Crustal magnetization and accretion at the Southwest Indian Ridge near the Atlantis II fracture zone, 0–25 Ma

Allegra Hosford,1,2 Maurice Tivey,1 Takeshi Matsumoto,3 Henry Dick,1 Hans Schouten,1 and Hajimu Kinoshita3

Received 9 May 2001; revised 15 June 2002; accepted 24 June 2002; published 25 March 2003.

[1] We analyze geophysical data that extend from 0 to 25-Myr-old seafloor on both flanks of the Southwest Indian Ridge (SWIR). Lineated marine magnetic anomalies are consistent and identifiable within the study area, even over seafloor lacking a basaltic upper crust. The full spreading rate of 14 km/Myr has remained nearly constant since at least 20 Ma, but crustal accretion has been highly asymmetric, with half rates of 8.5 and 5.5 km/Myr on the Antarctic and African flanks, respectively. This asymmetry may be unique to a ~400 km wide corridor between large-offset fracture zones of the SWIR. In contrast to the Mid-Atlantic Ridge, crustal magnetization amplitudes correlate directly with seafloor topography along the present-day rift valleys. This pattern appears to be primarily a function of along-axis variations in crustal thickness, rather than magnetic mineralogy. Off-axis, magnetization amplitudes at paleo-segment ends are more positive than at paleo-segment midpoints, suggesting the presence of an induced component of magnetization within the lower crust or serpentinized upper mantle. Alteration of the magnetic source layer at paleo-segment midpoints reduces magnetization amplitudes by 70–80% within 20 Myr of accretion. Magnetic and Ocean Drilling Program (ODP) Hole 735B data suggest that the lower crust cooled quickly enough to lock in a primary thermoremanent magnetization that is in phase with that of the overlying upper crust. Thus magnetic polarity boundaries within the intrusive lower crust may be steeper than envisioned in prior models of ocean crustal magnetization. As the crust ages, the lower crust becomes increasingly important in preserving marine magnetic stripes.

INDEX TERMS: 1550 Geomagnetism and Paleomagnetism: Spatial variations attributed to seafloor spreading (3005); 3035 Marine Geology and Geophysics: Mid-ocean ridge processes; 3045 Marine Geology and Geophysics: Seafloor morphology and bottom photography; 8158 Tectonophysics: Evolution of the Earth: Plate motions—present and recent (3040); 9340 Information Related to Geographic Region: Indian Ocean; KEYWORDS: Mid-ocean ridge, Southwest Indian Ridge, crustal magnetization, ridge segmentation, asymmetric spreading, Atlantis II fracture zone


1. Introduction

[2] The Southwest Indian Ridge (SWIR) is among the world’s slowest-spreading ridges, with an average full spreading rate of ~14 km/Myr [Patriat et al., 1997; Chu and Gordon, 1999]. Many inferences about crustal accretion at the ultraslow spreading SWIR have been based on analogies with the more extensively studied northern Mid-Atlantic Ridge (MAR), even though the northern MAR spreads 60% faster than the SWIR. Some aspects of crustal accretion at the MAR and the SWIR are likely to be similar.

For example, common observations such as rugged topography, deep axial valleys, asymmetric bathymetry across segment ends, and outcrops of serpentinized peridotite [Dick et al., 1992; Cannat, 1993] indicate that crustal accretion along both ridge systems probably occurs in the absence of ubiquitous and long-lived magma chambers [Detrick et al., 1990; Sinton and Detrick, 1992]. Additionally, seismic refraction experiments at the MAR and SWIR show similar P wave velocity structures within on-axis crustal sections [Tolstoy et al., 1993; Muller et al., 2000], indicating a gross similarity in lithologic structure at the two ridges. However, some fundamental aspects of mantle melting, melt migration, and ultimately, crustal accretion at the ultraslow spreading SWIR may differ fundamentally from that at the MAR because the mean crustal thickness decreases from a global average of 7 km to just 4 km at ultraslow spreading ridges [Sotin and Parmentier, 1989; Bown and White, 1994; White et al., 2001].
The combined effects of low crustal production and pervasive tectonism complicate the recording of marine magnetic anomalies in ultraslow spread crust. Hypotheses on the nature of the source of magnetic anomalies remain uncertain despite decades of work. Many studies conclude that all or most of the magnetic signal resides in the upper 500 m of extrusive crust, commonly associated with seismogenic layer 2A [Talwani et al., 1971; Atwater and Mudie, 1973; Fox and Opdyke, 1973; Bleil and Petersen, 1983; Arkani-Hamed, 1989; Gee and Kent, 1998]. However, compilations of intensity measurements from Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) drill-cores show that a purely basaltic magnetic source cannot explain the total amplitude of marine magnetic anomalies, especially away from the active zone of accretion [Dunlop and Prevot, 1982; Johnson and Pariso, 1993]. The lower crust may also contribute to magnetic anomalies, but estimates of the magnitude of this contribution range greatly from 20 to 75% [Blakely, 1976; Kidd, 1977; Dunlop and Prevot, 1982; Banerjee, 1984; Harrison, 1987; Pariso and Johnson, 1993a, 1993b; Tivey et al., 1998]. The magnetic contribution of diabase dikes and sills at intermediate crustal levels is uncertain as well; some studies deem them as significant contributors to magnetic anomalies [Dunlop and Prevot, 1982] while others assert that the natural remanent magnetization (NRM) of the intrusive middle crust is weak and unstable [Kent et al., 1978]. Finally, the large NRM intensity of serpentinite potentially makes the upper mantle as important as layer 2A in recording the geomagnetic field, if a sufficient volume of serpentinite exists and if it preserves a primary thermoremanent magnetization or a very early acquired chemical remanent magnetization [Dunlop and Prevot, 1982; Harrison, 1987].

The best evidence that lower crustal units formed at very low spreading rates can sustain a primary thermoremanent magnetization comes from ODP Hole 735B on the SWIR, where paleomagnetic measurements on a 1.5-km long gabbro section demonstrate an average magnetization of 2.5 A/m and a remarkably stable inclination of 71° [Dick et al., 2000]. Because the drill hole penetrates an uplifted crustal block adjacent to a large-offset fracture zone, it is unknown whether these magnetic properties are unique to a segment-end environment or if, in fact, a coherent remanent magnetization characterizes the lower crust across a full spreading segment. In this study, we use geophysical data collected between the Atlantis II and Novara fracture zones of 56°45′–58°40′E (Figure 1) on the SWIR to evaluate the source of magnetic anomalies in ultraslow spread ocean crust. These data extend to 25-Myr-old seafloor on both flanks of two ridge segments, constituting the most extensive off-axis survey of any portion of the SWIR. Comparable coverage of the MAR is available only for a single corridor north of the Kane fracture zone [Tucholke et al., 1997]. The new data provide an excellent opportunity to characterize crustal accretion and magnetic anomalies at ultraslow spreading ridges, and to make comparisons with the more extensively studied northern Mid-Atlantic Ridge.

2. Regional Setting

The segmentation patterns of the ~8000-km-long SWIR can be loosely subdivided into three provinces based on geophysical characteristics such as axial relief and spacing of transform offsets. The section that lies west of the Andrew Bain fracture zone at 30°E and the section that lies east of the Melville fracture zone at 61°E are both characterized by anomalously deep axial valleys, oblique spreading, poorly developed central magnetic anomalies and mantle Bouguer gravity anomalies, and unstable transform and nontransform offsets [Grindlay et al., 1992; Patriat et al., 1997; Rommevaux-Jestin et al., 1997; Cannat et al., 1999]. These properties suggest that in these two regions of the SWIR the mantle is cold, the lithosphere is strong and thick, and magma supply is substantially limited. In contrast, the central portion of the SWIR between 30°E and 61°E is generally characterized by long-lived transform and nontransform discontinuities (NTDs), ridge-perpendicular spreading, and well-developed central magnetic anomalies and mantle Bouguer anomalies [Rommevaux-Jestin et al., 1997; Cannat et al., 1999]. The study area between the Atlantis II (57°E) and Novara (58°30′E) fracture zones falls within the section of the SWIR that appears to be similar to much of the central and northern MAR.

