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A TECTONIC MODEL FOR RIDGE–TRANSFORM–RIDGE PLATE BOUNDARIES: IMPLICATIONS FOR THE STRUCTURE OF OCEANIC LITHOSPHERE

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ABSTRACT

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The first-order geologic and morphologic relationships at, along and proximal to ridge–transform–ridge plate boundaries are used to construct an empirical and speculative tectonic model. The distinctive but variable morphotectonic fabric and crustal structure of the transform domain are the product of a tectonic continuum that ranges from very slow rates of strain associated with very thick edges of lithosphere (slowly-slipping ridge–transfrom–ridge plate boundaries) to extremely high rates of strain associated with very thin edges of lithosphere (fast-slipping ridge–transform–ridge plate boundaries). The geometry of a ridge–transform intersection necessitates the juxtaposition of a relatively cold, thick edge of lithosphere against the truncated end of an accreting plate boundary. The cold face of lithosphere cools the adjacent wedge of asthenosphere rising beneath the axis of accretion and restricts the amount of partial melting thus attenuating the volume of basaltic melt segregated from the asthenosphere per unit time. The shallow level manifestation of this cold edge effect is a thinner oceanic crust. At depth, the evolving lithosphere thickens rapidly as the cold edge is approached creating changes in the properties of the young lithosphere and forming a weld of upper mantle material against the cold edge of truncating lithosphere. The ridge–transform weld of upper mantle material creates a shear couple in the lithosphere underlying the intersection resulting in the progressive reorientation of the maximum tensile stress from normal to the ridge axis, at some distance from the ridge–transform intersection, to an oblique angle near the boundary. The brittle carapace of oceanic crust that overlies the mantle weld will deform accordingly with the development of oblique, dip-slip faults. The model predicts that the geologic expressions of this cold boundary effect will become more dramatic with increasing thickness of the truncating edge. At very slow rates of accretion (< 4 cm/yr full rate) large offset transforms (> 100 km) juxtapose thick (30–50 km) edges of lithosphere against the accreting plate boundary all but nullifying the processes that lead to the emplacement of normal oceanic lithosphere. In this environment, a relatively strong mantle weld will form at the ridge–transform intersection and the thick edges of opposing lithosphere along the length of the transform will tend to confine the strike-slip tectonism to a relatively narrow and temporally stable principal transform displacement zone. In contrast, at very fast rates of accretion large offset transforms (> 100 km) place relatively thin edges of lithosphere (< 15 km) against axes of accretion and consequently the changes on accretionary processes will not be as profound: the crust will not thin dramatically; the mantle weld will be relatively weak; and the upper mantle will not be as heterogeneous.

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In addition, the thin edges of lithosphere and the higher strain rates associated with fast-slipping transforms create an environment that makes it relatively easy for complex and temporally unstable geometries to develop within the principal transform displacement zone.

INTRODUCTION: RIDGE–TRANSFORM–RIDGE PLATE BOUNDARIES

First-order relationships

Compilations of reconnaissance bathymetric data of the oceans (e.g. Menard and Chase, 1970; physiographic map of Heezen and Tharp, 1977) and earthquake epicenter maps (e.g. Barazangi and Dorman, 1969) document that transform faults routinely offset the axis of the world-encircling Mid-Oceanic Ridge System. The distribution of transforms along the axis of the ridge, however, is not uniform. For example, along the axis of the slowly accreting Mid-Atlantic Ridge south of the Azores, transforms offset the axis every 50 km to 100 km (Fox et al., 1969; Johnson and Vogt, 1973; Collette et al., 1974; Phillips and Fleming, 1978) whereas along the rapidly accreting East Pacific Rise transforms occur infrequently and are located several hundreds of kilometers apart (Mammerickx and Smith, 1978). The offset lengths of these transforms range from several kilometers (i.e. Kurchatov Transform —Searle and Laughton, 1977) to well over 100 km (i.e. Equatorial transforms—Heezen et al., 1964; Van Andel et al., 1967) and the transform domain is composed of linear topography that is parallel with the strike of the transform fault. (The term transform domain is used here as a morphotectonic term to describe all the topographic elements that create the distinctive transform morphology; strike-slip tectonism takes place within the domain but may be confined to a relatively narrow zone.) This distinctive transform topography can be traced away from the crest of the Mid-Oceanic Ridge across its flanks suggesting that the properties of oceanic lithosphere created within and proximal to a transform boundary contrast in some fundamental way with "normal" lithosphere.

Kinematic considerations mandate that strike-slip tectonism must be associated with transform boundaries (Wilson, 1965; Sykes, 1967). The total accretion rate of ridge segments varies from less than 2 cm/yr to approximately 18 cm/yr and, therefore, the slip-rate of transforms offsetting these ridge axes will vary accordingly. With slip rates varying by a factor of 10, the style, timing and intensity of deformation developed within and along the transform are likely to change in a progressive and systematic way as rates of slip range from low to high values (Fig. 1).

Geometric considerations of transforms coupled with the monotonic thickening of oceanic lithosphere with age (e.g. Parker and Oldenburg, 1973) indicate that a cold edge of lithosphere will be juxtaposed against the truncated axis of accretion at a ridge–transform intersection. The thickness of a truncating edge of lithosphere will

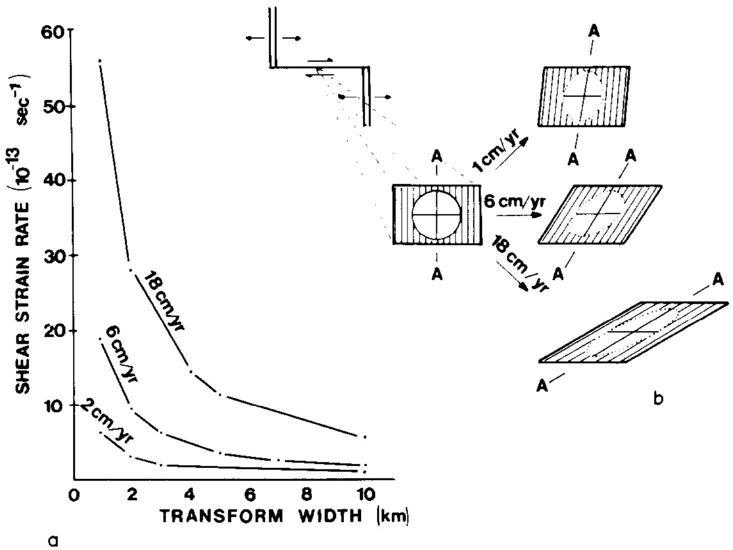
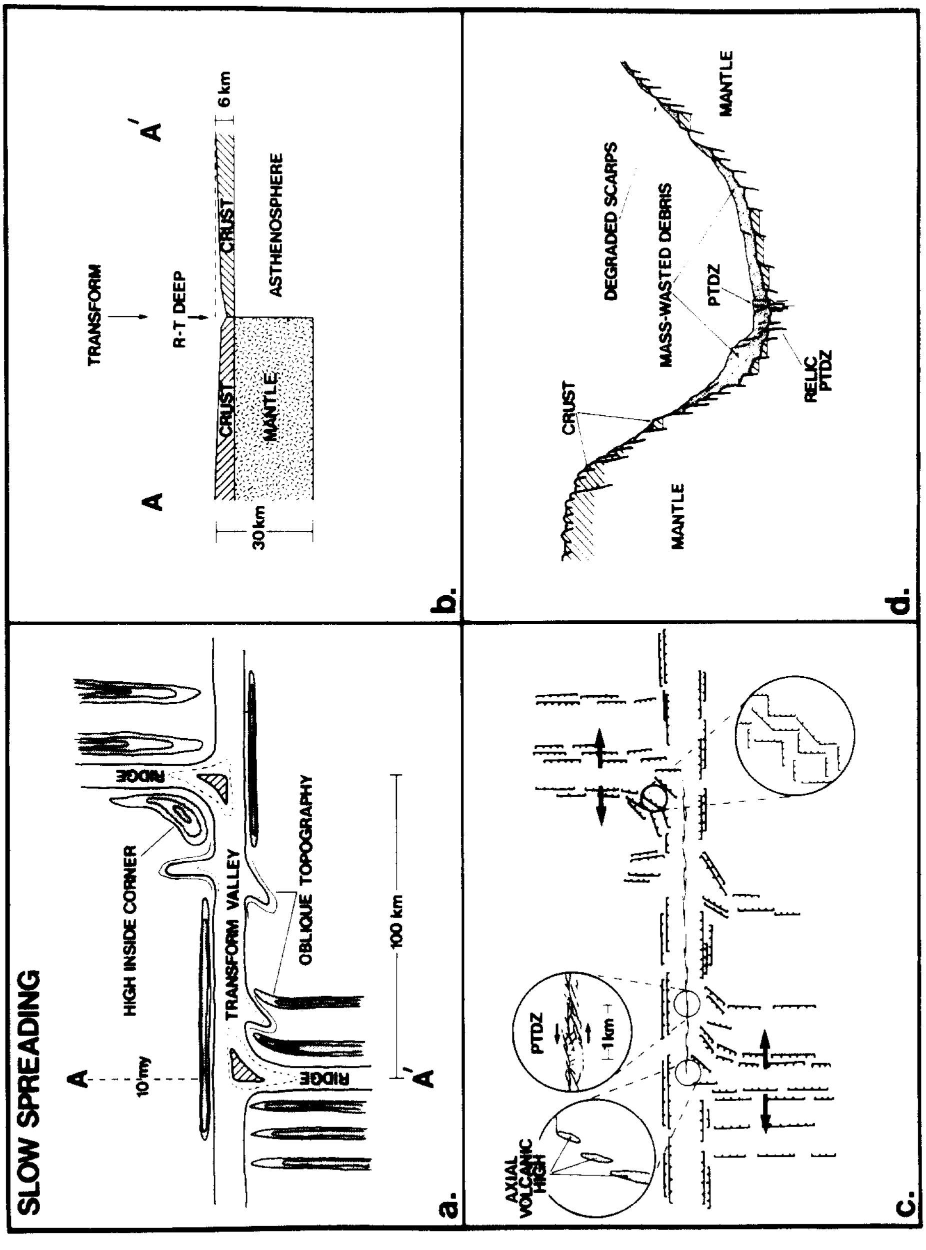


Fig. 1. a. Diagram depicting how shear strain rate changes as a function of the width over which shear displacement occurs for a range of transform slip-rates. Field studies suggest that shear displacements for oceanic transforms are confined to a narrow interval (2–4 km; e.g. Eittreim and Ewing, 1975; ARCYANA, 1975; Choukroune et al., 1978; CYAMEX and Pastouret, 1981; Castillo et al., 1982; OTTER, 1984).
b. Schematic illustration showing relative shear deformation of a rectangle during the same time interval with different slip-rates. Numbers at right give percentage of extension along line $A-A$.

depend on its age which will vary as a function of the transform slip rate and the length of the transform offset. Ridge axis segments with low accretion rates (1–2 cm/yr) that are terminated by large-offset transforms (80–300 km) will be juxtaposed against evolved lithosphere that is several tens of kilometers thick but the contrast in the thickness of lithosphere across a ridge–transform intersection will decrease markedly as the rate of accretion increases.

It seems intuitively clear that, in the simplest case, the distinctive morphotectonic fabric of the transform domain must be a product of the complex interplay between strike-slip tectonism and the effects of juxtaposing a cold edge of lithosphere against an accreting plate boundary. The structural and petrologic manifestations of this tectonic process will be spatially variable because at one end of the spreading-rate spectrum there are ridge–transform–ridge (R–T–R) plate boundaries that are characterized by very low slip rates (< 2 cm/yr) and that have relatively thick edges of lithosphere (> 30 km) at the ridge–transform intersections. In contrast, at the other end of the spreading spectrum, the R–T–R boundaries are slipping at 18 cm/yr and relatively thin edges of lithosphere (< 15 km) are juxtaposed against a ridge axis at a transform boundary. We suggest that the morphology and the constitution of the oceanic lithosphere within the transform domain change in a systematic manner and that this variation can be related to changes in the kinematic and geometric parameters outlined above.


