Magnetic polarity structure of the lower oceanic crust

Simon Allerton¹ and Maurice A. Tivey²

Abstract. A newly developed rock drill was used to take the first fully oriented samples from in-situ gabbros exposed on Atlantis Bank, which formed 11 m.y. ago on the Southwest Indian Ridge (SWIR). These paleomagnetic data locate the seabed outcrop of a normal polarity magnetic lineation within an outcrop of lower crust that correlates with overlying deep-towed and sea-surface magnetic field measurements. A systematic lateral offset between the seabed boundary and measured anomalies implies that the reversal boundary dips ~25° to the south, away from the ridge axis. Paleomagnetic results suggest that following magnetization acquisition, the platform tilted south by ~20° (range: 33° south to 7° north), implying an original reversal boundary dip of only ~5°.

Introduction

The age distribution of the ocean basins comes largely from magnetic polarity reversal patterns frozen into the crust during the seafloor spreading process [Vine and Matthews, 1963]. Analysis of marine magnetic lineations has also played a vital role in refining the Geomagnetic Polarity Time-Scale (GPTS) [Cande and Kent, 1995]. Thus, it is clearly important to understand the nature of the magnetic recording medium. Extrusive upper crust is often thought to form the primary source of the magnetic lineations [Smith, 1990], but it is now clear that the lower gabbroic layer is also capable of creating a significant contribution. Sea surface magnetic studies show magnetic stripes cross areas where extrusive crust is absent, suggesting that lower crust and/or upper mantle can record and preserve magnetic reversals (e.g. Schulz et al., 1988; Tivey and Tucholke, 1998). Rock magnetic studies and paleomagnetic studies in unoriented dredge samples [Fox and Opdyke, 1973; Kent et al., 1978], partially oriented Ocean Drilling Program (ODP) cores [Pariso and Johnson, 1993a,b; Kidwa and Pariso, 1991] and ophiolites [Banerjee, 1980] all illustrate that oceanic gabbros are capable of retaining a stable magnetization. Inherent in the Vine-Matthews hypothesis is the idea that oceanic crustal magnetization is acquired while cooling from magma close to the ridge axis, either by cooling through the Curie temperature or by high-temperature alteration soon after eruption or intrusion. In the rapidly quenched extrusives, reversals follow individual lava contacts and dip towards the axis of spreading [Macdonald et al., 1983; Tivey et al., 1998]. Lower crustal reversal boundaries have been modeled as cooling isotherms [Cande and Kent, 1976], dipping away from the spreading axis. Anomalous skewness of magnetic anomalies also implies that reversal boundaries in lower crust and upper mantle are inclined away from the axis [Arkani-Hamed, 1988]. There are, however, few observational constraints on the nature and geometry of magnetic polarity boundaries in lower oceanic crust. To address this point, we present the first systematic paleomagnetic investigation of reversals in lower oceanic crust and integrate these results with near-bottom magnetic-field measurements. The British Antarctic Survey research vessel James Clark Ross collected these data during cruise JR31 to Atlantis Bank on the SWIR in April 1998.

Geology and Tectonic Setting of Atlantis Bank

Atlantis Bank is a ~11 m.y. old shallow submerged platform, located on the east side of the Atlantis II fracture zone, which offsets the very slow spreading SWIR [Dick et al., 1991] (Fig. 1). The ~5 km wide platform extends 10 km in the direction of spreading (north-south) and currently lies at a depth of ~700 m below sea level. Initial uplift of Atlantis Bank is thought to have occurred by unroofing of lower crust due to low-angle detachment faulting at the inside corner of the SWIR ridge-transform intersection [Dick et al., 2000]. Recent drilling at ODP Site 735B [Robinson et al., 1989; Dick et al., 1991; 1999]. Hole 735B cored ~1.5 km of gabbros and paleomagnetic results show that they have a stable magnetization of 2.5 A/m with a consistent reversed polarity inclination and growth rate. This is consistent with the hypothesis that the gabbros were emplaced within a few hundred thousand years of the reversal boundary [Cande and Kent, 1976].

