Transitions in axial morphology along the Southeast Indian Ridge

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Abstract. Shipboard bathymetric and magnetic profiles across the Southeast Indian Ridge (SEIR) were analyzed in order to examine the nature of along-axis variations in axial morphology at this intermediate spreading rate ridge. Three types of axial morphology are observed along the SEIR: an axial high, a shallow (200–700 m deep) axial valley and a deep (>1000 m deep) axial valley. An axial high is found to the east of the Australian–Antarctic Discordance (AAD) (east of 128°E) and between 82°E and 104°E. A shallow rift valley is found from 104°E to 114°E and from 82°E westward past the Amsterdam/St. Paul hotspot (ASP) to about 30°S, 75°E. Deep rift valleys are found from 114°E to 128°E in the vicinity of the AAD and from the Indian Ocean Triple Junction (IOTJ) at 25°S, 70°E to about 30°S, 75°E. The transition near 30°S occurs in an area of constant zero-age depth and does not appear to result from an increase in mantle temperature. It could be the result of the rapid increase in spreading rate along that portion of the SEIR. The most likely cause of the other transitions in axial morphology is variations in mantle temperature. The transitions between the different types of axial morphology are well defined and occur over a limited distance. Transitions in axial morphology are accompanied by significant changes in ridge flank topographic roughness. The transitions from axial valleys to axial highs are also accompanied by changes in the amplitude of the seafloor magnetic anomalies. Our observations suggest that there are distinct modes rather than a continuum of axial morphology on the SEIR and that there appears to be a "threshold" mechanism for a rapid change between different states of axial morphology. The ASP has only a limited influence on the SEIR. The ridge axis is marked by an axial valley for the entire distance from the IOTJ up to and past the ASP. The ridge axis becomes shallower as the ASP is approached from the northwest but only by about 300 m over a distance of 800 km. In addition, the ridge continues to become shallower away from Amsterdam Island toward the transition to an axial high at 82°E, 350 km to the east of the ASP. The Kerguelen hotspot appears to exert a major influence on the morphology of the SEIR by feeding asthenospheric material to the ridge axis. A long, narrow finger-like gravity high extends ENE away from the Kerguelen Plateau for a distance of 500 km. Shipboard data show that the gravity high results from a large volcanic ridge. The ridge appears analogous to the Rodriguez Ridge extending from the Reunion hotspot toward the Central Indian Ridge. A series of lower and broader lineated gravity highs extend from the volcanic ridge toward the SEIR in the ridge segment between the 81°E and 85°E transforms, which is the westernmost segment with an axial high. The only region of significant off-ridge seismicity on the Antarctic flank of the SEIR is a diffuse band of epicenters extending from Kerguelen to the SEIR within the segment between the 81°E and 85°E fracture zones. The along-axis gradient in depth from 86°E to the AAD and the transitions in axial morphology at 104°E and 114°E most likely reflect along-axis variations in mantle temperature and melt production rate due to distance from the Kerguelen hotspot and the influence of the AAD.

Introduction

The morphology and dynamics of slow spreading and fast spreading ridges differ dramatically. Slow spreading ridges (e.g., the Mid-Atlantic Ridge, spreading at 20–30 mm/yr) are characterized by 1–2 km deep, 25–40 km wide rift valleys and rugged flanking topography. In contrast, fast spreading ridges (e.g., the East Pacific Rise, spreading at 65–150 mm/yr) are characterized by approximately 20 km wide and 200–400 m high ridge axis highs and smooth flanking topography [Heezen, 1960; Heezen *et al.*, 1959; Macdonald, 1986, 1989; Menard, 1964, 1967]. In addition, slow spreading ridges

exhibit a much greater along-axis variation in axial depth. While the axial depth and the amplitude of the axial topographic high at fast spreading ridges generally do not vary along strike within a ridge segment by more than about 200 m, the axial depth of slow spreading ridge axes frequently varies along strike by 1000 m or more within a single ridge segment [Fox and Gallo, 1984; Lin and Phipps Morgan, 1992].

Several models have been proposed for the origin of axial morphology. A class of models which have received considerable attention is the vertical conduit or "viscous head loss" model [Lachenbruch, 1973; Lachenbruch and Sass, 1975; Osmaston, 1971; Sleep, 1969; Sleep and Rosendahl, 1979] which depends on viscous interactions between the upwelling asthenospheric material and the walls of an idealized vertical "conduit" in which the upwelling material is assumed to flow.

At slow spreading rates (narrow conduit), viscous coupling results in a loss of hydraulic head and emplacement of new crustal material at a level below where it is in hydrostatic equilibrium, while the flanking areas are uplifted by viscous shear, resulting in a deep rift valley. At faster spreading rates (wide conduit), viscous interactions are less important, and the asthenospheric material is able to reach a level where it is in hydrostatic equilibrium. This model is intuitively appealing and simple. However, *Phipps Morgan et al.* [1987] have criticized this model on the grounds that the temperature field resulting from even very concentrated upwelling beneath a mid-ocean ridge does not correspond to a steep-walled conduit.

There are presently two competing models for the origin of axial valleys at slow spreading ridges which appear consistent with current understanding of the thermal structure and rheology of ridge axes. *Phipps Morgan et al.* [1987] have proposed an extension of the "steady state necking" model of *Tapponnier and Francheteau* [1978]. They argue that the thickening of the lithospheric plate away from the axis will result in horizontal extensional stresses effectively being applied at different depths with distance from the axis. This will create bending moments resulting in the deformation of the plate to form an axial valley. They show that the amplitude of the stress-created topography is inversely proportional to the distance over which the thickening of the lithosphere occurs. Thus it might be expected that this mechanism will produce an inverse relationship between spreading rate and axial valley depth with the axial valley depth decreasing steadily with increasing spreading rate.

Chen and Morgan [1990] have proposed a quite different model for the creation of an axial valley at slow spreading ridges. Their model based on the interplay between a potential failure zone in the strong (brittle) lithosphere resulting from high stress due to the divergence of mantle flow beneath the ridge axis and a weak (ductile) zone in the lower crust which has a significantly lower yield stress than does mantle material at the same temperature. They argued that at slow spreading rates, isotherms will be deeper at a given distance from the axis, resulting in a rigid crust or small region of ductile lower crust near the axis. As a result, the crust and upper mantle will fail brittly under extension, producing an axial rift valley. As the spreading rate increases, the region of ductile lower crust becomes wider and at some point can serve to decouple the brittle crust above from the viscously stretching area below. This mechanism has the potential of producing a very rapid transition with increasing spreading rate (or increasing mantle temperature) from a deep rift valley to no rift valley.