SLOW SPREADING
a.
A
A'
10my
HIGH INSIDE CORNER
TRANSFORM VALLEY
OBLIQUE TOPOGRAPHY
RIDGE
RIDGE
100 km
b.
A
A'
TRANSFORM
R-T DEEP
CRUST
CRUST
MANTLE
ASTHENOSPHERE
6 km
30 km
c.
AXIAL VOLCANIC HIGH
PTDZ
1km
d.
CRUST
MANTLE
MANTLE
DEGRADED SCARPS
MASS-WASTED DEBRIS
PTDZ
RELIC PTDZ

CHARACTERISTICS OF RIDGE–TRANSFORM–RIDGE PLATE BOUNDARIES

Slowing-slipping transforms

Bathymetric maps of slowly-slipping transform faults (full rate 1.5–5 cm/yr) with variable lengths (e.g. Heezen et al., 1964; Heezen and Tharp, 1965; Van Andel et al., 1967; Fox et al., 1969; Collette et al., 1974; Phillips and Fleming, 1978; Searle, 1981) demonstrate that these plate boundaries exhibit a similar morphotectonic fabric but the relief, scale and gradients of the terrain elements which characterize these tectonic domains vary systematically with changing length of transform offset: the larger the length of the transform the more dramatic the topography. All slowly-slipping transform faults are characterized by topographic lineaments that strike at a high angle to regional magnetic isochrons (Fig. 2a), and these topographic features, linear ridges and valleys, and aligned closed basins and conical peaks, fall within a transform domain; the scale of the topographic features and the width of the domain vary as a function of transform length. The most striking feature of a slowly-slipping transform domain is the broad, anomalously deep valley that is defined by inward-facing and opposing walls which rise up and away from an axis of maximum depth producing relief of 1000–5000 m. Small-offset (< 30 km) transforms like Transform A (Renard et al., 1975; Phillips and Fleming, 1978) and Kurchatov (Searle and Laughton, 1977) have relief of approximately 1500 meters contrasting markedly with the several thousands of meters of relief developed within the transform domain of large-offset (> 100 km) transforms like the Romanche (Heezen et al., 1964), Vema (Van Andel et al., 1971), Kane (Fox, 1972; Purdy et al., 1979) or Oceanographer (Fox et al., 1969; Schroeder, 1977; Fox et al., 1984).

The lower portions of the walls of slowly-slipping transforms are parallel with the regional transform fabric but at higher levels on the walls ridge-axis trends are often observed. Ridge-flank topography, made up of ridges and troughs, can be traced towards the transform over distances of tens of kilometers and as the transform is

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Fig. 2. a. First order morphotectonic elements associated with slowly-slipping (< 6 cm/yr) ridge–transform–ridge (R–T–R) plate boundaries.
b. Schematic cross-section of the oceanic lithosphere proximal to a 100 km offset slowly-slipping ridge–transform intersection. Location of cross-section *A–A* is given in Fig. 2a. Thickness of lithosphere is calculated as a function of age (e.g. Parker and Oldenburg, 1973); edge effect complexities are ignored.
c. Structural grain of slowly-slipping R–T–R plate boundary. Note that the ridge generated structures facing the transform become progressively more oblique with proximity to the transform over distances of 10–25 km (e.g. Searle and Laughton, 1977; Schroeder, 1977; Searle, 1979).
d. Schematic geologic cross-section across a slowly-slipping transform boundary based on submersible and high resolution bathymetric studies of North Atlantic transforms (OTTER, 1984; ARCYANA, 1975; Choukroune et al., 1978; VE = 4 : 1). Small throw, inward facing dip-slip faults create relief of transform valley; principal transform displacement zone (*PTDZ*) defined by a narrow belt of deformation centered about axis of maximum depth.

approached it is common, but not always the case, for the sea floor to deepen and for the orientation of the ridges and troughs to become oblique to the strike of the rise axis, swinging into the direction of strike-slip motion (Fig. 2a, c; Whitmarsh and Laughton, 1975, 1976; Searle and Laughton, 1977; Schroeder, 1977; Searle, 1979; 1981; Fox et al., 1984). At each ridge–transform intersection one flank of the ridge is accreted against the aseismic limb of a fracture zone never experiencing strike-slip related tectonism (the term fracture zone is used to include the transform and the aseismic limbs). Ridge-generated terrain in this environment is truncated abruptly by the aseismic limbs of the fracture zone, and the sea floor generally deepens, but there is little or no change in trend of the ridge axis parallel terrain as the fracture zone is approached (Fig. 2a, c; Searle and Laughton, 1977; Searle, 1979, 1981). Within some fracture zones long, linear ridges, trending parallel with the regional trend of the fracture zone, rise above the surrounding sea floor to create discontinuous ribbons of anomalously shallow topography; these ridges are thought to be composed predominantly of serpentinized ultramafic rocks and have probably evolved by diapirism of serpentizined blocks (Bonatti, 1976, 1978; Fox et al., 1976; Bonatti and Hamlyn, 1978; Bonatti and Chermak, 1981; Francis, 1981; OTTER, 1983).

The valley walls of slowly-slipping transforms develop regional gradients of 20°–30°, have relief on the order of a few to several thousand meters (e.g. Heezen and Tharp, 1965; Fox et al., 1969) and expose basalts, gabbros, ultramafics and their metamorphic equivalents (e.g. Miyashiro et al., 1969; Bonatti et al., 1970; Melson and Thompson, 1971). The relief of the transform valley is, however, not produced by a few, large-throw faults which provide tectonic windows into the foundation of the oceanic crust (ARCYANA, 1975; Francheteau et al., 1976; Fox et al., 1980) but rather these regionally steep slopes are the product of a very large number of inward-facing escarpments with relief of a few hundred meters or less that integrate spatially to create a deep valley (Detrick et al., 1973; ARCYANA, 1975; Francheteau et al., 1976; Choukroune et al., 1978; Searle, 1979, 1981; OTTER, 1980, 1984; D. Needham, unpublished SEA BEAM data). Deep-sea submersible investigations of the 3000 m high north and south walls of the Oceanographer Transform have found that the axis and the lower flanks of the transform valley are choked with an admixture of pelagic debris and talus creating steep (10°–30°) slopes that dip towards the axis (Fig. 2d; OTTER, 1980, 1984). At higher elevations on the walls, near-vertical, laterally discontinuous scarps expose predominantly gabbroic or ultramafic rocks, and have relief of several meters to a few hundred meters. These inward-facing outcrops have been badly incised and modified by mass-wasting; the scarps have orientations that are variable ranging from transform parallel to ridge-axis parallel (OTTER, 1984). The scarps are linked by steeply-dipping (20°–40°), semi-consolidated wedges of sediment creating a blanket that laps up, in an undisturbed fashion, against and around rock outcroppings.

Apart from the occasional rock fall and minor differential subsidence (Delong et

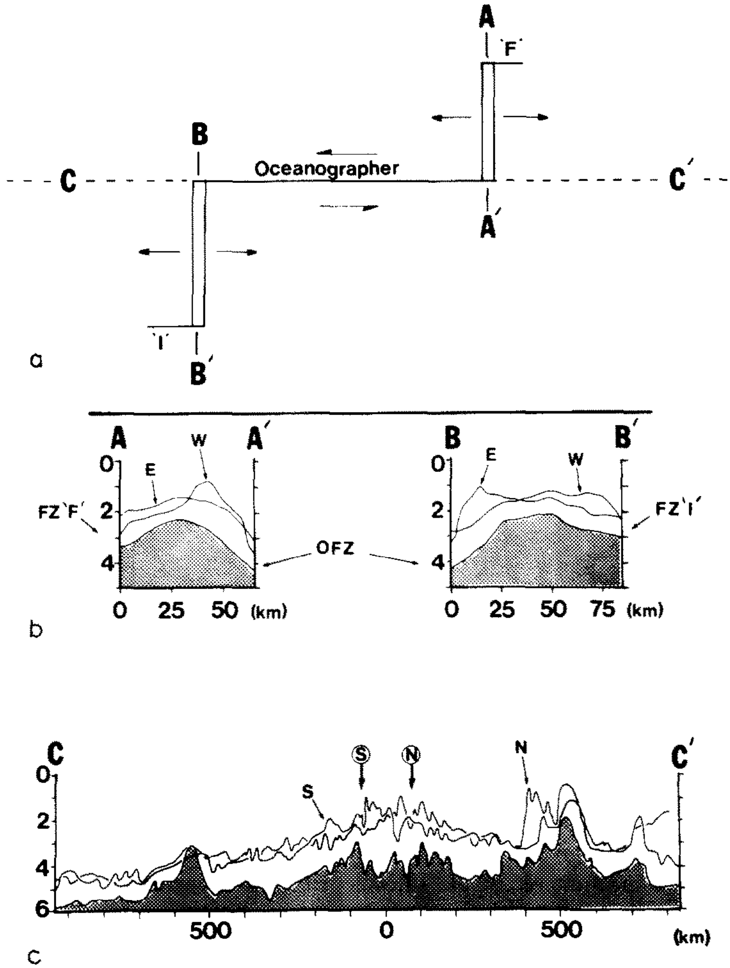

Fig. 3. a. Ridge–transform plate geometry shown for the Mid-Atlantic Ridge at 35°N including the Oceanographer, F and I transforms. Locations of cross-sections shown below are indicated by $A-A'$, $B-B'$ and $C-C'$.

b. Topographic cross-sections down the ridge axis segments to the north and south of Oceanographer. Shaded areas indicate depth to the floor of the rift valley along strike; E and W denote depths to crestal mountains on eastern and western side of ridge. Note that the change in depth toward the bounding transforms is greatest toward the large-offset (128 km) Oceanographer and less towards the shorter offset (< 30 km) transforms (Schroeder, 1977; Fox et al., 1984).

c. Profile $C-C'$ indicates maximum depth to basement along axis of the Oceanographer Fracture Zone (shaded). Lines N and S indicate depths to the crest of the bounding fracture zone walls (N for North and S for South); N and S (circled) give the location of the northern and southern ridge–transform intersections. Apart from two seamount complexes located approximately 500 km away from the ridge–transform intersections, the aseismic limbs of the fracture zone deepen with increasing age and the floor of the fracture zone is always deeper than flanking sea-floor.

al., 1977, 1979) slowly-slipping transform valley walls appear to be devoid of tectonic activity. Direct evidence for recent tectonic activity within both Transform A and the Oceanographer Transform, including recent near-vertical fault traces and fresh rubble, is confined to a narrow zone only a few hundred meters to a few kilometers wide that is centered about the axis of maximum depth (Fig. 2c; ARCYANA, 1975; Choukroune et al., 1978; OTTER, 1984). This narrow belt of complex and rapidly evolving terrain, disrupted by an anastomsing network of faults, is assumed to mark the location of the principal transform displacement zone (PTDZ) and represents the shallow level manifestations of strike-slip tectonism (Fig. 2c, Choukroune et al., 1978; OTTER, 1984).

Mapping of ridge segments proximal to slowly-slipping transform boundaries shows that over distances of 20–40 km the rift-valley floor slopes towards the closed-contour deep of the ridge–transform intersection with regional gradients of 1°–2° and elevation changes of 500–1500 m (Fig. 3; Needham and Francheteau, 1974; Macdonald and Luyendyk, 1977; Ramberg and van Andel, 1977; Schroeder, 1977; Phillips and Fleming, 1978; Fox et al., 1984). The larger the transform offset, the greater the change in the along-strike relief of the rift valley floor (Fig. 3). Furthermore, as the transform boundary is approached, the strike of the rift valley wall that is truncated by the transform becomes oblique with respect to the orientation of the rift valley (Fig. 2a, c), whereas the trend of the opposing rift wall parallels the strike of the rift valley and is unchanged with increasing proximity to the aseismic limb of the fracture zone (Whitmarsh and Laughton, 1975, 1976; Schroeder, 1977; Phillips and Fleming, 1978; Searle, 1979; Karson and Dick, 1983; OTTER, 1983a; K. Macdonald and P. Fox, unpublished Deep Tow data—Vema Transform).