Figure 1. Bathymetry map of Atlantis Bank (100 m contours) with the platform above 800 m shaded grey. Drill site locations with polarities are shown by boxes. The location of the deep-towed magnetic (DTM) profiles and sea surface lines are marked by bold lines. Inset map shows the location of Atlantis Bank and ODP Hole 735B on the Southwest Indian Ridge.
Paleomagnetic Sampling and Analysis

Two remotely-operated drilling systems developed by the British Geological Survey (BGS) were used for sampling. The BRIDGE drill (BR drill), specifically designed to take 1 m long fully-oriented cores for paleomagnetic and structural studies, collected 10 oriented gabbro cores and 1 diabase core. A second drill (BGS drill) has a 5 m core penetration capability but no azimuthal orientation. A total of 30 partially-oriented cores were taken using this drill; 12 were in situ gabbros, the rest were carbonates. A total of 23 cores were drilled along a 7 km north-south profile of the platform, parallel to the spreading direction (Fig. 1).

Table 1a. Paleomagnetic Data of BR Cores

<table>
<thead>
<tr>
<th>Site</th>
<th>D (°)</th>
<th>I (°)</th>
<th>α95</th>
<th>N</th>
<th>R</th>
<th>L</th>
<th>Location (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BR1</td>
<td>188.0</td>
<td>83.0</td>
<td>7.5</td>
<td>3</td>
<td>2.99</td>
<td>0.53</td>
<td>42.68/16.75</td>
</tr>
<tr>
<td>BR2</td>
<td>246.1</td>
<td>51.6</td>
<td>12</td>
<td>2</td>
<td>1.80</td>
<td>0.10</td>
<td>43.39/15.89</td>
</tr>
<tr>
<td>BR3</td>
<td>24.2</td>
<td>16.1</td>
<td>7.5</td>
<td>2</td>
<td>1.98</td>
<td>0.25</td>
<td>43.61/16.67</td>
</tr>
<tr>
<td>BR5</td>
<td>194.8</td>
<td>55.2</td>
<td>20.4</td>
<td>6</td>
<td>5.58</td>
<td>0.22</td>
<td>43.13/17.52</td>
</tr>
<tr>
<td>BR6</td>
<td>158.9</td>
<td>56.2</td>
<td>7.5</td>
<td>2</td>
<td>5.97</td>
<td>0.82</td>
<td>43.35/17.52</td>
</tr>
<tr>
<td>BR8</td>
<td>202.4</td>
<td>50.0</td>
<td>14.2</td>
<td>4</td>
<td>3.97</td>
<td>0.22</td>
<td>41.01/17.52</td>
</tr>
<tr>
<td>BR9</td>
<td>136.8</td>
<td>68.7</td>
<td>37.5</td>
<td>3</td>
<td>2.83</td>
<td>0.42</td>
<td>44.21/16.66</td>
</tr>
<tr>
<td>BR10</td>
<td>173.9</td>
<td>63.0</td>
<td>7.5</td>
<td>2</td>
<td>1.99</td>
<td>0.25</td>
<td>43.35/16.67</td>
</tr>
<tr>
<td>BR12</td>
<td>311.4</td>
<td>68.8</td>
<td>12.7</td>
<td>10</td>
<td>9.42</td>
<td>0.45</td>
<td>40.65/17.44</td>
</tr>
<tr>
<td>BR14</td>
<td>134.3</td>
<td>70.7</td>
<td>30.6</td>
<td>3</td>
<td>2.88</td>
<td>0.18</td>
<td>40.65/17.44</td>
</tr>
</tbody>
</table>

Paleomagnetic measurements, including thermal and alternating field (AF) demagnetization, were made on mini-cored samples taken from the drill cores (Table 1). The rocks contained a dominant, stable high-temperature and high-field component which is removed by cleaning at temperatures of ~590°C and AF fields of 80 mT, consistent with previous studies which have identified pseudo-single domain magnetite as the predominant carrier in oceanic gabbros [Pariso and Johnson, 1993a] (Fig. 2). A few samples had a minor low temperature (<400°C) and low field (<20 mT) component, but in contrast to many ODP cores [e.g. Kikawa and Pariso, 1991; Aundunnson and Levi, 1989], no dominant drilling induced component was evident, and the stable high-temperature high-field components were easily identifiable. The age of the magnetization cannot formally be established, but the measured normal and reversed directions are approximately antipodal (Fig. 2, Table 1). The dominance of reversely magnetized directions supports the petrological evidence that the magnetization is either an original thermoremanence, or a very early, chemical remanence acquired by high temperature alteration of the gabbros as they cooled [Pariso and Johnson, 1993a].

