A Submersible Study of the Western Intersection of the Mid-Atlantic Ridge and Kane Fracture Zone (WMARK)

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Abstract

In 1994, a joint Japanese-American dive program utilizing the worlds deepest diving active research submersible (SHINKAI 6500) was carried out at the western ridge-transform intersection (RTI) of the Mid-Atlantic Ridge and Kane transform in the central North Atlantic Ocean. A total of 15 dives were completed along with surface-ship geophysical mapping of bathymetry, magnetic and gravity fields. Dives at the RTI traced the neovolcanic zone up to, and for a short distance (2.5 km) along, the Kane transform. At the RTI, the active trace of the transform is marked by a narrow valley (< 50 m wide) that separates the recent lavas of the neovolcanic zone from the south wall of the transform. The south wall of the transform at the western RTI consists of a diabase section near its base between 5000 and 4600 m depth overlain by basaltic lavas, with no evidence of gabbro or deeper crustal rocks. The south wall is undergoing normal faulting with considerable strike-slip component. The lavas of the neovolcanic zone at the RTI are highly magnetized (17 A/m) compared to the lavas of the south wall (4 A/m), consistent with their age difference. The trace of the active transform changes eastwards into a prominent median ridge, which is composed of heavily sedimented and highly serpentinized peridotites. Submersible observations made from SHINKAI find that the western RTI of the Kane transform has a very different seafloor morphology and lithology compared to the eastern RTI. Large rounded massifs exposing lower crustal rocks are found on the inside corner of the eastern RTI whereas volcanic ridge and valley terrain with hooked ridges are found on the outside corner of the eastern RTI. The western RTI is much less asymmetric with both inside and outside corner crust showing a preponderance of volcanic terrain. The dominance of low-angle detachment faulting at the eastern RTI has resulted in a seafloor morphology and architecture that is diagnostic of the process whereas crust formed at the WMARK RTI must clearly be operating under a different set of conditions that suppresses the initiation of such faulting.

Introduction

Ridge-transform intersections (RTI) not only mark the transition from the seafloor spreading regimes of the slow spreading Mid-Atlantic Ridge (MAR) to the tectonic regime of the transform domain, but also delineate the boundaries between tectonically and lithologically distinct terrain found on the inside corner (IC) and outside corners (OC) of the RTI [Karson and Dick, 1983; Tucholke and Lin, 1994]. IC crust is bounded by the spreading axis and the active transform and typically has elevated, smooth and rounded topography with high residual mantle Bouguer gravity anomaly (RMBA) [Severinghaus and Macdonald, 1988; Tucholke and Lin, 1994]. High RMBA gravity can be interpreted as indicating thin or denser crust. OC crust is bounded by the inactive transform and the spreading center and typically has low-relief topography with well-developed, ridge-parallel abyssal hill trends and no anomalous RMBA gravity. In situ observations and sampling at IC locations reveal that IC terrain is dominated by lower crustal and upper mantle lithologies [Miyashiro et al., 1969; 1971; Karson and Dick, 1983; Karson et al., 1987; Auzende et al., 1993; 1994; Cannat et al., 1995a]. It has been suggested that this lower crust has been unroofed by low-angle detachment faulting [Karson and Dick, 1983; Brown and Karson, 1988; Karson, 1991; Tucholke and Lin, 1994; Cann et al., 1997]. Lower crustal rocks are exposed in the footwall of the low-angle fault on the inside corner of the RTI, while the hanging wall of the fault comprising the volcanic carapace is transferred and carried off on the outside corner of the RTI [Karson and Dick, 1983; Cannat et al., 1995a]. Further evidence of deep crustal exhumation through low-angle faulting has recently been observed in high-resolution multibeam bathymetric surveys that have found shallow dipping corrugated topography or “megamullions” on the massifs that have formed on the inside corners of both RTIs and non-transform offsets [e.g. Cann et al., 1997; Tucholke et al., 1998].

The Kane Fracture Zone (KFZ) is a large left-stepping transform zone located in the central North Atlantic at 23°40'N that offsets the MAR by about 150 km (Figure 1). Over the past 10 m.y. the crust to the north of the Kane transform has been spreading asymmetrically, with a faster rate to the west (~13 mm/yr) compared to the east (~10 mm/yr) [Hussenoeder et al., 1996]. South of the Kane transform, the MAR is also spreading asymmetrically, with faster rates to the west (~14 mm/yr) than to the east (~11 mm/yr) [Schultz et al., 1988; Pockalny et al., 1995]. The KFZ trends orthogonal to the spreading direction at ~100°, although past changes in plate motion are clearly visible in the older fracture zone trace [Collette and Roest, 1992]. The most recent change in the transform direction occurred at Chron 6 (~20 Ma) [Tucholke and Schouten, 1988] and resulted in a reorganization of the spreading geometry north of the transform. This reorganization created a distributed zone of non-transform offset between 24°30'N and 25°N and precipitating a series of repeated rift propagation events that migrated towards the transform [Sempere et al., 1990; 1993; N. Seama, personal communication 1996].

While the eastern RTI of the KFZ shows a clear morphologic and lithologic contrast between IC and OC crust, the western RTI is dominated by volcanic lavas on both IC and OC terrain [Karson and Dick, 1983; Zonenshain et al., 1989; Lawson et al, 1996]. The south wall of the Kane transform is IC terrain created at the eastern RTI and is dominated by rounded smooth topographic massifs [Auzende et al., 1993; 1994] that are indicative of lower crustal exposures (Figure 1). The north wall of the Kane transform is IC terrain created at the western RTI and appears to be dominated by ridge and valley morphology, which is more indicative of upper crust and volcanics. The two RTIs of the Kane transform thus display contrasting styles of evolution of crust both today and over the past 10 m.y. To investigate this dichotomy in crustal structure and tectonics, a joint oceanographic expedition was undertaken by the Japan Marine Science and Technology Center (JAMSTEC) in collaboration with Woods Hole Oceanographic Institution (WHOI) in June 1994 [Fujimoto et al., 1994; 1995; 1996a]. The dive program focused on the less-studied western intersection of the MAR and Kane transform (WMARK) and utilized the deep diving capabilities of the Japanese submersible SHINKAI 6500 (i.e. depth rating of 6500 m). From an historical perspective, this dive program marked the first time a Japanese submersible operated in the Atlantic Ocean.

The Ridge Transform Intersections of the Kane Transform

Eastern RTI

The eastern intersection of the MAR and the Kane transform (MARK) and the adjacent ridge segment have been studied in detail by several submersible dive programs [Karson and Dick, 1983; Karson et al., 1987; Mevel et al., 1991; Gente et al., 1991, Auzende et al., 1993, 1994], deep-towed camera surveys [Karson and Dick, 1983], side-looking sonar surveys [Kong et al., 1988; Karson et al., 1992; Gao et al., 1998], crustal drilling [Detrick et al., 1990; Cannat et al., 1995b; Karson et al., 1997] and geophysical cruises (e.g. Detrick and Purdy, 1980; Cormier et al., 1984; Detrick et al., 1984; Schulz et al., 1988; Kong et al., 1988; Pockalny et al., 1988; Morris and Detrick, 1991; Deplus et al., 1992; Gente et al., 1995; Maia and Gente, 1998). The axis of spreading is defined by the neovolcanic zone (NVZ), which is marked by a narrow elongate axial volcanic ridge typically composed of young lavas with little or no sediment cover [Karson and Dick, 1983; Pockalny et al., 1988; Gao et al., 1998]. These lavas are also usually highly magnetic resulting in a strong magnetic anomaly peak or central anomaly magnetic high (CAMH) (e.g. Hussenoeder et al., 1996). In the MARK area, the NVZ is located along the eastern edge of a single, well-developed nodal deep and crosses the transform valley to terminate against the north wall of the transform [Pockalny et al., 1988] (Figure 1). Submersible observations and sampling of the north wall of the transform revealed exposures of ultramafic rock at ~5900 m depth [Auzende et al., 1994]. The IC crust of the eastern Kane RTI has elevated topography with smooth and rounded morphology and exposures of lower crustal rocks [Karson et al., 1987; Auzende et al., 1994; Cannat et al., 1995a]. This edifice was drilled during Ocean Drilling Program (ODP) Leg 106 (Site 669) and although not entirely successful, serpentinized peridotites were recovered from a nearby drill site located on the rift valley wall [ODP Site 670 - Detrick et al., 1990]. Further drilling was attempted during ODP Leg 153 [Cannat et al., 1995b] with drill Sites 921-924 located just east of Site 669 on the upper slopes of the IC high at the RTI. These drill sites penetrated to depths varying between 14 to 82 m below seafloor and recovered primarily gabbro and olivine gabbro with a remarkably wide range of igneous, metamorphic and deformational textures [Cannat et al., 1995b; Karson et al., 1997; Karson and Lawrence, 1997].

