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J. M. Lorenzo · P. Wessel

Flexure across a continent–ocean fracture zone: the northern Falkland/Malvinas Plateau, South Atlantic

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Abstract Bathymetry, satellite-derived gravity, and interpreted seismic reflection data across the northern Falkland/Malvinas Plateau fossil continent–ocean transform rim may record the degree of mechanical coupling across the boundary after ridge–transform intersection time. The rim comprises a broad microcontinental block in the east and a continental marginal fracture ridge 50–100 km wide elsewhere. Free-air gravity anomalies tentatively suggest that the fracture ridge is locked against oceanic elastic lithosphere both to the north (Argentine Basin) and south (Central Falkland Basin).

Introduction

Continent–ocean fracture zones (COFZs) constitute the fossil transform offsets of passive margins. Kinematic models of COFZs distinguish at least two principal stages during their evolution (Scrutton 1979; Mascle and Blarez 1987). In the first stage, continent–continent shearing is the dominant process controlling the tectonic development of a long, narrow region in which the transform fault will eventually rupture. During the second stage, as seafloor spreading proceeds, the younger oceanic block slides along the active transform, heating the older continental side, possibly inducing thermal uplift and accompanying denudation. Early thermal–rheological models assumed that the oceanic and continental sides had no mechanical strength and either remained uncoupled mechanically at all times [Southwest Newfoundland Transform Margin (Todd and Keen 1989)], or weakly coupled during only the active transform stage [Southwest Newfoundland Transform Margin (Reid 1989), Senja Fracture Zone (Vågnes, this issue, or southern Exmouth Plateau (Lorenzo and Vera 1992)]. The potential influence of coupled flexure on the stratigraphy and structure of these margins has only been proposed recently (Gadd and Scrutton, this issue) and remains undocumented.

J. M. Lorenzo (✉)
Department of Geology and Geophysics,
Louisiana State University, Baton Rouge, LA 70803-4101, USA

P. Wessel
School of Ocean and Earth Science and Technology,
University of Hawaii at Manoa, Honolulu, HI 96822, USA

Continental margin offsets gained early recognition during the development of plate tectonic theory due to their ability to constrain the geometry and movement between drifting continents and, to some extent, the origin of fracture zone and oceanic transforms (Wilson 1965; Le Pichon and Hayes 1971; Le Pichon and Fox 1971; Francheteau and Le Pichon 1972). Herein, the term fracture zone is defined as the inactive trace of transform faults (Wilson 1965) that extends beyond the junction of seafloor spreading centers and transforms, known as the ridge–transform intersection (RTI) (Kastens 1987) (Fig. 1). Although variable in nature, COFZs are often marked by high-standing continental marginal ridges (50–100 km wide) bounding deep sedimentary basins (Scrutton 1979). Analogous high-standing ridges have been identified in modern extensional settings at zones of accommodation between large half-graben rift basins systems (e.g., Rosendahl 1987). In addition, the history of vertical motion at COFZs throughout the continent–continent shearing stage and into the ocean–continent stage (Scrutton 1979; Mascle and Blarez 1987) has been explained through transpressional tectonic activity (Mascle and Blarez 1987) and permanent isostatic uplift caused by crustal density changes at depth through underplating directed from the adjacent oceanic ridge (Lorenzo et al. 1991).

Thermomechanical models that assume full mechanical coupling across fracture zones were first developed for oceanic studies. After the RTI passes (Sandwell and Schubert 1982), the transform fault is thought to become locked as the blocks on either side become coupled. As the blocks on either side of the fracture zone are of different