Deep-towed camera and submersible investigations of the relationships developed at the eastern intersection of the Oceanographer Transform with the MAR indicate that the morphology of the rift valley is shaped by a distinctive fault pattern and by vigorous mass wasting of basement outcroppings (OTTER, 1983). Along the floor of the rift valley, over a distance of 10–20 km, the orientation of faults, fissures and volcanic ridges becomes increasingly more complex with proximity to the transform. Although the ridge-axis parallel orientations are still apparent, transform parallel structures and structures oblique to both the ridge and transform are recognized. The oblique structures are concentrated on the transform side of the rift floor and become more oblique with respect to the ridge axis as the transform boundary is approached mimicking the morphologic grain shown on surface-ship derived bathymetric maps (Fig. 2a, c). The structural fabric of the rift floor is rapidly modified, and obscured by degradation of bedrock exposures as the products of mass wasting, talus wedges and debris fans fill depressions, abut rock escarpments and smooth the terrain.

As parcels of crust and associated sediment leave the rift floor to create the distinctive inward facing steps of the rift valley walls the polarized structural fabric developed on the rift floor is preserved. The morphology of the rift flank that is terminated by the aseismic limb of the fracture zone is the product of inward facing scarps that approximately parallel the ridge axis along with broad, sedimented slopes that dip steeply (10°–40°) towards the rift axis (Karson and Dick, 1983; OTTER, 1983). The morphology of the opposite wall, the transform side of the accreting plate boundary, is made up of a series of inward facing escarpments, exhibiting a range of orientations, that is linked by steeply dipping sediment slopes. Oblique trending escarpments as well as E–W and N–S striking scarps intersect to create the overall oblique trend of the rift wall (Fig. 2c). Irrespective of orientation, however, no evidence for strike-slip displacement is observed on the escarpments and evidence for normal (dip-slip) displacement is ubiquitous regardless of scarp orientation (Karson and Dick, 1983; OTTER, 1984).

Geophysical evidence indicates that the velocity structure and distribution of mass of the oceanic crust and upper mantle within the slowly-slipping fracture zone domain contrast with the properties of "normal" oceanic lithosphere. Gravity data indicate that large mass anomalies are often associated with fracture zones (Cochran, 1973; Robb and Kane, 1975; Sibuet and Veyrat-Peiney, 1980; Louden and Forsyth, 1982). These data have been interpreted in contrasting ways. In one instance, mass anomalies are thought to reflect the existence of upper mantle rocks at relatively shallow crustal levels (Cochran, 1973; Robb and Kane, 1975) but, in another model, mass anomalies are not required if mantle density is allowed to change with age across the fracture zone (Sibuet and Veyrat-Peiney, 1980). The fundamental unconstrained nature of gravity data makes it difficult to arrive at a unique solution but it seems clear that the mass distribution in and around fracture zones is heterogeneous disrupting the gravity anomaly pattern characteristic of "normal" oceanic lithosphere (Louden and Forsyth, 1982).

Seismic refraction experiments along the large offset fracture zones (Kane—Detrick and Purdy, 1980; Cormier et al., 1983; Oceanographer—Sinha, 1981; Sinha and Louden, 1981, 1983; and Vema—Detrick et al., 1982) reveal anomalous crustal structure. The Kane fracture zone is characterized by crustal thicknesses of only 2–3 km and the crust exhibits low compressional wave velocities. Proximal to the Oceanographer fracture zone refraction data indicate that the crust thins towards the fracture zone, and within the fracture zone there is evidence for very thin crust with anomalously low velocities (Sinha and Louden, 1983). The thickness of the oceanic crust within the Vema fracture zone is not anomalously thin but the compressional wave velocities observed for the crustal arrivals are anomalously low suggesting that the seismically defined crust is comprised predominantly of sheared and hydrated upper mantle rocks (Detrick et al., 1982). A seismic refraction experiment designed to investigate the velocity structure of a small offset (15 km) fracture zone suggests that anomalously high velocities (7.3–7.6 km/s) are found at shallow crustal levels (< 2 km) along the fracture zone (White and Matthews, 1980).

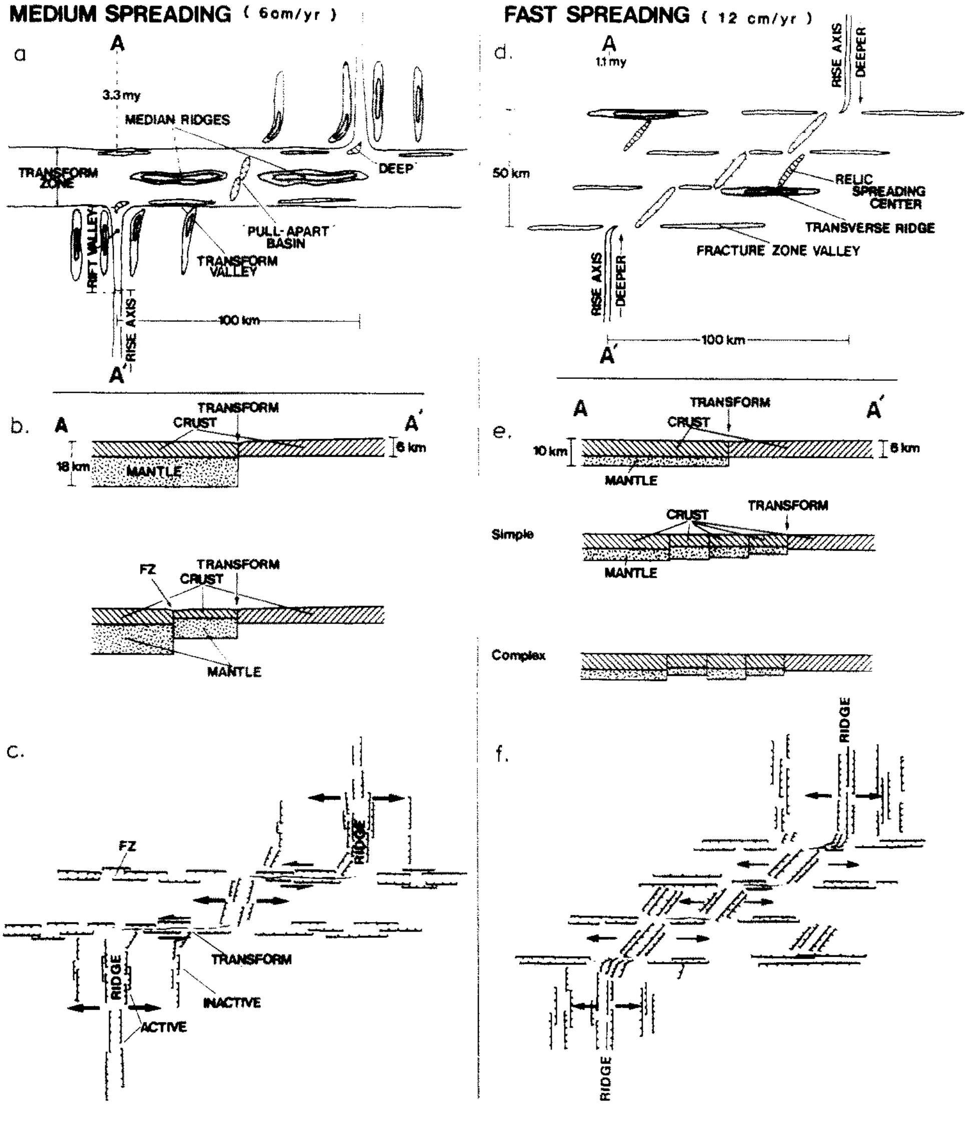

Fig. 4. a. First-order morphotectonic elements associated with a medium slip-rate (6–12 cm/yr) ridge–transform–ridge plate boundary.
b. Schematic cross-section of the oceanic lithosphere proximal to a medium slip-rate ridge–transform intersection. Location of cross-section $A-A'$ is given in Fig. 4a. Two possible cross-section geometries are shown: one depicts the simplest case where the transform is defined by a single shear zone (this geometry not shown in 4a); the other is a more complex geometry that would evolve if sea floor within the transform domain is created along a short extensional relay zone (reference Fig. 4a and 4c). Thickness of lithosphere is calculated as a function of age (e.g. Parker and Oldenburg, 1973); edge effect complexities are ignored.
c. Structural grain of a medium slip-rate R–T–R plate boundary. Ridge axis swings sharply into

Medium slip-rate transforms

The regional, morphotectonic fabric of medium slip-rate transforms (6–12 cm/yr) and adjoining ridge segments have not been as well investigated as their slowly-slipping counterparts. Recconnaissance studies, however, of the Tamayo (Kastens et al., 1979), Rivera (Reid, 1976; Prothero and Reid, 1982), Inca (Embley et al., 1984), Orozco (Trehu, 1982) and Siqueiros (Crane, 1976) transforms demonstrate that the transform domains of these plate boundaries exhibit many of the same morphotectonic relationships as their slowly-slipping counterparts, but the scale of the features is diminished and in some ways the tectonic pattern appears to be more complex when viewed along strike of the transform. The depth to the ridge axis increases towards the transform and the axis of the ridge sometimes terminates in a small, closed-contour nodal basin at the ridge–transform intersection (Fig. 4a; Macdonald et al., 1979; Tamayo Tectonic Team, 1983) but, in contrast to slowly-slipping ridge-transform intersections, the change in relief along the ridge axis is only a few hundred meters and the nodal basin is small (Fig. 4a). In the case of the Tamayo Transform, which is only slipping at 6 cm/yr, the morphology of the axis of the East Pacific Rise, over a distance of 30 km, evolves from a broad, gently-rifted swell into a shallow rift valley as the Tamayo Transform is approached (McClain and Lewis, 1980; Macdonald et al., 1979). The same structural fabric and relationships that are characteristic of slowly slipping R–T intersections are found to be associated with the medium slip rate R–T intersections (compare Fig. 2c with Fig. 4c; Tamayo Tectonic Team, 1983). The development of well defined rift valley morphology, however, is not recognized at faster-slipping ridge–transform intersections (> 8 cm/yr full rate; K. Macdonald and P. Fox, unpublished SEA BEAM data). Furthermore, at these faster slipping ridge–transform intersections structures on both sides of the ridge axis develop oblique trends progressively as the transform

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transform over a distance of several kilometers; principal transform displacement zone is characterized by two shear zones linked by an extensional relay zone.

d. First-order morphotectonic elements associated with a fast-slipping (12–18 cm/yr) ridge–transform–ridge plate boundary.

e. Schematic cross-section of the oceanic lithosphere proximal to a fast-slipping ridge–transform intersection. Location of cross-section $A-A'$ is given in Fig. 4d. Three plate geometries are shown: the first assumes that the transform is defined by a single shear zone that links the two offset ridge axes (this geometry is not shown in Fig. 4d); the second assumes that the transform is characterized by a shear zone with a complex but stable geometry made up of strike-slip strands linked by right stepping extensional relay zones; the third assumes that the shear zone geometry of strike-slip strands and extensional relay zones is not stable in time and therefore the thickness of the lithosphere across the transform zone is variable.

f. Structural grain of a fast-slipping R–T–R plate boundary. Ridge axis swings very sharply into transform over a distance of several kilometers; principal transform displacement zone is characterized by several shear zones linked by extensional relay zones.

boundary is approached but only over distances ranging from several kilometers to 10 km; the larger the transform offset and/or the greater the slip-rate, the more apparent the development of oblique trends (Fig. 4f; Crane, 1976; K. Macdonald and P. Fox, unpublished SEA BEAM data for Rivera, Orozco, Clipperton and Siqueiros transforms). Deep-tow and submersible investigations at the Tamayo–East Pacific Rise intersection indicate that the oblique trending faults found at the ridge–transform intersection accommodate only dip-slip motion and that the transition from accretion related tectonics to strike-slip dominated processes is very narrow (< a few km; Macdonald et al., 1979; Tamayo Tectonic Team, 1983). At the present time we have no constraints on the motion of the oblique faults found at faster slipping ridge–transform intersections.