In addition to the IC high at the RTI itself there are a number of older IC massifs located along the south wall of the Kane transform [Auzende et al., 1994], including one that shows distinct corrugations in topography at Chron 2A (Figure 1) [Tucholke et al., 1998]. The OC crust, east of the RTI and nodal deep, is dominated by a series of hooked ridges and lineated abyssal hill terrain, oriented parallel to the spreading axis [Tucholke and Schouten, 1988] (Figure 1). This terrain is primarily volcanic in morphology and lithology [Karson and Dick, 1983; Cannat et al., 1995a]. The hooked ridge terrain extends east along the KFZ to approximately 44°W, where a major relict nodal basin occupies the majority of the fracture zone valley.

Western RTI

The WMARK RTI (Figures 1, 2 and 3) has only received scant investigation with dredging and near-bottom camera tows in 1978 and 1980 by [Karson and Dick, 1983] and by a more recent dredging, near-bottom sidescan sonar and magnetic survey [Allerton et al., 1995; Lawson et al., 1996; Hussenoeder et al., 1996]. Two MIR submersible dives traversed the WMARK nodal deep and IC high [Zonenshain et al., 1989]. The IC high forms the eastern boundary of the MAR axial rift valley and consists of a steep scarp rising to less than 2000 m depth (Figures 1 and 2). Dredging recovered only basalt and greenstone from this scarp [Karson and Dick, 1983]. Two MIR dives on this scarp (Figure 2) also revealed well-sedimented debris and talus slides, indicating no recent tectonic activity [Zonenshain et al., 1989]. The MIR dives encountered very few outcrops, but pillow basalts were observed at 4680 m, and greenstone, altered gabbro and dikes were reported between 4200 and 3500 m depth. Above this depth only pillow basalts and basaltic talus were found.

The MAR spreading axis, north of the Kane transform is marked by a discontinuous NVZ that becomes a morphologically more distinct axial volcanic ridge towards the RTI [Lawson et al., 1996]. This axial volcanic ridge separates two deep basins to the east and west and can be traced up to the south wall of the transform before it curves eastward into the active trace of the transform (Figure 2). The active trace of the transform or principal transform displacement zone (PTDZ) [Macdonald et al., 1986] is defined as the present-day zone of active strike-slip displacement and is typically marked by a narrow zone of interconnected faults which are aligned parallel to the transcurrent plate motions [Karson and Dick, 1983]. The eastern basin is the nodal basin of the WMARK RTI, reaching greater than 6000 m depth (Figure 2). This nodal basin is bisected by a narrow ridge parallel to the spreading axis, which has been interpreted as either an incipient IC high [Zonenshain et al., 1989] or a relict neovolcanic ridge [Pockalny et al., 1988; Hussenoeder et al., 1996]. The shallower (5200 m) western basin is elongated along the western fracture zone trace, and shows evidence of N-S lineated terrain, characteristic of constructional volcanism (Figure 2). In general, the terrain west of the neovolcanic ridge shows relatively low-lying ridge and valley topography, but no obvious hooked ridges are observed to bend into the transform domain, as is seen east of the eastern RTI [Pockalny et al., 1988] (Figure 1). At the WMARK RTI, the south wall of the transform is marked by a remarkably steep cliff that forms the northern end of a large, bathymetrically distinct edifice, directly adjacent to PTDZ (Figures 1 and 2). The edifice shows evidence of lineated abyssal hill topography perpendicular to the spreading direction, which is uncharacteristic for IC crust created at the eastern RTI over the past 10 Ma. Just east of the WMARK RTI, the south wall of the transform steps southwards (approx. 8 km) away from the trace of the active transform and lines up with the younger edifices that form the south wall of the Kane transform (Figure 1). The intervening basins between the PTDZ and the south wall are interpreted to be relict nodal deeps and were suggested by Pockalny et al. [1988] to reflect a temporal variation in magmatic activity.

The axial volcanic ridge of the MAR north of the WMARK RTI is marked by a very prominent magnetic anomaly; the CAMH, which reaches a maximum amplitude at the RTI (Figure 1). The CAMH follows the summit of the axial volcanic ridge and was resolved in finer detail by a recent deeptow magnetic survey [Hussenoeder et al., 1996]. Magnetic anomaly highs at the tips of ridge segments are a common feature of the MAR, although the precise reason for the enhanced magnetic intensity is not well-known. Various hypotheses for this behavior have been suggested including thicker extrusive crust [Hussenoeder et al., 1996], deeper Curie isotherm depth [Grindlay et al., 1992] or greater enrichment of Fe and Ti in basaltic lavas at segment ends [Weiland et al., 1996].

In addition to the contrasting morphologies and lithologies of the eastern and western RTIs, the transform valley itself is marked by a distinctive narrow median ridge that begins near the southeastern corner of the nodal deep and extends almost 30 km east along the north side of the transform valley (Figures 1 and 2). A second ~50 km long median ridge is also found extending from the eastern RTI nodal deep west along the northern edge of the transform valley. The western median ridge is shorter and discontinuous in places, while the eastern median ridge is longer and more continuous along the transform.

SHINKAI WMARK Cruise

The WMARK cruise took place in June and July 1994 using the Japanese research vessel M/S Yokosuka and the deep diving submersible SHINKAI 6500 [Fujimoto et al., 1995; 1996a]. The SHINKAI 6500 is equipped with two manipulators, a sample basket, a pan and tilt color video camera and a fixed video camera. The submersible is operated by a pilot and co-pilot and carries one science observer. Submersible navigation utilized conventional long-baseline navigation with a nominal accuracy of ±10 m. The ship used global positioning satellite (GPS) navigation to provide the geographical reference for the submersible and shipboard operations.

The submersible carried two 3-axis fluxgate magnetometers: a three-axis fluxgate sensor and a motion reference unit built by the University of Tokyo, Ocean Research Institute (ORI) to collect oriented vector field measurements [Sayanagi et al., 1995] and the ALVIN three-axis fluxgate magnetometer [Tivey et al., 1993] for total field measurements. The magnetometers were operated continuously during all 15 dives. Seafloor gravity measurements were also made inside the personnel pressure sphere using a LaCoste and Romberg land gravity meter. The SHINKAI 6500 submersible dives focused on two specific sites. The first area encompassed the western RTI region, where the MAR axis of spreading directly abuts the Kane transform and the crust that formed south of the Kane (Figure 3). The second area encompassed the active transform zone domain with its prominent median ridge, the south wall of the transform, and the IC high, just to the north of the Kane transform and east of the MAR axis (Figure 3). Eight dives were devoted to the RTI study area, while the remaining 7 dives focused on the second area objectives within the transform valley, south wall of the transform and inside corner high north of the transform.

Submersible Dive Observations : Western RTI Site

At WMARK, near-bottom sidescan TOBI imagery shows that the MAR NVZ appears to curve rapidly (within 1 km) into the PTDZ [Lawson et al., 1996]. Several dives crossed the axial volcanic ridge of the NVZ as it approached and intersected the transform domain and confirmed the TOBI interpretation [R. Searle, personal communication] of a curving NVZ into the transform (Figure 3). SHINKAI dive 201 traversed across the NVZ from north to south, directly adjacent to the transform wall, and found the zone to be comprised of several prominent constructional pillow lava ridges that trend ~140°, oblique to both the spreading and transform direction (Figures 3 and 4). The pillow ridges are composed of relatively recent lavas with little to no sediment cover and are observed to flow both to the north and to the south off the tops of the pillow ridges. The southern boundary of the NVZ is abruptly defined by a narrow valley (~50 m wide), trending east-west along the base of the south wall of the transform (Figure 4). This valley is floored by a thick sediment blanket (i.e. no outcrops) and most likely defines the PTDZ. The south wall consisted of a steep outcrop of massive diabase with a sediment veneer extending between 4950 m and the end of the dive at 4600 m.