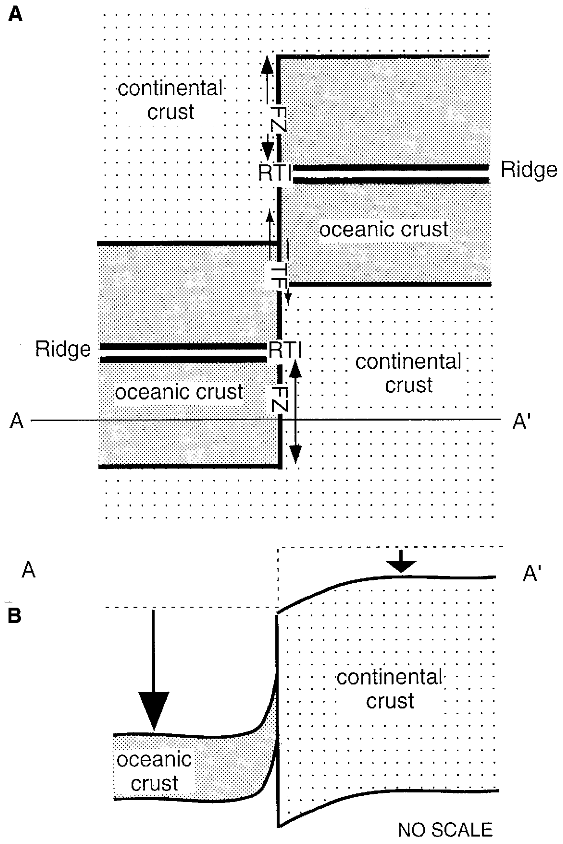

Fig. 1 **A:** Schematic plan view of continent–ocean active and passive transform boundaries. RTI: ridge–transform intersection, FZ: fracture zone, TF: transform fault. **B:** Profile through the fracture zone, landward of the RTI shows effect of assuming a mechanical model where lithosphere on either side of fracture zone subsides at different rates but remains mechanically coupled. Only crustal blocks used for gravity modeling are shown. Total amount of subsidence implied by the length of the arrows is greater on the oceanic side. A dashed line marks the initial top of the crust at RTI time. Predicted bathymetry has a thermally younger (oceanic) side that bows up whereas the thermally older (continental) side bows down

thermal age, they cool, become denser, and subside isostatically at different rates. Forces and moments set up by differential subsidence across the fracture zone (Wessel and Haxby 1990) lead to a bathymetry where the younger side bows up several hundred meters over a distance of about 50 km approaching the fracture zone, whereas the older side bows down (Fig. 1). At continent–ocean fracture zones the continental block is considered to be the thermally older block. In some instances the fracture zone may continue to slip vertically for a few million years after passing the RTI (Wessel and Haxby 1990; Bergman and Solomon 1992; Christeson and McNutt 1992) if it is mechanically weakened by intraplate volcanism (Bonneville and McNutt 1992; Lowrie et al. 1986).

Bathymetry, shipboard free-air gravity, and satellite-derived deflection of the vertical data have been employed (Bonneville and McNutt 1992; Christenson and McNutt 1992) to examine the degree of coupling across oceanic fracture zones after RTI time. In this paper we examine existent shipboard bathymetric and multichannel seismic reflection data across the northern Falkland Plateau escarpment, a continent–ocean fracture zone in the South Atlantic (Fig. 2). We focus on two very different N–S gravity profiles and adjacent seismic data, at 42°W and 52.5°W, at right angles to the fracture zone.

Mechanical coupling of the elastic lithosphere across the Falkland Plateau escarpment should bend the sea floor down on the continental side and raise it on the oceanic side, relative to the overall thermal subsidence. At 42°W a thermomechanical model is used to explain the bathymetry and free-air gravity field. The Moho, a surface within the elastic lithosphere, is expected to deform almost parallel to the bathymetry and will not be at its expected local isostatic depth. Reliable deep seismic information on the shape of the Moho is unavailable, but models of Moho depths can be tested by forward modeling the free-air gravity field using available bathymetry and limited crustal structural information. In this manner at 52.5°W a modeled Moho shape is used to determine the degree of mechanical coupling across the COFZ.

Net vertical shear stresses can be estimated from the isostatic gravity anomaly profiles across continent–ocean fracture zones to determine whether vertical slip has occurred across a fracture zone (Bonneville and McNutt 1992). The local isostatic anomaly is the difference between the observed free-air anomaly and the free-air anomaly calculated from bathymetry assuming local isostasy. Departure from local isostasy changes the gravity field because adjacent rock columns of equal thickness no longer have equal masses. The anomalous lateral mass distribution implies a horizontal variation in vertical forces away from the fracture zone; this variation can be one of the causes of isostatic gravity anomaly residuals. The local isostatic anomaly is used to calculate the excess vertical shear stresses at the fracture zone that are then compared to theoretical estimates of vertical shear stress.

Falkland Plateau geology

Seismic refraction experiments (Ewing and Ewing 1959; Ewing et al. 1971), Deep Sea Drilling Project (DSDP) results (Barker et al. 1977; Ludwig 1983a), and reflection seismic analysis (Lorenzo and Mutter 1988) reveal that the Falkland Plateau is a foundered complex of continental and oceanic blocks. Two major tectonic episodes affected the Falkland Plateau: (1) In the Middle Jurassic, early Gondwanide breakup of Antarctica from South America and Africa created the Falkland Basin, currently centered between the continental shelf to the west and the Maurice Ewing Bank, a microcontinental block (Figs. 2 and 3) (Barker et al. 1977). The Falkland Basin contains as much as 7 km of synrift, transitional, and hemipelagic sediments (Fig. 3) and is probably floored in parts by overthickened

