect on lithosphere strength during continental breakup. *Tectonophysics,* **249**, 155–171.

WALSH, J. J. & WATTERSON, J. 1988. Analysis of the relationship between displacements and dimensions of faults. *Journal of Structural Geology,* **10**, 239–247.

WHITE, R. S. & WILLIAMS, C. A. 1986. Introduction: oceanic fracture zones. *Journal of the Geological Society, London,* **143**, 737–742.

WILCOX, R. E., HARDING, T. P. & SEELY, D. R. 1973. Basic wrench tectonics. *American Association of Petroleum Geologists Bulletin,* **57**, 74–96.

WILLEMSE, E. J. M. & POLLARD, D. D. 1998. On the orientation and patterns of wing cracks and solution surfaces at the tips of a sliding flaw or fault. *Journal of Geophysical Research,* **103**, 2427–2438.

WILSON, T. J. 1965. A new class of faults and their bearing on continental drift. *Nature,* **207**, 343–347.

WILSON, T. J. 1995. Cenozoic transtension along the Transantarctic Mountains-West Antarctic Rift boundary, Southern Victoria Land, Antarctica. *Tectonics,* **14**, 531–545.

WOODCOCK, N. H. 1986. The role of strike-slip fault systems at plate boundaries. *Philosophical Transactions of the Royal Society of London,* **A317**, 13–29.

WOODCOCK, N. H. & SCHUBERT, C. 1994. Continental strike-slip tectonics. *In*: HANCOCK, P. L. (ed) *Continental Tectonics*, Pergamon Press, Oxford, pp. 251–263.

WRIGHT, T. L. 1991. Structural geology and tectonic evolution in the Los Angeles Basin, California. *In*: BIDDLE K. T. (ed) *Active margin basins*. American Association of Petroleum Geologists Memoir, **52**, 35–135.