The transform domain of a medium slip-rate transform is often 20–30 km wide and is comprised of elongate ridges and troughs that strike at a high angle to the ridge axis and that develop relief, in general, of a few hundred to a thousand meters (Fig. 4a). These ridges and troughs are discontinuous and are sometimes interrupted by anomalously deep (3–4 km) basins and scarps striking at an angle oblique to the trend of the transform (e.g. Tamayo—Macdonald et al., 1979; Kastens et al., 1979; Orozco—Trehu, 1982). Submersible investigations at the Tamayo Transform document that evidence for recent tectonism within the transform domain is confined to a very narrow belt less than a kilometer wide and this interval is taken to represent the principal transform displacement zone (CYAMEX and Pastouret, 1981; Tamayo Tectonic Team, 1983). This zone is offset along strike, however, appearing to be displaced by an extensional relay zone or pull-apart basin that is located along the center of the transform (Fig. 4c). It is in this area that piercement structures (diapirs) are seen to protrude up and, in one case, through, the sedimentary fill of the transform (Macdonald et al., 1979). In addition, extensional relay zones have been predicted to exist within the Rivera (Prothero and Reid, 1982) and Orozco (Trehu, 1982; K. Macdonald and P. Fox, unpublished SEA BEAM data) transforms based on a distinctive pattern defined by earthquake epicenters and topography.

The Clipperton Transform (12 cm/yr slip rate), which offsets the East Pacific Rise 80 km in a right-lateral sense (in contrast to the other major transforms of the East Pacific Rise which all offset the axis in a left-lateral sense), is the one transform that does not exhibit a morphotectonic pattern suggestive of strike-slip zones linked by an extensional relay zone. Although the transform domain is typically wide (20 km) along most of its length, recently acquired SEA BEAM and SEA MARC I side scan data (P. Fox, K. Macdonald and W. Ryan, unpublished data) indicate that the truncated rise axis tips have propagated towards each other narrowing the transform domain. SEA MARC I data also define a narrow (< 1 km) belt of disturbed terrain disrupted by discontinuous and anastomosing transform parallel faults that can be traced in a direct line from one ridge tip to another.

With the exception of the deep basins that are oblique to the grain of the transform, the relief of medium slip-rate transforms is generally only several hundred meters to approximately one thousand meters and *in situ* investigations of this terrain indicate that small-throw faults (< 100 m) integrate spatially to create steep regional slopes (Macdonald et al., 1979; CYAMEX and Pastouret, 1981; Tamayo Tectonic Team, 1983; Embley et al., 1984). Fine-grained extrusive basaltic rocks are recovered in abundance from these slopes (e.g. Batiza et al., 1977; Fornari et al., 1983; Tamayo Tectonic Team, 1983) suggesting that the processes associated with the development of medium slip-rate transforms do not lead to the routine exposure of deep-seated rocks (e.g. gabbroic and ultra-mafic rocks). In one instance, however, gabbros, serpentinized ultramafic rocks and mafic breccias have been recovered from fracture zone escarpments that offset the Galapagos Rise (Nishimori and Anderson, 1973). Seismic refraction experiments show that the oceanic crust associated with the Tamayo Transform (McClain and Lewis, 1980) and Orozco (Trehu, 1982; Ouchi et al., 1982) is 1–3 km thinner than normal oceanic crust, but these results do not indicate the extreme crustal thinning thought to be typical of slowly-slipping transforms (Detrick and Purdy, 1980; Fox et al., 1980; Stroup and Fox, 1981).

Fast-slipping transforms

The morphology and salient tectonic elements of fast-slipping transforms (12–18 cm/yr) have been poorly known until very recently. Compilations of reconnaissance data (Mammerickx and Smith, 1978; Lonsdale, 1978; Kureth and Rea, 1981; Rea, 1981) reveal that several large-offset transforms (Quebrada–Gofar System between 3°30′S and 4°40′S; Wilkes at 9°S; Garret at 12°S) are found to disrupt the axis of the East Pacific Rise along the Nazca–Pacific plate boundary, the fastest accreting ridge segment in the world. These transforms are composed of wide zones of discontinuous ridges and troughs that generally develop several hundred to a few thousand meters of relief (Figs. 4d, f). The Quebrada–Gofar transform system offsets the ridge axis a total of 390 km and can be best described as a 150 km-wide shear zone that is made up of a large number of short transforms that are linked together by short, oblique trending, extensional relay zones (Fig. 4d, Lonsdale, 1978; Searle and Francis, 1982; Searle, 1983). The Wilkes Transform offsets the rise axis 200 km and defines a 100 km wide interval of complexly arranged ridges and troughs; Kureth and Rea (1981) present a tectonic model based on bathymetric and magnetic data proposing that the transform domain is characterized by two shear zones linked by an extensional relay zone.

A Deep Tow survey of a 15 km^2 portion of the Quebrada Transform near its eastern intersection with the EPR axis defines a single narrow (< 200 m) principal transform displacement zone lying along the north side of a 7 km wide transform trough (Lonsdale, 1978). The east–west trending grain of the transform trough is disrupted by structures that are 45° oblique to the transform and are interpreted to represent normal faults (Lonsdale, 1978). A reconnaissance map of the intersection east of the Deep-tow survey shows that the axis of the rise crest bends sharply into the transform over a distance of a several kilometers and slopes into the transform trough (Lonsdale, 1978).

The morphology of the Garret Transform, the fastest-slipping large-offset (130 km) transform in the world, was essentially unmapped until recently when a reconnaissance SEA BEAM survey (P. Lonsdale, pers. commun., 1982) followed by a detailed SEA BEAM investigation (P. Fox, unpublished data) showed that the Garret is characterized by a 20–30 km wide transform domain made up of several short strike-slip segments that are linked by oblique-trending ridges interpreted to be extensional relay zones. These results as well as SEA BEAM crossings (P. Lonsdale, pers. commun., 1982) and GLORIA data (Searle, 1983) from the Wilkes and Quebrada–Gofar transforms suggest that fast-slipping transforms have a distinctive and diagnostic morphotectonic character, defined as a broad region of complex topography composed of swaths of transform-parallel ridges and troughs that are bounded by short oblique ridges and basins (Figs. 4d, f).

GEOLOGIC MODEL

The morphotectonic character exhibited by ridge–transform–ridge plate boundaries varies in a systematic way: the relief and scale of the terrain associated with the transform domain becomes better defined as the thickness of the truncating edge of lithosphere at the ridge–transform boundary increases (compare Figs. 2b, 4b, e). For example, at the eastern ridge–transform intersection of the Oceanographer Transform, lithosphere that is approximately 35 km thick is juxtaposed against the ridge axis (Fig. 5); in this tectonic environment, the depth to the nodal basin approaches 5000 m and the relief developed at the intersection approaches 4000 m. The Tamayo Transform, on the other hand, juxtaposes lithosphere that is only 15 km thick against the axis of the East Pacific Rise and, in this tectonic environment, the nodal basin is only 3400 m deep and the relief developed at the intersection is about 1000 m (Fig. 5). Well constrained bathymetric data do not exist for many ridge–transform intersections and, in the absence of these data, we have plotted the maximum depth measured at a number of ridge–transform intersections against the estimated thickness of lithosphere at that intersection (Fig. 6). The correlation is striking and supports our working hypothesis that the processes responsible for the formation and evolution of oceanic lithosphere proximal to ridge–transform plate boundaries are sensitive to the plate-thickness geometry (Fox et al., 1981).

Based on theoretical considerations that investigate the thermal conditions governing the accretion and evolution of oceanic lithosphere, investigators have proposed that slowly accreting plate boundaries (1–2 cm/yr full rate) are approaching a lower limit below which sea-floor spreading (i.e. the emplacement of basaltic melts at shallow crustal levels) can no longer take place (Sleep, 1975; Kusznir and Bott, 1976). Bottinga and Allègre (1978) have suggested that the slower advection

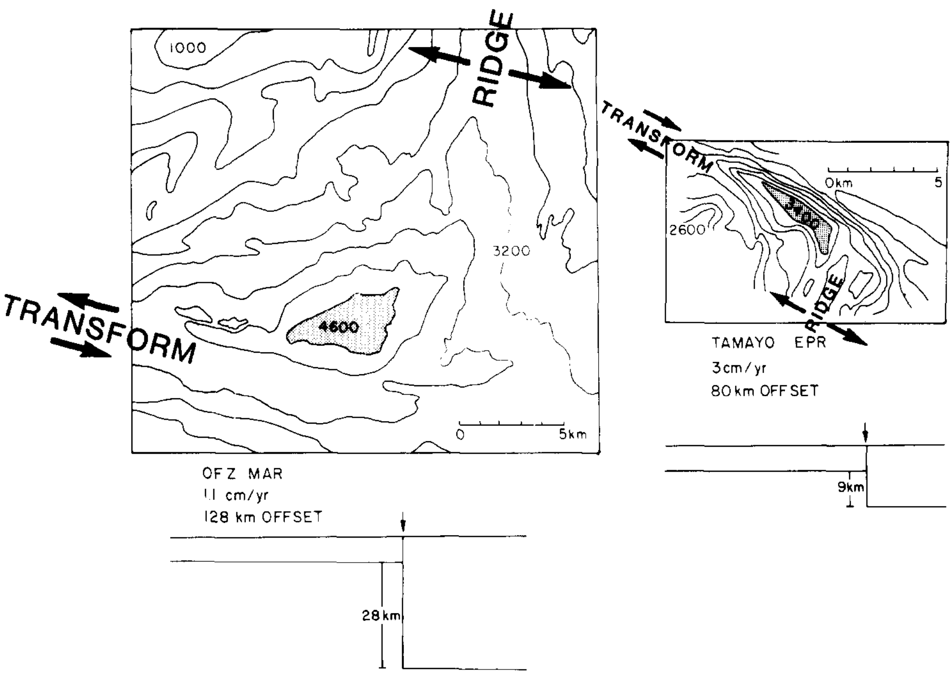

Fig. 5. Bathymetric comparison of two ridge–transform intersections (same scale) characterized by contrasting thicknesses of truncating lithosphere at the transform boundary. Oceanographer transform shown at left; Tanayo transform shown at right. Although the morphotectonic character of the two ridge–transform intersections is the same, the scale and relief of the topographic elements is much greater for the Oceanographer than for the Tamayo. Depth in meters.

beneath slowly spreading ridges results in lower temperatures and, as a result, less partial melt migrates to the surface to form oceanic crust. A compilation of seismic refraction data for segments of the Mid-Oceanic Ridge that exhibit different accretion rates suggests a correlation between total crustal thickness and spreading rate, with slower rates of spreading leading to the development of thinner crust (Reid and Jackson, 1981; Jackson et al., 1982). The results of these investigations are all consistent with the notion that at slow rates of accretion the processes leading to the generation and emplacement of basaltic melts are modified because of the cooler thermal environment governing the generation of oceanic lithosphere.