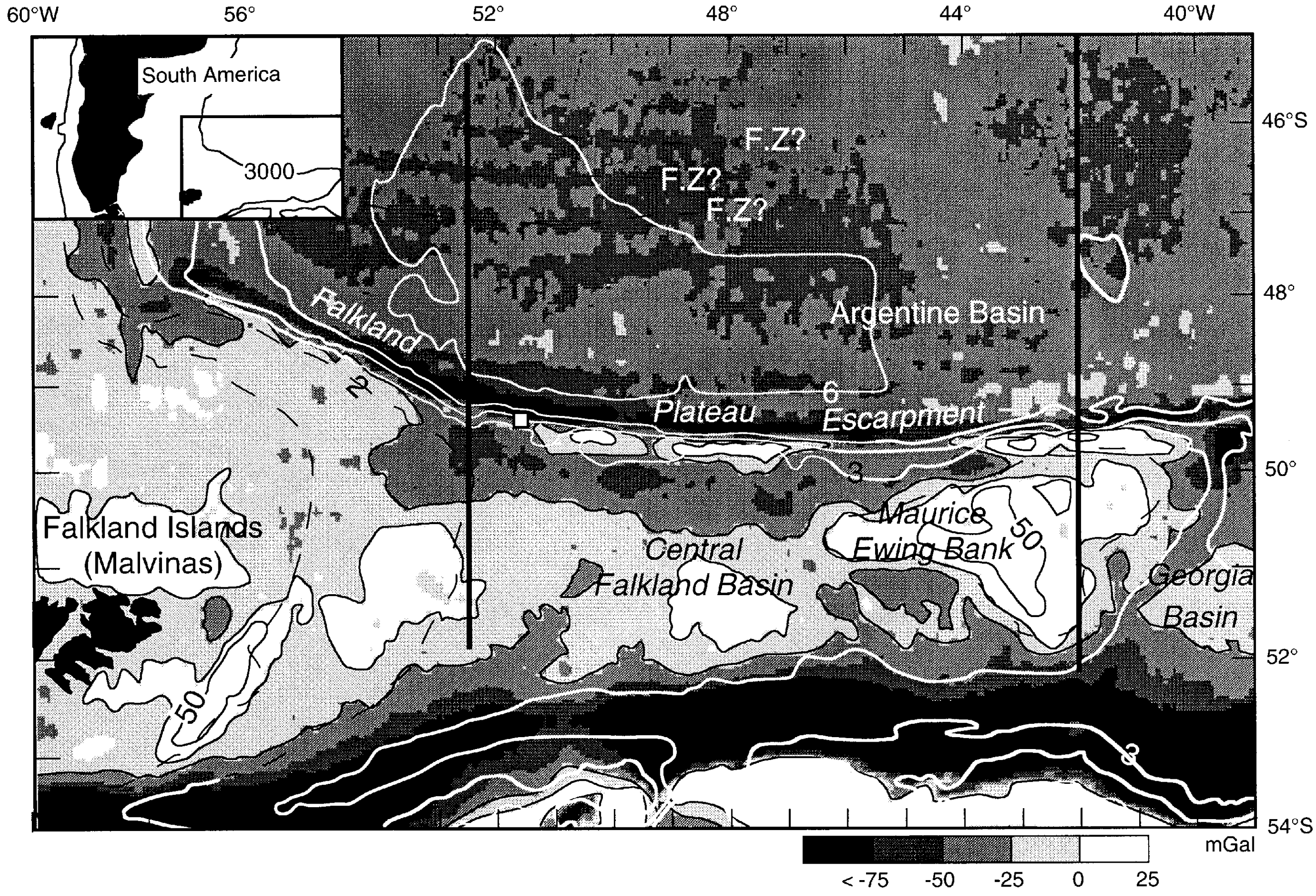

Fig. 2 Satellite-derived free-air gravity anomaly map (Sandwell et al. 1995) over the Falkland Plateau study area; bathymetry contours are superposed. Study area is located west of South America (inset). Two thick black lines locate the extent of north–south sections used for modeling and seen in Fig. 5A–E. A negative gravity anomaly lineament in the Argentine Basin, running along the 6000-m isobath and parallel to the Falkland escarpment corresponds to a linear negative magnetic anomaly modeled by Rabinowitz and LaBrecque (1979) as an oceanic fracture zone (FZ). By extension, parallel lineations further north by also overlie fracture zones (FZ). White square locates dredge haul of basaltic rock from Falkland Plateau (Lorenzo and Mutter 1988)

oceanic crust (Ludwig 1983a; Lorenzo and Mutter 1988). (2) In the Early Cretaceous, final Gondwanide breakup of Africa from South America created a NE–SW-trending passive rifted margin along the eastern side of the Maurice Ewing Bank parallel to the oldest ocean magnetic anomalies M10 and M11 (Rabinowitz and LaBrecque 1979) in the adjacent Georgia Basin. Furthermore, a 1200-km-long east–west-trending continent–ocean transform margin, the focus of this study, was created just north of the Falkland Basin. Transform activity ceased when the RTI cleared the northeastern Maurice Ewing Bank at about M34 ($\sim$83 Ma) (Cande and Kent 1995).