The geophysical data in this study encompass two ridge segments that lie between the Atlantis II and Novara fracture zones (Figure 1). To simplify the discussion of each ridge segment, we devise a naming convention that identifies a segment by the first initials of the nearest transform offsets and the position of the segment between those offsets, following similar nomenclature used for the MAR [e.g., Detrick et al., 1995; Thibaud et al., 1998]. In this convention, the western segment in the study area is AN-1 and the eastern segment is AN-2 (Figure 1c). For comparison, segments AN-1 and AN-2 are referred to as segments 19 and 18, respectively, by Cannat et al. [1999] and segments S2 and S3, respectively, by Rommevaux-Jestin et al. [1997]. Similarly, the segment east of AN-2 is referred to as NM-1; this segment is referred to as segment 17 by Cannat et al. [1999] and segment S4 by Rommevaux-Jestin et al. [1997].

The Atlantis II fracture zone formed at anomaly 24 time (56 Ma) [Patriat and Segoufin, 1988], when the Rodriguez Triple Junction marking the eastern terminus of the SWIR propagated eastward [Fisher and Sclater, 1983]. All three discontinuities in the study area offset the SWIR axis in a left-lateral sense. The north–south striking Atlantis II and Novara fracture zones are 200 and 45 km long, respectively, representing age offsets of ~25 and ~7 Myr. A 10–15 km wide NTD offsets the two intervening ridge segments. Satellite gravity maps [Sandwell and Smith, 1997] show that the off-axis traces of the three discontinuities are marked by significant and continuous lows that extend north and south to the boundaries of the propagation wake of the Rodriguez Triple Junction. This is strong evidence that both fracture zones and the NTD have existed since the initiation of the SWIR plate boundary.

The Atlantis II fracture zone and Atlantis Bank, the site of Hole 735B, flanking the fracture zone’s eastern wall have been the focus of numerous studies, including a site survey cruise (RC2709) in 1987 in preparation for drilling [Dick et al., 1991], two ODP legs (118 and 176) [Dick et al., 1991, 2000], a geophysical survey of the SWIR axis between 55°E and 70°E [Mendel et al., 1997; Patriat et al., 1997], a rock coring and remotely operated vehicle
Figure 1. (a) Multibeam bathymetry data gridded at 0.15 km resolution, contoured at 0.5 km (black and white version), and artificially illuminated from the north (color version). The Atlantis II and Novara fracture zones form the western and eastern margins of the study area, respectively. Filled star marks location of ODP Hole 735B on Atlantis Bank. (b) Regional location map of survey area (filled star) on the SWIR. (c) Sketch of plate boundary configuration in the study area. AN-1 denotes the western segment and AN-2 denotes the eastern segment between the Atlantis II and Novara fracture zones, respectively. NM-1 denotes the westernmost segment between the Novara and Melville fracture zones. See color version of this figure at back of this issue.
expedition [Allerton and Tivey, 2001], a seismic refraction and reflection experiment [Muller et al., 1997, 2000], and two Japan Marine Science and Technology Center (JAMSTEC) cruises. Data from these latter two expeditions form the basis of this study. Atlantis Bank, a plutonic complex consisting of gabbro and serpentinitized peridotite, formed between 9.5 and 13 Ma [Dick et al., 1991] at the intersection of the Atlantis II transform fault and segment AN-1. Atlantis Bank exhibits characteristics that are typical of seafloor adjacent to active transform faults, in inside-corner settings, including anomalous elevation, positive residual gravity anomaly, and the absence of basaltic upper crust [e.g., Karson and Dick, 1983; Severinghaus and Macdonald, 1988; Tucholke and Lin, 1994].

3. Magnetic Analysis

3.1. Data Collection and Processing

[6] The majority of the data used in this study derive from two JAMSTEC cruises to the Atlantis II fracture zone region of the SWIR, one on board R/V Yokosuka in October–November 1998 and one on board R/V Kairei in September 2000. Both cruises were principally designed for submersible operations on Atlantis Bank with supplementary geophysical programs that collected multibeam bathymetry, underway magnetics and gravity, and continuous GPS navigation. The surveying program on the first cruise encompassed nearly 35,000 km² of seafloor; track lines were oriented north–south and spaced 1.5–7 km apart (solid lines, Figure 2a). Seafloor bathymetry was collected with the HS-10 multibeam surveying system manufactured by Furuno Electronics. This system generates depth values resulting from each measurement the regional magnetic field predicted by the best available International Geomagnetic Reference Field [International Association of Geomagnetism and Aeronomy (IAGA), 2000]. Magnetic data from the Yokosuka cruise contained significant high-frequency noise, and every track line anomaly contained spikes of up to 250 nT. Application of a median filter with a window size of nine samples removed all of the spurious data without degrading the anomaly resolution. Data quality was higher on the Kairei cruise. As with the bathymetry data, magnetic data from cruise RC2709 supplemented data from the two recent cruises to produce the final magnetic data set. After applying a constant shift of 43 nT to the RC2709 data, the root-mean-square error for 409 magnetic anomaly crossover points was 33 nT. Much of this error likely derives from daily secular variation, from large gradients in the magnetic field associated with the extreme and shallow topography near Atlantis Bank, and from navigational errors.

[11] To correct the magnetic anomaly data for the effects of topography, which varies by 6 km over the study area, and for phase shift due to latitude, a gridded version of the magnetic anomalies was inverted for source magnetization using the method described by Parker and Huestis [1974] and extended for grid analyses by Macdonald et al. [1980]. The assumptions required by this iterative, Fourier-based inversion procedure include: a constant thickness magnetic source layer that follows the seafloor topography and is uniformly magnetized with depth; a direction of magnetization consistent with the geocentric dipole; and filled and periodic grids of bathymetry and magnetic field anomaly. Specific parameters used for the inversion included a source layer thickness of 0.5 km, a geocentric magnetic inclination of −51.1°, and 1 km resolution grids of bathymetry and magnetic field anomaly. To stabilize the inversion, a band-pass filter tapered short wavelengths between 3.5 and 7 km and long wavelengths between 50 and 100 km.

[12] A common step in finalizing crustal magnetization estimates is adding to the inversion solution the annihilator, a magnetization distribution for the assumed magnetic source layer that produces no external magnetic field at the observation level. The addition of any multiple of the annihilator produces a nearly constant shift of magnetization amplitudes and is typically utilized to balance the reversal amplitudes of the main anomalies on the flanks of the ridge axis. Our final solution, shown in Figure 3a, did not require the addition of any annihilator.

[13] One factor omitted in the inversion scheme is the presence of sediment. Neglecting a sediment section of significant thickness means that the depth to the top of the magnetic source layer is greater than assumed in the inversion calculation, causing magnetization amplitudes to be underestimated. Although sediment accumulates in isolated ponds within the study area [Dick et al., 1991; Muller et al., 2000], we do not include a sediment layer in the inversion because the thickness of those deposits is unknown. This is likely to have little effect in the resulting
magnetization structure, however, because the effect of ignoring ~300 m of sediment is estimated to be <1 A/m in the final result.