Although the morphology of both fast and slow spreading ridges has been extensively studied, the nature of the transition between an axial valley and an axial high is poorly understood. *Macdonald* [1986] compiled data on the depth of the axial valley as a function of spreading rate and argued that there is a systematic decrease in the valley depth with increasing spreading rate and a gradual transition from a valley to an axial high. *Malinverno* [1993] also found that relief of the axial valley generally decreased with increasing spreading rate but that at a given axial depth (which can be taken as an indicator of crustal thickness and/or mantle temperature) the transition from axial valley to axial high occurs over a small range of spreading rates. *Small and Sandwell* [1989] analyzed Geosat altimetry gravity profiles across mid-ocean ridges and concluded that there is an abrupt change in the nature of the ridge axis gravity anomaly at spreading rates of 60-70 mm/yr. At slower

spreading rates, the axis is marked by a variable high-amplitude gravity trough, while at higher spreading rates the axis is associated with a uniform low-amplitude gravity high. *Small and Sandwell* [1994] suggest that ridge-axis morphology is controlled by a threshold phenomenon that is sensitive to relatively small changes in temperature and conclude that the location of the transitions in axial morphology along these two ridges may be controlled by proximity to hotspots and the availability of melt at the ridge axis. They conclude that there is an abrupt change in the compensation mechanism and that there are two dynamically distinct modes of seafloor spreading.

It is of importance in understanding ridge axis processes to determine how and in what circumstances one mode of axial morphology passes into the other. Acquiring this knowledge requires examination of intermediate spreading rate ridges, ideally along a continuous length of spreading center where the transition in morphology can be observed along strike. Our present knowledge of intermediate spreading rate ridges is primarily based on studies of a number of relatively short ridge segments in the eastern Pacific, in particular the northernmost East Pacific Rise between the Tamayo and Rivera fracture zones, the Juan de Fuca Ridge and the Galapagos (Cocos-Nazca) Spreading Center. The Chile Ridge is also spreading at intermediate rates of from 58 to 63 mm/yr [*Herron et al.*, 1981], but very little shipboard data are available across the Chile Ridge axis. It also appears to be associated with one form of axial morphology (a deep rift valley) along its entire length.

The two Southern Ocean ridges (Pacific-Antarctic Ridge and Southeast Indian Ridge) have spreading rates that span the range over which the transition from axial high to axial valley is expected to occur, and transitions in axial morphology along these ridges have been inferred from Geosat altimetry data [*Marks et al.*, 1991a; *Small and Sandwell*, 1992, 1994]. There are very few shipboard data available from the southwestern portion of the Pacific-Antarctic Ridge, where the change in axial morphology is observed in the satellite gravity data. In this paper, we will analyze available shipboard magnetic and bathymetric profiles across the Southeast Indian Ridge (SEIR) from the Indian Ocean Triple Junction (IOTJ) at 25°S, 70°E to the George V fracture zone at 51°S, 139°E in order to examine variations in morphology along an intermediate spreading rate ridge axis, the nature of transitions in axial morphology, and the factors which control such transitions.

Tectonic Setting

The Southeast Indian Ridge forms the boundary between the Australian and Antarctic plates from the IOTJ located near 25°S, 70°E to the Macquarie Ridge complex south of New Zealand near 63°S, 165°E (Figure 1). Rifting between Australia and Antarctica began in the Late Cretaceous with seafloor spreading beginning south of Australia by Anomaly 34 time [*Cande and Mutter*, 1982], probably at about 96 Ma [*Veevers*, 1986]. Very slow seafloor spreading (~4.5 mm/yr half rate) continued to the south of Australia until about Anomaly 18 time (mid-Eocene 40-45 Ma) [*Cande and Mutter*, 1982] and appears to have been accompanied by extension through faulting and block rotation farther to the west on the Kerguelen Plateau and in the deep-water Diamantina Zone and Labuan Basin [*Fritsch et al.*, 1992; *Munsch et al.*, 1992a, 1992b;

Figure 1. Map of the Southeast Indian Ridge (SEIR) showing the locations of the four areas (boxes) discussed in the text, Kerguelen hotspot (KER), Amsterdam and St. Paul hotspot (ASP), and the Australian-Antarctic Discordance (AAD). Solid line shows the ridge axis location as digitized by *Small and Sandwell* [1992]. Dots indicate the actual locations at which the profiles used in this study crossed the ridge axis.

Rotstein et al., 1991; *Royer and Sandwell*, 1989; *Royer and Coffin*, 1992]. In the mid-Eocene, about Anomaly 18 time, the spreading full rate increased to 45-50 mm/yr to the south of Australia [*Cande and Mutter*, 1982; *Royer and Sandwell*, 1989] and seafloor spreading propagated to the west, splitting Broken Ridge off of the Kerguelen Plateau [*Munschy et al.*, 1992a; *Royer and Sandwell*, 1989]. Spreading rates increased again about Anomaly 13 time to intermediate rates of about 55-65 mm/yr rate [*Royer and Sandwell*, 1989].

The Southeast Indian Ridge to the west of Kerguelen and Broken Ridge formed in the Cretaceous as the result of seafloor spreading between India and Antarctica. In the late Cretaceous, the seafloor now making up the Central Indian and Crozet Basins was generated by very fast (>140 mm/yr) spreading on this ridge [*McKenzie and Sclater*, 1971; *Norton and Sclater*, 1979; *Patriat and Segoufin*, 1988; *Schlich*, 1981]. This western portion of the proto-SEIR was linked by the Ninetyeast transform to a spreading center in the Wharton Basin which continued north of Australia [*Sclater and Fisher*, 1974]. Thus a triple junction had to exist near the southern end of Ninetyeast transform.

The mid-Eocene increase in opening rate between Australia and Antarctica was accompanied by a major reorganization of spreading throughout the Indian Ocean. The Wharton Basin spreading center became inactive at that time with Anomaly 20 as the last identifiable magnetic lineation [*Liu et al.*, 1983]. Spreading rates in the Central Indian Basin decreased from about 110 mm/yr to about 60 mm/yr [*Schlich*, 1981] and the direction of spreading changed from nearly north-south to N45°E [*Sclater et al.*, 1976]. Spreading on the SEIR was essentially constant in both direction and rate from the mid-Eocene (40-45 Ma) [*Patriat and Segoufin*, 1988; *Royer and Sandwell*, 1989] until about 10 Ma, when rates increased by about 15-20% [*Royer and Sandwell*, 1989; *Royer and*

Schlich, 1988; *Vogt et al.*, 1984; *Weissel and Hayes*, 1971, 1972].