Free-air gravity anomaly patterns

Satellite-derived gravity data coverage (Sandwell et al. 1995) far exceeds currently available shipboard gravity data (Smith and Sandwell 1994). In our study area, satellite track spacing is 2–4 km (Smith 1993), whereas on average less than one ship track per 100 km crosses the Falkland escarpment.

The degree of mechanical coupling across the Falkland escarpment may be indicated by gravity anomaly patterns. A marked free-air anomaly edge effect is observed in the extreme east of the Maurice Ewing Bank. The edge effect is a negative–positive couplet of gravity anomalies, expected from the juxtaposition of oceanic against continental crust in conditions of local isostasy. Whereas a band of negative free-air anomaly gravity values marks the base of the Falkland Plateau fossil transform escarpment (Fig. 2), the positive half of the free-edge couplet is less apparent over the plateau rim. Instead, the northwest (e.g., at 52.5°W) and central plateau escarpment display a broad negative band of anomalies over continental crust punctuated by local highs parallel to the escarpment. Only in the extreme east (e.g., at 42°W) are gravity values positive over the plateau rim. If this difference is attributed wholly to lithospheric mechanical models, the northern

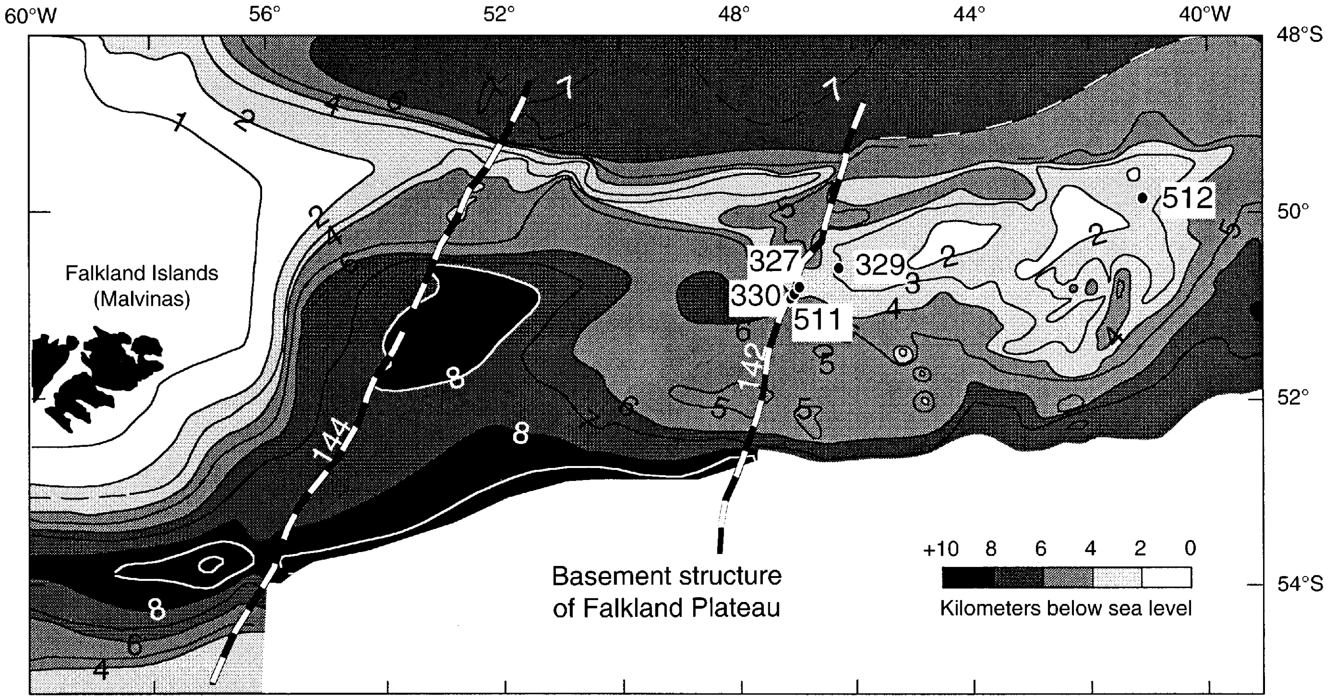

Fig. 3 Depth-to-basement map adapted from Lorenzo and Mutter (1988) and rechecked against the original seismic data along our study profiles. A conical edifice in oceanic crust north of the Maurice Ewing Bank has now been removed. Deep Sea Drilling Project site numbers (filled circles) are shown on the Maurice Ewing Bank. Long thick lines crossing northern Falkland escarpment at an angle locate nearest high-quality seismic lines, interpreted in Fig. 4

marginal ridge may not be in local isostatic equilibrium but may be flexurally depressed into the mantle as the result of coupling to the adjacent denser oceanic lithosphere.

