Cannat, M., Karson, J.A., Miller, D.J., et al., 1995
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1. INTRODUCTION[1]

Shipboard Scientific Party[2]

INTRODUCTION

Accretion of oceanic lithosphere along the mid-ocean ridge system is a first-order tectonic process. Approximately 60% of Earth's present lithosphere was created at mid-ocean ridge spreading centers through the interplay of magmatic construction, mechanical extension, and hydrothermal alteration. At spreading centers, plate tectonic processes have generated a vast volume of oceanic lithosphere that is recycled back into the mantle at subduction zones. Despite the fundamental importance of oceanic lithosphere, little is known of its detailed internal structure, constitution, and modes of creation.

In the past two decades, extensive studies of the morphology and surficial geology of mid-ocean ridge spreading centers have been conducted. Samples of coarse-grained mafic and ultramafic rocks thought to have formed in the middle to lower oceanic crust and upper mantle have been sampled from seafloor outcrops. However, the geologic setting and vertical dimensions of the rock bodies from which these samples are derived are poorly constrained. Indirect seismic investigations of oceanic crust and studies of ophiolite complexes, considered to be on-land exposures of oceanic crust and upper mantle, provide the foundation of our understanding of the deeper levels of the oceanic crust and upper mantle. However, only direct sampling can provide unambiguous data pertaining to the internal composition and structure of the oceanic lithosphere. Accordingly, the Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) have made the investigation of the oceanic lithosphere a high-priority thematic objective, following the recommendations of the Conferences on Scientific Ocean Drilling (COSOD I and II reports).

Drilling the hard-rock foundation of the ocean floor has proven to be a daunting engineering challenge. The first holes were attempted during DSDP Leg 37 and continued on several subsequent legs. The goal of drilling through the entire thickness of the crust has been difficult to attain because of the drilling problems posed by the highly fractured nature of the basaltic carapace that typically characterizes surface exposures. After seven legs, Hole 504B has reached a depth of just over 2100 meters (m) into the oceanic basement through variably fractured and altered basaltic pillow lavas and sheeted diabase dikes. Present hole conditions make further penetration doubtful.

Following success in drilling plutonic rocks exposed in locations where the oceanic crust had been stripped of its overlying fine-grained basaltic layers on Legs 109 and 118, a strategy for drilling a series of relatively shallow, closely spaced holes that could be correlated laterally (offset drilling) was proposed. The report of the Offset Drilling Working Group recommended a program directed toward collecting long, continuous sections of deep crustal rocks that could be used to erect composite sections of the oceanic crust created in various spreading environments.

The first holes of this program were drilled during Leg 147 in the walls of the Hess Deep Rift of the Equatorial Pacific. This exposure of the plutonic foundation of 1 Ma crust formed at the fast-spreading (132 mm/yr) East Pacific Rise is the result of amagmatic rifting at the westward-propagating tip of the Cocos/Nazca plate boundary. Drilling at two nearby sites yielded cores of coarse-grained gabbroic and serpentinized ultramafic rocks, some of the first substantial samples of deep crustal and upper mantle rocks from any fast-spreading ridge. Subsequent legs are planned to further develop the goal of a complete, albeit composite, section through all of the mafic and ultramafic units exposed in this area.

Leg 153 represents a second ODP offset drilling leg, in this case devoted to drilling the plutonic foundation of slow-spreading (~25 mm/yr) oceanic lithosphere in the Mid-Atlantic Ridge (MAR) at the Kane Transform (MARK) area (Figs. 1 and 2). Exposures of coarse-grained mafic and ultramafic rocks on the median valley walls are considered to be <1 m.y. old and to have resulted from mechanical extension of the axial lithosphere during a period of very limited magmatic construction. The sites drilled during Leg 153 are located in the footwalls of major detachment faults cutting ultramafic and gabbroic rocks. The primary goals of drilling include the investigations of the mantle flow and partial melting history of the ultramafic rocks, the igneous crystallization history of the gabbroic rocks, and the deformation and metamorphism attending seafloor spreading in this particular environment.