Present-day total spreading rates increase rapidly from 57.5 mm/yr at the IOTJ to 68 mm/yr at Amsterdam and St. Paul Islands and then more slowly to a maximum of 75.6 mm/yr near 50°S, 114°E (Figure 2). Spreading rates then decrease slowly to 72 mm/yr at the George V fracture zone near 50°S, 139°E [*DeMets et al.*, 1990]. Spreading on the SEIR has generally been asymmetric. The overall sense of the asymmetry is such that more crust has been created to the north of the ridge axis to east of Amsterdam Island and more crust has been created to the south of the axis to the west of Amsterdam Island [*Hayes*, 1976; *Royer and Schlich*, 1988; *Weissel and Hayes*, 1971]. The degree of asymmetry has varied in time but has been as great as 30%.

The dominant morphologic structures of the southeast Indian Ocean include the Amsterdam/St. Paul massif, the Kerguelen Plateau, Broken Ridge, and the Australian Antarctic Discordance. Amsterdam and St. Paul Islands, located near 38°S, 77.5°E and 39°S, 78.5°E, occupy a volcanic ridge roughly 250 km by 150 km in extent, much of which lies at a depth of about 1500-2000 m [*Royer and Schlich*, 1988]. The islands themselves are small (<10 km across) and consist of young tholeiitic basalts. The SEIR ridge axis appears to cross the northern portion of the massif, approximately 50-100 km north of the two islands. The volcanic massif has been built within the past 4 m.y. since the SEIR axis passed over the Amsterdam/St. Paul hotspot (ASP) [*Royer and Schlich*, 1988]. Prior to that time, the hotspot was under the Australian plate, and its trace can be seen in the satellite gravity field [*Sandwell and Smith*, 1992, 1994] as a discontinuous series of small gravity highs, presumably representing seamounts, which extend from the ridge axis toward the western end of Broken Ridge.

Figure 2. Variation of spreading rate along the SEIR from the Indian Ocean Triple Junction (IOTJ) to 140°E. Spreading rates were calculated from the NUVEL-1 pole [DeMets *et al.*, 1990].

The Kerguelen hotspot is a major and long-lived feature which apparently has been responsible for the creation of both the Ninetyeast Ridge and the Kerguelen Plateau. Schilling [1991] concludes that Kerguelen has the highest flux of any near-ridge hotspot, and Storey *et al.* [1989] suggest that Kerguelen material has contaminated the entire Indian Ocean asthenosphere giving Indian Ocean mid-ocean ridge basalt (MORB) its particular isotopic composition. The Kerguelen Plateau extends for over 2500 km from about 46°S, 65°E to 64°S, 85°E near the Antarctic margin. The northern portion of the plateau (north of about 54°S) is relatively shallow with large areas at depths of less than 1000 m, while the southern Kerguelen Plateau is deeper, reaching depths of 1500–2000 m [Schlich *et al.*, 1987]. The Kerguelen Plateau was primarily built by volcanic activity prior to about 110 Ma at a near-ridge oceanic setting in the region between Greater India and Australia/Antarctica [Royer and Sandwell, 1989; Royer and Coffin, 1992; Whitechurch *et al.*, 1992]. Much of the Kerguelen Plateau was near sea level until the late Cretaceous rifting [Fritsch *et al.*, 1992; Munschy *et al.*, 1992b] and it may have formed a feature very similar to present-day Iceland and the aseismic ridge between Greenland and the Faroe Islands. Volcanic activity related to the Kerguelen hotspot has continued up to the present. Royer and Coffin [1992] conclude that the northeastern portion of the Kerguelen Plateau (north of the islands) dates from the early Cenozoic, and Royer and Sandwell [1989] suggest that the northwestern end of the Kerguelen Plateau is Oligocene or younger. The oldest isotopic age for volcanism on the Kerguelen islands is middle Eocene [Giret and Lameyre, 1983], and recent volcanism has occurred at both Kerguelen and at Heard Island, farther south near 53°S, 73° 30'E.

The Australian Antarctic Discordance (AAD) is a region of anomalously deep and rough bathymetry centered on the SEIR axis between 120°E and 128°E [Hayes and Conolly, 1972; Weissel and Hayes, 1974]. The AAD is characterized by a morphology typical of slow spreading ridges with a deep rift valley, rough bathymetry, and closely spaced fracture zones [Hayes and Conolly, 1972; Palmer *et al.*, 1993; Sempere *et al.*,

1991]. Klein *et al.* [1988] found a sharp geochemical boundary between isotopically distinct mantle provinces characteristic of the Pacific and Indian Oceans occurs in the AAD. Klein *et al.* [1991] interpreted major element variations along the SEIR as indicating a colder mantle and lower melt production beneath the AAD than to the east or west. Forsyth *et al.* [1987] showed that the AAD has unusually high surface wave velocities also suggesting that the upper mantle beneath the AAD is unusually cold. These observations support an early hypothesis that the AAD represents a “cold spot” or area of mantle down flow [Hayes and Dimm, 1972; Vogt and Johnson, 1973; Weissel and Hayes, 1974]. Marks *et al.* [1990, 1991b], following Vogt and Johnson [1973], suggested that hotspot-driven subaxial mantle flow along the SEIR from the Amsterdam hotspot in the west and the Balleny and Tasmanid hotspots to the east is responsible for convergence and downwelling beneath the AAD.

Data

All shipboard profiles across the SEIR axis were extracted from the Lamont Doherty Earth Observatory marine geophysical database. To minimize the possibility of contamination with transform topography, profiles that crossed or were close to a transform fault or fracture zone observed on the satellite altimetry-derived gravity map [Sandwell and Smith, 1992; 1994] were eliminated, as were profiles where the topography was obviously affected by transform morphology. We also excluded data from the AAD because it has been the subject of a number of recently published studies [Palmer *et al.*, 1993; Pyle *et al.*, 1992; Sempere *et al.*, 1991] based on much more detailed data than are available from the rest of the SEIR. We will assume the results of these studies in our discussion.