Table 1b. Paleomagnetic Data of BGS Cores

<table>
<thead>
<tr>
<th>Site</th>
<th>I</th>
<th>σ</th>
<th>N</th>
<th>Location (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BGS1</td>
<td>74.5</td>
<td>7.8</td>
<td>2</td>
<td>42.71/16.70</td>
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<tr>
<td>BGS10</td>
<td>-67.75</td>
<td>2.2</td>
<td>4</td>
<td>41.53/17.45</td>
</tr>
<tr>
<td>BGS16</td>
<td>46</td>
<td>2</td>
<td>1</td>
<td>42.57/16.65</td>
</tr>
<tr>
<td>BGS17</td>
<td>77</td>
<td>4.2</td>
<td>2</td>
<td>42.45/16.72</td>
</tr>
<tr>
<td>BGS20</td>
<td>-67.5</td>
<td>10.6</td>
<td>1</td>
<td>40.82/17.54</td>
</tr>
<tr>
<td>BGS21</td>
<td>52.5</td>
<td>24.7</td>
<td>2</td>
<td>41.79/17.43</td>
</tr>
<tr>
<td>BGS23</td>
<td>55</td>
<td>1.4</td>
<td>2</td>
<td>42.92/17.65</td>
</tr>
<tr>
<td>BGS24</td>
<td>56</td>
<td>2.0</td>
<td>3</td>
<td>42.54/16.86</td>
</tr>
<tr>
<td>BGS27</td>
<td>79.5</td>
<td>4.9</td>
<td>2</td>
<td>43.45/16.57</td>
</tr>
<tr>
<td>BGS29</td>
<td>41</td>
<td>1</td>
<td>1</td>
<td>44.19/16.95</td>
</tr>
<tr>
<td>BGS30</td>
<td>25.5</td>
<td>0.7</td>
<td>2</td>
<td>41.96/15.65</td>
</tr>
<tr>
<td>BGS31</td>
<td>60</td>
<td>11.3</td>
<td>2</td>
<td>43.22/17.00</td>
</tr>
</tbody>
</table>

Results

All BR cores have reversed (steeply down-directed) directions, apart from BR14, which has a normal (steeply upward) component of magnetization, and BR 3 a diabase core with an ambiguous, shallow, northerly direction (Table 1) (Fig. 2b). Similarly, BGS cores have dominantly reversed stable mag-
netization vectors, with the exception of BGS10 and BGS20, which record steeply inclined, upward-directed normal polarity components. The overall BR core site mean magnetization direction (inverting reversed directions to normal) has an declination of 354.6° and an inclination of -69.9° (α95 = 12.7°), somewhat steeper than the geocentric axial dipole (GAD) of -52°. The mean inclination of the partially oriented BGS cores is -58.6° (α95 = ±13.4°) using the McFadden and Reid [1982] method. These bounds encompass the inclination determined from Hole 735B (i.e. -71.4° ±0.3/-3.1°, Dick et al., 1999). The errors of the BR and BGS core means gives a maximum range of tilt relative to the GAD of 33° to -7°, which overlaps the ODP Hole estimate of 23° to 19°. These results are consistent with a steepening of the recorded magnetization by ~20° southward tilting of Atlantis Bank. There is no evidence for any internal differential tilting or folding over the platform, which suggests a coherent tectonic unit with deeper structural levels exposed to the north than to the south. Close correspondence of BR core declinations with the axial dipole field also suggests that the lower crustal sequence has not been subjected to any significant vertical axis rotations, in contrast to some studies of ophiolites [Allerton, 1989].