Previous drilling in this area included penetration into the basaltic upper crust of Serocki Volcano about 60 km south of the Kane Transform (Site 648, Leg 106; Fig. 3). During the same leg, several holes were drilled into sulfide mineralized deposits near active black-smoker vents at the Snake Pit Hydrothermal Area (Site 649; Fig. 3). Site 649 is located at the crest of a neovolcanic ridge that lies near the center of the median valley and extends more than 40 km southward from the eastern intersection of the Mid-Atlantic Ridge and the Kane Transform.

During Leg 109, two sites were drilled in geologic settings similar to those of Leg 153. Site 669 (Fig. 3) was drilled near the top of the western median valley wall of the MAR just south of the Kane Transform. This area lies on the eastern flank of a broad uplifted area referred to as a ridge/transform intersection massif (Fig. 3). Variably deformed gabbroic rocks crop out extensively from depths of >6000 m to about 2000 m on a 30°–35° slope, essentially a dip-slope of a major detachment fault. Site 669 (Figs. 3 and 4) is located in highly fractured and degraded basaltic rocks that crop out across the summit of the massif (1980 meters below sea level [mbsl]). An unsupported spud-in at this site penetrated only 4 m of basaltic rubble and intermixed pelagic sediments. After the penetration rate dropped dramatically, the site was abandoned. Leg 153 Sites 921, 922, and 923 are located on gabbroic rocks cropping out lower on the slope at a water depth of approximately 2500 m (Figs. 3 and 4). Site 924 is approximately 1.5 km downslope from Sites 921 and 922, at approximately 3150 mbsl (Fig. 4).

Site 670 is also situated on the western rift valley wall about 35 km south of the Kane Transform. It is some 18 km south of Site 920 (Figs. 3 and 5), in a 2-km-wide belt of serpentinized peridotite that extends at least 20 km along the western median valley wall. The serpentinites were discovered during *Alvin* dives that took place during Leg 109. The last four days of Leg 109 were devoted to an unsupported spud-in and drilling at Site 670 in the serpentinites. Penetration rates were good, and a depth of nearly 100 meters below seafloor (mbsf) was reached. Serpentinized harzburgite and lherzolite, with

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[2] Shipboard Scientific Party is as given in the list of participants preceding the Table of Contents.

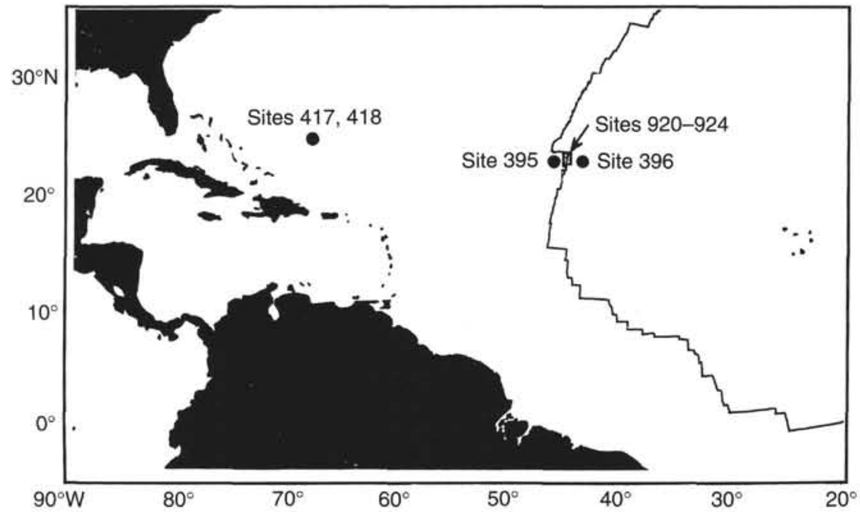

Figure 1. Box shows location of the Leg 153 MARK sites on the Mid-Atlantic Ridge (Fig. 2). Also shown are Sites 417 and 418 (DSDP Legs 51, 52, and 53) and Sites 395 and 396 (DSDP Legs 45 and 78B and ODP Legs 106 and 109).