In the models that follow we take the continent–ocean boundary as the base of the Falkland Plateau escarpment. Rabinowitz and LaBrecque (1979) position the continent–ocean boundary generally at the base of the Falkland escarpment, although sometimes a few tens of kilometers further north.

Composition of continental fossil transform basement

The northern escarpment of the Falkland Plateau is a heavily sedimented fossil transform rim across which the possible effects of flexure can be tested by examining the stratigraphic and structural history in seismic profiles. Drilling at DSDP site 330 on the Maurice Ewing Bank (Fig. 3) encountered a granitic continental basement (Barker et al. 1977). By correlating the seismic basement reflector from the DSDP sites to the nearby escarpment and considering that the majority of dredge hauls at the top of the escarpment (Lorenzo and Mutter 1988) have a gneissic composition, the basement is also assumed to have a continental composition. The depth-to-basement map of Lorenzo and Mutter (1988) shows that the fossil transform basement marginal fracture ridge is 50–100 km wide west of the northern Maurice Ewing Bank (Fig. 3).

We note that a basalt sample was dredged from an unsedimented portion of the sea floor at the top of the Falkland Plateau escarpment (Lorenzo and Mutter 1988) (Fig. 2). The sample is apparently not representative of the whole basement composition but could indicate some measure of constructional volcanism and magmatic intrusion at depth (Lorenzo et al. 1991). The age of the basalt sample is unknown, and seismic data are lacking where the basalt was dredged.

Seismic structural style across the continental fossil transform segment

Two line interpretations of seismic cross sections across the northern Falkland escarpment (Figs. 4A and 4B) display two different stratigraphic and tectonic styles, perhaps related to vertical movements across the COFZ. Major faulting along the continental segment of the COFZ ends below unconformity U2 (Late Jurassic–Neocomian), which marks the end of rifting and onset of Early Cretaceous sea-floor spreading in both the Argentine and Georgia basins. This and other regionally significant unconformities were correlated to biostratigraphic ages at DSDP sites 327, 329, 330, 511, and 512 (Barker et al. 1977) on the Maurice Ewing Bank.