[14] To determine the history of plate spreading in the study area, the magnetization solution (Figure 3a) was sampled along the ship track lines and the peak amplitude of each seafloor spreading anomaly was identified (Figure 3b). Each peak was assigned an age based on the middle of the corresponding polarity epoch of the geomagnetic time-scale of Cande and Kent [1995]. Because of the very slow spreading rates in this area, identification of normal and reversed polarity chronos was straightforward for chronos of long duration such as 2An, 5n, 5AC-ADn, and 6n, but was more difficult for chronos of shorter time span. Forward models of crustal magnetization profiles (Figure 4) and solutions from poles of rotation for the SWIR [Patriat and Segoufin, 1988] confirmed the isochron identifications and their consistency on each flank. Between the two fracture zones, the magnetic coverage extends to anomaly 8n (26 Ma) north of the ridge axis and to anomaly 7n south of the axis (25 Ma) (Figure 3b).

3.2. Spreading Rate Results

[15] Both the magnetic anomaly (Figure 2b) and crustal magnetization (Figure 3b) profiles show well-developed magnetic lineations that are orthogonal to the north–south spreading direction. Spreading rates are determined by fitting a line, in a least squares sense, to crustal age versus distance measurements (Figure 5), where the distance corresponds to the distance in the spreading direction from the SWIR axis and the age corresponds to the peak age for each anomaly. Figure 5 shows a long history of steady spreading on each flank but significantly asymmetric crustal accretion at both segments. The mean northern spreading rate is 5.5 km/Myr at AN-1 (Figure 5a) and 5.7 km/Myr at AN-2 (Figure 5b), while the mean southern spreading rate is 8.5 km/Myr at both segments (Figures 5c and 5d) (see also Table 1). The magnitude of the spreading asymmetry, 21%, is defined as the difference between the mean half rates (3 km/Myr) divided by the full spreading rate (14 km/Myr). Spreading rate magnitudes agree with those reported by Dick et al. [1991] from cruise RC2709 when the previous rates are corrected for refinements of the geomagnetic timescale, which lowered global spreading rate estimations by nearly 10% [DeMets et al., 1994; Cande and Kent, 1995].

[16] Changes in slope between individual anomalies in Figures 5a–5d indicate variations in crustal accretion that occur within an individual ridge segment. In contrast, changes in slope in a plot of total distance (distance between conjugate anomalies on both flanks) versus age indicate larger-scale changes in plate motion. Figures 5e–5f show that the total spreading rate for each segment has remained relatively uniform and virtually identical at the two segments since 20 Ma (see also Table 2). The least squares fit yields a total opening rate of 14.0 km/Myr at both segments. These results agree with spreading rate calculations by Chu and Gordon [1999] for this section of the SWIR.

4. Tectonic Results

4.1. Present-Day Segmentation

[17] High-resolution bathymetric data allow a detailed examination of the seafloor expression of present-day crustal accretion. Figure 6 illustrates the morphology and segmentation structure of segments AN-1 and AN-2. The neovolcanic zone of each segment was defined by assuming that bathymetric closed-contour highs with relief >50 m are constructional volcanic features. The western ridge segment, AN-1, trends east–northeast over a distance of 48 km. Eleven volcanoes define the primary locus of eruptive activity in the axial valley of AN-1 (Figure 6b). Of these, eight form an en echelon array that follows the curvilinear trend of the inner valley and three lie within an extinct basin that abuts the northern valley wall. The volcanoes range from 50 to 180 m in height and from 1 to 3 km in length. All of the identified volcanoes at AN-1 are confined to the eastern two thirds of the segment; no obvious volcanic edifices are observed between the western edge of the neovolcanic ridge and the nodal deep.

[18] The eastern ridge segment, AN-2, trends N5°E along its 75 km length. At the resolution of the multibeam bathymetry, terrain within the inner valley of AN-2 appears more volcanic in character than that within the inner valley of AN-1 (Figure 6). The neovolcanic zone is defined by 31 volcanoes with heights ranging between 50 and 230 m (Figure 6b). In contrast to segment AN-1, the volcanoes at segment AN-2 are higher, less faulted, and extend over the entire width and length of the inner valley. A distinct change in volcano size and aspect ratio occurs near the segment midpoint of AN-2 (58°E); west of the segment midpoint the seamounts form a quasi-continuous ridge at the center of the axial valley, whereas east of the segment midpoint the seamounts are smaller, more circular, and appear to be randomly distributed. Along both segments, the neovolcanic zones defined by the volcanic ridges align with the peak of the Brunhes normal polarity epoch (Figure 6b).

[19] The NTD between segments AN-1 and AN-2 spans a distance of 9 km along axis and 12.5 km across axis.
defining a diffuse transfer zone between the two segments. Seafloor depths within the NTD average 0.5 km deeper than adjacent seafloor at either segment. The rift valley walls of each segment terminate at the NTD except for the northern bounding fault of AN-2, which overshoots the offset and penetrates the crust created at segment AN-1 (Figure 6). Shallow-level volcanism appears to be robust at the segment ends bordering the NTD; the volcanoes within the inner valley of each segment extend all the way to the NTD, and three volcanoes lie within the shallow valley of the transfer zone itself.

[20] The rift valleys of segments AN-1 and AN-2 form hourglass shapes in plan view, and seafloor depths average 1–1.5 km shallower near segment centers relative to segment ends (Figure 6). Evidence of large-scale mass wasting is observed on the southern boundary wall of AN-1 at 57°15′E, where concave bathymetry contours define a 10-km wide scarp interpreted as a landslide headwall. At the base of the inferred AN-1 slump-scar, the slightly bulging topography of the valley floor may signal the presence of a debris field that extends across the entire width of the inner valley. A debris field of significant volume may explain the absence of any obvious volcanoes within the western third of the segment. Scallop-shaped bathymetry contours suggestive of an additional landslide are also observed on the southern boundary wall of AN-2 at 57°45′E. Rather than indicating an intrinsic property of the axial valley, the hourglass shape of the two rift valleys may result from

Figure 4. Representative magnetization profiles and corresponding synthetic models for the (a) southern and (b) northern flanks of the survey area. Models are created by generating a magnetization square wave at a constant spreading rate, assuming a 0.5 km thick source layer with a constant magnetization of ±6 A/m. The depth to the top of the source layer was chosen as the mean depth along each profile. To simulate a finite zone of emplacement, the square waves are smoothed with a Gaussian distribution of specified standard deviation. A standard deviation of 2 km provides a good visual match to the data. Shaded lines connect the identified chron on the data with the predicted chron on the model. “5ACADn” denotes the combination of anomalies 5CA and 5ADn, which merge together at these low-spread rates and “5ACADr” is the reversed period following 5ACADn.
Figure 5. Magnetic anomaly distance versus age from the axis of spreading for the (a) north flank of AN-1, (b) north flank of AN-2, (c) south flank of AN-1, (d) south flank of AN-2, (e) both flanks of AN-1, and (f) both flanks of AN-2. The thick shaded line in each panel is the line that best fits the age-distance data in a least squares sense. The slope of this line is the quoted spreading rate. The correlation coefficient is greater than 0.997 for all line fits. Spreading rate data extend to longer times in Figures 5c and 5d due to better resolution on the south flank than on the north flank. For age uncertainties of 0.25 Myr, spreading half rates have uncertainties of ±0.25 km/Myr. The anomalously large accretion rates at segment AN-2 (Figures 5d and 5f) between anomalies 2An and 3r result from crustal deformation associated with a short-lived propagating rift. Labeled chrons correspond to the /C243 Myr intervals used for the calculation. Tables 1 and 2 list interval spreading rates for each panel. Note that the range of the vertical axis differs for each pair of panels.
massive slope failure near their endpoints, effectively widening the valleys in plan view.