It seems intuitively reasonable to infer that the thermal environment beneath an accreting plate boundary will become even colder as a ridge–transform plate boundary is approached and a thick truncating edge of lithosphere, of variable thickness, is juxtaposed against an axis of accretion. We suggest that the result is the emplacement of oceanic lithosphere that contrasts markedly with lithosphere accreted at some distance from a ridge–transform intersection. The distance over which the properties of the oceanic lithosphere change with respect to a fracture zone will

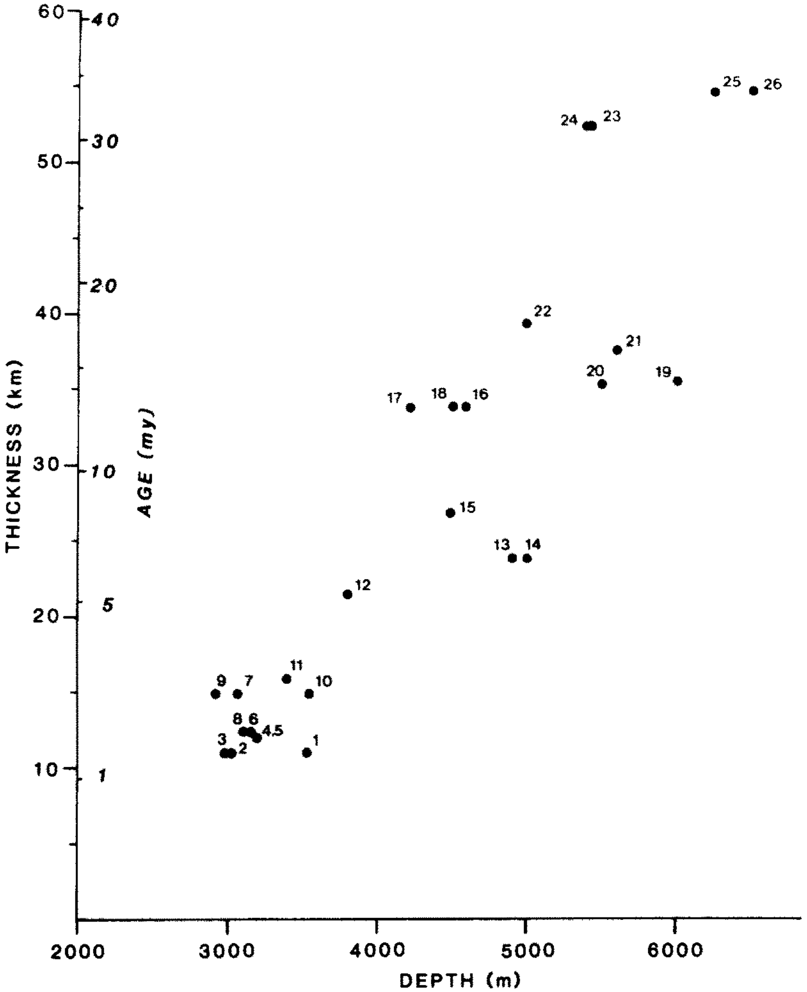

Fig. 6. Thickness of truncating lithosphere ($Z(t) = 9.4t^{1/2}$; Z = thickness in km and t = age in m.y.—Parker and Oldenburg, 1973) at various ridge–transform intersections plotted against the maximum depth observed at ridge–transform nodal deeps. Note the systematic increase in depth with increasing age of the truncating edge of lithosphere. Each dot defines a ridge–transform intersection, number by each dot keys these data to a list of locations and sources: *1*—eastern intersection of the 33°N Transform on the MAR (Naval Oceanographic Office, unpublished data, C. Smoot, pers. commun., 1982); *2*, *3*—Transform C (Phillips and Fleming, 1978); *4*, *5*-Kuchatov Transform (Searle and Laughton, 1977); *6*, *7*—Transform A (Phillips and Fleming, 1978); *8*, *9*—Transform B (Phillips and Fleming, 1978); *10*—Tamayo-EPR intersection (Tamayo Tectonic Team, 1983); *11*—Transform F (Phillips and Fleming, 1978); *12*—Blanco Transform eastern intersection with Gorda Ridge (Office of Naval Research unpublished data; C. Fox, pers. commun., 1982); *13*, *14*—Atlantis Transform on the MAR (Naval Oceanographic Office, unpublished data; C. Smoot, pers. commun., 1982); *15*—13°N Transform (Collette et al., 1974); *16*, *17*—Oceanographer Transform (Fox et al., 1969, Schroeder, 1977); *18*—Hayes Transform western intersection (Naval Research Lab, unpublished data; H. Fleming, pers. commun., 1980); *19*, *20*—Kane Transform (Fox, 1982; Purdy et al., 1978; Karson and Dick, 1983); *21*—Spitzbergen Transform

depend on the interplay of two factors: thickness of the edge of lithosphere that was juxtaposed against an axis of accretion at the time of formation and the thermal structure of the ridge segment in question (i.e. high accretion rate vs. a slow accretion rate). The cold edge of lithosphere that faces the axis of accretion across the transform lowers the temperature in the wedge of asthenosphere rising beneath the axis of accretion proximal to the plate boundary perturbing the processes of accretion.

Sleep and Biehler (1970) argued that the deep intersection depressions are the surface manifestation of viscous head loss experienced by the rising wedge of asthenosphere ascending beneath slowly-accreting plate boundaries immediately proximal to ridge–transform intersections. After emplacement this lost head is recovered as the depressed crust rises isostatically as it ages. Indeed, bathymetric data of well-mapped intersections (FAMOUS area—Phillips and Fleming, 1978; Oceanographer Transform—Fox et al., 1969; Schroeder, 1977; Kane Transform—Fox, 1972; Detrick and Purdy, 1980; Vema Transform–Needham, unpublished data; R. Prince, unpublished data) show that these nodal basins are short-lived and depths to basement initially decrease with distance from the depresion. An along-strike profile down the axis of the Oceanographer Fracture Zone illustrates how basement depths shallow rapidly away from these nodal basins (Fig. 3, profile *C*). Other deep basins are recognized along the transform axis of the Oceanographer Transform and these probably reflect expressions of vertical tectonism caused by aspects of strike-slip deformation (extensional and compressional relay zones) and/or differential hydration of blocks resulting in variable sepentinization of ultramafic rocks (Fox et al., 1976; Fox et al., 1984). The sides of the nodal basins flanking the aseismic limbs of the Oceanographer fracture zone reach a sill depth a few tens of kilometers from the ridge–transform depression and then the basement of the fracture zone floor deepens continuously with increasing age (Fig. 3, profile *C*). Independent of age, however, the depth to the axis of the fracture zone is always deeper than the adjacent oceanic lithosphere of equivalent age suggesting that some process, other than the isostatic considerations of viscous head loss, is operative along the transform.

Whitmarsh and Laughton (1976) suggested that fracture zone valleys associated with slowly-accreting plate boundaries are caused by a perturbation of volcanic processes proximal to the ridge–transform intersection due to the inhibiting effect of shear stresses focussed at the intersection and the reduced size of the magma chamber caused by the opposing edge of lithosphere. The recovery of gabbroic rocks

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eastern intersection (Naval Research Lab., unpublished data; P. Vogt, pers. commun., 1983); *22*—15°N Transform (Collette et al., 1974); *23*, *24*—Vema Transform (unpublished SEA BEAM data; H. Needham, pers. commun., 1980); *24*—Oriente Transform–Cayman Ridge intersection (CAYTROUGH, 1979); *25*—Swan Transform–Cayman Ridge intersection (CAYTROUGH, 1979).

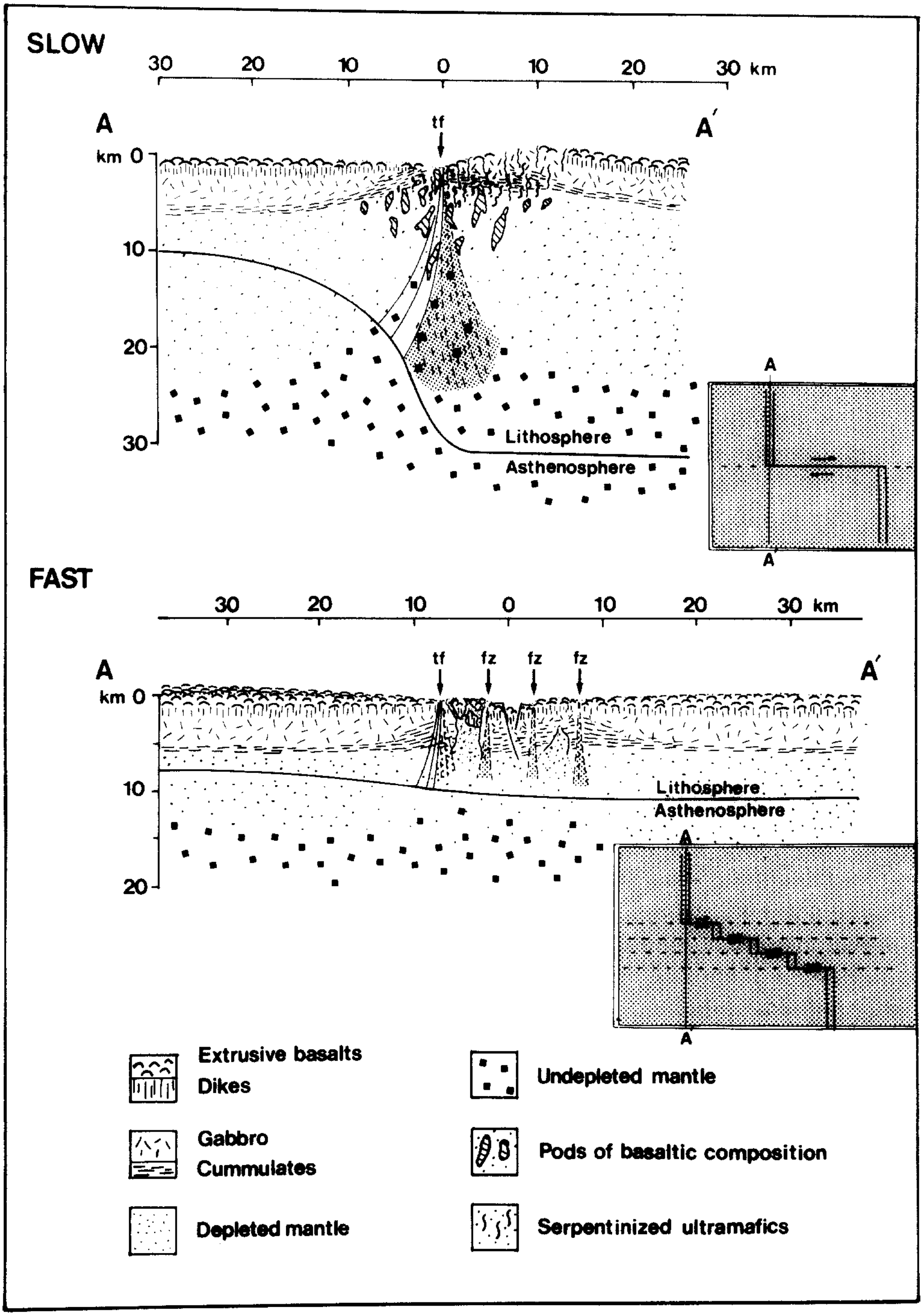

Fig. 7. Generalized geologic model of the structure of oceanic lithosphere at and proximal to a 100 km offset slowly-slipping (2 cm/yr) ridge–transform intersection (top) and a 100 km offset fast-slipping (18

from localities high on the walls of transform valleys along the Mid-Atlantic Ridge suggested to Francheteau et al. (1976) that the shallow intrusive and extrusive carapace (seismic Layer 2) of the oceanic crust is anomalously thin proximal to transforms and they proposed that this characteristic reflects a diminished volcanic budget caused by the thick, cold plate on one side of the intersection. Seismic refraction investigations of the Kane Transform (Detrick and Purdy, 1980) and the documentation by submersible sampling that oceanic crust is very thin along rift valley segments proximal to cold, thick transform edges (CAYTROUGH, 1979; Stroup and Fox, 1981; Karson and Dick, 1983) and along the bounding walls of fracture zones (Oceanographer Transform; OTTER, 1980, 1981, 1984) led to the formulation of more generalized models predicting that the occurrence of thin crust is a diagnostic and characteristic property of slowly-slipping ridge–transform–ridge intersections (Fox, 1978; Gallo and Fox, 1979; Detrick and Purdy, 1980; Fox et al., 1980; Stroup and Fox, 1981). The thickness of the oceanic crust is dependent upon the volume of basaltic melt that is segregated at depth from the asthenosphere and delivered to shallow level magma chambers. Proximal to the thick, cold wall of transform boundaries, lateral conductive heat flow reduces the ambient temperature and disrupts the upward migration of the hot upwelling asthenosphere diminishing the volume of basaltic melt generated and impairing the ability of the melt that is segregated to reach the surface. As a result, the total magmatic budget supplied to shallow level magma chambers decreases as the transform boundary is approached and the crust thins correspondingly (Fig. 7). At very large transforms, the edge effect may be so profound that the emplacement of basaltic magma at shallow levels is discontinuous with periods of accretion being punctuated by the emplacement of solely peridotites laced with stringers of basaltic rocks.