SHINKAI dive 202 traversed parallel to dive 201, but 2 km to the east (Figures 3 and 4). Here a large, and sedimented debris fan covers the south wall of the transform. The base of the scarp, and presumably the PTDZ, was marked by a zone of fresh basaltic talus that quickly gave way to the heavily sedimented terrain of the south wall (Figure 4). No volcanic basement or evidence of recent volcanic activity was observed at the base of the scarp. This observation is consistent with the TOBI sidescan data [Lawson et al., 1996], which suggests that the eastern termination of the NVZ is located between dives 201 and 202. Dive 202 observed evidence of active tectonic activity on the south wall with debris flows of basaltic material with grain size varying from gravel to cobbles and little or no sediment cover. A sedimented terrace with sparse talus was encountered between 4850 m and 4580 m depth. Above 4580 m to about 4472 m, the slope steepened and exposed the first significant outcrops of altered basalt and basaltic breccia.

Dives 203 and 204 (Figure 3) traversed the steepest part of the south wall of the Kane transform to complete the traverses begun in dives 201 and 202. Dive 203 crossed very large talus blocks (~10 m in size) and mixed grain size (i.e. gravel to meter-sized fragments) debris fans along the east-west trending scarp between 4750 m and 4600 m depth and ascended near vertical cliffs of brecciated pillow basalts between 4350 m and 4100 m depth. The cliffs probably formed as a consequence of steep fault scarps with slopes of 70° to 90°N and trends with azimuths between 30° and 100°.

Dive 204 traversed similar terrain as dive 203 and encountered an almost vertical outcrop of massive basalt between 4000 and 3800 m depth (Figure 4) and recovered basaltic samples from the scarp. Slickenlines, measured on a north-facing, steep (70° to 90°) fault surface, were found to be plunging west at 60° indicating oblique-slip movement, with a normal and right-lateral displacement. The summit of the south wall was covered with pelagic sediments that almost completely covered occasional outcrops of low-lying basement rock (Figure 4).

Dive 205 began on the east side of the axial volcanic ridge of the NVZ, about 3 km north of the western RTI (Figure 3). An outcrop of pillow lava was sampled at 5240 m depth, and a talus slope of pillow lava fragments was traversed until approx. 4899 m depth, where in situ lavas and lava tubes were seen. These lavas marked the summit region of a small volcano complex that consisted of two small volcanic centers. Rock samples collected along the dive traverse consisted of pillow basalt with thin manganese coatings (<1 mm) and spalled glass surfaces. The volcanic ridge showed evidence of disruption by shear stress, with the observed opening of randomly-oriented fractures. No evidence of ridge-parallel tension fractures was observed. The summit area is an elongated collapse pit or graben several tens of meters across with broken basalt talus and very little sediment cover (Figure 3). No evidence of recent eruption was observed within the summit graben. Sediment cover becomes more prevalent on the lower slopes of the neovolcanic ridge.

Dive 206 continued on from the end of dive 205 and traversed along the crest of the axial volcanic ridge towards the south wall (Figure 3). The end of dive 206 encountered the same narrow valley or graben as in dive 201, which marks the southern boundary of the NVZ and the locus of the PTDZ. This valley was "dammed" by a fresh-looking lava flow that had flowed south from the NVZ into the valley. The floor of the valley was blanketed with sediment with occasional sink holes in the sediment. Sink holes are often formed when basement collapses or subsides due to tectonic movement.

Dive 207 explored the WMARK nodal basin to the east of the RTI, beginning at its deepest point at ~6024 m depth (Figures 2 and 5). The nodal basin was thinly sedimented (~1 m) with several small constructional volcanic ridges oriented north-south separated by tectonically disrupted zones consisting of pillow basalt talus. The flank of a large and robust constructional volcanic ridge was reached towards the end of the dive (Figure 5). This ridge was described by Zonenshain et al. [1989] as being an extinct axial volcanic ridge that has been rafted away from the current axis of spreading. SHINKAI magnetic data collected during dive 207 [Fujiwara and Fujimoto, 1998] and a TOBI near-bottom magnetic profile across the same area [Hussenoeder et al., 1996] both show a magnetic anomaly high associated with the relict ridge suggesting that the ridge is relatively recent in age.

Finally, SHINKAI dive 215 traversed the southern boundary of the nodal deep (Figure 2), climbing a scarp composed of a series of rubble terraces, talus ramps and debris fans, between 5300 m and 4800 m depth that presumably defines the PTDZ. The lack of sediment cover and considerable size range of the debris from gravel to large decameter-sized blocks suggests a vigorous and tectonically active zone consistent with the proximity to the PTDZ. Recovered rock samples were primarily weathered basalt with rare dolerite and gabbro cobbles.

Submersible Dive Observations : Western Median Ridge and Transform Domain Study Area

Two dives (208 and 212) traversed the median transform ridge located along the trace of the transform just east of the WMARK RTI (Figures 2 and 6). These dives found a heavily sedimented ridge with sporadic outcrops of blocks composed of peridotite and gabbro. Both dives traversed the northern side of the ridge and crossed a series of small fault scarps (~ 1m throw) that form linear features cutting the sediments, producing small ridges and valleys that trend approximately parallel to the PTDZ (azimuth ~100°). Near the top of the ridge at 4378 m depth and between 4144 and 4119 m depth, several white, round, mound-like features were encountered approx. 3 m in diameter and a few meters high (Figure 6). The mounds have aprons of whitish clay material and are cored by serpentinized peridotite. Larger rounded structures are also visible in the TOBI sidescan data over the area [R. Searle, personal communication, 1994] and probably represent a macro-scale view of the features observed from the submersible.

Dives 209 and 214 also traversed the median ridge, where it abuts the south toe of the IC high (Figure 2). Dive 209 began in the sedimented plain at the base of the IC high and just south of the active transform trace. Relatively undisturbed sediment was observed up to a small scarp at 4400 m depth beyond which talus and debris slides were encountered. This scarp appears to mark the beginning of the active transform tectonized zone. The region consisted of faults, fissures, talus ramps and debris flows forming an undulating terrain of small ridges and troughs. Another steep scarp face was reached at 4300 m depth and altered basalt was sampled from this area. Above the scarp, the terrain was dominated by talus and debris slides that cover any active faults that may be present. Stratified slope sediments with intercalated talus and volcanic breccia characterized the upper part of the slope. A ~50 m high scarp was reached at 4075 m depth beyond which only a smoothly sedimented slope was observed with no further signs of tectonic activity. The scarp most likely marks the northern extent of the transform tectonized zone making a total width for the zone of about 900 m.

Dive 214 also crossed the median ridge and foot of the inside corner high (Figure 2). The dive began at 4272 m depth, south of the crest of the median ridge. Small troughs or gullies less than 10 m deep and a few meters wide, and oriented at an azimuth of 105° marked the summit of the median ridge at 3975 m. At the northeastern termination of the median ridge, a steep scarp sloping 70° SSW and trending ~105° was encountered at 4071-4054 m depth. This scarp appears to be the trace of the active transform fault. Rocks exposed in the scarp were found to be composed of fault gouge and white clay. North of the scarp, the terrain slopes gently upwards. The dominant lithology observed and sampled on this scarp was basalt with no evidence of peridotite or gabbro.

Dive 211 ascended the upper part of the eastern side of the IC high at the WMARK RTI (Figure 2). The dive began at 2877 m depth in calcareous sediment and immediately ascended several steep (30-50°) cliffs, including one over 1000 m high that exposed pillow lava with thin manganese coatings (1-3 mm). The cliffs were interspersed with small talus ramps and intercalated calcareous sedimentary units. A wide bench of basaltic cobbles was reached at 1850 m depth. East-west trending fissures were also observed. The peak of the summit ridge of the IC high was reached at 1480 m depth and was covered with sediment and weathered basaltic rock outcrop.