### 4.2. Paleosegmentation

[21] The off-axis trace of the NTD between segments AN-1 and AN-2 defines the tectonic evolution of these segments during the last 25 Myr. Using a combined analysis of seafloor morphology and magnetic isochron trends, the off-axis expression of the NTD was traced to the limits of bathymetric coverage on each flank (Figure 7). Between 25 and 12.5 Ma, the discontinuity consisted of a 20–30 km wide (~2.4 Myr) zone characterized by deep seafloor (Figure 1) and obliquely trending magnetic isochrons (Figure 3b). Beginning 12.5 Ma, the discontinuity zone narrowed to its present-day width of 10–15 km (~1.2 Myr). At the same time, the trace of the discontinuity in satellite gravity data becomes undetectable, indicating a diminished effect of the discontinuity within the upper mantle.

[22] A secondary discontinuity is observed in the off-axis bathymetry of segment AN-2 as V-shaped fabric that is symmetric about the ridge axis (Figure 1). This feature is interpreted as the trace of a propagating rift that initiated at the western end of segment AN-2 at anomaly 3n-old time (5.5 Ma) and propagated eastward until it terminated near the segment midpoint at anomaly 2An time (3 Ma) (Figure 3b). Its seafloor expression consists of a series of ridge-parallel, elongate basins that link along strike and offset the magnetic anomalies between those chronos. The anomalously high accretion rate for segment AN-2 between anomalies 2An and 3An (Tables 1 and 2) is a direct consequence of crustal deformation associated with the propagator. This inferred propagating rift differs from the abundant propagators observed at the MAR [e.g., Brozena and White, 1990] because it is wholly confined to one segment and it affects a relatively small volume of crust in its immediate vicinity.

### 4.3. Off-Axis Morphology

[23] Seafloor fabric on the southern flank of segment AN-2 is dominated by topographic highs within the inside-corner corridor adjacent to the Atlantis II fracture zone (Figure 1). With the exception of Atlantis Bank, the inside-corner massifs are quasi-circular in plan view, varying in diameter from 9 to 25 km and in cross-axis spacing from 2 to 10 km. The massifs rise from 0.7 to 1.5 km above the surrounding seafloor. Atlantis Bank differs from the other basement highs in the inside-corner corridor in both its shape and size; it is elongated in the direction of spreading, implying a long formation time, and it rises 3 km above the surrounding seafloor to its shallowest point at just 0.7 km water depth.

[24] Emanating from the eastern edges of the inside-corner highs are several narrow, curvilinear ridges that extend across AN-1 and terminate at the paleotrace of the NTD (Figure 1). On young seafloor where sediment cover is sparse, these ridges rise nearly as high as the adjacent inside-corner highs and are separated by small intervening basins. On older seafloor, the ridges are increasingly buried by sediment, making the basins appear larger. Aside from the large faults associated with the inside-corner massifs and these large ridges, the remainder of the seafloor lineaments at AN-1 are short, ridge-parallel and obliquely trending features that probably consist of small faults and relict volcanic ridges (Figure 7).

[25] Seafloor on the southern flank of segment AN-2 is dominated by wide-spread rift mountain terrain (Figure 1). Between the spreading axis and ~32°30’S, a large highland rises 2 km above the surrounding seafloor. The highland extends for ~60 km along axis, from the trace of the NTD to the deeper seafloor of the outside corner of the Novara fracture zone. Abyssal hill fabric is more pervasive south of segment AN-2 than south of AN-1, and a greater proportion of those hills parallel the present-day ridge axis (Figure 7).

[26] Seafloor texture north of the ridge axis appears smoother than to the south. Flanking both segments, seafloor lineaments lie parallel to the ridge axis, are regularly spaced, and commonly reach 20 km in length (Figure 7). The structure of the inside-corner massifs near the Novara fracture zone differs significantly from those near the Atlantis II fracture zone. In contrast to the discrete, quasi-circular domes within the Atlantis II inside-corner corridor, elevated topography within the Novara inside-corner corri-

### Table 1. Average Half-Spreading Rates for Each Ridge Segment

<table>
<thead>
<tr>
<th>Anomaly Interval</th>
<th>Time Interval, Myr</th>
<th>AN-1, North, km/Myr</th>
<th>AN-1, South, km/Myr</th>
<th>AN-2, North, km/Myr</th>
<th>AN-2, South, km/Myr</th>
</tr>
</thead>
<tbody>
<tr>
<td>1n–2An</td>
<td>0–2.9</td>
<td>5.6</td>
<td>8.8</td>
<td>6.0</td>
<td>8.1</td>
</tr>
<tr>
<td>2An–3An</td>
<td>2.9–6.2</td>
<td>5.5</td>
<td>8.2</td>
<td>6.3</td>
<td>11.1</td>
</tr>
<tr>
<td>3An–4Ar</td>
<td>6.2–9.4</td>
<td>7.3</td>
<td>7.0</td>
<td>4.5</td>
<td>6.2</td>
</tr>
<tr>
<td>4Ar–5Ar</td>
<td>9.4–12.7</td>
<td>5.6</td>
<td>10.6</td>
<td>6.5</td>
<td>8.4</td>
</tr>
<tr>
<td>5Ar–5ACADn</td>
<td>12.7–15.5</td>
<td>4.2</td>
<td>9.1</td>
<td>5.2</td>
<td>9.2</td>
</tr>
<tr>
<td>5ACADn–5Dr</td>
<td>15.5–17.9</td>
<td>4.0</td>
<td>8.4</td>
<td>2.8</td>
<td>7.4</td>
</tr>
<tr>
<td>5Dr–6An</td>
<td>17.9–20.9</td>
<td>...</td>
<td>8.2</td>
<td>...</td>
<td>8.7</td>
</tr>
<tr>
<td>6An–6Dr</td>
<td>20.9–24.3</td>
<td>...</td>
<td>5.3</td>
<td>...</td>
<td>7.8</td>
</tr>
</tbody>
</table>

*a*For age uncertainties of 0.25 Myr, spreading half rates have uncertainties of ±0.25 km/Myr.

*b*The calculation extends only to ~18 Ma due to a lack of isochron identifications at older crustal ages on the north flank.

*c*Anomaly 4r (8.3 Ma) is used for north flank calculations.

*d*5ACADn” refers to the reversed period following chron 5An, which merge together at these low spreading rates.
Figures 1) and little off-axis data existed for the Novara fracture zone. Contrasting morphology at the two inside-corner regions suggests different magnitudes of extension and patterns of stress distribution adjacent to the fracture zones.

4.4. Fracture Zone Structure

[27] A major result of the field program in this region is the complete mapping out to 25 Myr of two fracture zones formed at ultraslow spreading rates. While most of the structure of the active Atlantis II transform was surveyed during RC2709 [Dick et al., 1991], its inactive trace on the northern flank was mapped only to anomaly 5n (~10 Ma) and little off-axis data existed for the Novara fracture zone. The new bathymetry data show that the 10° counterclockwise change in plate motion described by Dick et al. [1991] is also well recorded in the trace of the Novara fracture zone (Figure 1). However, the expression of the change in plate motion differs between the fracture zones. Whereas the inactive trace of the Atlantis II fracture zone changes orientation from northeast to due north–south over a distance of ~100 km, the inactive trace of the Novara fracture zone bends sharply at the time of the reorientation (58°25′E/32°45′S) (Figure 1). The abrupt change in the trend of the shorter Novara fracture zone thus allows the timing of the reorientation event to be more precisely determined than the 17–20 Ma age determined previously using just the trend of the Atlantis II fracture zone [Dick et al., 1991]. Magnetic isochron identifications and fracture zone morphology pinpoint the time of the plate motion change at anomaly 6n (19.5 Ma) (Figures 1 and 3b). This timing is consistent with tectonic reconstructions of the Indian Ocean basin, which indicate that the last major change in spreading direction occurred between 20 and 10 Ma [Patriat and Segoufin, 1988]. The highly detailed data of this study suggest that the reorientation was rapid and occurred at the beginning of this interval.