The first estimate of the location of the ridge axis on each profile was taken as the intersection of the profile with the digitized ridge axis of Small and Sandwell [1992]. The first estimates were modified to the actual position of the axis using the magnetic anomaly and bathymetric profiles. We were able to use the central anomaly magnetization high [Klitgord,

1976] over the active intrusion zone to locate the ridge axis when it was not entirely clear from the bathymetry. Most corrections from the *Small and Sandwell* [1992] axis were less than 20 km, with a maximum of 40 km. In most cases, the axis is located at the deepest point of a valley or shallowest point of a bathymetric high. The profiles were projected perpendicular to the local ridge axis trend [DeMets *et al.*, 1990]. The projected bathymetric profiles used in this study are shown in Figure 3, and the corresponding magnetics profiles are shown in Figure 4. The ridge crossing location of each profile is shown in Figure 1.

To quantify axial morphology and determine along-axis variations in parameters, we measured the axial depth, zero-age depth, and roughness from the projected bathymetric profiles. The axial depth is the depth actually observed at the axis on the original profile. The zero-age depth is estimated by fitting a thermal subsidence trend on 175-km-long windows extending from 25 to 200 km from the axis on each flank using local subsidence rates from *Hayes and Kane* [1994] averaged for 5° along the axis. Although the use of 175-km-long profiles reduces the number of estimates, it gives more reliable estimates than do shorter profiles. The roughness is defined as the square root of the average squared deviation about a linear trend measured on 100-km-long windows of the profile for each flank. The use of 100-km-long profiles provides roughness estimates that are independent of the profile length [Malinverno and Cowie, 1993]. Only topographic points at least 25 km away from the axis were used to avoid the effects of axial morphology. Since the estimation of zero-age depth and roughness can be severely biased by the presence of large seamounts, seamounts were excluded from calculation. The estimates of zero-age depth and roughness for each profile are plotted as a function of position along the axis in Figure 5.

Anderson et al. [1980] observed an abrupt change in the amplitude of the magnetic anomaly across the eastern boundary of the AAD, which has been attributed to a difference in the Fe and Ti content of the basalts in the two areas [Anderson *et al.*, 1980; Klein *et al.*, 1991]. *Cochran* [1991] suggested that magnetic anomaly amplitude varies systematically with the form of the axial topography along the entire SEIR. We quantitatively investigated variations in magnetic anomaly amplitude using a procedure similar to that used with the bathymetric data. Since the "roughness" of the magnetic anomaly depends on the portion of the geomagnetic reversal sequence analyzed, we used the same age range for all profiles. The profiles were divided into 2.6-m.y. segments, and the standard deviation about the mean was determined for each segment. Since we wish to investigate the region near the ridge axis, only the axial segment extending to 1.3 Ma on both sides of the axis and the two segments extending in each direction from the axis for 2.6 m.y. were used. The 2.6-m.y. segments vary in length from 98 km at 114°E, where the spreading rate is a maximum of 75.6 mm/yr, to 75 km near the IOTJ with the minimum spreading rate of 58.0 mm/yr. The results are plotted in Figure 5. The data presented graphically in Figure 5 may be obtained in tabular form from the authors upon request.

Results

Along-Axis Variations in Axial Morphology

The SEIR axis is marked by a rift valley for the entire distance from the IOTJ to the Amsterdam/St. Paul massif

(Figure 3a). From the IOTJ to 29.5°S, the ridge is marked by a deep (>850 m), well-developed axial valley. Near 29.5°S, there is a transition from a deep rift valley to a shallow axial valley (200–400 m deep). The shallow valley persists to a fracture zone near 34°S which marks the limit of the geochemical signal of the ASP [Dosso *et al.*, 1988; Michard *et al.*, 1986]. A somewhat deeper (~600 m) axial valley is observed immediately to the southeast of the fracture zone. A shallow valley (~250 m) is again observed on two profiles (jc1320 and ggal1b, Figure 3) which cross the axis near 36°S, 79°E, close to the hotspot massif. Profiles presented by *Royer and Schlich* [1988] (their profiles GA1-11 and MD37-16) suggest that a shallow axial valley may persist onto the massif.

There are very few shipboard data across the ridge axis between Amsterdam Island and 90°E. *Royer and Schlich* [1988] show a short profile (their profile GA3-35) across a broad 500-m-deep valley near 78.1°E, 39°S about 55 km ESE of St. Paul's Island, which the magnetics indicate is the present spreading center. The depth at the bottom of the valley is about 2150 m. The satellite gravity map (Plate 1) shows a very pronounced axial low extending to the southeast away from the hotspot massif near 78°E, 40.5°S. *Royer and Schlich* [1988] show a bathymetric and magnetic profile across the eastern part of this feature (their profile MD37-13) which shows it to be a nearly 1-km-deep valley which the magnetics identify as the spreading center. The depth at the axis is just below 3000 m. Our westernmost profile across this portion of the SEIR (ggal1) is located in an area between 78°E and 80°E where the ridge axis can not be easily located on the satellite gravity map (Plate 1). The shipboard profile shows the axis to consist of a shallow valley with axial relief of 120 m.

Our next profile to the east, c1105, also shows a shallow axial valley (Figure 3b). This profile crosses the axis at 81.25°E, 41.29°S, just to the east of a transform (Figure 1 and Plate 1). The satellite gravity shows that the axis changes from an axial valley to an axial high slightly east of profile c1105, about 350 km east of the ASP. There is a gap in surface ship profiles from 81°E to nearly 88°E. The gravity map (Plate 1) shows the ridge axis to be marked by a gravity high through this interval. The axis is associated with a bathymetric high on all of the shipboard profiles from 88°E to 101°E except perhaps for profile c0802 at 93.8°E, on which the axis is difficult to recognize from the bathymetry. This is probably due to a propagating rift near 94°E (Plate 1).

All bathymetry profiles from 105°E to the AAD show a valley at the ridge axis. The ridge axis is not easily recognized between 101°E and 104°E on the satellite-derived gravity map. A recent cruise to this portion of the SEIR [Cochran *et al.*, 1995; Sempere *et al.*, 1995] found that the transition from an axial high to an axial valley occurs near 104°E. The valley is relatively shallow (200 m–700 m) from 104°E to 114°E but deepens to over 1 km as the AAD is approached (Figure 5).

All of our profiles across the SEIR to the east of the AAD show a well developed axial high (Figure 3). This morphology persists westward to the 175-km offset Birubi transform which marks the eastern boundary of the AAD [Palmer *et al.*, 1993]. Immediately west of the Birubi transform, the ridge axis is characterized by an axial valley and the valley deepens from 400 m to over 1500 m relief within a single ridge segment (<50 km) away from the transform [Sempere *et al.*, 1991].