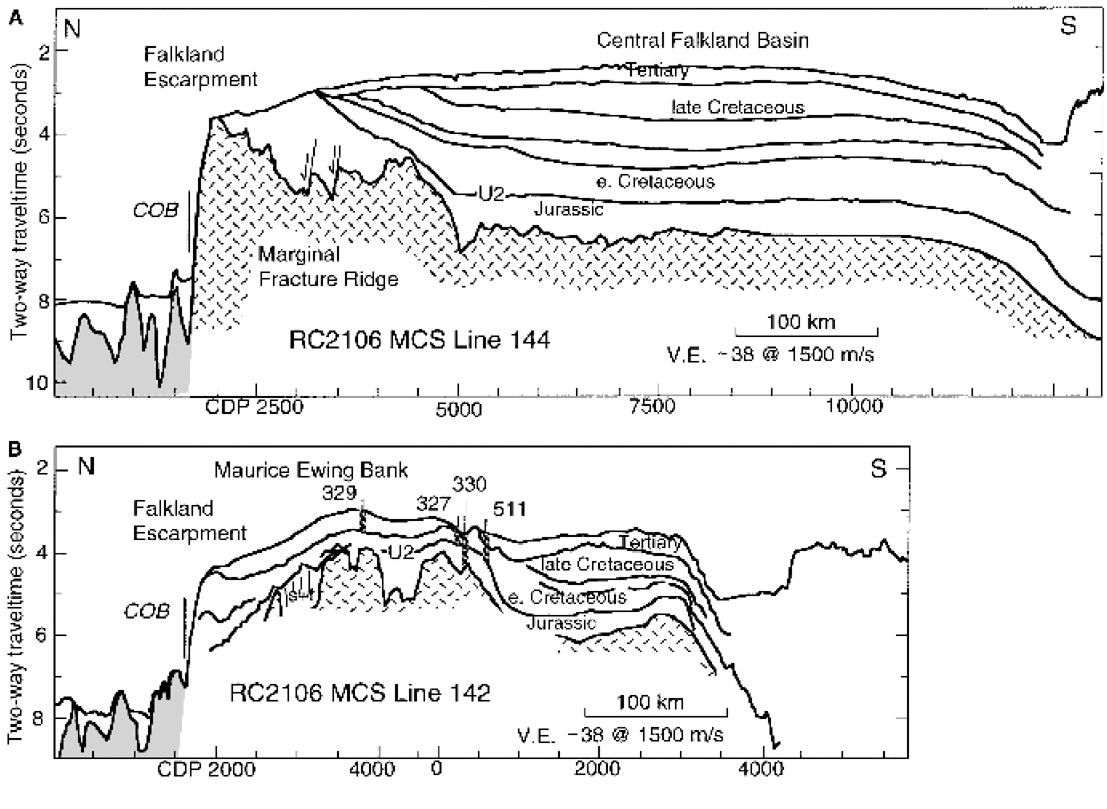

Fig. 4 A: Line 144 – line interpretation of multichannel seismic profile across central Falkland Plateau. Arrows indicate fault movement. Stratigraphic lines represent major depositional sequence boundaries; U2, regional unconformity. COB is the estimated continent–ocean boundary, shown as 0 km in Fig. 5A–E. CDP, common depth point. **B:** Line 142 – line interpretation of multichannel seismic profile across Maurice Ewing Bank; s+f marks area of transpressional strike-slip faulting and folding (Lorenzo and Mutter 1988); U2, regional unconformity. DSDP locations have been projected obliquely onto line 142 and are indicated by shaded vertical rectangles. COB is the estimated continent–ocean boundary, shown as 0 km in Figs. 5A–E. CDP, common depth point through Maurice Ewing Bank. Line interpretation of multichannel seismic profiles are adapted from Lorenzo and Mutter (1988)

Along line 144 (Figs. 3 and 4A), deep north-dipping normal faults below U2 imply a component of extension across the marginal fracture ridge probably related to the first rifting stage in the Middle Jurassic. However, along the eastern profile (Figs. 3 and 4B, line 142) there is more evidence for strike-slip faulting and transpressional folding (Lorenzo and Mutter 1988) more typical of transform margin development (e.g., Mascle and Blarez 1987; Mascle et al. 1987). Along line 144 (in the west) sedimentary reflectors dip south into the Central Falkland Basin whereas along line 142 (across the northwest Maurice Ewing Bank) they dip north toward the Argentine Basin (Ludwig 1983b). Along both these lines only a small portion of oceanic crust is shown, but the top of the oceanic basement appears to rise slightly on approaching the fracture zone.

Modeling gravity, bathymetry, and vertical shear stress

Simple crustal density models at right angles to the Falkland escarpment and near existent multichannel seismic lines (Fig. 4A at 42°W, and Fig. 4B at 52.5°W) are derived by two different means. First, we employ a differential thermal subsidence model with mechanical coupling across the COFZ (Sandwell and Schubert 1982) to estimate topography and best match the observed free-air gravity field (Figs. 5A and B). In order to account for rifted continental lithosphere, we assign a thermal age for the continental block at the time of RTI. The effective elastic thickness of the lithosphere is calculated from the depth to the 450°C isotherm. Final modeled flexed bathymetric profiles are the summed result of incremental flexure calculated in 20 time steps. This method produced better results at 42°W than at 52.5°W.

Second, we start with an isostatistically (only local) balanced crustal section using sea-floor depths and proceed by adjusting the Moho so as to best match the free-air gravity data (Fig. 5C and D). This approach does not assume mechanical coupling. No local detailed seismic velocity estimates are available from which to derive empirical density values for modeling gravity. Densities for average continental and oceanic crust are similar to those in modeling analogous tectonic settings (Lorenzo and Vera 1992). Over the Falkland Plateau we derive total sediment thicknesses by subtracting known sea floor from basement depths (Lorenzo and Mutter 1988). In the