As the overlying crust thins towards the transform edge, the thickness of the underlying lithosphere will increase as newly formed lithosphere is plated against the base of the cold transform wall (Fig. 7). This thickening wedge of lithosphere will be very heterogeneous because some fraction of the melt produced during partial melting will never successfully pass through the upper mantle and contribute to the development of the crust but will be trapped within the upper mantle creating pods of rock with basaltic composition (Fig. 7). The magnitude of the cold-edge effect at ridge–transform intersections will vary as a function of the thickness of the lithosphere at the ridge axis termination; the effect will be profound at large-offset,

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cm/yr) ridge–transform intersection (bottom). The models predict that the crust will thin and the underlying lithosphere will thicken as the ridge–transform boundary is approached. Lateral heterogeneities in composition, layering and mass distribution are predicted for both cases but will be dramatic at a slowly-slipping boundary where the cold edge of the transform is large. However, given the relatively thin edges of lithosphere characterizing fast-slipping transforms, a relatively wide and complex shear zone evolves developing complex structural and compositional relationships. See text for a more complete explanation.

slowly-slipping transforms (Fig. 7a) and will be increasingly less important as the thickness of the truncating cold edge diminishes (Fig. 7b). This model predicts that as a fracture zone is approached along an isochron, the crust and underlying mantle will become increasingly more heterogeneous in terms of lithology and distribution of mass; the degree of heterogeneity and the distance over which the heterogeneity is recognized along an isochron will depend on the plate-edge geometry that characterized the ridge–transform intersection at the time of lithosphere generation. The great relief associated with those ridge–transform intersections that juxtapose relatively thick edges of lithosphere against an accreting plate boundary (i.e. slowly-slipping transforms) reflects an isostatic response to the marked changes in the mass distribution within the oceanic lithosphere caused by thinning oceanic crust and the changing composition of the upper mantle due to a decrease in partial melting. These lateral variations in mass distribution are likely to result in a change in isostasy for the lithosphere. Furthermore, the thin crust proximal to transforms will allow seawater to penetrate into the underlying upper mantle and hydrate ultramafic rocks; serpentinite diapirs will be created leading to vertical tectonism and the creation of the large serpentinite ridges that are so often the hallmark of slowly-slipping, large-offset fracture zones (Aumento and Loubat, 1971; Bonatti, 1976; 1978; Bonatti and Honnorez, 1976; Francis, 1981). Serpentinization of mafic and ultramafic rocks will further complicate and contribute to the crustal and subcrustal heterogeneity predicted for ridge–transform–ridge plate boundaries in general and slowly-slipping ones in particular. Given this complex geologic environment, it is not surprising that gravity data suggest that the mass anomalies associated with fracture zone troughs are variable and not systematic (Louden and Forsyth, 1982). Seismic refraction experiments within and proximal to the transform domain typically record anomalously low mantle velocities (>7.6–7.9 km/s; Detrick and Purdy, 1980; McClain and Lewis, 1980; Ouchi et al., 1982; White, 1983) and we suggest that these low values reflect the contamination of the upper mantle with pods and veins of rocks with basaltic compositions that have been trapped at depth (A. Nicolas, pers. commun., 1983 reports finding basaltic rocks impregnating peridotites in ophiolite terrains interpreted to be preserved remnants of transform faults).

Investigators have suggested that the oblique trends observed proximal to ridge–transform intersections reflect a change in the state of stress along the accreting plate boundary caused by a shear couple at the ridge–transform boundary (Crane, 1976; Lonsdale, 1978; Searle, 1979, 1981). We have documented that the oblique faults in this environment do not accommodate significant amounts of strike-slip motion (OTTER, 1983; Tamayo Scientific Team, 1983; Karson and Dick, 1983) and, therefore, these oblique trends are not analagous to the shear structures that are known to develop in continental shear zones and clay box experiments during the initial stages of deformation (e.g. Tchalenko, 1970; Tchalenko and Ambraseys, 1972; Wilcox et al., 1973; Courtillot et al., 1974).

We propose that the thick wedge of lithosphere that forms beneath the axis of accretion proximal to the transform boundary creates a weld with the cold, thick lithosphere at the ridge–transform interface (Fig. 8). Part of the weld links newly formed lithosphere across the aseismic limb of the fracture zone and has little effect on intersection tectonics because the sense and magnitude of the relative motion

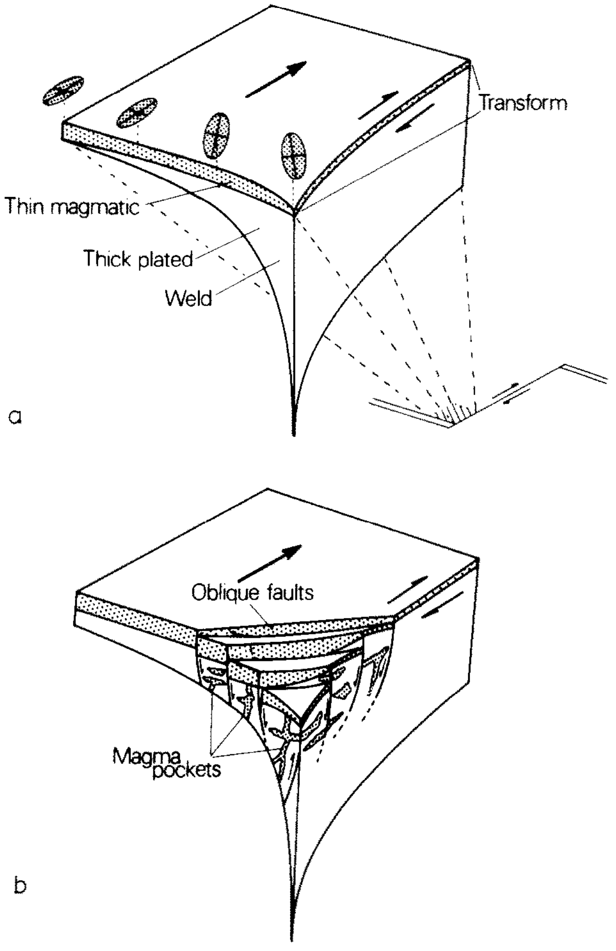


Fig. 8. a. Schematic model outlining the change in the thickness of the oceanic lithosphere as a ridge–transform plate boundary is approached; the magmatic component (crust) thins and the underlying mantle thickens. The newly formed mantle welds against the cold truncating edge of the transform creating a shear couple. Ellipses along the ridge axis show progressive reorientation of the direction of maximum tensile stress due to the shear couple generated during deformation of a mantle weld.
b. Geological consequences due to deformation of the mantle weld: proximal to the ridge–transform intersection normal faults form that strike oblique to the direction of plate motion (obliquity of the faults depend on the strength of the weld and the strain rate); batches of melt are trapped in the mantle proximal to the cold truncating edge (melt entrapment is dependent on the thickness of the cold transform edge).

across this boundary is the same. The weld, however, that bonds the newly formed lithosphere with cold lithosphere across the transform portion of the fracture zone creates a shear couple in the underlying lithosphere resulting in the progressive reorientation of the maximum tensile stress from normal to the ridge axis, at some distance from the ridge transform intersection, to an oblique angle near the boundary (Fig. 8; Gallo et al., 1980; Fox and Gallo, 1982). The brittle carapace of oceanic crust that is created proximal to the R–T intersection will not be an effective stress guide because of its disrupted and fragmented nature but, in response to strain in the underlying mantle weld, the basaltic carapace will deform accordingly with the development of oblique, dip-slip faults. The shear couple will be an intermittent phenomenon in that the mantle weld will be continually broken as the strain induced by transform related strike-slip motion exceeds the strength of the upper mantle rocks. For some period of time after the rupture of the mantle weld the orientation of the maximum tensile stress will be normal to the ridge axis. This temporal variation in the behavior of the mantle weld explains why, proximal to ridge–transform intersections, we observe dip-slip faults and fissures that exhibit a range of orientations (Fig. 2c; Tamayo Tectonic Team, 1983; OTTER, 1983).

The distance over which the transform-generated shear couple affects the tectonics of the accreting plate boundary will depend on the size of the mantle weld which in turn will vary as a function of the thickness of the truncating edge of the lithosphere at the transform boundary (Fig. 9). The number of well-mapped intersections is small but the distance over which oblique structures are recognized along the ridge axis seems to increase as the truncating edge of lithosphere becomes thicker: in the FAMOUS area or at the Tamayo Transform the thickness of the lithosphere at the ridge–transform intersection is only 10–15 km and oblique bathymetric and structural trends along the transform side of the rift valley are recognized over distances of only a few kilometers (Transform A—Detrick et al., 1973; Renard et al., 1975; Tamayo Transform—Tamayo Tectonic Team, 1983) whereas oblique structures develop over much greater distances (20–30 km) along rift wall segments terminated by a thick edge of lithosphere (> 30 km) like the Oceanographer (Schroeder, 1977; Fox et al., 1984) or Vema (K. Macdonald and P. Fox, unpublished Deep-Tow data) Transforms.

Another predicted manifestation of the development of anomalously thick lithosphere proximal to ridge–transform intersections is that at shallow levels these thick edges of lithosphere facing each other across the transform fault will tend to confine the deformation associated with the relative motion to a narrow, highly-strained principal transform displacement zone. Direct observations along Transform A (ARCYANA, 1975; Choukroune et al., 1978) and the Oceanographer Transform (OTTER, 1984), and a Deep-Tow study of a 60 km-long segment of the Vema Transform (Castillo et al., 1982) document that surficial evidence for recent deformation within the transform domain is located along a very narrow (< 2 km) linear belt that is centered about the axis of maximum basement depth. The Vema

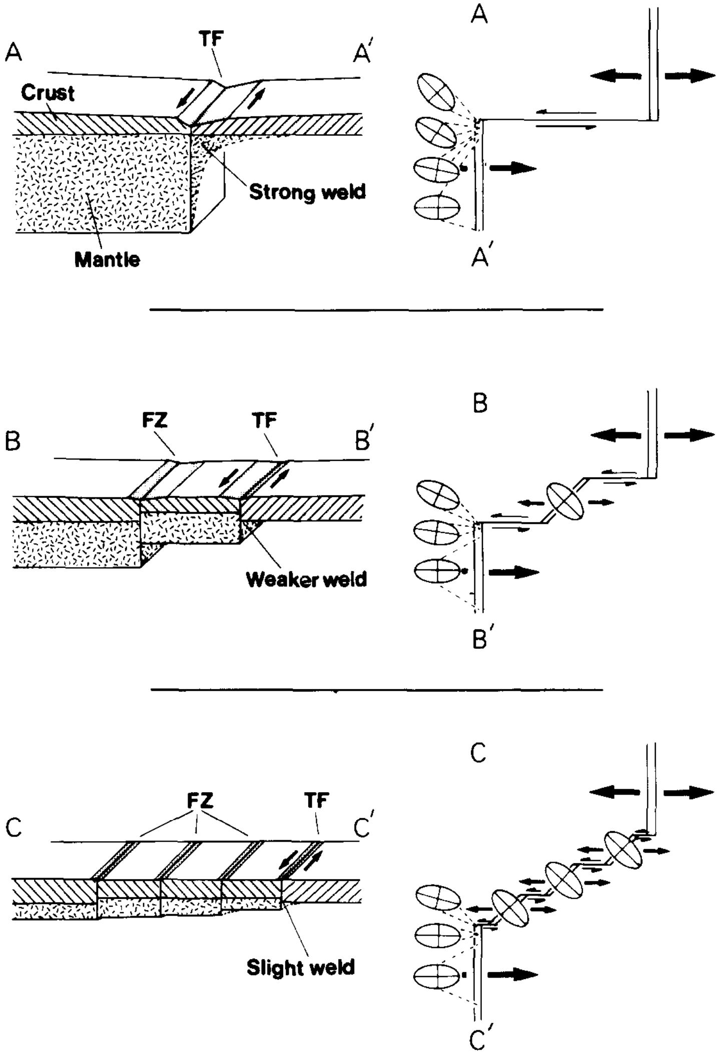


Fig. 9. Schematic models of the lithosphere at three ridge–transform intersections illustrating how contrasting thicknesses in these transform edges effect the tectonics of slowly-slipping (2 cm/yr; *A*), moderate slip-rate (6 cm/yr; *B*) and fast slip-rate (15 cm/yr; *C*) R–T–R plate boundaries (same total offset length for all three examples; ~ 100 km).