There are thus several transitions in the form of the axial morphology along the SEIR. The ridge axis is characterized by an axial high in two regions: from 82°E to 104°E to the

Figure 3a. Bathymetric profiles across the SEIR from the IOTJ to the ASP. The tick marks on the vertical axes are 1000 m apart. The name of each profile is at the tick mark corresponding to 3000 m depth for that profile. The ridge axis is at zero km on the horizontal axes.

Figure 3b. Bathymetric profiles across the SEIR from the ASP to 140°E. See Figure 3a caption.

Figure 4a. Magnetic profiles across the SEIR from the IOTJ to the ASP. The tick marks on the vertical axes are 1000 nT apart. The name of each profile is at the tick mark corresponding to 0 nT amplitude for that profile. The ridge axis is at 0 km on the horizontal axes.

west of the AAD, and for the entire region from the eastern boundary of the AAD near 128°E to the eastern end of our study area at 140°E. The transition from a high to a valley at the eastern boundary of the AAD is quite sharp and occurs across the Birubi transform. The transitions from an axial high to a shallow (200-700 m deep) axial valley near 82°E and 104°E appear to occur within a ridge segment (Plate 1) [Cochran *et al.*, 1995; Sempere *et al.*, 1995]. The transition from a shallow valley to a deep (>1000 m) valley at 114°E is sharp and occurs across a 70-km offset transform (Plate 1). A similar set of transitions occurs at the western end of the region with an axial high. The transition from an axial high to a shallow valley near 82°E occurs within a single ridge segment (Plate 1). There is also a transition from a shallow to a deep rift valley near 29.5°S. However, we do not have sufficient data to determine the nature of that transition.

Along-Axis Variations in Depth, Bathymetric Roughness, and Magnetic Anomaly Amplitude

The observed axial depths decrease slightly from the IOTJ to 29.5°S, although with a great deal of scatter. With the excep-

tion of one extremely deep value of 4180 m found at 29.1°S and one shallow value of 3415 m at 27.7°S, axial depths are in the range of 3540-3870 m (Figure 5). Axial depths are significantly shallower to the southeast of 29.5°S and are in the range of 2930-3270 m from 29.5°S to the immediate vicinity of the Amsterdam massif.

Zero-age depth remains nearly constant from the IOTJ to a transform offset at about 31.5°S, 76.5°E. Zero-age depths along this portion of the ridge are in the range of 2700-2800 m, although there is one deeper value of 2885 m at 30.5°S (Figure 5). The change in axial depth near 29.5°S is not accompanied by a change in the zero-age depth. Zero-age depths begin to decrease southeastward from 31.5°S reaching about 2550 m at 35°S. A zero-age depth of about 2471 m was determined on a profile at 36.3°E in the last segment to the northwest of the ASP. One profile across the ridge axis within the Amsterdam-St. Paul's segment presented by Royer and Schlich [1988] (their profile EL48-3) shows the ridge axis to be located within a shallow valley at a depth of 1700-2000 m.

Zero-age depth is 2569 m on profile ggal1 just east of the ASP and remains in the range of 2350-2550 m east to about

Plate 1. Satellite gravity maps of the SEIR for the four areas shown on Figure 1. Color changes are at 5 mGal intervals. Axial highs appear as linear gravity highs, and axial valleys are marked by linear gravity lows.

Figure 4b. Magnetic profiles across the SEIR from the ASP to 140°E. See Figure 4a caption.

Figure 5. Variation of (top) axial depth and zero-age depth, (middle) bathymetric roughness and (bottom) magnetic anomaly amplitude with latitude from the IOTJ to the ASP and with longitude from the ASP to 140°E. Horizontal scale is nearly linear in distance. Axial highs and axial valleys are represented by solid and open symbols, respectively. Zero-age depth was calculated by fitting ridge flank subsidence trend, and axial depth is depth measured at the ridge axis for each profile. Manner in which bathymetric roughness and magnetic anomaly amplitude are defined is discussed in text.

92°E. The shallowest profile is cel54 at 87.7°E with a zero-age depth of 2366 m, 200 m shallower than on the profile closest to the ASP. There are insufficient data to determine the variation in depth of the SEIR between 78°E and 90°E in any detail. However, it is clear that the ridge axis does not become systematically shallower toward the ASP and that the crest in zero-age depth is some distance to the east of the ASP.

Zero-age depth increases regularly from 2366 m near 88°E to 3192 m at 119.7°E near the boundary of the AAD. There is a slight steepening in the slope of the increase in zero-age depth to the east of 101°E. To the east of the AAD, zero-age depth increases monotonically to the west from 2444 m on profile v1606 at 138.4°E to 3016 m on profile v3302 at 129.8°E near the boundary with the AAD.

Plate 2. Satellite gravity map showing the region between Kerguelen Plateau and the SEIR axis. Color changes are at 5 mGal intervals. Note the prominent ridge extending from the Kerguelen Plateau at 48°S, 72°E to 47°S, 77°E and the broad, low-amplitude gravity highs extending from the ridge to the SEIR axis. Black line shows the locations of the bathymetric profile mmd3. Dots show the locations where teleseismic earthquake epicenters have been located. KER indicates Kerguelen Plateau and ASP indicates Amsterdam and St. Paul massif.

Bathymetric roughness determinations to the northwest of the ASP appear to fall into two distinct groups. To the northwest of 31°S, roughness shows the largest values measured along the entire SEIR (151-294 m) and a great deal of scatter, while to the southeast of 31°S the range in roughness is 89-180 m with all but one estimate less than 150 m (Figure 5). The transition in axial morphology observed at 29.5°S does not coincide exactly with the change in bathymetric roughness. However, the transition in the amplitude of the satellite gravity anomaly over the axis appears to be near 30.5°S rather than at 29.5°S (Plate 1). Thus, if the satellite anomalies give a truer picture of the along-axis variations in morphology than do the two profiles (jc132c and jc132d) available to us between 29.5°S and 30.5°S, then the transitions in axial valley depth and bathymetric roughness may more nearly coincide.

Magnetic anomaly amplitude also shows no change at 29.5°S and no systematic variation from the IOTJ to 31°S with a mean of 131 ± 36 nT. These are the lowest-amplitude anomalies recorded along the SEIR (Figures 4 and 5). There is a data gap between 31°S and 35°S. Magnetic anomaly amplitude increases very rapidly from 35°S toward the ASP on the few profiles available in that region.