Dives 210 and 213 completed a transect up the south wall of the transform at 46°04'W, which is located a considerable distance (approx. 8 km) from the PTDZ (Figures 2 and 7). The transform wall at this location shows a subdued and gentle topography with a slope of only 18° compared to 35° for the south transform wall at the western RTI site. The topography of the south wall is also typical of IC terrain with rounded and smooth topography. The age is approx. 7.5 Ma (Chron 4) (Figure 1). Dive 210 began at 4828 m depth in the transform valley and reached 4154 m depth (Figure 7). Thick sediment cover predominates (> 50 cm : length of heat flow probe). Sparsely distributed rounded rock fragments dot the terrain and samples reveal them to be highly altered serpentinized peridotite. A small outcrop at 4159 m depth formed a small ridge and from samples appears to be composed of serpentinized peridotite. Dive 213 continued the traverse of the south wall beginning at 4149 m depth and reaching 3474 m depth (Figure 7). The entire traverse covered sedimented terrain with abundant benthic organisms suggesting a benign and tectonically inactive region. Three talus samples, collected between 4047 and 3852 m, were found to be highly serpentinized peridotite.

Submersible Gravity Measurements

On-bottom gravity measurements were collected at 20 seafloor stations (see Figure 2 for locations and Table 1 for values). The gravity stations were located in several tectonically and lithologically distinct regions within the WMARK study area. Dives 201 through 207 form one gravity transect at the WMARK RTI site from the nodal basin across the axial volcanic ridge to the top of the south wall of the transform wall (Figure 2). Dives 209, 210, 212 and 213 form a second gravity transect across the transform valley, median ridge and up the south wall (Figure 2). The gravity data were first corrected for latitude and water depth to give a free-water anomaly (Table 1) following the procedure of Luyendyk [1984]. To correct for local topographic effects a Bouguer anomaly was calculated using a three-dimensional terrain correction based on the multibeam bathymetry (Figures 8 and 9). For each region, an average crustal density can be obtained by estimating the slope of the regression line for the terrain-corrected Bouguer anomaly as a function of the water depth. For the WMARK RTI, nodal deep and adjacent region, we obtain an average crustal density of 2784 ± 78 kg/m^3, consistent with the dominantly basaltic terrain (Figure 9). Likewise, the WMARK IC measurements give a density value of 2658 kg/m^3 consistent with basaltic crust. For the south wall of the transform valley, an average crustal density in the range of ~3000 kg/m^3 (2947 to 2999 kg/m^3) is obtained (Figure 9) which is significantly higher than the RTI region and consistent with the observed occurrence of gabbro and serpentinized peridotite in this area. Thus, to a first approximation, these gravity measurements confirm the submersible observations that predominantly lower crust and mantle rocks are found at the surface in the transform valley and basaltic crust dominates at the WMARK RTI and IC high.

Submersible Magnetic Measurements

Oriented three-component magnetic field data were obtained on most of the SHINKAI dives using the ORI magnetometer and these data were analyzed for vector magnetic information and are discussed separately in Fujiwara and Fujimoto [1998]. Magnetic field data collected using the ALVIN magnetometer sensor mounted to SHINKAI 6500 were used to investigate the vertical magnetic structure of the crust exposed in the wall of the Kane transform. These magnetic field data were first calibrated for the effect of the submersible by having the submersible spin during descent and ascent and minimizing total field variations using a Nelder-Meade method (see Tivey et al., 1993 for details). The magnetic field measurements were then processed using the vertical magnetic profiling approach [Tivey, 1996], which employs a rotation of the survey geometry, such that the scarp face is rotated into a horizontal plane by an amount equal to the scarp angle. This procedure is merely to facilitate the mathematical analysis and has no physical meaning for crustal rotation. In this rotated geometry, the crust is treated mathematically as a series of tabular bodies dipping at an angle equal to the slope of the scarp. The magnetic field data within this rotated reference frame can be interpreted in terms of dipping, semi-infinite tabular bodies for which there are analytical and Fourier transform solutions [e.g. Gay, 1963; Pederson, 1978; Tivey, 1996]. To demonstrate the feasibility of magnetic measurements on a scarp face we construct a forward model where we make no inferences about crustal structure, but merely show what we expect from a single horizontally-oriented magnetized body (Figure 10). The combination of scarp slope angle (~35°) and geomagnetic field inclination (45°N), means that the effective magnetic field has a shallow inclination (10°), but this still provides enough of a contrast to generate anomalies on the order of 2000 nT at 5 m distance from the scarp for a 5 A/m magnetization contrast (Figure 10). The magnetic field data collected up the scarp face can also be inverted for crustal magnetization, assuming that the dipping bodies extend to effectively infinite depth.

The magnetic effect of a single magnetized block while useful is perhaps too simple, thus we constructed a second model where we attempt to model a series of slipped fault blocks. This is particularly important for the scarps that form the walls of the Kane transform, because they are known to be regularly interrupted by normal faults, that step down into the transform [e.g. Karson and Dick, 1983; Wilcock et al., 1990; Auzende et al., 1994] (Figure 4). Normal faults produce a repeated crustal section as the blocks form a progressively down-dropped series of blocks towards the transform valley, akin to a "slipped deck of cards" [Francheteau et al., 1976]. We compute the magnetic field for a conceptual model, where an upper, highly-magnetized crustal section is faulted three times, producing three slipped blocks (Figure 11). The resultant magnetic field is shown for the upper magnetized unit alone and then for all the units (Figure 11). As can be seen from the forward model, each slipped block produces a short-wavelength magnetic anomaly with anomaly highs at the toe of each block and anomaly lows at the top of each block.

The intact block at the top gives an overall longer wavelength signature. The forward model magnetic field was inverted for crustal magnetization to simulate our data analysis steps (Figure 11). The overall positive magnetization of the upper block is recognizable as a long-wavelength positive zone, the slipped blocks give a less intuitive result. The short-wavelength magnetization highs reflect the variable thickness of the source layer due to the overlapping of the blocks, rather than any amplitude difference in magnetization. Note also that the profile shows zones of reversely magnetized crust where there are normal polarity blocks (e.g. Figure 11 at 4450 m). These models (Figures 10 and 11) are meant to serve as a guide to interpreting our observed magnetic profiles rather than to explicitly fit them.

We concentrated our analysis on two pairs of dives from the WMARK dive program that form relatively complete transects up the south wall of the Kane transform at the two contrasting study sites. Dives 202 and 204 provide a relatively continuous transect of the south wall of the transform in the WMARK RTI region. Submersible observations and sampling suggest that the scarp is composed primarily of basaltic lavas and diabase. Sea surface magnetic anomaly identifications suggest that the crust in this region is approx. 10 m.y. old (Chron 5). The scarp face in the RTI region has a slope angle of ~35° and the regional magnetic field inclination is ~45°. The magnetic data were projected along a vertical cross-section of the scarp (Figure 12) and then inverted for crustal magnetization with the top of the scarp acting as the zero reference level (i.e. seawater is non-magnetic). The profile shows a peak near 3 A/m at 5300 m depth but for crust shallower than this magnetization appears to oscillate about the zero line (Figure 12). The magnetization amplitude is compatible with both inversion results of sea surface magnetic data over the same region (Figure 1) and rock magnetic measurements discussed later.

We use the faulted block model (Figure 11) to help understand the measured profiles. Using magnetic field lows and highs in the field, we identify individual blocks or zones in the dive 202/204 transect (Figure 12). At the top of the scarp extending to 3900 m depth (Zone 1 of Figure 12) two blocks are interpreted with an intervening bench. Below 3900 m (Zone 2, Figure 12) a repetitive sequence of short-wavelength anomalies is observed suggestive of a highly faulted section. This view is supported by the submersible observations, which find a series of alternating outcrops with debris and talus fans between 3900 and 4450 m (Figure 4). Below 4450 m (Zone 3, Figure 12) we interpret a section of two main blocks with a bench or fault at ~4700 m. Below

4900 m to the base of the scarp at 5300 m (Zone 4, Figure 12) we interpret a single intact block. The interpretations of magnetic data are nonunique and a variety of models could satisfy the data. For example, some of the short-wavelength, low-amplitude magnetic variations between 4500 and 3600 m could simply be due to topographic noise due to the submersible traversing undulating outcrop and faults. The variation in altitude of the observations was not measured and thus not taken into account in the inversions allowing some topographic noise to leak through into the solution. This topographic noise is not applicable to longer wavelength anomalies like those near the bottom of the transect (Zone 3 and 4). These strong anomalies suggest to a first order, that the deeper crust is more intact with larger blocks and fewer faults than the shallower crust.