[28] A striking feature discovered in the new bathymetry data is a region of highly lineated terrain that formed within an inside-corner corridor at segment NM-1 (Figures 1 and 7). The east–west trending lineations consist of narrow (<1 km wide) ridges with crests that are studded with circular, closed-contour highs interpreted as volcanoes. The ridges are regularly spaced and at least 20 km long, and they average several hundred meters in relief. Following the reorientation of the spreading direction at 19.5 Ma, the formation of highly lineated topography ceased and elevated topography; the largest magnetization amplitudes occur where topography is elevated, and the smallest values occur where seafloor topography is depressed (Figure 8c). At segment AN-2, coincident peaks in seafloor topography and crustal magnetization occur at the segment center, while at segment AN-1, these peaks are aligned with the volcanic ridge that dominates the eastern half of the axial valley.

5. Magnetic Results

5.1. Axial Magnetization

[29] The most prominent feature in the crustal magnetization map is the large-amplitude Brunhes anomaly centered over the present-day ridge axes of segments AN-1 and AN-2 (Figure 8a). Axial magnetization values are greater than anywhere else in the study area, with a peak to trough amplitude of 25 A/m (assuming a 0.5 km thick source layer). Along both segments AN-1 and AN-2, the Brunhes anomaly consists of two closed-contour highs, or magnetic subsegments (Figure 8a). At AN-1, the eastern magnetic subsegment dominates the axial magnetization and spatially coincides with the pronounced volcanic ridge. The magnetic subsegments are more equal in length and magnitude at segment AN-2. The general correspondence between the morphologic axis defined by the valley floor volcanoes and the magnetic axis defined by the peak of the Brunhes anomaly is commonly observed at the Mid-Atlantic Ridge [Grindlay et al., 1992; Tivey and Tucholke, 1998].

[30] An along-axis profile of the peak of the Brunhes anomaly highlights segment-scale trends in axial magnetization (Figure 8b). The dominant trend is a direct correlation between magnetization amplitude and seafloor topography; the largest magnetization amplitudes occur where topography is elevated, and the smallest values occur where seafloor topography is depressed (Figure 8c). At segment AN-2, coincident peaks in seafloor topography and crustal magnetization occur at the segment center, while at segment AN-1, these peaks are aligned with the volcanic ridge that dominates the eastern half of the axial valley.

5.2. Off-Axis Magnetization

[31] The extensive data set collected in this region of the SWIR allows a detailed study of temporal variations in crustal magnetization over the past 25 Myr. The most obvious difference between the magnetization of zero age and older crust is the rapid decay in amplitude within the first few million years (Figure 3a). At the MAR, a rapid reduction in magnetization intensity is observed both in magnetization inversion solutions [Pockalny et al., 1995; Pariso et al., 1996] and in NRM values of dredged samples [Johnson and Atwater, 1977], and this is commonly attributed to the alteration of titanomagnetite to less-magnetic titanomaghemite in the magnetic source rocks [Irving, 1970; Johnson and Atwater, 1977]. In the SWIR area, the decay from high values on axis (15–25 A/m) to much lower values (~5 A/m) off axis is complete by anomaly 2An time (~3 Ma) on both flanks (Figure 3a). This result seems to concur with the documented rapid reduction of the magnetic signal in young crust [e.g., Gee and Kent, 1994]. However, Tivey and Tucholke [1998] suggest that trends between the Brunhes anomaly and anomaly 5n do not accurately record true variations in magnetization amplitude at low spreading rates because the frequent polarity reversals produce narrow spreading stripes that destructively interfere when observed at the sea surface.

[32] If off-axis magnetization amplitudes are investigated using only chron of long duration (1n, 5n, and 6n), crustal magnetization amplitudes decay at a more modest rate than implied by the large loss of amplitude between the axis and anomaly 2An (Figure 9). Further, the rate of intensity decay decreases with time, with a loss of intensity in young crust (0–10 Ma) that exceeds that of older crust (10–20 Ma) by an order of magnitude (Figure 9). Twenty million years after accretion, the total reduction in crustal magnetization intensity is 70–80%, or about 0.6 (A/m)/Myr, at both segments. Because geomagnetic intensity fluctuations on timescales...
Figure 6. (a) Detailed bathymetry of axial valleys of segments AN-1 and AN-2. Grid resolution is 0.15 km and contour interval is 0.3 km (color version) and 0.2 km (black and white version). Illumination is from the north (color version). (b) Interpretive drawing depicting the major features of the inner valley floor. IC denotes inside-corner tectonic setting, OC denotes outside-corner tectonic setting, and LS denotes location of inferred landslide headwall. Segment boundaries are as in Figure 3a. See color version of this figure at back of this issue.
less than ~0.5 Myr cannot be resolved due to the very low spreading rates and because the latitude of the ridge axis has changed by only 5° since 20 Ma [Patriat and Segoufin, 1988], the significant decrease in magnetization intensity during this time period likely derives from local effects such as low-temperature alteration of the upper crust [e.g., Zhou et al., 2001], rather than from changes in geomagnetic intensity or geographic position. 

[33] Variations in magnetization amplitude along isochron are superimposed on cross-axis changes. To investigate segment-scale variations in magnetization amplitude off axis, peak magnetization amplitudes along individual

Figure 7. Line-drawing of major tectonic features in the study area. Segment boundaries are as in Figure 3a. Thin shaded lines denote lineaments identified in seafloor morphology. Lightly shaded polygons denote topography that is elevated above 2.75 km water depth and darkly shaded polygons denote topography that is deeper than 4.75 km water depth. Axial valley volcanoes identified in Figure 6 are denoted by filled polygons. PM emphasizes the change in regional plate motion at 19.5 Ma. LT marks the region of highly lineated terrain observed at segment NM-1 between the southern limit of bathymetry coverage and 19.5 Ma. Filled star marks location of ODP Hole 735B on Atlantis Bank.
Isochrons are plotted relative to their mean (Figure 10). The results for each isochron are normalized to unit magnetization to remove the larger-amplitude decay with age, which would mask smaller-amplitude, along-isochron variations. Figure 10a shows that the axial magnetization pattern of relatively low magnetization at segment ends and relatively high magnetization at segment midpoints is present for some normal polarity chron (5n, 6n, and 7n on the south flank of AN-2 and 5n on the north flank of AN-1), but is not observed elsewhere. Rather, segment ends predominantly tend toward higher magnetization during epochs of normal polarity, particularly on the north flank of AN-2 adjacent to the NTD and on the south flank of AN-1 bordering the Atlantis II fracture zone (Figure 10a). In contrast, reversed polarity isochrons behave more like the axial anomaly, with more intense magnetization at the segment midpoints than at the segment offsets (Figure 10b). The apparent discrepancy between normal and reversed polarity magnetization amplitudes off axis is resolved if the isochronal variations are interpreted in terms

Figure 8. (a) Detail of crustal magnetization solution at ridge axis. Axial valley volcanoes identified in Figure 6 are denoted by filled polygons. Contour interval is 2.5 A/m. Color scale is divided into quartiles. (b) Maximum magnetization of the Brunhes anomaly along axis, measured at each axis-crossing ship track (crosses). The mean magnetization for AN-1 is 16.2 A/m and the mean value for AN-2 is 11.1 A/m. (c) Seafloor depth at the locations of the Brunhes anomaly maxima in Figure 8b. Note the direct correlation between topography and magnetization amplitude. VE = 8.
of how positive these amplitudes are. That is, both normal and reversed polarity anomalies exhibit systematically more positive magnetization at segment ends than at segment centers.