The bathymetric roughness is low and quite uniform in the region with an axial high to the east of the ASP. Roughness determinations from 88°E to 101°E are consistently in the range of 51-83 m with the exception of one larger value of 105 m on profile eel49 near 95°E which runs near a 150-km offset multiple transform system. The bathymetric roughness in the region with an axial valley between 105°E and the AAD is consistently greater than to the west ranging from 76 to 218 m with a regular increase toward the AAD (Figure 5). In addition to the change in bathymetric roughness, the transition in the form of the axial morphology near 104°E is accompanied by a significant change in the magnetic amplitude (Figure 5). The magnetic anomaly amplitudes to the west of 104°E are in the range of 183-348 nT with a mean of 266 ± 49.8 nT, in contrast to the range of 128-238 nT with a mean of 187 ± 26.8 nT from 104°E to the AAD. Bathymetric roughness to the east of the AAD is low with values in the range of 66-135 m, similar to those observed from 82°E to 104°E where the ridge also has an axial high. The roughness may increase slightly to the west toward the AAD, but if so, the trend is not well developed (Figure 5). The magnetic anomaly amplitude pattern to the east of the AAD is also similar to that observed from 82°E to 104°E with high amplitudes (mean value of 332 ± 58.7 nT), a great deal of scatter and no geographic trend.

Examination of Figure 5 shows that both the bathymetric roughness and magnetic anomaly amplitude are correlated with the form of axial morphology on the SEIR. In general, areas with axial highs have uniform, lower bathymetric roughness and larger magnetic anomaly amplitude, while regions with axial valleys have larger and more variable roughness and lower magnetic anomaly amplitude.

Discussion

Influence of the Amsterdam-St. Paul and Kerguelen Hotspots on the SEIR

The Amsterdam/St. Paul volcanic edifice is one of the dominant morphologic features of the SEIR. However, although located less than 100 km from the ridge axis, the ASP has only limited influence on the SEIR. The SEIR axis is characterized

by an axial valley for the entire distance from the IOTJ to the ASP and on beyond it to 82°E. The western region with an axial high (82°E to 104°E) does not include the hotspot but rather begins 350 km to the east and extends away from the ASP.

Zero-age depths on the SEIR decrease to the southeast toward the ASP from about 31°S. However, the change in depth toward the ASP is not profound. Zero-age depths decrease from about 2750 m to the northwest of 31°S to a measurement of 2471 m in the segment immediately northwest of the Amsterdam fracture zone, a change of about 280 m over a distance of 800 km. However, the ASP is not at the shallowest portion of the SEIR axis. The ridge axis continues to become shallower to the east away from the ASP. Our shallowest zero-age depth of 2366 m and axial depth measurement of 2279 m are both on profile eel49, which crosses the ridge axis at 87.79°E, about 800 km east of the ASP.

Spreading rates along the SEIR increase steadily to the southeast away from the IOTJ (Figure 2). *Small and Sandwell* [1992] analyzed deflection of the vertical profiles obtained from satellite altimetry across mid-ocean ridges with a wide range of spreading rates. They found a relatively rapid change from a pattern characteristic of a gravity low to a pattern characteristic of a gravity high at spreading rates of 55-80 mm/yr. A similar transition has been noted in compilations of bathymetric profiles [Macdonald, 1986; Malinverno, 1993; Small, 1994]. The entire SEIR falls within that range of spreading rates. In particular, the transition from an axial valley to an axial high near 82°E occurs at a spreading rate of 70.5 mm/yr (Figure 2). It could thus be argued that the transition from an axial valley to an axial high observed at 82°E is simply caused by the increase in spreading rates to the east. However, spreading rates continue to increase to the east and the fastest spreading portion of the SEIR is actually near 114°E [DeMets *et al.*, 1990] (Figure 2), where the ridge has a deep axial valley. In addition, the portion of the ridge axis between 82°E and 104°E with an axial high is also the shallowest section of the ridge. These observations imply that the morphology of the SEIR to the east of Amsterdam is not governed solely by spreading rate dependent effects and that large lateral variations in the temperature of the underlying mantle must play an important role.

Ridge axis depths along the SEIR increase steadily to the east for nearly 3000 km from near 88°E to the AAD. The great depth and rough bathymetry of the AAD led quite early to the hypothesis that the AAD represents a "cold spot" or area of mantle downflow [Hayes and Conolly, 1972; Vogt and Johnson, 1973; Weissel and Hayes, 1974]. Klein *et al.* [1988] found that a sharp boundary occurs in the AAD between isotopically distinct mantle provinces characteristic of the Pacific and Indian Oceans. They propose that this boundary results from downward flow resulting from the convergence of two convective flows, one bringing Indian Ocean mantle from the west and the other Pacific mantle from the east. Marks *et al.* [1990, 1991b], following Vogt and Johnson [1973], have suggested that hotspot-driven subaxial mantle flow from the Amsterdam hotspot in the west and the Balleny and Tasmanid hotspots in the east is responsible for the convergence and downwelling beneath the AAD.

However, the shallowest section of the ridge axis is located considerably to the east of the ASP, making it unlikely that the ASP is the source of the mantle material flowing eastward toward the AAD. This raises the possibility that it is Kergue-

len which is feeding hot asthenospheric material to the ridge axis to the east of 82°E. A long, narrow 90-mGal gravity high flanked by gravity lows extends ENE away from the Kerguelen Plateau for a distance of about 500 km (Plate 2). A bathymetric profile across this gravity feature at 74°E shows it to be a broad, steep-sided topographic ridge, more than 2000 m high and 30 km wide at its base, which rises to within 820 m of the sea surface (Figure 6). On other oblique crossings, the central portion of the ridge reaches depths of 633 m. The ridge ends at about 46° 45'S, 77°E, but a series of broad lineated gravity highs and seamounts continue to the northeast up to the ridge axis between 81°E and 85°E (Plate 2). These gravity highs could represent areas of thicker and therefore shallower crust resulting, for example, from underplating.