While the majority of the drill sites record reversed polarity across Atlantis Bank, three are normally magnetized (BR14, BGS10, BGS20) and define a narrow but coherent normal polarity zone towards the northern end of the bank (Fig. 1 and 3). The northern boundary of this normal chron lies between BR15 and BGS20, and the southern boundary between BGS10 and BGS21, giving a width range of ~2 km to 1.25 km. This normal polarity zone is overlain by a narrow short-wavelength sea surface magnetic anomaly [Dick et al., 1991], which we correlate to chron C5r.2n (11.476 to 11.531 Ma). This anomaly is also seen in two deep-tow magnetic profiles collected during the JR31 cruise over the top of Atlantis Bank from south to north, at a survey height of ~100 m above the seafloor (Fig. 3). These profiles were navigated by calculating a layback behind the ship using the ship position, wire out and fish depth. Position was confirmed by matching deep-tow bathymetry with ship bathymetry. The shallow water depths (~700 m) and short wire lengths resulting location errors <100 m. The deep-tow magnetic data were calibrated to remove the effects of the permanent magnetic field of the towing harness and reduced to anomaly by removing the regional field based on the 1995 IGRF interpolated to 1998 [IAGA, 1996]. Profiles were continued upward to a level plane, 0.5 km below sea level [Guspi, 1987] and inverted for crustal magnetization [Parker and Huestis, 1974] assuming an inclination of -71° and a 1.5 km constant thickness source layer (Fig. 3). The maximum gradient in the anomaly defines the zero crossing of the magnetization profile and sufficient annihilator (~1) was added to each profile to zero the profiles at these points. Inversion results indicate that most of Atlantis Bank is reversely magnetized, but normal polarity magnetization is present between about 32° 41’S and 32° 43’S (Fig. 3). The overall amplitude of the magnetization anomalies is ±1.5 A/m, similar to the 735B drillcore average [Dick et al., 2000].

We find a systematic offset in the position of the normal chron identified from paleomagnetic sampling relative to the deep-tow inversion boundary. The simplest explanation for this offset is that the normal polarity body dips to the south, so that the surface outcrop boundary sampled by our drilling is offset to the north compared to the inversion boundary which averages over the entire thickness of the magnetized layer. The southerly dip of the boundary, away from the ridge axis, is opposite to the sense of dip of magnetic reversals in extrusive crust [Atwater et al., 1973; Macdonald et al., 1986; Tivey et al. 1998]. Further constraints on the dip of the normal chron is provided by Hole 735B, ~3.25 km south of the southern limit of the normal magnetic lineation. Uniformly reversely magnetized Hole 735B extends to 1.5 km beneath the platform [Dick et al., 2000], so that any normal polarity boundary must have a dip >25° so as not to intersect the hole. We forward modeled the observed deep-tow anomaly using the polygon algorithm of Won and Bevis [1987]. A constant magnetization was assumed for the polygons (1.5 A/m) and the seafloor polarity boundaries between polygons was fixed in lateral position by the drilling results. Only the dip of the polygons was varied (inclination was fixed at -71°). We find that a model (Fig. 3) with a dip angle of ~25° to the south i.e. away from the spreading center, reproduces the observed anomaly quite well (RMS misfit of 100 nT) any shallower dip would intersect Hole 735B. Steepening the dip of the boundary progressively worsens the RMS misfit of the model (Fig. 3), suggesting that the true dip of the boundary is not significantly greater than our estimate of 25°.

**Discussion and Conclusions**

Our observation of a shallowly inclined polarity reversal boundary dipping away from the spreading axis within lower oceanic crust is consistent with early models of polarity.
boundary shape [Cande and Kent, 1976] which envisioned boundaries controlled by conductive cooling through the Curie temperature of magnetite (580°C). These early models assumed that midocean ridge systems were supplied by a large, continuously replenished magma chamber, even at slow-spreading rates [Cann, 1970]. These models did not include the effect of hydrothermal circulation, however, which may be several times the conductive cooling rate [Lister, 1974]. Recent observational and modeling studies (e.g. Cannat, 1993; Phipps-Morgan and Chen, 1993; Wolfe et al. 1995) show that at slow spreading rates, magma chambers are sporadic and ephemeral, so that the axis cools significantly between intrusive events. We estimate the present dip of the boundary to be 25°, but the paleomagnetic evidence from this study and from ODP drilling suggests that the rocks of the platform have been tilted ~20° south. Thus, the original dip of the boundary would have been equivalently less i.e. ~5°. Such a shallow angle implies a slow cooling rate with conductively cooled isotherms. This presents somewhat of a paradox because the short polarity chron recorded by the gabbros (C5r, 55 kyr long) and the relatively sharp polarity boundaries suggest a more rapid cooling rate. In addition, the seafloor width of the 55 kyr chron is 1.6 km, which is ~4 times that calculated using the average half-spreading rate of 8 km/m.y. Clearly, the interpretation of polarity chron is problematical over such short timescales. Further evidence is needed to define the relationship between the polarity boundaries and the igneous units. For example, slow cooling would infer that isothermal surfaces and thus polarity boundaries would traverse intrusive contacts, but faster cooling would imply that reversals follow the edges of discrete intrusions. Further evidence is needed to test these ideas.

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