6. Discussion
6.1. Prolonged Asymmetric Spreading

A significant result of this study is the documentation of a systematic and prolonged spreading asymmetry at the SWIR between the Atlantis II and Novara fracture zones. While the magnitude of the spreading asymmetry is similar to that observed at other oceanic spreading centers [Rea, 1981; Carbotte et al., 1991; Cormier and Macdonald, 1994; Sempère et al., 1995; Weiland et al., 1995], the stability and longevity of the SWIR spreading asymmetry is distinctive. Along several segments of the southern MAR, for example, the magnitude of spreading asymmetry varies in both intensity and sign from one anomaly interval to the next and between adjacent ridge segments [Carbotte et al., 1991]. These asymmetries typically diminish, however, when spreading half rates are averaged over periods of several millions of years and greater [Stein et al., 1977; Carbotte et al., 1991; Weiland et al., 1995]. In contrast, this SWIR study area exhibits a 21% asymmetry in spreading half rates for at least 25 Myr at both AN-1 and AN-2 segments.

Long-term asymmetric spreading is probably driven by deep-seated processes within the asthenosphere, such as the migration of a ridge with respect to a fixed or slowly moving mantle [Stein et al., 1977] or the propagation of a ridge toward one or more mantle plumes [Small, 1995; Müller et al., 1998]. Both mechanisms explain observations on scales approaching the size of an ocean basin [e.g., Müller et al., 1998]. If the prolonged asymmetric spreading observed in the study area is driven by a dynamic mechanism, the spreading asymmetry should extend over a portion of the SWIR axis that is greater than the 150-km-long section encompassed by the Atlantis II and Novara fracture zones. Although spreading half rates calculated using a compilation of isochrons for the southwest Indian Ocean [Patriat and Segoufin, 1988] suggest significant asymmetry over more than 1000 km of the eastern SWIR, recent high-resolution magnetic studies show otherwise. Both west of the Atlantis II fracture zone and east of the Melville fracture zone, no evidence exists for asymmetric spreading over the past 10 Myr [Bralee et al., 2002; V. Mendel, personal communication, 2002].

These observations do not rule out a regional driving force for asymmetric spreading. Profiles of axial bathymetry, mantle Bouguer gravity anomaly, and central magnetic anomaly exhibit fundamental differences west of the Atlantis II fracture zone and east of the Melville fracture zone [Rommevaux-Jestin et al., 1997]. These observations, combined with the absence of asymmetric spreading in those same regions, strongly suggest that the SWIR between the Atlantis II and Melville fracture zones (including segments AN-1 and AN-2) constitutes an independent and distinct section of the plate boundary that may behave as a long-lived tectonic corridor. Distinct tectonic corridors are similarly observed along the southern MAR and other mid-oceanic spreading centers, and appear to be delimited by large-offset fracture zones [Kane and Hayes, 1992; Hayes and Kane, 1994]. Thus asymmetric spreading within the ∼400 km wide corridor between the large-offset Atlantis II and Melville fracture zones may be sustained by dynamic forces that

![Figure 9. Peak magnetization amplitude as a function of age. Each data point represents the mean of magnetization maxima for tracks at paleo-segment midpoints of AN-1 and AN-2. Error bars are one standard deviation on either side of the mean. Dashed lines connect anomalies 1n, 5n, and 6n to illustrate long-term decay of crustal magnetization. Nominal distance axis is obtained by assuming constant half-spreading rates of 8.5 and 5.5 km/Myr for the south and north flanks, respectively. Filled arrows centered on 0 km distance denote spreading direction with lengths scaled by the ratio of the half-spreading rates.](image)
operate on scales $<10^3$ km. A thorough test of this distinct corridor hypothesis requires detailed off-axis surveying both within and adjacent to the proposed Atlantis II-Melville corridor to definitively determine spreading rates through time.

### 6.2. Source of Axial Crustal Magnetization

[37] The magnetic inversion solution (Figure 8a) shows that the central anomaly diminishes from high values at the shallow neovolcanic zones of segments AN-1 and AN-2 to low values at the deeper segment offsets. Several factors could explain this direct correlation between seafloor topography and crustal magnetization along axis. Perhaps the simplest explanation is a change in the thickness of the magnetic source layer along strike. If the extrusive layer dominates the magnitude of the axial magnetic signal [e.g., Tivey, 1996], the observations require an extrusive section that is relatively thick where magnetization amplitudes are elevated and relatively thin where amplitudes are low. While there is no detailed information on basalt thickness along axis, ridge-parallel seismic refraction data collected on 12-Myr-old crust at segment AN-1 suggest that the whole of layer 2 maintains a constant thickness along axis [Muller et al., 1997, 2000]. If the upper crust does not change thickness along axis, then the observed variation in axial magnetization may derive instead from variations in the thickness of the lower crust.

[38] To evaluate the relative contributions of basaltic upper crust and gabbroic lower crust to the observed axial magnetization, we devised two-dimensional forward models that predict the magnetic anomaly produced by a variable thickness source layer of constant magnetization. The mean of the peak axial magnetization at segment AN-1, ~15 A/m (Figure 8a), was used for the basaltic source; the mean magnetization of gabbros from Hole 735B, 2.5 A/m [Dick et al., 2000], was used for the gabbroic source. Both the variable thickness upper crustal (Figure 11a) and lower crustal (Figure 11b) sources produced a good match between the modeled and observed magnetic anomaly (Figure 11c). However, the basaltic source model (Figure 11a) required a high-magnetization extrusive layer up to 2.1 km thick. Because seismic models in this area show that the entire upper crust (both the extrusive and intrusive components) is only 2–2.5 km thick [Muller et al., 1997, 2000], the extrusive layer thickness predicted by magnetic modeling is inconsistent with the imaged crustal structure at segment AN-1. In contrast, the crustal thickness variation required for the gabbroic source model (Figure 11b) agrees well, within error, with both seismic models [Muller et al., 2000] and crustal thickness predictions from sea-surface gravity data [Hosford, 2001]. These results imply that on axis, the extrusive crust contributes a large but near-constant amplitude to the crustal magnetization, while the lower crust produces the variation in that signal.

In any magnetic modeling, a trade-off exists between source layer thickness and intrinsic magnetization due to magnetic mineralogy. Some constraints on intrinsic magnetization in the study area are provided indirectly by the chemistry of rift valley basalts from segment AN-1 between

![Figure 11. Along-axis profiles of the (a) extrusive upper crust, and the (b) intrusive lower crust required to fit the axial magnetic anomaly observed at segment AN-1. Constant source intensities of 15 and 2.5 A/m were used for the upper and lower crust, respectively. VE = 3. (c) The observed magnetic field anomaly of segment AN-1 (filled circles) and the calculated anomaly profiles for the upper (light shading) and lower (dark shading) crustal sources. Both predictions fit the observed anomaly amplitude to within a root-mean-squared difference of 14 nT.](image-url)

Figure 10. (opposite) Along-isochron variation in crustal magnetization for (a) normal and (b) reversed polarity chron's. Magnetization amplitudes for each isochron are plotted relative to the mean value and normalized to lie between +1 and −1. In both panels, shaded wiggle fill denotes weaker magnetization than isochron mean and black wiggle fill denotes stronger magnetization than isochron mean. Segment boundaries are as in Figure 3a. Ridge axis corresponds to 0 Myr. Filled star marks location of ODP Hole 735B on Atlantis Bank. Filled arrows centered on 0 Ma denote spreading direction with lengths scaled by the ratio of the half-spreading rates.
57°29′E and 57°36′E. These samples exhibit relatively constant iron and titanium contents [Robinson et al., 1996], suggesting that titanomagnetite contents may be relatively constant over that ~11 km section of the axial valley. Further, an empirical relationship between FeO content and NRM [Gee and Kent, 1998] yields predicted NRM values that agree remarkably well with axial magnetization amplitudes obtained by magnetic inversion. Taken together, these results suggest that the assumed thickness of the upper crust (0.5 km) is appropriate; that the extrusive layer maintains a uniform intrinsic magnetization along the rift valley of segment AN-1; and that variations in the thickness of the magnetic source layer, rather than magnetic mineralogy, control magnetization amplitudes along axis.