There is also a diffuse band of moderate ($M_b = 4.7-5.5$) earthquake epicenters extending from the Kerguelen Plateau to the SEIR axis within the segment between the 81°E and 85°E fracture zones (Plate 2). This is the only such area of significant off-ridge earthquakes on the Antarctic flank of the SEIR. *Bergman et al.* [1984] determined normal faulting mechanisms for three of the earthquakes, all with T axis parallel to the ridge axis. They also concluded that these earthquakes appear to be causally different from a broad band of off-ridge earthquakes observed on the northern flank of the SEIR, which they attributed to a plate-wide stress regime resulting from the collision of India and Asia. *Bergman et al.* [1984] concluded that the most likely explanation of the Antarctic plate earthquakes between the 81°E and 85°E fracture zones is thermal and bending stresses in the lithosphere overlying a thermal anomaly resulting from channeled flow from the Kerguelen hotspot to the SEIR ridge axis.

Schilling [1985; 1991] and *Sleep* [1990] have investigated interactions between hotspots and ridge axes. *Schilling* [1985, 1991] has proposed that plume flux entrained by a mid-ocean ridge migrating away from a hotspot will establish a sublithospheric conduit through which material is transferred from a plume source to a migrating ridge sink and that plumes may maintain connections with migrating spreading centers over distances of more than 1400 km. *Schilling* [1991] found evidence for Kerguelen influence on the SEIR but assumed that Kerguelen flux is channeled to the ASP.

We infer that asthenospheric material from the Kerguelen plume is being channeled to the SEIR axis between the 81°E and 85°E fracture zones. The long finger-like ridge extending out from Kerguelen is analogous to the Rodriguez Ridge extending from the Reunion hotspot toward the Central Indian Ridge and the Darwin-Wolf lineation extending from the Galapagos hotspot toward the Cocos-Nazca spreading center which *Morgan* [1978] inferred to be volcanic ridges that formed over an asthenospheric channel. The distance from Kerguelen to the ridge axis (about 1300 km) is only slightly greater than the distance from Reunion to the Central Indian Ridge axis (about 1100 km).

A possible reason why asthenospheric material from the Kerguelen hotspot is channeled to the SEIR axis near 85°E rather than toward the ASP is that until very recently, the Amsterdam hotspot was on the opposite side of the ridge axis. The track of the Amsterdam/St. Paul hotspot is a series of seamounts extending from the western tip of Broken Ridge to the southwest between the Amsterdam and St. Paul fracture zones. The SEIR ridge axis passed over the hotspot at about 4-5 Ma [*Royer and Schlich*, 1988]. Until that time, the ASP was

Figure 6. Along-track bathymetric and satellite gravity profiles for cruise mmdu3. The ship track is shown on Plate 2. The satellite gravity data were sampled at the location of the shipboard bathymetric data; 0-km on the profiles is located at 73°E, 46.6°S. The shipboard data present a bathymetric profile across the linear large-amplitude gravity high extending to the east from the Kerguelen Plateau.

isolated from the Kerguelen hotspot by the ridge axis and the two hotspots could not interact.

Transitions in Ridge Axis Morphology

There are three distinct styles of axial morphology displayed on the SEIR. An axial high is found to the east of the AAD (128°E–140°E) and from 82°E to 104°E on the shallowest portion of the ridge axis. A shallow rift valley (200–700 m deep) is found from 104°E to 114°E and from 82°E westward past the ASP to about 30°S, 75°E. A deep rift valley (>1000 m deep) is found from 114°E to 128°E (essentially the AAD but extending some distance to the west) and from the IOTJ at 25°S, 70°E to 30°S, 75°E.

The boundaries of the region centered in the AAD which is characterized by a deep rift valley are sharp and occur across transforms. The transition from an axial high to the east of the AAD and the axial valley within the AAD occurs across the 175 km offset Birubi transform [Palmer *et al.*, 1993]. The ridge axis to the east of the transform is marked by an axial high. Immediately west of the Birubi transform in the AAD, the ridge axis is characterized by a 400-m-deep rift valley which quickly deepens within the first ridge segment (<50 km) to have 1500-m relief [Sempere *et al.*, 1991]. The boundary between a deep rift valley and a shallow rift valley at 114°E to the west of the AAD appears to occur across a 75-km offset transform and to be abrupt; there is a shallow (~200 m deep) valley to the west of the transform and a deep (>1000 m deep) valley to the east (Figures 3 and 5 and Plate 1). The boundary between an axial high and a shallow valley occurs within a ridge segment near 104°E [Cochran *et al.*, 1995; Sempere *et al.*, 1995]. The other transition from an axial high to a shallow axial valley near 82°E also appears to occur within a single ridge segment based on satellite altimetry data (Plates 1 and 2) and the limited shipboard data available (Figure 3). It is difficult to characterize the westernmost transition, from a deep to a shallow axial valley near 30°S, since satellite and shipboard data disagree on its location by nearly 100 km.

The transition in axial morphology from a deep to a shallow rift valley near 30°S is not accompanied by a change in zero-age depth, implying that the change in morphology is probably not caused by a change in mantle temperature [e.g., Klein and Langmuir, 1987]. However, it does occur in a region where the spreading rate is increasing rapidly to the southeast and may represent a spreading rate induced change in the axial morphology.

The most likely cause of the other transitions in along-axis morphology is variations in mantle temperature. Klein *et al.* [1991] presented $\text{Si}_{8.0}^{\text{G}}$ and $\text{Fe}_{8.0}^{\text{G}}$ data from the SEIR axis between 115°E and 140°E which show consistently higher values to the east of the AAD than from within the AAD. Klein *et al.* [1991, p. 2103] interpret their data as indicating that “the coolest mantle temperatures reside beneath the AAD” and that “the hottest temperatures in the study area” occur to the east of the AAD. The two transitions from an axial high to a shallow rift valley at 82°E and 104°E flank the shallowest portion of the ridge axis and may reflect the variation in temperature and magma supply of the Kerguelen asthenosphere with distance from Kerguelen. The difference in the distance from the shallowest portion of the SEIR axis at 85°E–88°E and the two transitions (at 82°E and 104°E) may reflect the channeling of Kerguelen asthenospheric material along the axis toward the AAD.

The observation that there are distinct modes rather than a continuum of axial morphology on the intermediate spreading SEIR and that the transitions between the different modes are well defined and occur over a limited distance supports models for the formation of axial morphology which involve a “threshold” mechanism for a rapid change between two different states. This conclusion is supported by the observation that changes in axial morphology are accompanied by large abrupt changes in the bathymetric roughness and in the magnetic anomaly amplitude. These changes imply that the changes in axial morphology are accompanied by significant changes in the accretion mechanism which affects the local bathymetric relief (probably through the relief on faults) and the magnetization of the basalts erupted at the axis.