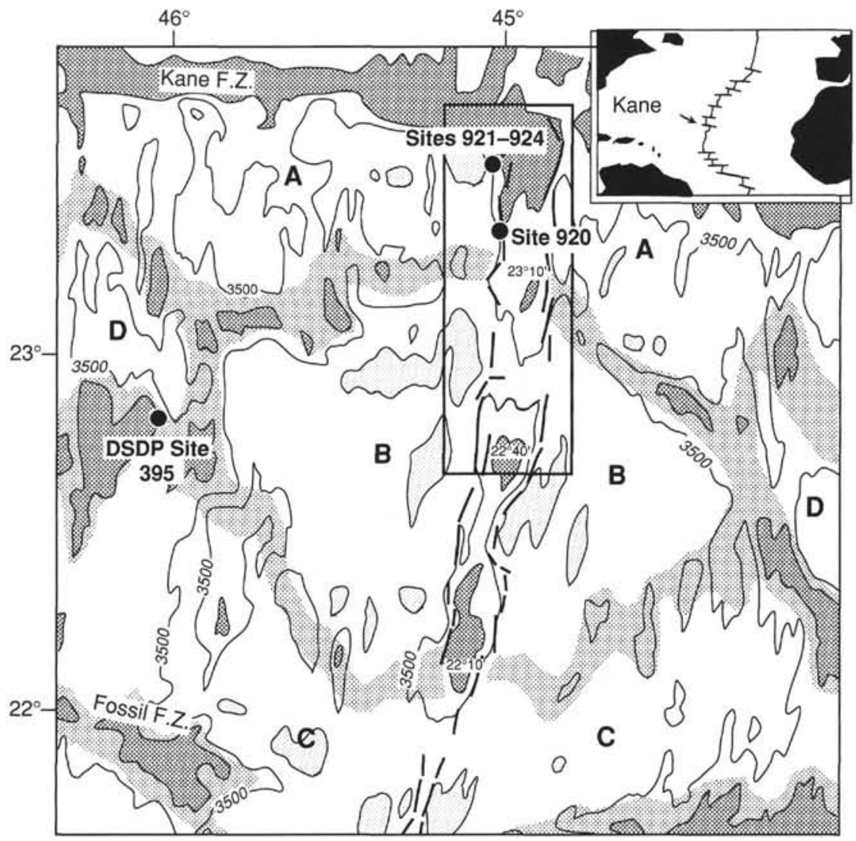

Figure 2. Generalized bathymetric map of the 22°–24° N region, after Gente et al. (1991a). The main scarps bounding the axial valley walls are shown in bold lines. F.Z. = fracture zone; dark gray stippling = areas deeper than 4000 m; light stippled pattern = areas shallower than 2500 m. Off-axis alignments of basins, thought to have formed at the tip of magmatic segments, are shaded (medium stippled pattern). They bound four off-axis rhombohedral domains (A to D), which may have formed at spreading segment centers. DSDP Site 395 is located in one of the basins bounding these domains. Present-day axial segmentation in the region is associated with the Kane Transform and with nontransform axial discontinuities (23°10′N, 22°40′N, and 22°10′N), two of which (23°10′N, 22°10′N) have off-axis traces. The trace of a fossil fracture zone at 22°N is deduced from bathymetry and magnetics. The box shows the location of Figure 3.

rare intervals of dunite and pyroxenite, were cored, with poor recovery rates (7% average).

Although they do not lie in the MARK area, several other DSDP and ODP holes lie along a seafloor spreading flow line from those of the MARK area. Basaltic lavas were drilled at Site 396 holes in 10 Ma crust of the African Plate, east of the MARK area (Melson, Rabinowitz, et al., 1979; Dmitriev, Heirtzler, et al., 1979). At Site 395, in 7 Ma crust of the North American Plate, about 100 km west of the MARK area (Figs. 1 and 2), basaltic pillow lavas and flows were found intercalated with sedimentary breccias containing clasts of gabbroic and ultramafic rocks. Even farther west, DSDP Sites 417 and 418 (Fig. 1) penetrated 108 Ma basaltic rocks along the same flow line from the MARK area (Donnelly et al., 1979).