Transform is filled with over two kilometers of a flat bedded, laterally continuous, acoustically reverberant assemblage of Pliocene to Recent turbidites (Eittreim and Ewing, 1975; Supko and Perch-Nielsen, 1977). Seismic reflection profiles across the transform valley indicate that the acoustic stratigraphy is undisturbed across its width except for a narrow (< 2 km) zone that cuts vertically through the assemblage intersecting the basement along the axis of the transform valley. At the surface this discordant zone is defined by disrupted reflectors, and is coincident with the 60 km-long disturbed zone mapped by Deep-Tow and can be traced as a continuous feature linking the two ridge–transform intersections (Eittreim and Ewing, 1975; D. Needham, unpublished data). We interpret these data to indicate that the strains associated with transform tectonism along the Vema Transform have remained fixed for as long (a few million years) as the turbidites have been deposited along the valley. Submersible traverses up the walls of the Oceanographer Transform (OTTER, 1984) find no evidence to suggest that strike-slip faulting has been important in shaping this terrain. Although more work must be done to establish the location and behavior of principal transform displacement zones in time and space, the fragmentary record is consistent with the notion that, except for major changes in the pole of relative motion, the thick edges of lithosphere associated with large-offset transform faults confine the PTDZ to a narrow zone centered along the plate edges.

The geologic model outlined in the preceding paragraphs is based largely on our understanding of slowly-slipping ridge–transform–ridge plate boundaries. The geologic constraints on the properties of medium and fast slip-rate transforms and the adjoining ridge segments are not as numerous thereby making it difficult to generalize with confidence, but a few first-order morphotectonic themes emerge and these can be tested against predictions of our model. As the slip-rate of a transform increases, the thickness of a truncating edge of lithosphere will decrease rapidly per unit length of transform when compared with a more slowly-slipping transform (compare Fig. 2b with Figs. 4b, 4e). Therefore, except for exceptionally long offset transforms, ridge segments displaced by medium and fast slip-rate transforms will be bounded at the transform intersection by relatively thin lithosphere (Figs. 4b, 4e). As a consequence, the edge effect of the cold wall on the thermal structure of the adjoining asthenosphere–lithosphere will not be as radical; the migration of basaltic melts and fractionated products to shallow levels will not be substantially inhibited (Fig. 7); the wedge of newly-formed lithosphere plated against the cold edge will not be large; and the significance of the shear-couple formed by intermittently welding the plated lithosphere to the cold edge will be reduced (Fig. 9b). As a result, the distribution of mass proximal to the transform is not as laterally heterogeneous as it is in the lithosphere created adjacent to a slowly-slipping ridge-transform plate boundary and, consequently, the relief of rapidly slipping transforms will become increasingly subdued with increasing rates of accretion. Furthermore, the relatively weak mantle weld predicted to characterize medium slip-rate intersections implies that the shear couple formed in the upper mantle will be restricted to the intersection with oblique structures forming only in close proximity to the transform boundary (Fig. 9).

Given the higher strain rates and the small age contrast in the lithosphere across a medium slip-rate transform it seems intuitively reasonable to assume that in this tectonic environment the location and geometry of the plate boundary would be unstable. Along this “soft” plate boundary the principal transform displacement zone may remain fixed for relatively short periods of time (10^3 to 10^5 yrs. before it migrates to a new location. As a result, the geometry of the strike-slip zone is constantly changing with extensional or compressional relay zones forming to accommodate the strike-slip motion (Fig. 9). An extensional relay zone, depending on how long it remains active, could evolve from an anomalously deep, small rifted basin to a zone of serpentinized upper mantle diapirs and then to a zone of accretion (processes of accretion likely to be abnormal given the tectonic setting); a compressional relay zone would lead to the creation of ridges and shallow crustal flakes by thrust faulting. Indeed, based on various lines of evidence extensional relay zones have been evoked to explain the present-day tectonic settings of the Tamayo (Macdonald et al., 1979; CYAMEX and Pastouret, 1981), Rivera (Reid, 1976; Prothero and Reid, 1982) and Orozco (Trehu, 1982) transforms. Furthermore, if the PTDZ migrates within a broad, 10–30 km wide transform domain, the location of the ridge–transform intersections will not be stable under some conditions and the accretionary ridge tip may propagate into or retreat from the transform domain. It has been suggested that the morphology and magnetic pattern at the Cobb propagator–transform intersection located along the Juan de Fuca–Pacific plate boundary can be best explained by ridge–transform intersections that migrate back and forth (J. Delaney, pers. commun., 1982). The morphology of the Clipperton plate boundary and the present-day location of the truncated rise tips and linking PTDZ suggest that this plate geometry has not been stable and the rise tips have migrated into the transform domain modifying the pre-existing geometry (strike-slip strands linked by a short relay zone; K. Macdonald and P. Fox, unpublished data).

Given the very high slip-rates governing fast-accreting ridge–transform–ridge plate boundaries, the thickness contrast in lithosphere across a transform (i.e. 100 km offset length) will be very small (several kilometers; Fig. 4e) and the strain rates will be very high (Fig. 1). At these fastest-slipping plate boundaries the effect of the truncating edge of lithosphere on accretionary processes proximal to the intersection will be minimized and, as a consequence, the changes in accretionary processes (i.e. effect on partial melting, structure of the upper mantle and strength/size of the mantle weld) that we predict to be a product of this boundary will be subtle and will be apparent only in close proximity to the intersections (Fig. 7). The transform domain, however, will be wide and the tectonic geometry within the domain will be complex because the strains associated with a fast-slipping ridge–transform–ridge plate boundary will not be constrained by thick edges of lithosphere but will be free to migrate across the edges of thin lithosphere that bound the offset rise axis creating a broad zone of shear (Fig. 7; Rea, 1981). The shear zone appears to be characterized by short, strike-slip fault strands that are linked by oblique trending extensional relay zones (Fig. 9; Searle and Francis, 1982; Searle, 1983; P. Lonsdale, pers. commun., 1982; P. Fox, unpublished data). It is unlikely that the processes of accretion that take place along these short extensional relay zones will lead to the development of normal oceanic crust and upper mantle; rather accretion is likely to be discontinuous with crustal sections thinned tectonically by extension and perhaps intruded by diapirs of serpentinized ultramafic rocks (Fig. 7). It is interesting to note that gabbroic and ultramafic rocks have been recovered from a small, anomalously deep, closed-contour basin (4500 m) within the transform domain of the Garret transform (Bideau et al., 1983). We suggest that these rocks have been recovered from terrain that is the product of discontinuous accretion and that has been thinned by extension. Since there are no accurate high-resolution maps at a regional scale of fast-slipping fracture zones, we do not have an accurate description of the time-integrated morphotectonic fabric and, therefore, we cannot use the topography to infer how the strike-slip, relay-zone geometry has behaved in time. We suspect, however, that the geometry will be temporally unstable in this tectonic environment characterized by high strain-rates and thin lithosphere and the time integrated product of this tectonic regime will create a swath of disrupted and anomalous oceanic lithosphere with variable properties and internal characteristics (Fig. 7).

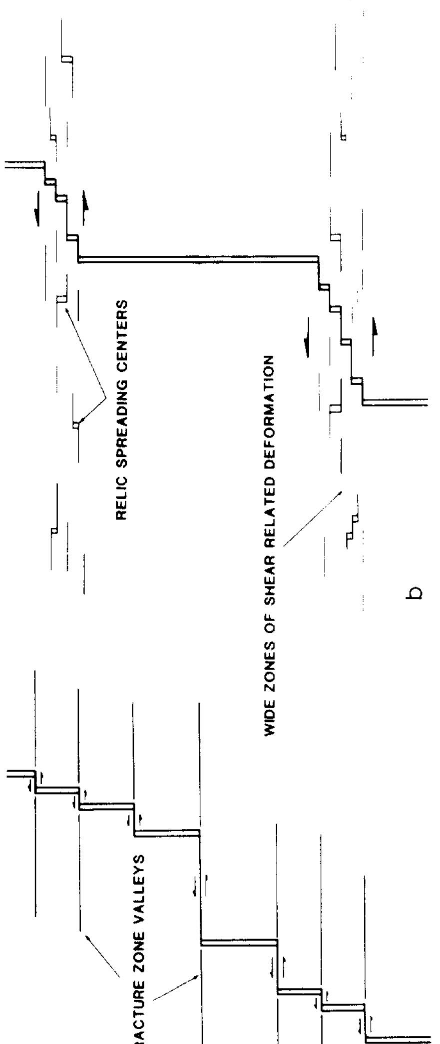
DISCUSSION

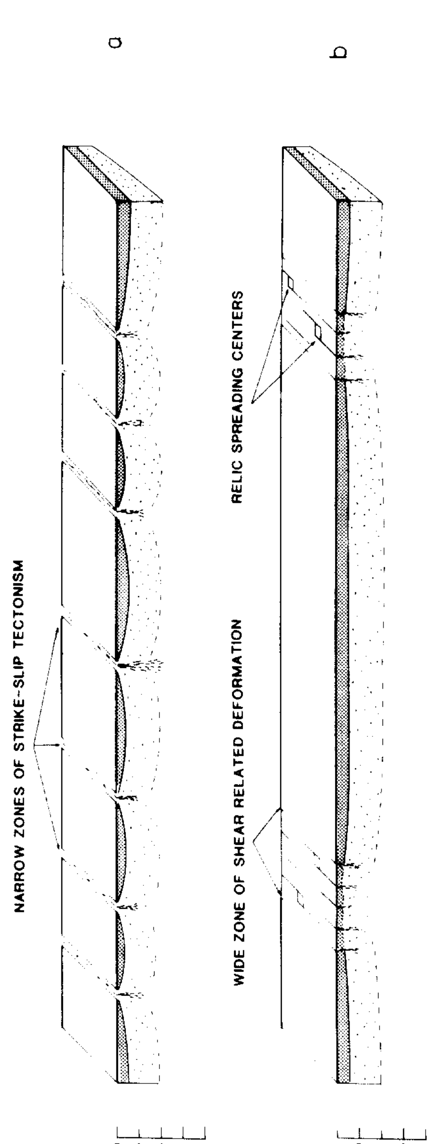
In the preceding section we used the first-order geologic and morphologic relationships at, along and proximal to ridge–transform–ridge plate boundaries to construct an empirical and speculative tectonic model. We suggest that the distinctive but variable morphotectonic fabric and crustal structure of the transform domain are the product of a tectonic continuum that ranges from very slow rates of strain associated with very thick edges of lithosphere (slowly-slipping ridge–transform–ridge plate boundaries) to extremely high rates of strain associated with very thin edges of lithosphere (fast-slipping ridge–transform–ridge plate boundaries). The central theme of the model is that the cold edge of lithosphere juxtaposed against the truncated end of an accreting plate boundary perturbs the processes of accretion leading to the production of anomalous lithosphere.

This model implies that the cold edge of a ridge–transform plate boundary should condition the petrology of basalts created proximal to these boundaries. In order to determine the petrologic consequences, a suite of glassy basalts was collected at sixteen localities over a distance of 50 km down the axis of the East Pacific Rise south of the Tamayo Transform (Tamayo Tectonic Team, 1983). Twenty-nine samples of basalt glass were analyzed by Bender et al. (1984) for major and trace elements including the rare-earth suite and they found that there are systematic differences between samples collected in close proximity to the transform and those samples collected at distances greater than 20 km from the transform. Detailed modeling of these data by Bender et al. (1984) suggests that this geochemical signal can be explained both by lower extents of melting and more variable, but generally greater, extents of crystal fractionation near the transform because of the cooler thermal environment. They propose that the residual mantle beneath a transform will be less refractory and that small scale heterogeneities in the mantle will be more frequently observed near transform faults because of autometasomatism, smaller degrees of partial melting and possibly greater extents of crystal fractionation. These same investigators are presently modeling geochemical data for basalt collections located proximal to other ridge–transform intersections and these results suggest that basalts close to a transform are generally derived by lower extents of partial melting compared to basalts far from the transform along the same ridge segment (Langmuir and Bender, 1984).

In terms of overall structure and composition, the geologic character of oceanic lithosphere is largely conditioned by the processes that govern the generation and emplacement of the predominantly mafic and ultramafic rocks along the Mid-Oceanic Ridge system in both time and space. Along any given ridge axis segment there will certainly be important, but as yet unresolved, temporal and spatial changes in the processes of plate accretion that will lead to non-trivial changes in oceanic lithosphere. These variations, however, are of second-order significance when compared to the changes in the structure and composition of oceanic lithosphere predicted to occur as the fracture zone is approached. Our model suggests, moreover, that there will be some important differences between the geology of lithosphere created along a slowly accreting plate boundary versus lithosphere created along rapidly accreting plate boundary.