Dives 210 and 213 form a similar transect of the south wall of the transform, but in younger-aged crust (approx. 7.5 Ma; Chron 4). Submersible observations and sampling at this location suggest that the wall is primarily composed of lower crustal material and serpentinized peridotite (Figure 7). The slope angle is only about 18 degrees, which is at the limit of the assumptions for the Fourier analysis techniques used in the vertical magnetic approach [Pederson, 1978; Tivey, 1996] (Figure 13). The top of the scarp was not attained during the dives so that no zero reference could be obtained for the composite profile. The amplitude of magnetization is generally less than 2.5 A/m. A noticeable gradient in the magnetic field occurs between 4300 and 4200 m depth (Figure 13). This gradient is not attributable to any overlap errors between the two dives. In the inversion, the gradient in magnetic field translates into a transition from apparently reversely magnetized crust to normal polarity crust (Figure 13). This result suggests a polarity reversal could have been encountered along this transect. However, because of a lack of zero reference, it remains ambiguous at this stage whether this transition is from near zero to strongly positive magnetization, or from truly reversed polarity to weakly positive magnetization. From sea surface magnetic maps (Figure 1) the dive transect does appear to be at the old edge of polarity Chron 4, so that a polarity reversal is not unexpected at this location. Regardless of whether the magnetization pattern is a polarity reversal or a difference in magnetization intensity, it is clear that the lower crustal material has a magnetization. It is speculated that at least part of the magnetization is remanent rather than induced because the overlying sea-surface magnetic anomaly lineations are clear and relatively well-defined. Dive 210/213 observations (Figure 7) show that outcrops of serpentinized peridotite were found at or near the transition in magnetic field at 4200 m depth. If these peridotite sequences are recording a magnetic polarity reversal this has important consequences for the source of the marine magnetic anomaly signal. Further studies are needed to fully define this result however.

Paleomagnetic Studies

Rock samples collected by SHINKAI 6500 were measured for their paleomagnetic parameters, which included: natural remanent magnetization (NRM), magnetic susceptibility (χ), and stability with step-wise alternating frequency (AF) demagnetization. Results are shown in Table 2. The basaltic rocks collected from the neovolcanic zone of the RTI region (dives 201, 205, 206) give a mean NRM of 17.5 A/m ± 8.6 A/m, which is consistent with estimates of zero-age crustal magnetization obtained from deeptow magnetic profiles [Hussenoeder et al., 1996]. The mean susceptibility of the axial volcanic ridge basalts is 0.21 x 10^6 SI, which gives a Koenigsberger ratio (Q) of remanent to induced magnetization of ~2 indicating remanent magnetization dominates. The basaltic rocks from the south wall of the RTI region (dives 210 and 213) give a mean NRM of 4.1 ± 3.8 A/m, which is less than the neovolcanic value, but consistent with rock magnetic values of similar-aged crust [Johnson and Pariso, 1993] and magnetic anomaly analysis (e.g. Pockalny et al., 1995; Tivey and Tucholke, 1998). Susceptibility is 0.32 x 10^6 SI and the Q ratio is less than one (0.3) indicating that induced magnetization dominates. One basalt sample from the nodal deep (dive 207-2-1) has an NRM of 7.5 A/m, intermediate between the old crust value and neovolcanic zone value. In contrast to the basaltic rocks, the sampled gabbro and serpentinized peridotites have a mean NRM of approximately 1 ± 0.9 A/m and a susceptibility of 0.73 x 10^6 SI, giving a Q value of 0.033 indicating induced magnetization dominates.

Discussion

How do the observations at the WMARK RTI compare with similar observations at the eastern RTI of the Kane and RTI's at other transforms? On a broad scale, the asymmetry in seafloor geology, lithology and morphology at the IC and OC of the RTI, that is so well developed at the eastern RTI of the Kane [Karson and Dick, 1983] and other transforms like the Atlantis [OTTER, 1984; Cann et al., 1997], is clearly not seen at WMARK. Each side of the spreading axis of the WMARK segment (i.e. both IC and OC settings) is dominated by volcanic terrain with lineated topography oriented parallel to the spreading axis. This contrasts sharply with the rounded, elevated, smooth topography and lower crustal exposures found at IC settings and the lineated, dominantly volcanic terrain found at OC settings of the eastern RTI of the Kane and Atlantis RTI [Karson and Dick, 1983; OTTER, 1984; Cann et al., 1997]. While low-angle detachment faulting appears to have been active at the IC of these latter sites, volcanic accretion and normal faulting tectonics appear to have dominated the recent history (at least for the last 5 m.y.) of the WMARK segment.

On a more local scale, the NVZ of the WMARK spreading segment appears to almost run into the south wall of the transform. In fact, it curves quite rapidly into the transform domain within about 1 km of the PTDZ. This contrasts with the eastern RTI of Kane which curves gently around the nodal deep over a distance of several kilometers [Karson and Dick, 1983]. Observations at the eastern RTI of the Vema transform also show a rather gentle curvature to the NVZ as it curves into the transform over about 5 km [Macdonald et al., 1986]. The volcanic terrain along the WMARK NVZ appears to be relatively recent and this is consistent with earlier observations [Karson and Dick, 1983] that suggested the western NVZ was younger than the eastern NVZ. Also, magnetic measurements of NVZ basalts show that they are strongly magnetized (17 A/m) consistent with a relatively young age. At WMARK, volcanic construction appears to continue right up to the transform and for a short distance (~2.5 km) along it (Figure 3). A small volcanic complex at the southern end of the NVZ (23°50.2'N 46°19.2'W) marks the last distinct volcanic center before the NVZ curves into transform domain. The two small volcanoes characterize the typical mode of accretion along the majority of the MAR i.e. discrete volcanoes that coalesce to form an axial volcanic ridge [Smith and Cann, 1992]. South of this volcanic center, the NVZ curves eastward along the transform trace and evolves into a series of narrow constructional volcanic ridges that are oriented at an oblique angle (~140°) between the spreading and transform trends (Figure 3). Assuming the lavas are fed by diking events, this orientation presumably reflects the dominating influence of the direction of least compressive stress as the stress trajectories rotate from the spreading regime into the transform stress regime [Macdonald et al., 1986]. Our observations suggest that while the NVZ reaches the transform wall it is not able maintain a robust ridge that curves into the transform for any significant distance or length of time. This view is consistent with Lawson et al. [1996] who find that the lavas of the

NVZ at the WMARK RTI have multiple parental magma batches suggestive of episodic and discontinuous magma supply.

The large bathymetric high that forms the south wall at WMARK is similar in some respects to the intersection highs found at the Clipperton RTI's (e.g. Barth et al., 1994). Both are large bathymetric features that dominate the surrounding terrain. They are high standing with an obviously volcanic carapace based on morphology. Barth et al. [1994] make a convincing case that the origin of these intersection highs at the Clipperton RTI's is due to magmatic leakage from the adjacent ridge axis and possibly magma from deeper sources as well. Rocks recovered from the Clipperton edifices reveal relatively young basalts with lava chemistries closely related to the adjacent ridge [Barth et al., 1994]. The WMARK south wall edifice does not fit this view for a number of reasons. The rocks exposed on the top of the edifice are thickly sedimented and clearly old where they outcrop, which is consistent with their age of approximately 10 m.y. The edifice also has distinct ridge parallel abyssal hill morphology that shows little or no curvature towards the transform. The magnetic anomaly pattern over the edifice is also remarkably well lineated and clearly correlatable to the geomagnetic polarity timescale. Measured rock magnetic values and submarine profile inversions both reveal relatively low magnetization intensities for the crust of the south wall at the RTI which is compatible with an age of about 10 m.y. The origin of the south wall edifice is thus more likely to be due to a short period of abundant magmatic supply and/or brief cessation of low-angle detachment faulting at the IC of the eastern RTI allowing a relatively intact crustal section to be transported away from the spreading axis.