Thus, despite the low recovery rates achieved during previous drilling efforts in the MARK area, valuable scientific results and drilling experience were obtained. Whereas drilling of the poorly consolidated hydrothermal deposits of the Snake Pit provided new insights into black smoker hydrothermal systems, no basement penetration was achieved. Attempts to drill highly fractured pillow lavas at Serocki Volcano and on the crest of the intersection massif made it clear that new techniques would be necessary to drill these lithologies. However, the rapid penetration of MARK area serpentinized

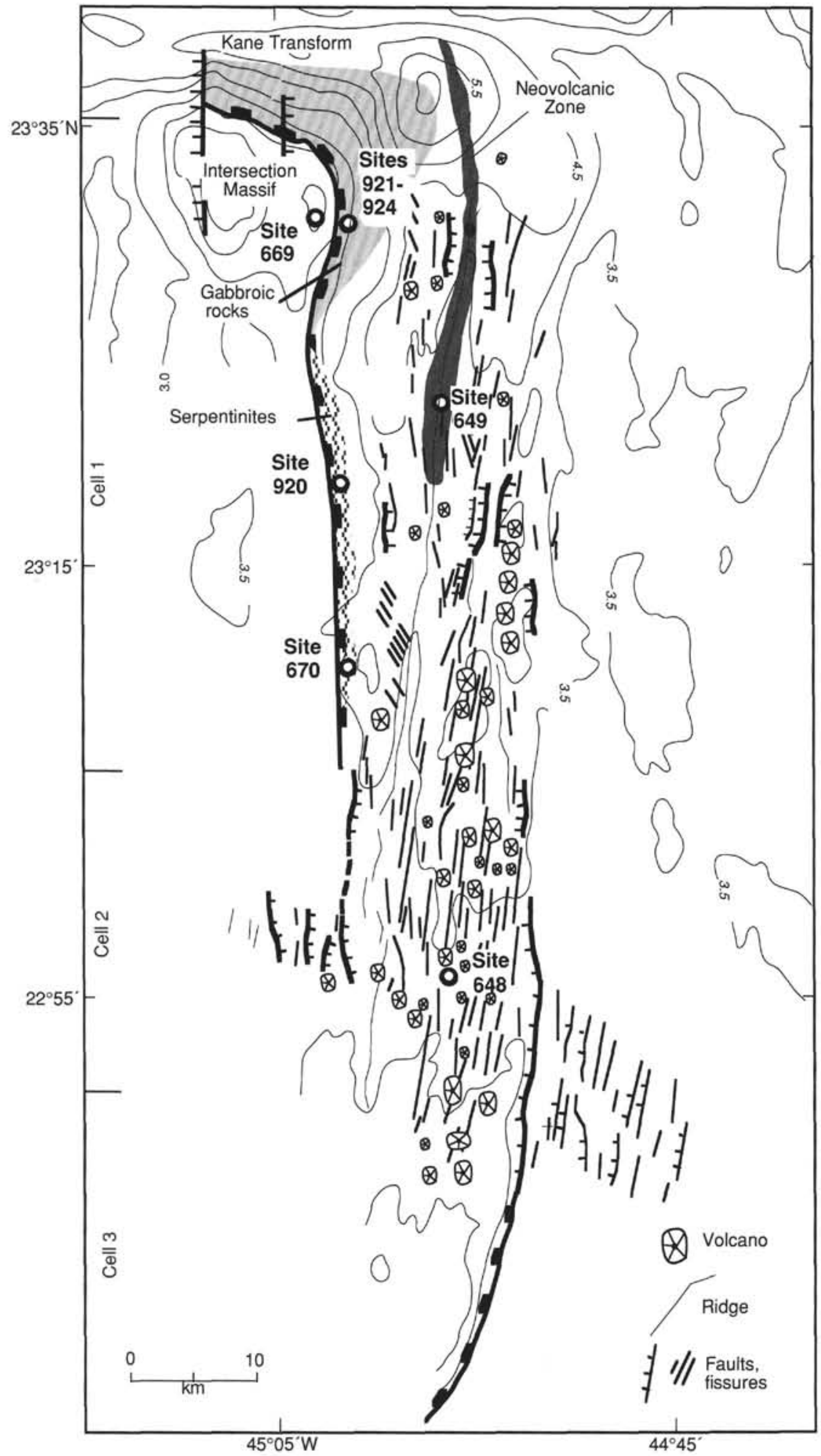

Figure 3. General bathymetric map of the MARK area. ODP sites shown as open circles. Shading represents area of outcrop of gabbroic rocks, zig-zag line represents area of serpentinite outcrops, identified from submersible studies and dredging. Discrete spreading segments, or cells, described in text. Contours in kilometers below sea level.

peridotites during Leg 109 demonstrated that hard-rock drilling was possible in this rock type. In addition, drilling on the Southwest Indian Ridge during Leg 118 showed that good penetration and recovery rates could also