[40] The direct correlation between axial depth and magnetization amplitude observed in the study area is also observed at 15°E–25°E on the SWIR [N. R. Grindlay, personal communication, 2000] and at the ultraslow spreading Mohns Ridge [Géli et al., 1994]. This trend is opposite, however, to that observed at the MAR, where rift valley rock samples and numerical inversions of magnetic anomalies both show an inverse relationship between crustal magnetization and axial depth [Grindlay et al., 1992; Pariso et al., 1996; Weiland et al., 1996; Ravilly et al., 1998]. Because seismic models indicate that large-scale crustal structure along the rift valleys of the MAR, the SWIR, and the Mohns ridge are similar [e.g., Tolstoy et al., 1993; Klingelhofer et al., 2000; Müller et al., 2000], the contrasting axial magnetization relationship at ultraslow and slow spreading centers must be explained in some other manner.

[41] Two principal factors that may differ considerably between the two classes of ridges are thickness of the constituent layers and ambient lower crust and upper mantle temperatures. A difference in basalt layer thickness at segment ends may explain the difference in axial magnetization near segment offsets at the MAR and SWIR. Smith and Cann [1999] suggest that the upper crust at the MAR is constructed by lava fed from dikes that propagate for distances of tens of kilometers along axis, from the segment midpoint to the distal ends. Multichannel seismic images at 35°N on the MAR show a thicker extrusive crust near the fracture zone than near the segment midpoint [Husseyoeder et al., 2002], consistent with that model. At ultraslow spreading ridges, however, down-axis flow of melt within the shallow crust may be inhibited by a combination of relatively low melt volume, infrequent melt injections, and a thick lithospheric lid. Indeed, at segments AN-1 and AN-2, the seafloor appears distinctly nonvolcanic near the nodal deeps. While elevated magnetization at the ends of MAR segments may derive from a thicker extrusive pile, low magnetization at the ends of SWIR segments may simply derive from relatively thin basalt and gabbro sections.

[42] A difference in the ambient lower crust and upper mantle temperature may explain the contrasting axial magnetization at MAR and SWIR segment midpoints. At the MAR, where mantle upwelling is presumed to occur at the midpoints of individual segments [Lin et al., 1990], temperatures within the lower crust and upper mantle may be maintained above the Curie point for magnetite. Such elevated temperatures would prevent a coherent magnetic signal from locking in to the axial crust and thereby produce locally diminished crustal magnetization. This thermal demagnetization effect is the likely cause of low magnetization at the TAG hydrothermal mound on the MAR [Tivey et al., 1993]. In contrast, because magma chambers at the SWIR are probably ephemeral features, and because mantle melting ceases at greater depths at the SWIR relative to the MAR [Robinson et al., 2001], the lower crust and upper mantle may be colder at the midpoints of SWIR segments relative to the midpoints of MAR segments, thereby allowing relatively rapid cooling of newly emplaced crust and the preservation of thermoremanent magnetization. Elevated magnetization at the midpoints of SWIR segments may simply be the combined effect of a high-magnetization upper crust and a locally thick lower crust. Unlike at the MAR, then, the observed magnetization pattern at SWIR segments appears to be a function of first-order crustal thickness variations along axis.

6.3. Off-Axis Isochronal Trends in Crustal Magnetization

[43] As crustal sections are lifted out of the rift valley and transported off axis, the magnetic source layers are physically modified through fracturing and block rotation and chemically altered by hydration reactions. A significant change in the structure of the magnetic source layer occurs near segment offsets, where isochron-parallel magnetization profiles tend toward positive values regardless of the polarity of the Earth’s field at a prior time (Figure 10). This pattern, which is also observed along discontinuity traces at the MAR [Pockalny et al., 1995; Pariso et al., 1996; Tivey and Tucholke, 1998], strongly suggests the presence of an induced component of magnetization within the lower crust and/or upper mantle.

[44] Induced magnetization is a by-product of the serpentinization of the lower crust and upper mantle because hydrous alteration of olivine and pyroxene produces secondary magnetite that is highly susceptible to inducement by the present ambient field [Dunlop and Prevor, 1982; Gee et al., 1997]. Tivey and Tucholke [1998] derived an expression for the magnitude of induced magnetization that is based on a simple model in which the remanent magnetization component is invariant along axis. In this formulation, induced magnetization equals the sum of normal and reversed polarity amplitudes divided by two. Because of the dense magnetic isochron identifications in this study area, the calculation for induced magnetization is performed on gridded versions of the profiles shown in Figure 10. Where data coverage extends to the discontinuity traces, the result (Figure 12) indeed shows a concentration of inferred induced magnetization near the two transform offsets and within the zone of the NTD.

[45] Paleomagnetic studies on rocks from ODP Hole 735B provide constraints, in one area, on whether the induced magnetization component resides in the lower crust or in the upper mantle. The ratio of remanent to induced magnetization of a rock sample is quantified by the Koenigsberger ratio, Q. Nearly 700 samples from the 1.5-km-long drill-core exhibit Q values greater than one [Kikawa and Pariso, 1991; Dick et al., 1999], providing strong evidence that Hole 735B gabros record a primary remanent magnetization component and that any induced component resides at a deeper level. Because near-bottom studies show that serpentinite crops out extensively on the western wall
of Atlantis Bank [Dick et al., 2002], serpentinized peridotite is the likely carrier of an induced magnetization component in the study area.

6.4. Atlantis Bank

The conclusion that the lower crust may contribute significantly to marine magnetic anomalies is based on inference from forward magnetic modeling, seismic observations, and isochronal magnetization patterns. The coincidence of a dense magnetic survey and the 1.5-km-long drill hole at Atlantis Bank allows this idea to be tested by examining the thermal conditions under which the lower crust formed. A key issue in determining the significance of the contribution of the lower crust to marine magnetic anomalies is the shape of magnetic polarity boundaries within intrusive layers. In the traditional model of the

**Figure 12.** Inferred induced component of magnetization, defined as the sum of the normal and reversed polarity isochronal variations divided by two [Tivey and Tucholke, 1998]. The calculation was performed on gridded versions of the data presented in Figure 10. Note the tendency for the inferred induced component to align with the segment discontinuities, possibly indicating the presence of serpentinized lower crust and/or upper mantle. Segment boundaries are as in Figure 3a. Filled star marks location of ODP Hole 735B on Atlantis Bank.
magnetic structure of the lower oceanic crust, magnetic polarity boundaries dip shallowly away from the ridge axis along the Curie isotherm for magnetite as a result of slow, purely conductive cooling [Blakely, 1976; Cande and Kent, 1976; Kidd, 1977]. Polarity transitions occur over longer length scales and timescales in slowly cooled lower crust relative to quickly cooled upper crust, causing the magnetic contribution of the lower crust to sea-surface magnetic anomalies to be smaller than, or even out of phase with, the magnetic contribution of the overlying upper crust [Cande and Kent, 1976; Arkani-Hamed, 1989].