Dives on the south wall of the Kane transform at the WMARK RTI (Figure 4) observed primarily basalt lavas and diabase and no evidence of lower crustal lithologies. While the south wall of the KFZ at the RTI is directly adjacent to the PTDZ, just 10 km east of the RTI, the south wall of the transform steps south from the PTDZ by about 8 km and forms a more gently sloped wall more typical for the south wall of the Kane transform for the last 8 m.y. (Figures 1 and 2). Submersible dives on this part of the south wall at 46°10'W (Figure 7) recovered only deep crustal lithologies and traversed low-relief hills, more typical of that now observed at the eastern RTI. Submersible observations on the south wall of the Vema FZ [Auzende et al., 1989] found a relatively intact crustal section ranging from ultramafic rocks through gabbros to dikes and extrusives. The south wall of the Kane fracture zone is clearly not as straightforward as this.

Dives on the lower relief scarps of the south wall (Figure 7) and further east along the transform wall [Auzende et al., 1994] towards the eastern RTI find lower crustal and upper mantle rocks. Drilling on the IC high at the eastern RTI also penetrated gabbros and highly deformed rocks [Cannat et al., 1995b; Karson et al., 1997]. This evidence is consistent with the macroscopic observations of seafloor morphology and RMBA gravity that these rounded edifices are unroofed and uplifted lower crustal and upper mantle crust exposed by low-angle detachment faulting at the IC of an RTI. The south wall edifice at the WMARK RTI is different because of its abyssal hill morphology and dominance of upper crustal lithologies. It apparently has not been unroofed and remains a relatively intact crustal section. While the submersible observations at the Vema transform [Auzende et al., 1989] found good exposure of upper and lower crustal units over a depth range of 3000 m, the WMARK RTI south wall has only half the depth range, 1500 m, and consequently only exposes the upper crustal units. Mass wasting and associated normal faulting would further reduce the possibility of exposing gabbro or other lower crustal units at the WMARK south wall.

Finally, median ridges are a common feature of many transform valleys, e.g. Vema, [Macdonald et al., 1986], Atlantis II fracture zone [Dick et al., 1991]; and these features have often been cited as being formed by serpentinite diapirism [Bonatti, 1978]. Much of the PTDZ of the Kane transform is marked by a long narrow median ridge that is split into a western and eastern part [Pockalny et al., 1988] (Figure 1). The western median ridge consists of a narrow linear ridge that merges to the east with the north wall of the transform and the crust of the inside corner high of the WMARK ridge segment (Figure 2). Our dives reveal that this median ridge is heavily sedimented, but that white mounds ~1-3 m in diameter occur near the top of the ridge and are almost certainly the alteration products of highly serpentinized peridotite that were sampled at the cores of these features (Figure 6). The slopes of the ridge are also occasionally interrupted by series of small faults or steps ~1 m high, and oriented parallel to the transform. Our observations are thus consistent with the view that these ridges are not volcanic constructional features but zones of highly altered upper mantle that have been uplifted and emplaced at the seafloor along the active trace of the transform fault [Bonatti, 1978; Macdonald et al., 1986].

In terms of crustal magnetization, the relatively strong NRM of the neovolcanic zone rock samples at the WMARK RTI are consistent with the strong intensity in crustal magnetization calculated from sea-surface magnetic anomaly data [Hussenoeder et al., 1996] and the vertical magnetic profiles. One goal of the rock magnetic measurement program was to investigate the hypothesis that basalts are more chemically evolved towards segment ends on the MAR, with higher Fe and Ti contents and stronger magnetization intensities compared to basalts from the central portion of the ridge segment [Weiland et al., 1996]. Christie and Sinton [1981] suggested that the enhanced Fe and Ti in lavas are the result of increased fractionation within small magma bodies that supply the tips of propagating rifts at fast spreading ridges. For the slow-spreading MAR, however, Sinton and Detrick [1992] suggested that fractionation would take place during melt migration and thus significant along-axis migration would be needed to produce the required fractionation. Recent petrological studies on the MAR suggest only a limited range of fractionation and do not support significant along-strike magma transport [Niu and Batiza, 1994; Michael et al., 1994]. Based on analyses of dredged rock samples from the ridge segment immediately north of the Kane fracture zone, Lawson et al. [1996] suggested moderate fractionation along the ridge segment and lateral dike injection to explain the multiple parental magma chemistries that were obtained at the RTI. In this study, we selected samples of basalt with intact glass to analyze for Fe and Ti content (see Table 3). We normalized the NRM measurements to equatorial values and plotted them versus Fe content (Figure 14). We compared the WMARK data with similarly normalized NRM values from the south MAR [Weiland et al., 1996]. We find the general trend of increasing NRM for greater Fe content [Weiland et al., 1996]. The WMARK rocks fall within the middle range of Fe content and have relatively low NRM values compared with the South MAR data (Figure 14). What is not considered here, however, is that in addition to chemistry, NRM depends on a number of factors such as grain size and age, which can have large effects upon the NRM value of a rock [e.g. Gee and Kent, 1994]. The WMARK RTI rocks are not particularly enriched in Fe or Ti (Table 3) and thus we do not believe the increased fractionation of basalts is responsible for the magnetization high at the end of the WMARK segment. Recent imaging of the seismic layer 2A near the western RTI of the Oceanographer fracture zone on the MAR suggests that layer 2A i.e. the extrusive layer, does thicken towards the transform thus suggesting this is partly the reason for higher magnetism at ridge segment ends [S. Hussenoeder, personal communication 1998].

Conclusions

SHINKAI submersible dives at the western intersection of the Mid-Atlantic Ridge and Kane Transform have provided a set of observations that can be contrasted to the eastern RTI:

The WMARK neovolcanic zone is found to sharply turn into the transform within 1 km of the PTDZ and extend along the transform for ~2.5 km. At the RTI the NVZ transitions from discrete volcanoes into a series of narrow constructional volcanic ridges at an oblique angle to both the transform and spreading direction. These ridges presumably form as a consequence of the rotation of stress trajectories from the spreading regime into the transform regime.

At the RTI, the PTDZ or active trace of the transform is defined by a narrow valley (< 50 m wide) separates the neovolcanic zone to the north from the south wall of the transform. Further east, the PTDZ is marked by a narrow median ridge that skims with the northern wall of the transform. Dives show that the median ridge is marked by mound-like structures cored by highly altered serpentinized peridotites. No evidence was found for any volcanic activity along the western median ridge.

The south wall of the transform at the RTI forms a massif high that exposes diabase and extrusive basalts but no deeper crustal lithologies. This massif is similar to the large intersection highs of other RTI's like the Clipperton [Barth et al., 1994] but in this case, the WMARK massif appears to have formed at the spreading axis during a period of magmatic robustness and while low-angle detachment faulting was not operating. East of the WMARK RTI, the south wall of the transform is displaced ~8 km south of the PTDZ and shows relatively low-angle slopes with rounded topography exposing only lower crustal material. These characteristics typify the remainder of the south wall of the transform wall up to the present day spreading axis and suggest that uplift and unroofing of lower crustal material by low-angle detachment has dominated the seafloor spreading process at the eastern RTI over the last ~10 m.y.

The north wall of the transform formed at the IC of the WMARK segment contrasts quite markedly from the south wall. Dives show that basaltic lithologies are found in exposures from the transform valley to the crest of the IC high. The bathymetry shows a ridge and valley morphology consistent with the volcanic carapace being intact. Crustal densities estimated from submersible gravity measurements and ship-based RMBA gravity [Fujimoto et al., 1996b] both reveal that the north wall of the transform have anomalies consistent with an intact crustal section.