be achieved in gabbroic rocks, at least in that anomalous shallow-water site. For Leg 153, ODP returned to the MARK area with new tools and techniques that had been developed over the course of several legs of hard-rock drilling. With these new approaches, it was hoped that significant progress could be made in drilling the major rock units exposed in the area and insights into slow-spreading ridges could be gained.

SCIENTIFIC OBJECTIVES

The goals of drilling in the MARK area were to investigate the composition, structure, magnetic and other physical properties of gabbroic and ultramafic rocks exposed on the rift valley walls of the

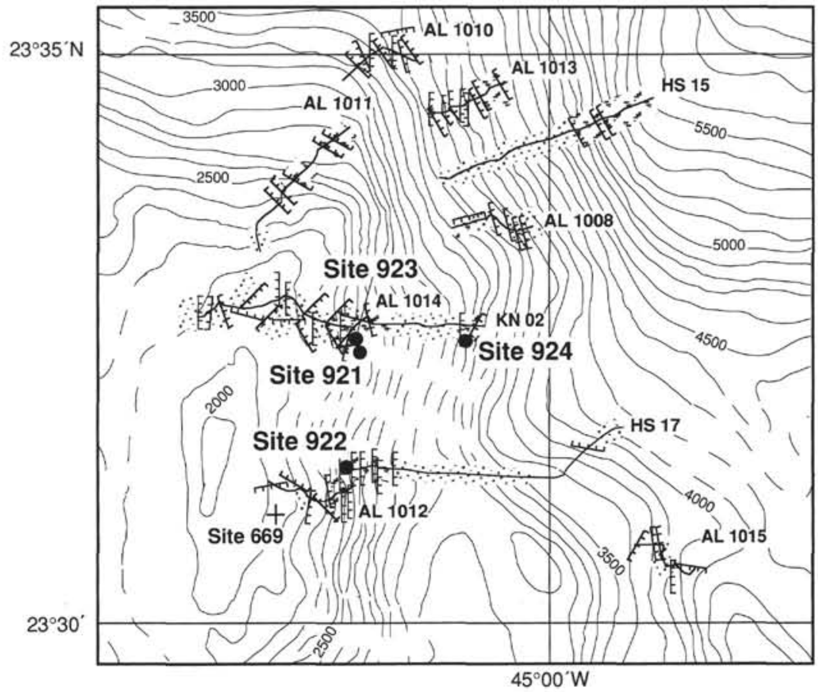

Figure 4. Bathymetric map of the western wall of the MAR in the MARK area about 5 km south of the Kane Transform. Dive tracks of *Alvin* (AL 1008, AL 1010–1014) and *Nautile* (HS 15, 17, and KN 02) show numerous east-facing escarpments and extensive exposures of variably deformed and metamorphosed gabbroic rocks (shading) and rubble derived from these exposures (stippled pattern). Sites 921–923 are located among a series of cliffs of gabbroic rock at about 2500 m water depth. Site 924 is deeper (about 3400 mbsl).