One recent study at Atlantis Bank corroborates this traditional view of the magnetic structure of the lower oceanic crust. After correcting for a 20° southward tilt to the entire complex, Allerton and Tivey [2001] obtained a dip angle for a reversal boundary at Atlantis Bank of only 5°, suggesting that Atlantis Bank gabbros cooled very slowly. However, this inference is rejected if the sloping boundary proposed by Allerton and Tivey [2001] is not an isotherm, but rather a contact between intrusions, or if postemplacement faulting rotated the polarity boundary from an initial high angle to a lower angle.

Indeed, abundant direct and indirect evidence exists from Hole 735B for rapid cooling of the Atlantis Bank gabbros, suggesting that magnetic polarity boundaries may be steep and that the 20° tilt correction of Allerton and Tivey [2001] does not restore the magnetic data to the paleohorizontal. First, the drilled section is magnetized with a remarkably stable inclination downhole [Kikawa and Pariso, 1991; Dick et al., 2000], indicating that the entire section cooled rapidly below the Curie point for magnetite, perhaps when the crustal section was very near the ridge axis [Pariso and Johnson, 1993a]. Second, alteration assemblages in the core formed under granulite-to-amphibolite-facies conditions [Stakes et al., 1991; Dick et al., 2000]. Fluid temperatures would only have been high enough to produce these facies when the crust was still within the active rifting zone. In addition to rapidly cooling the newly formed lower crust at Atlantis Bank, high-temperature fluid flow had the added effect of sealing the gabbro within the active rifting zone. In contrast to rapidly cooling from adjacent cold lithosphere across segment offsets and rapid uplift along large-throw faults at inside corners, Rapid cooling of the lower crust at segment midpoints may occur via hydrothermal circulation. Alternatively, the lower crust and upper mantle at segment midpoints may be intrinsically cold because of the long time span between magma injections and an excessively thick lithospheric lid. As the crust ages, hydrothermal alteration of magnetic minerals significantly attenuates the intrinsic magnetization of the upper crust, but its amplitude remains constant along strike (Figure 13c). In contrast, the remanent magnetization intensity of the lower crust remains unchanged from its axial value (Figure 13c). In off-axis settings, upper and lower crustal units contribute equally to the preservation of marine magnetic stripes.

Figure 13 summarizes the remanent magnetization structure of oceanic crust formed at ultraslow spreading mid-ocean ridges, based on this study. The observed magnetic signal is produced by two sources, a constant thickness upper crust and a variable thickness lower crust (Figure 13a). On axis, the extrusive crust quickly locks in the magnetic field due to rapid cooling upon emplacement. The magnetization intensity is high, but is invariant along axis if variations in iron and titanium content and thickness are small (Figure 13b). Heterogeneity in magnetization intensity along axis derives instead from thickness variations in the lower crust (Figure 13b). The magnetic field is quickly locked into the lower crust at segment ends due to cooling from adjacent cold lithosphere across segment offsets and rapid uplift along large-throw faults at inside corners. Rapid cooling of the lower crust at segment midpoints may occur via hydrothermal circulation. Alternatively, the lower crust and upper mantle at segment midpoints may be intrinsically cold because of the long time span between magma injections and an excessively thick lithospheric lid. As the crust ages, hydrothermal alteration of magnetic minerals significantly attenuates the intrinsic magnetization of the upper crust, but its amplitude remains constant along strike (Figure 13c). In contrast, the remanent magnetization intensity of the lower crust remains unchanged from its axial value (Figure 13c). In off-axis settings, upper and lower crustal units contribute equally to the preservation of marine magnetic stripes.

Summary

Tectonic and magnetic analyses suggest the following:

1. A 21% asymmetry in half-spreading rates is maintained for at least 25 Myr, with a half rate of 8.5 km/Myr on the Antarctic plate and 5.5 km/Myr on the African plate. This asymmetry may be unique to the ~400 km wide corridor between the large-offset Atlantis II and Melville fracture zones of the SWIR.

2. Crustal accretion is well organized even at very low spreading rates. Lineated marine magnetic anomalies are consistent and identifiable, even over a plutonic complex (Atlantis Bank) known to consist solely of gabbro and serpentinized peridotite.

3. A ~10° counterclockwise change in plate motion known previously to have occurred between 17 and 20 Ma is pinpointed by magnetic isochron data and fracture zone morphology at ~19.5 Ma.

4. Axial crustal magnetization is directly correlated with seafloor topography, with the largest amplitudes observed at shallow neovolcanic zones, and the smallest amplitudes observed at deep segment offsets. The observed axial magnetization is well explained by a two-layer magnetic source that consists of a constant thickness upper crust underlain by a variable thickness lower crust. In contrast to faster spreading ridges, the observed magnetization pattern at SWIR segments appears to be a function of first-order crustal thickness variations along axis.

5. Crustal magnetization decays by 70–80% between the ridge axis and 20-Myr-old crust. This decay occurs more rapidly during the first 10 Myr following emplacement than during the subsequent 10 Myr. At segment offsets, more positive magnetization off axis suggests the presence of an
induced component of magnetization. Large Koenigsberger ratios of drill hole gabbros suggest that this induced component probably resides at a deeper level, within serpentinized peridotites.

6. Qualitative features of the magnetic lineations combined with quantitative drill hole data from Hole 735B suggest that the lower crust cooled below the Curie point for magnetite while still within the active rifting zone. Both the basalt and gabbro layers thus cooled quickly enough to lock in a primary thermoremanent magnetization, and thus reflect the same magnetic age. The magnetic contributions of the upper and lower crust appear to be in phase, and both units contribute equally to sea-surface magnetic anomalies in off-axis settings.

7. Rift valley morphology appears similar to that at the MAR. elongated, en echelon volcanoes define the neovolcanic zones within the inner valleys at the segment midpoints, while circular and isolated volcanoes concentrate near the nodal deeps. The peak of the Brunhes anomaly coincides with the trend of the axial volcanic ridges, indicating a spatial and temporal association of the magnetic and morphologic axes.

[52] Acknowledgments. We thank the Japan Marine Science and Technology Center and the crews of Yokosuka and Kairei for helping to make the geophysics programs so successful. Discussions with Bobbie John, Greg Hirth, Brian Tucholke, Debbie Smith, and Dan Scheirer were helpful. Many thanks to Dave Carress for adding the HS-10 format to MB-System. The comments of Steven Cande, Tim Minshull, and an anonymous reviewer improved the manuscript. A.H. was supported by Joint Oceanographic Institutions Subcontract JSC1-00 to the Woods Hole Oceanographic Institution. Woods Hole Oceanographic Institution contribution 10527.

References


Carbotte, S., S. M. Welch, and K. C. Macdonald, Spreading rates, rift propagation, and fracture zone offset histories during the past 5 my on

Figure 13. Schematic illustration summarizing the conclusions from this study regarding the remanent magnetization structure of the upper and lower crust. (a) Ideal model for axial crust. The model incorporates a seafloor relief of 1 km, an upper crustal thickness of 1.5 km, and a lower crustal thickness of 1.5 km at the segment endpoints and 3 km at the segment midpoint. White line within the upper crustal layer marks the base of the extrusive crust, assumed to be 0.5 km thick. (b) Apparent magnetization expected along axis, defined as the magnetization due to variations in source layer thickness alone. The magnetization contribution from the upper crust (shaded line) is invariant along axis while that from the lower crust (solid line) produces the variation in observed amplitudes (dotted line). (c) Same as in Figure 13b but for along-isochron profiles off axis. An induced magnetization component is not shown as its magnitude and source remain speculative and we emphasize the remanent component that is intrinsic to the structure of the ocean crust at the time of accretion. The remanent magnetization of the lower crust has not changed while that of the upper crust has decayed significantly relative to its on-axis value. The lower crust is at least as important as the upper crust in preserving magnetic stripes in older, ultraslow spread crust.