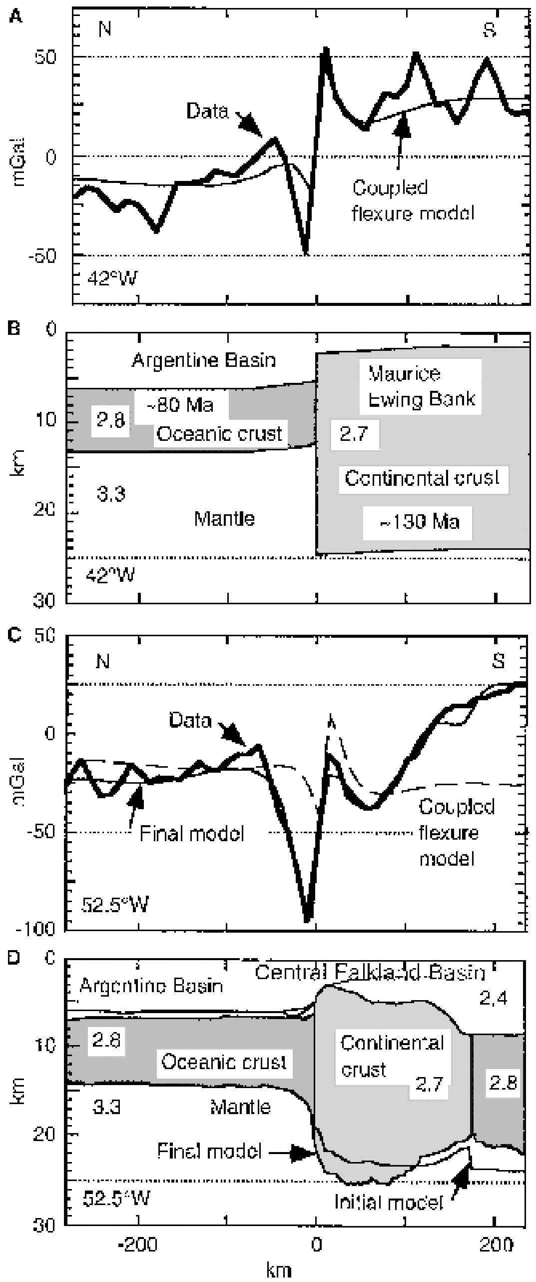

Fig. 5 Free-air gravity profiles (thick lines) across the Falkland Plateau and the gravity response of the best-matching crustal models (thin lines) following Sandwell and Schubert (1982). At RTI time, the top of the continental block is assumed to be at sea level. **A:** At 42°W the coupled flexure model best matches the gravity data. **B:** Thermal ages are in millions of years and densities in grams per cubic centimeter. Initial depth to oceanic ridge axis = 3 km; thickness of oceanic crust = 7 km. **C:** At 52.5°W the coupled flexure match (dashed line) is poorer than at 42°W. Modeling parameters: current thermal age of oceanic side = 114 Ma; current thermal age of continental side = 150 Ma; initial depth to oceanic ridge axis = 3 km; thickness of oceanic crust = 7 km; thickness of continental crust = 24.45 km. **D:** The best match to the observed gravity low over the Falkland Plateau marginal ridge is suggested by a crustal model derived by adjusting Moho depth and oceanic crustal thickness in a local isostatic starting model

Argentine Basin, oceanic basement depths are obtained by subtracting sediment isopach values (Hayes 1991) from known bathymetric values.

We use ETOPO-5 digital bathymetry (NGDC 1988), a 5×5 arcmin gridded data set, that corresponds to about 6×9 km at the latitude of the Falkland escarpment. In general, ETOPO-5 is a readily available data set that should be interpreted with care (Smith 1993) but which nevertheless proved similar to available ship-sounding data (NGDC 1992).

Following Bonneville and McNutt (1992), the average shear stress is computed from the ratio of the net vertical force per unit distance away from the fracture zone, distributed over the height of the vertical transform contact. Because shearing stress is symmetric across the boundary and does not change sign (Hetenyi 1979), we integrate the isostatic gravity anomaly only on the oceanic side, to a distance of 200 km. The isostatic gravity anomaly is used instead of the free-air anomaly, which has been used in oceanic settings (Christeson and McNutt 1992) because across continent–ocean fracture zones there are large variations in crustal thickness and composition. Ideally, these variations are removed in the isostatic anomaly.

Results

Although gravity models do not uniquely define the subcrustal structure, they can provide first-order models for consideration in terms of flexure. At 42°W the observed gravity field can be adequately matched using a fully coupled thermomechanical model (Fig. 5A and B). However, at 52.5°W, the missing free-edge effect is not as well matched using a similar approach (Fig. 5C). At 52.5°W, one possible crustal model suggests that the marginal ridge root is considerably deeper than an initial isostatic model (Fig. 5C and D). The best-matching model also suggests that oceanic crustal thickness is variable. Regionally, the top of the oceanic crust is also seen to rise on approaching the COFZ. Currently, we do not have adequate seismic data with which to test these models or distinguish between complex cases established at other continent–ocean fracture zones on the basis of seismic studies (Todd et al. 1988; Lorenzo et al. 1991; Edwards et al. this issue).

Both at 42°W and 52.5°W, the crustal models derived from modeling gravity profiles imply greater net vertical shear stresses than can be explained by the coupled differential subsidence of lithosphere across the fracture zone. At 42°W the ratio of vertical shear stress calculated from the observed gravity to those calculated from the best flexurally coupled model is greater than about 1.3. At 52.5°W the same ratio is considerably larger (~ 3), affected by the very large negative free-air gravity anomalies observed on the oceanic side of the Falkland Plateau escarpment. The simple assumption of a coupled flexural model provides a conservative, lower estimate for stress across the Falkland Plateau fracture zone.