The mechanism proposed by Phipps Morgan *et al.* [1987] produces axial morphology through deformation of the plate due to the bending moment created by the thickening of the plate away from the axis. It thus predicts that axial relief should vary continuously with variations in spreading rate or mantle temperature. A model such as that proposed by Chen and Morgan [1990], in which there is a mechanism to cause a rapid transition between two different forms of axial morphology, is in better agreement with the observed variations in morphology along the SEIR axis. In their model, the extent of a “decoupling zone” in the ductile lower crust serves to control which of two very different forms of tectonics occurs at the ridge axis. If the ductile zone is small, the brittle crust will fail in extension due to stresses arising from the divergence of mantle flow beneath the ridge axis, resulting in a deep rift valley. As the spreading rate or mantle temperature increases, the width of the region of weak, ductile lower crust also increases and eventually serves to isolate the brittle upper crust from the area of mantle flow and viscous stretching below. It appears that on the SEIR the transition is from a deep rift valley to a shallow rift valley rather than to no rift valley, perhaps implying that the brittle portion of the crust is not completely isolated from extensional stress.

The third form of axial morphology observed on the SEIR is an axial high. Both the Phipps Morgan *et al.* [1987] and Chen and Morgan [1990] mechanisms are models for the rift valley–no rift valley transition. They do not attempt to explain the formation of axial highs. Madsen *et al.* [1984] and Kuo *et al.* [1986] have argued that the axial high is a flexural feature supported by a region of slightly lowered density in the crust and mantle beneath the axis.

A relatively rapid transition along axis from an axial high to a shallow rift valley is observed on the SEIR which again suggests that there is a temperature-related mechanism which controls whether an axial high or a shallow rift valley is found along the axis. A possible mechanism is the presence or absence of a steady state or quasi-steady state magma chamber. Phipps Morgan and Chen [1993] have argued that the presence or absence of a steady-state magma chamber will greatly reduce the vertically integrated yield strength of the lithosphere at the axis. Since large tensile stresses cannot be supported in a weakened lithosphere, the formation of normal faults with significant throw will be inhibited near the axis, perhaps preventing the creation of a shallow rift valley and resulting in lower bathymetric roughness.

A relationship between a magma chamber and the presence of an axial high is suggested by the observation that the amplitude of the seafloor spreading magnetic anomalies is significantly greater in areas with an axial high than in areas

with either a shallow or a deep rift valley (Figure 5). The greater magnetization of the axial high basalts on the SEIR could result from increased fractionation within an axial magma chamber. *Sinton and Detrick* [1992] noted that basalts erupted at ridges with axial highs are more fractionated than those from ridges with rift valleys which could result from increased residence time of the melt in a steady-state magma chamber.

Conclusions

1. Three very distinct types of axial morphology are found along the intermediate spreading rate SEIR. An axial high is found from the eastern boundary of the AAD at 128° to the eastern end of our study area at 140°E and from 82°E to 104°E. The latter area is the shallowest portion of the SEIR ridge axis. A shallow rift valley (200-700 m deep) is found from 104°E eastward to 114°E and from 82°E westward across the Amsterdam/St. Paul massif to about 30°S, 75°E. A deep rift valley (>1000 m deep) is found from 114°E to 128°E and from the IOTJ at 25°S, 70°E to 30°S, 75°E.

The three forms of axial morphology each have a different pattern of ridge flank bathymetric roughness and amplitude of seafloor spreading magnetic anomalies. Areas with an axial high are characterized by large-amplitude magnetic anomalies and smooth ridge flank topography. Areas with an axial valley have lower-amplitude magnetic anomalies and rougher ridge flank topography. Ridge flank topography is rougher in areas with a deep axial valley than in areas with a shallow axial valley.

Transitions between the three types of morphology are well defined and occur over short lengths of ridge axis. The boundaries of the region with a deep axial valley between 114°E and 128°E are abrupt and occur across transforms. The transitions from an axial high to a shallow rift valley at 82°E and 104°E both appear to occur within a ridge segment. Our data are not adequate to characterize the transition from a deep valley to a shallow valley near 30°S, 75°E. The observation that there are distinct modes rather than a continuum of axial morphology and that the transitions between the modes are well defined and occur over a limited distance supports models for the formation of axial morphology which involve a "threshold" mechanism for a rapid change between two different states. The change from a deep rift valley to a shallow rift valley may result from the formation of a sufficiently large region of ductile lower crust which serves to decouple the brittle crust to ductile flow in the mantle as proposed by *Chen and Morgan* [1990]. The change from a shallow axial valley to an axial high may result from the formation of a steady state magma chamber.

2. The Amsterdam-St. Paul hotspot has only a limited influence on the SEIR. The ridge axis is marked by an axial valley for the entire distance from the IOTJ up to and past the Amsterdam/St. Paul massif. The transition to an axial high occurs near 82°E, 350 km southeast of the ASP. The ridge axis becomes shallower as the ASP is approached from the northwest, but only modestly. Zero-age depths decrease by about 300 m over a distance of 800 km. In addition, the ridge continues to become shallower away from the ASP to the east toward the transition to an axial high at 82°E. If the AAD is due to downwelling resulting from the collision of asthenospheric flows channeled along the ridge axis [*Marks et al.*, 1990, 1991b], then the ASP does not appear to be the source of the asthenospheric flow from the west.

3. The Kerguelen hotspot exerts a major influence on the morphology of the SEIR by feeding asthenospheric material to the ridge axis. A long, narrow finger-like gravity high extends ENE away from the Kerguelen Plateau for a distance of 500 km. Shipboard data show that the gravity high results from a large volcanic ridge. The ridge appears to be analogous to the Rodriguez Ridge extending from the Reunion hotspot toward the Central Indian Ridge. A series of lower and broader lineated gravity highs extend from the end of the volcanic ridge to the SEIR axis in the ridge segment between the 81°E and 85°E fracture zone. This is the westernmost ridge segment with an axial high. This same area, between the 81°E and 85°E fracture zones is the only segment on the Antarctic flank of the SEIR with extensive off-ridge seismicity [*Bergman et al.*, 1984]. The region where the asthenospheric flow reaches the SEIR axis is the shallowest portion of the ridge axis and is characterized by an axial high. The along-axis gradient in depth from 86°E to the AAD and the transition to an axial valley at 104°E most likely reflect along-axis variations in mantle temperature and melt production rate due to distance from the Kerguelen hotspot and the influence of mantle downflow beneath the AAD.

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