In contrast, the south wall of the transform has high RMBA gravity anomalies [Fujimoto et al., 1996b] and densities from submersible gravity measurements are consistent with lower crust being exposed there. Thus these observations suggest that unlike the eastern RTI, low angle detachment faulting has not been active at the WMARK ridge segment over the last 10 m.y.

Magnetic anomalies over the south wall RTI edifice have a weak yet well-defined magnetic anomaly pattern as might be expected with a relatively intact crustal section. What is surprising is the equally remarkably well-preserved magnetic anomaly patterns that are preserved all along the southern wall of the Kane transform east of the WMARK RTI. The SHINKAI dives on the south wall find lower crustal lithologies exposed at the seafloor with upper crustal units apparently missing. Magnetic anomalies over the south wall remain relatively well preserved however, suggesting that these lower crustal units are capable of preserving the magnetic reversal history with sufficient fidelity to provide a good record of seafloor spreading. The lower crust thus may play a more important role in the preservation of remanent magnetization than previously thought.

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Figure List

Figure 1. a) Bathymetry map of Kane fracture zone (500 m contour interval) shaded from the north showing location of the western and eastern ridge transform intersections. Black box defines the WMARK study area and is shown in more detail in Figure 2. The transform trace and axis of spreading of the MAR, north and south of the transform are shown by bold black line. Dashed line marks fracture zone trace. Map made from WMARK shipboard data [Fujimoto et al., 1994; 1995] and Sea Beam data of Pockalny et al. [1988]. b) Regional magnetization map of WMARK region (1 A/m contour interval) computed using a Fourier transform technique assuming a constant 0.5 km thick source layer whose upper surface is the bathymetry. Polarity chrons are identified based on Cande and Kent [1995] timescale.

Figure 2. Bathymetry contour map of WMARK area (contour interval 100 m) with SHINKAI dive tracks shown by bold lines. White boxes mark the location of submersible gravity stations. The approximate location of the two MIR dives [Zonenshain et al., 1989] are also shown as MIR1 and MIR2. North-south bold line denoted by NVZ marks the neovolcanic zone (NVZ) and axis of spreading. Dashed line marks the principal transform displacement zone (PTDZ) i.e. the presumed trace of the active transform. Black box shows location of detailed bathymetry and geology map of Figure 3.

Figure 3. Detailed bathymetry site map of WMARK RTI region (25 m contour interval) showing SHINKAI dive tracks and geology mapped along dives. Bold line denoted by NVZ marks the neovolcanic zone. Dashed line marks the principal transform displacement zone (PTDZ) i.e. the presumed trace of the active transform.

Figure 4. Geological cross-sections of SHINKAI dives 201, 202 and 204 in the RTI region shown in approximate position relative to each other with no vertical exaggeration. See Figure 3 for locations. Thin lines labeled with depth in meters show bathymetric contours. Dashed line marks the principal transform displacement zone (PTDZ). Arrows and SY station numbers identify sample locations.

Figure 5. Geological cross-section of SHINKAI dive 207 which began in the WMARK nodal deep and traversed west to the relict volcanic ridge mentioned by Zonenshain et al., [1989]. See Figure 2 for location. Arrows and SY station numbers identify sample locations.

Figure 6. Geological cross-sections of SHINKAI dives 208 and 212 which traversed the north side of the western Kane median ridge and shown in approximate position relative to each other with no vertical exaggeration. See figure 2 for dive locations. Thin lines labeled with depth in meters show bathymetric contours. Dashed line marks the principal transform displacement zone (PTDZ). Arrows and SY station numbers identify sample locations.

Figure 7. Geological cross-section of SHINKAI dives 210 and 213 which traversed the south wall of the transform valley at 46° 05’W. Dives are shown in relative position to each other with no vertical exaggeration. Arrows and SY station numbers identify sample locations. Thin lines labeled with depth in meters show bathymetric contours.

Figure 8. Free water gravity anomaly computed from SHINKAI submersible gravity measurements (Table 1) plotted versus water depth (see Luyendyk, 1984). See Figure 2 and Table 1 for station locations. The different symbols are identified in the legend and separate the data into geographically distinct regions. Lines through the data points represent linear regression fits to each set of these geographical data with adjacent numbers identifying the resultant density (i.e. slope) and standard deviations in units of kg/m^3. A slope template for differing density values is also shown.

Figure 9. Terrain corrected Bouguer gravity anomaly computed from SHINKAI submersible gravity measurements (Table 1) plotted versus water depth. See Figure 2 and Table 1 for station locations. Bouguer correction parameters are shown in the inset. Symbols are as in Figure 8. Straight lines through the data points represent linear regression fits to each set of geographically distinct data with adjacent numbers identifying the resultant density (i.e. slope) and standard deviations where more than two points are present in units of kg/m^3. A slope template for differing density values is also shown. The transform valley and median ridge measurements consistently show a greater density than the RTI, Nodal deep and IC high measurements.

Figure 10. Forward model of a north-facing scarp at the Kane transform showing the magnetic field anomaly effect of a 5 A/m magnetized horizontal layer that would be measured along a vertical traverse of the scarp at an altitude of 10 m. Inclination of field is assumed to be 45° and the slope angle is assumed to be 35°.

Figure 11. Forward magnetic model showing the effect of faulted blocks upon the magnetic signal. Right panel shows the model with an upper intact block and three slipped blocks, all with 5 A/m normal polarity magnetization. Left panel shows the calculated magnetic field: solid line is the field due to a single intact block without the slipped blocks, the dashed line is the field due to the integrated effect of all blocks. Note that magnetic lows align with the top of blocks and magnetic highs with the base of blocks. Center panel shows the effect of inverting the calculated magnetic field for crustal magnetization: solid line is for the single intact block, the dashed line is for the integrated effect of all the blocks. Note that there is a complex relationship between magnetization and the blocks. Magnetization highs correlate to where the fault blocks overlap reacting to the effective thickness of the magnetized layer.

Figure 12. Composite vertical magnetic profile and inversion for SHINKAI dives 202 and 204 collected up the steep face of the WMARK RTI southern wall. See Figure 3 and 4 for locations and geology. Left panel shows observed magnetic field. Center panel shows computed magnetization assuming horizontal layers that extend to infinite depth with the zero level defined by the water column above the top of the scarp. Right panel shows the depth profile with discrete zones interpreted from the magnetic data. Zone 1 coincides with the upper faulted block, Zone 2 is a highly faulted section, Zone 3 is a faulted block and Zone 4 is an intact block. These zones correlate with the geological observations shown in Figure 4.

Figure 13. Composite vertical magnetic profile and inversion for SHINKAI dives 210 and 213 collected up the gently dipping scarp of the southern wall of the Kane transform at 46° 05'W. Left

panel shows observed magnetic field. Center panel shows computed magnetization assuming horizontal layers that extend to infinite depth. The top of the scarp was not reached so no absolute magnetization can be assigned, however, a magnetic gradient is seen at 4300 m which could be interpreted as a polarity reversal signal. Right panel shows depth profile with a template of slope angle for reference.

Figure 14. Plot of weight percent Fe content versus NRM normalized to the equator for the WMARK basalt data shown in Table 3 (star symbols). For comparison, we also show data (open circles) from the Southern Mid-Atlantic Ridge [Weiland et al., 1996] where we have also normalized the NRM values to the equator.

Table 1. On-bottom gravity measurements and corrections in mGals. Depth correction for free-water anomaly calculation in seawater is -0.2222 mGal/m. Depth of gravity meter is depth + 2 m. Bouguer gravity correction is based on a three-dimensional terrain correction with a correction crustal density of 2670 kg/m^3 and seawater density of 1030 kg/m^3.