Discussion

Our flexural models assume a homogeneous oceanic crust of constant thickness and assume only one continent–ocean boundary. On the oceanic side of the Falkland Plateau escarpment at the 52.5°W section, local crustal heterogeneities may be responsible for the low free-air gravity anomalies that are observed. North of the plateau escarpment there is also a marked increase in basement roughness (Fig. 4A and B). Fracture zones are recognized as zones of hydrothermal alteration, and crustal heterogeneity (Detrick et al. 1993) can alter the normal magnetic signature and density structure of oceanic crust.

Over the plateau rim itself one tentative explanation for the missing positive half of the free-edge gravity couplet can be found in the complex geological structure of the Falkland Plateau. Oceanic crust possibly abuts the marginal fracture ridge both to the south (Central Falkland Basin, Middle Jurassic) as well as to the north (Argentine Basin, Cretaceous). If both oceanic blocks are coupled flexurally to the marginal ridge, then both will tend to hold the block down in the mantle without the need for a single large vertical shear stress developed along the Falkland Fracture Zone.

One limitation of previous thermomechanical models for continent–ocean fracture zones is that coupling has not been considered (Todd and Keen 1989), with the result that previous estimates of uplift and ensuing erosion may have been overestimated. Coupling of subsiding oceanic crust with continental crust would decrease the maximum elevation experienced by the continental side during a phase of thermally induced uplift. As a result, the amount and patterns of subaerial erosion when the top of the crust is above sea level are expected to change. Lower land elevation would diminish the amount of denudation and the locus of maximum amount of erosion would be expected to move landward of the continent–ocean boundary. Hence, the amount of flexural coupling across these boundaries could influence erosion patterns. Unfortunately, it remains unclear whether the observed depositional sequences across the Falkland Plateau owe their variation in thickness to nondeposition or erosion. Therefore, the predictions of coupled flexure cannot be conclusively tested. Adequate seismic stratigraphic profiles and biostratigraphic age control are currently lacking.

If flexural coupling exists across sections of the Falkland Plateau Fracture Zone then we may expect similar behavior across other analogous margins. However, it remains to be determined how well thermomechanical models developed for oceanic fracture zones are applicable to continental crust. Continental lithosphere is generally older and more likely to have been affected by rifting and orogenic episodes. Therefore, continental lithosphere has the potential to be more heterogeneous mechanically (e.g., Lowry and Smith 1995). Quartz and olivine flow laws used to predict rock strength distribution for lower continental crust and mantle, respectively (Brace and Kohlstedt 1980; Kohlsdedt et al. 1995), imply that inelastic yielding occurs more readily at lower differential stresses in continents compared to the oceanic lithosphere (e.g., Vink et al. 1984).

Across several other continent–ocean fracture zones [southern Grand Banks (Todd and Keen 1989), southern Exmouth Plateau (Lorenzo and Vera 1992), and the Senja Fracture Zone (Vågnes, this issue)], several kilometers of sedimentary sections apparently have been eroded during transient thermal uplift. The extent of erosion is greatest near the continent–ocean boundary and the oceanic crust does not seem to rise noticeably on approaching the boundary. These characteristics do not require mechanical coupling across the boundary and imply a very weak transform fault. Admittance studies across the fossil transform boundary of the southern Grand Banks conclude that this COFZ is a weak suture that accommodates local compensation.

The varying sedimentary patterns and estimated stresses along the Falkland Plateau may reflect the flexural as well as the rifting history of the region. Southward-dipping reflectors in the western Falkland Plateau margin are most easily explained as the result of greater subsidence of the oceanic crust that probably floors the deep Central Falkland Basin. In contrast the reflectors along the western Maurice Ewing Bank show a northerly dip as might be expected from a flexurally coupled block. However, normal faults near the boundary in the Maurice Ewing Bank (Fig. 4) imply that local crustal thinning may also have led to local tectonic subsidence.

Conclusions

Both the up-warped shape of the oceanic crust on approaching the Falkland Plateau and the best-matching gravity models suggest that mechanical coupling is possible after RTI time between continental and oceanic crust. Future independent testing of these models with seismic studies to determine the crustal velocity structure is needed. Modeling within the scope of this paper is limited to considering only the effects of differential subsidence across the margin and implies that the narrowest portion of the marginal ridge in the western Falkland Plateau is being held down coupled to oceanic crust both to the north in the Argentine Basin and south (Central Falkland Basin).

Flexural coupling predicts that the locus of maximum uplift and potential erosion should be displaced away from the continent–ocean boundary. Previous thermomechanical models along other margins (southern Grand Banks and southern Exmouth Plateau, Senja Fracture Zone) have not addressed coupled flexure explanations of sedimentary patterns. The current study suggests that flexural coupling should be considered to understand the tectonic and sedimentary evolution of other well-known continent–ocean fracture zone margins such as the southern Agulhas Plateau (conjugate margin of the Falkland Plateau) and Ghana margin (Gulf of Guinea) where extensive published deep seismic and gravity studies are concurrently available.

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