Slowly accreting plate boundaries (< 4 cm/yr) are typically characterized by the frequent occurrence of variably-sized transform faults that offset the ridge axis every 30–80 km (see GEBCO map series for North and South Atlantic). Although the relative length of the transform may change with time due to phases of asymmetric spreading, the ridge–transform geometry is remarkably stable in time surviving numerous changes of the pole of relative motion (Schouten and White, 1980; Schouten and Klitgord, 1982). Given the validity of our model and the temporal stability of slowly-slipping ridge–transform–ridge plate boundaries, it means that ribbons of anomalous lithosphere can be traced, rib-like, from the ridge axis across the ocean basin to the margins (Fig. 10a). Each ribbon of anomalous lithosphere will be composed, in first-order terms, of two parts. There will be a relatively narrow band of highly-strained terrain, only a few kilometers wide, made-up of blocks of thin oceanic crust and partially serpentinized ultramafic rock (for example, see Karson and Dewey, 1978; Prinzhoffer and Nicolas, 1980; Karson, 1982; and Karson, 1983 for a discussion, based on ophiolite studies, of the complexities thought to characterize this interval) representing the paleo-principal transform displacement zone. At shallow levels (zone of brittle deformation) this high-strain zone is generally narrow because the strike-slip tectonism of the transform is constrained to a relatively narrow crush zone by the relatively thick edges of opposing lithosphere that face each other across the transform fault. It is interesting to note, however, that for small-offset (< 20 km) slowly-slipping ridge–transport–ridge plate boundaries, the thickness of the lithosphere may be thin enough (< 15 km) to permit the development of complex plate-boundary geometries in response to


a
WELL-DEFINED FRACTURE ZONE VALLEYS
b
WIDE ZONES OF SHEAR RELATED DEFORMATION
RELIC SPREADING CENTERS


NARROW ZONES OF STRIKE-SLIP TECTONISM
a
(km) 0
10
20
30
40
WIDE ZONE OF SHEAR RELATED DEFORMATION
RELIC SPREADING CENTERS
b
(km) 0
10
20
30
40

minor adjustments in relative motion. The maps of the FAMOUS area (Phillips and Fleming, 1978) reveal a confused ridge axis-transform geometry showing overlapping ridge axes and discontinuous transform trends which have been interpreted to reflect an unstable geometry involving the migration of transform fault zones with small offsets (< 20 km) and the change in length of intervening echelon ridge segments (Ramberg et al., 1977; Schouten et al., 1980). Independent of plate edge thickness considerations, times of major changes in the pole of relative motion are likely to induce reorganization of the ridge–transform geometry and, depending on the tectonic adjustments, the zone of strike-slip tectonism will widen appreciably (Menard and Atwater, 1968, 1969; Dewey, 1975). Flanking the paleo-principal transform displacement zone will be a wide (10–30 km) band of anomalous oceanic lithosphere that will become increasingly more heterogeneous with proximity to the fracture zone (Figs. 7, 10a). The width of these ribbons of anomalous lithosphere will depend on how cold the accretionary environment was at the time of lithosphere formation which in turn will be primarily a function of the thickness of the truncating edge of lithosphere at the ridge–transform intersection. The frequent occurrence of ridge–transform intersections along slowly accreting plate boundaries (< 4 cm/yr full rate), the temporal stability of this plate tectonic geometry, and the predicted manifestations of the cold-edge effect at ridge–transform intersections (Fig. 7) mean that the time integrated product of accretion will lead to a very heterogeneous oceanic lithosphere (Fig. 10a). It is no surprise that at six localities within the North Atlantic, chosen on the basis of standard marine geological and geophysical data to be removed from fracture zones, the Deep Sea Drilling Project has recovered gabbroic and ultramafic rocks at shallow crustal levels (< 200 m; Ryan et al., 1973; Aumento et al., 1977; Melson et al., 1978; Leg 82 Scientific Party, 1982).

At rates of accretion greater than 6 cm/yr the number of transforms that offset the axis of accretion diminish markedly and occur every several hundred kilometers or more along the ridge axis (Mammerickx and Smith, 1978). These transforms are

Fig. 10. Schematic diagrams contrasting the different kinds of oceanic lithosphere that are likely to be created along a slowly accreting plate boundary with many transforms (a) and a rapidly accreting plate boundary with a few widely spaced transforms (b). The numerous, relatively thick edges of lithosphere found at slowly-slipping ridge–transform–ridge intersections disrupts the accretion process leading to the creation of heterogeneous lithosphere at a regional scale. The relatively thick edges of lithosphere, however, constrain the strike-slip tectonism to a relatively narrow interval within the heterogeneous swaths of oceanic lithosphere. Along rapidly accreting plate boundaries (> 6 cm/yr) large offset transforms are widely spaced. The accretionary processes are not radically perturbed at ridge–transform–ridge intersections because the contrast in the thickness of the lithosphere at the intersection is not large. Therefore, the changes in the properties of the oceanic lithosphere are not profound as the boundary is approached. The thin edges of lithosphere along fast-slipping transforms allow wide shear zones with unstable geometries to develop creating swaths of disrupted lithosphere (see text for details).

generally long-lived surviving changes in the pole of relative motion (Mammerickx and Klitgord, 1982; Klitgord and Mammerickx, 1982) and have offset magnitudes of approximately 100 km or more. The extensions of these relatively fast-slipping transforms can be traced across the Pacific Basin, but the swath width of fracture zone terrain is variable along strike (Mammerickx and Smith, 1978) suggesting that the ridge–transform geometry is temporally unstable. There is, however, little or no evidence for long-lived, short-offset transforms (< 50 km) and, given the very thin edges of lithosphere involved at a fast-slipping short-offset transform, this geometry is not likely to be stable for long periods of time because the short transforms will be abandoned as the offset accreting plate boundaries propagate across the shear zone into relatively thin lithosphere and link up (Macdonald and Fox, 1983). In time one limb of the overlapping spreading center will be isolated, the transform plate boundary will be abandoned and the new accreting geometry will describe a gentle sinuous bend.

The oceanic lithosphere within and proximal to the widely-spaced but long-lived fracture zones that offset fast accreting plate boundaries will exhibit geological relationships that distinguish this terrain from its tectonic counterpart found along slowly accreting plate boundaries (Fig. 10b). The more robust thermal regime that characterizes a fast accreting plate boundary (Sleep, 1975; Kusznir and Bott, 1976) coupled with the relatively small contrast in the thickness of lithosphere across a transform boundary means that the cold edge effect will diminish with increasing spreading rate and the lithosphere will not exhibit profound heterogeneity as the transform is approached (Fig. 7). The strain history of the rock bodies within the principal transform displacement zone will be more temporally and spatially complex than a slowly-slipping transform because the high rates of strain that characterize fast-slipping transforms (Fig. 1) will not be as constrained by the relatively thin edges of opposing lithosphere and, in response to relatively small changes in the pole of relatively motion, the shear zone along a transform will be free to migrate over a wide interval (10–20 km; Figs. 9, 10b). Depending on the sense of transform offset relative to a given alteration in relative motion, a transform would experience a component of extension or compression (Menard and Atwater, 1968, 1969; Dewey, 1975). A component of extension would facilitate the development of a wide shear zone with several strike-slip strands linked by extensional relay zones whereas a component of compression would tend to collapse the shear zone. As discussed in the preceding section, all the transforms south of the Gulf of California that offset the rise axis in a left-lateral sense (Tamayo, Rivera, Orozco, Siqueiros, Quebrada, Gofar, Wilkes and Garret) are characterized by wide shear zones with short strike-slip strands linked by extensional relay zones. The one transform that offsets the rise in a right-lateral sense, the Clipperton Transform, is presently characterized by a narrow zone of shear. Whether this distinction is fortuitous or indicative of very recent changes in the poles of relative motion (Gallo and Fox, 1981; Searle, 1983) remains to be established but it seems clear that the thin edges of fast-slipping transforms will accommodate adjustments to small changes in plate kinematics relatively easily. Furthermore, the geometry within these fast-slipping shear zones is likely to be continually adjusting and changing and the time-integrated product of this evolving tectonic behavior will be the creation of oceanic lithosphere that is complex in terms of composition and structure. Accretionary processes within a 10–20 km wide shear zone will be disrupted by the unstable behavior of the shear zone; strike-slip faults will bound rock bodies of contrasting age; extensional relay zones can lead to tectonic thinning, hydration and uplift of the crust, or the development of a small zone of accretion. On the other hand compressional relay zones will create ridges through crustal thickening and under thrusting (Figs. 7, 10b).

The variation in the thickness of the truncating edge of lithosphere at a ridge–transform intersection will not only control the thickness of the oceanic crust but will also condition the structure of the oceanic crust and upper mantle. Bender and Langmuir have recently argued, based on geochemical data, that at ridge–transform intersections not only are smaller volumes of melt produced but those melts experience greater degrees of fractionation in shallow level reservoirs proximal to the transform termination (Bender et al., 1984; Langmuir et al., 1984). A similar conclusion was reached by Fornari et al. (1983) after collecting highly fractionated basalts from terrain proximal to the Inca Transform along the Galapagos Ridge. Enhanced fractionation suggests the accumulation of thicker sequences of cumulate ultramafics and/or olivine rich gabbros. Based on field mapping in the Bay of Islands, Newfoundland, Karson has defined the relationships that are developed along the contact between the Coastal Complex and the Bay of Island ophiolite (Karson and Dewey, 1978; Karson, 1982). Karson suggests that the relationships developed along this contact are the product of a relatively fast-slipping ridge–transform intersection and shows that as the contact is approached ultramafic cumulates (cumulate dunites and megalenses of wehrlite and gabbro) thicken and lap up against the highly strained rocks of the transform boundary. This relatively shallow level expression of the cold transform edge would be perhaps most dramatic at faster slipping ridge–transform intersections where the volume of partial melt remains relatively high and magmatic processes are relatively continuous. The development of a thicker basal cumulate member should develop at a slowly slipping ridge–transform intersection but this assemblage will not be as well developed because of the smaller volume of melt produced and the very discontinuous nature of accretion in this environment.

Our discussion has largely focused on the structural differences predicted to exist between ridge–transform intersections with contrasting tectonic characteristics (i.e. slip-rate and truncating edge thickness). Superimposed against these tectonic variables are temporal variations in accretionary processes that can create pronounced changes in crustal structure. Recently it has become clear that the migration of basaltic melts into shallow level reservoirs and the extrusion of these products onto the sea floor is an episodic process with magmatic pulses migrating in time along a ridge axis segment (e.g., Macdonald and Fox, 1983). This temporal variability of ridge axis accretion has important implications for crustal structure in general and, in terms of this discussion, ridge–transform intersections in particular. As the magmatic pulses near a given ridge–transform intersection wax and wane the thickness of the magmatic component and the constitution of the crust and the upper mantle will vary in complex ways. Consequently, when the the crustal structure of a fracture zone is viewed along strike the magmatic component should thicken and thin creating a lumpy profile.

CONCLUDING REMARKS

The preceding discussion has summarized the morphologic, geologic and geophysical characteristics of ridge–transform–ridge plate boundaries. Based on these data we suggest that the distinctive morphotectonic character exhibited by these plate boundaries is the surficial expression of what happens when a cold edge of lithosphere, of variable thickness, is juxtaposed against an accreting plate boundary (full spreading rate can vary from 2 to 18 cm/yr). The model that we propose (Figs. 7, 8, 9, 10) is speculative and must be tested against the acquisition of high resolution data that can be obtained by carefully designed experiments that involve multi-narrow beam echo sounders, deep-towed geophysical and photographic packages, submersibles and geophysical equipment (e.g. ocean bottom seismometers, multi-channel seismic systems). It is critical that maps of the principal transform displacement zones of both fast-slipping and slowly-slipping transforms be made so that we can obtain a better description of how these contrasting tectonic environments behave temporally and spatially. More information is needed on structural relationships developed at ridge-transform intersections as well as a better definition of the change in properties of the oceanic lithosphere that occur proximal to ridge–transform–ridge plate boundaries.

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