Station	Latitude	Longitude	Depth (m)	Absolute Gravity (mGal)	Depth corrected gravity (mGal)	Free Water Anomaly (mGal)	Bouguer Gravity (mGal)
SY-204-G1	23°48.517'N	46°19.266'W	3640	979638.0	809.3	-45.3	223.6
SY-205-G1	23°50.465'N	46°18.743'W	5240	979871.5	1164.8	-169.5	205.5
SY-206-G1	23°49.682'N	46°18.833'W	4970	979835.9	1104.8	-144.2	214.5
SY-207-G1	23°51.665'N	46°15.422'W	6024	979983.5	1339.0	-233.0	217.3
SY-207-G2	23°51.454'N	46°16.352'W	5695	979939.8	1265.9	-203.4	207.9
SY-207-G3	23°51.269'N	46°17.124'W	5445	979905.8	1210.3	-181.5	207.6
SY-208-G1	23 48.406'N	46°07.594'W	4497	979793.6	999.7	-80.0	238.9
SY-208-G2	23°47.636'N	46°08.295'W	4029	979726.1	895.7	-42.6	240.8
SY-209-G1	23°47.024'N	46°03.401'W	4520	979782.5	1004.8	-94.6	226.6
SY-209-G2	23°48.042'N	46°03.509'W	3816	979685.6	848.4	-36.3	235.8
SY-210-G1	23°43.874'N	46°02.811'W	4828	979835.2	1073.2	-106.8	237.1
SY-210.G2	23°42.741'N	46°03.511'W	4154	979745.2	923.5	-45.9	247.2
SY-211-G1	23°55.844'N	46°01.080'W	2877	979572.7	639.7	50.9	260.4
SY-211-G2	23°56.039'N	46°02.501'W	1481	979335.9	329.5	124.0	259.7
SY-212-G1	23°48.709'N	46°09.154'W	4707	979818.4	1046.3	-102.1	231.9
SY-213-G1	23°42.349'N	46°03.700'W	3853	979702.0	856.6	-21.7	249.7
SY-213-G2	23°41.899'N	46°04.176'W	3690	979681.5	820.4	-5.5	253.4
SY-213-G3	23°41.374'N	46°04.311'W	3507	979654.3	779.7	8.5	255.0
SY-213-G4	23°41.141'N	46°04.646'W	3475	979651.0	772.6	12.6	256.3
SY-215-G1	23°50.271'N	46°14.235'W	5353	979897.7	1189.9	-168.1	218.0

Table 2. Paleomagnetic measurements of WMARK rock samples collected by SHINKAI 6500. Koenigsberger ratio (Q) calculated based on local magnetic field intensity of approx. 40000 nT.

Sample	Latitude	Longitude	NRM (A/m)	Standard Deviation (% of NRM)	χ ($x10^6$SI)	Q ratio	Rock Type
SY-201-1-1	23° 50.36'N	46°18.48'W	36.4	0.84	.52	1.7	Basalt
SY-201-2-1A	23° 50.33'N	46°18.51'W	21.8	0.039	.14	3.8	Basalt
SY-201-2-1B	23° 50.33'N	46°18.51'W	20.5	0.048	.08	6.0	Basalt
SY-201-3-1	23° 49.90'N	46°18.89'W	6.8	0.015	.26	.66	Basalt
SY-201-3-2	23° 49.90'N	46°18.89'W	7.9	0.012	.20	.99	Basalt
SY-201-4-1	23° 49.75'N	46°19.00'W	10.5	0.028	.17	1.5	Basalt
SY-201-5-2	23° 49.71'N	46°19.02'W	1.0	0.025	.84	3.0	Dolerite
SY-201-6-1	23° 49.56'N	46°19.10'W	0.597	0.25	.69	2.2	Dolerite
SY-202-2-3	23° 49.30'N	46°17.47'W	2.2	0.36	.05	.96	Pillow Brec.
SY-203-1-1	23° 49.43'N	46°19.49'W	2.8	0.44	.05	1.2	Basalt
SY-203-3	23° 49.24'N	46°19.76'W	2.9	0.10	.02	2.5	Basalt Brec.
SY-203-4-1	23° 49.23'N	46°19.89'W	5.4	0.32	.08	1.6	Basalt
SY-203-5-1	23° 49.17'N	46°19.91'W	5.7	0.41	.08	1.6	Basalt
SY-203-6-1	23° 49.12'N	46°19.92'W	5.4	0.0056	.08	1.5	Basalt
SY-204-4-1	23° 49.02'N	46°18.72'W	13.7	0.31	.66	.52	Dolerite
SY-204-5-1	23° 48.99'N	46°18.76'W	0.12	0.022	0.00079	3.8	Basalt
SY-204-7-1	23° 48.82'N	46°18.98'W	4.8	0.017	.92	.13	Basalt
SY-205-1-1	23° 50.45'N	46°18.77'W	20.2	0.046	.26	1.9	Basalt
SY-205-2-1	23° 50.35'N	46°18.90'W	18.5	0.10	.32	1.4	Basalt
SY-205-3-1	23° 50.24'N	46°19.04'W	18.3	0.10	.40	1.1	Basalt
SY-205-4-1	23° 50.21'N	46°19.14'W	23.3	0.16	.32	1.8	Basalt
SY-205-5-1A	23° 50.18'N	46°19.16'W	24.3	0.044	.14	4.2	Basalt
SY-205-5-1B	23° 50.18'N	46°19.16'W	18.3	0.026	.17	2.6	Basalt
SY-205-6-1	23° 50.16'N	46°19.18'W	26.1	0.26	.23	2.8	Basalt
SY-205-7-1	23° 50.13'N	46°19.22'W	15.6	0.085	.14	2.7	Basalt
SY-205-9-1	23° 50.09'N	46°19.22'W	21.1	0.069	.14	3.7	Basalt
SY-206-1-1	23° 49.67'N	46°18.86'W	5.6	0.0033	.02	4.9	Basalt
SY-206-1-2	23° 49.67'N	46°18.86'W	2.5	0.18	.14	.45	Basalt
SY-207-2-1	23° 51.44'N	46°16.04'W	7.5	2.0	.43	.44	Basalt
SY-208-3-1	23° 48.24'N	46°07.72'W	2.5	0.16	.20	.32	Serp. perid.
SY-208-4-1	23° 48.01'N	46°07.72'W	0.45	0.27	.49	2.3	Serp. perid.
SY-208-6-1	23° 47.59'N	46°07.97'W	1.5	0.0054	.95	4.0	Serp. perid.
SY-209-1-1	23° 47.37'N	46°03.43'W	7.4	0.028	.40	.46	Basalt
SY-209-2-1	23° 47.64'N	46°03.46'W	1.3	0.10	.81	4.2	Dolerite
SY-210-1-1A	23° 43.75'N	46°02.90'W	0.62	0.10	2.3	6.6	Serp. perid.
SY-210-3-1B	23° 42.82'N	46°03.48'W	2.2	0.54	1.3	4.1	Serp. perid.
SY-211-1-1	23° 55.88'N	46°01.36'W	10.5	4.4	.05	4.6	Basalt
SY-211-5-1	23° 55.98'N	46°01.87'W	10.8	5.3	.08	3.1	Basalt
SY-211-7-1	23° 56.04'N	46°02.65'W	9.5	1.3	.08	2.7	Basalt
SY-212-2-1	23° 47.53'N	46°08.99'W	0.11	0.096	.05	4.6	Meta. gab.
SY-212-3-1	23° 47.40'N	46°09.08'W	0.11	0.010	.11	2.3	Meta. gab.
SY-214-1-1	23° 47.18'N	46°05.71'W	0.071	0.16	.40	4.4	Harzburgite
SY-214-5-1	23° 47.64'N	46°05.78'W	0.56	0.49	.55	2.6	Basalt
SY-215-2-1	23° 50.06'N	46°13.99'W	5.7	1.0	.086	1.6	Basalt/Dol.

Table 3 Geochemical values for selected basalts in weight percent total iron content (Fe0*) and titanium content ($Ti0_2$).

Sample	NRM (A/m)	Fe0*	$Ti0_2$
SY-201-2-1B	20.5	10.4	1.85
SY-201-2-1A	21.8	10.3	1.85
SY-201-3-1	6.8	9.1	1.58
SY-201-3-1B	6.8	9.0	1.60
SY-203-2-1	-	9.4	1.59
SY-203-6-1	5.4	9.75	1.72
SY-204-5-1	0.12	10.0	1.68
SY-205-1-1	20.2	9.0	1.50
SY-205-6-1	26.1	9.0	1.50
SY-206-1-1	5.6	9.6	1.65
SY-207-2-1	7.5	9.4	1.60