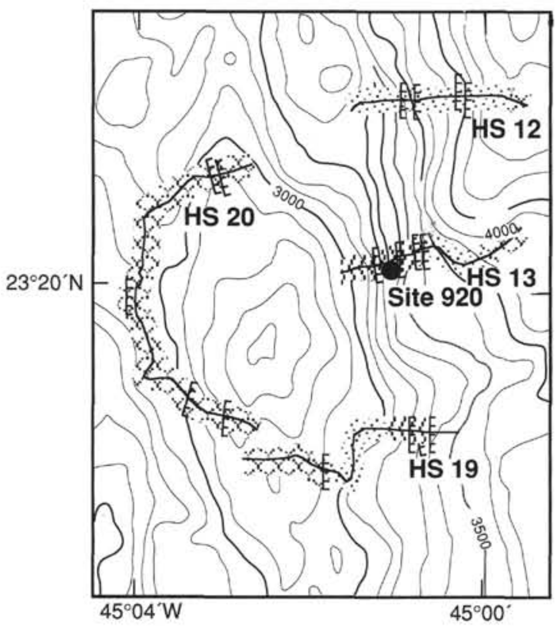

Figure 5. Bathymetric map of the western wall of the MAR in the area surrounding Site 920. Tracks crossing contours are *Nautile* dives (HS 12, 13, 19, and 20; Mével et al., 1991). Contours are in meters below sea level. Zig-zag lines = serpentinized peridotite outcrops; scalloped pattern = pillow basalt outcrops; stippling = pelagic sediments ± blocks; hatched lines = fault scarps (hatches on downthrown side).

slow-spreading Mid-Atlantic Ridge. In these rock types, a complete description of the igneous, metamorphic, and structural features is required to elucidate the processes attending seafloor spreading. Detailed, continuous sampling of these rock types can only be accomplished by high-recovery drilling.

It was anticipated that the ultramafic rocks would yield information on mantle flow under both asthenospheric and lithospheric conditions; melt generation, segregation, and transport within residual upper mantle material beneath the ridge axis; and high- to low-temperature hydration and deformation of ultramafic rocks along fault zones associated with seafloor spreading. Studies of gabbroic rocks were expected to reveal the igneous evolution of mantle melts that reside in the middle to lower crust and the retrograde deformation and hydrothermal alteration associated with major oceanic fault zones. Detailed paleomagnetic and other physical properties studies of these rock types were expected to shed light on the magnetic, gravity, and seismic expressions of the oceanic crust and upper mantle.

Long, continuous sections of mafic and ultramafic rocks have the potential to be reoriented by integrating the results of logging and paleomagnetic studies. Thus, it was anticipated that structures found in cores could be reoriented with respect to geographic coordinates, allowing them to be evaluated in the context of the slow-spreading ridge axis and outcrop-scale features previously mapped by camera sleds and manned submersibles.

GEOLOGICAL BACKGROUND

Previous Work in the MARK Area

The Mid-Atlantic Ridge near 23°N and the Kane Fracture Zone, known as the MARK area, is the most comprehensively studied portion of the MAR (Table 1). Investigations in the area date back to initial conventional wide-beam echo sounding surveys and dredging in the late 1960's (Miyashiro et al., 1969, 1970; van Andel et al., 1969; Fox, 1972). These and subsequent geophysical studies (Purdy et al., 1979; Schouten et al., 1985) demonstrated the general character of the ridge in this area and the diverse rock types that are exposed in the median valley and transform fault walls. Geophysical studies in the 1980's documented large variations in crustal structure in the Kane Transform (Detrick and Purdy, 1980; Louden and Forsyth, 1982; Cormier et al., 1984) and along the axis of the median valley south of the ridge-transform intersection (Purdy and Detrick, 1986). Additional dredging (Bryan et al., 1981), deep-towed *ANGUS* camera surveys and *Alvin* submersible investigations (Karson and Dick, 1983) provided constraints on the distribution of major rock units in the median valley near the ridge-transform intersection. *Alvin* dives documented extensive gently dipping normal fault surfaces in gabbroic

Table1. Geological and geophysical investigations in the MARK area.

Study	Technique	Area	References
Bathymetry	Single channel echo-sounder	22°–24°N on the MAR	Purdy et al., 1979
	SeaBeam	MAR axis 22°–24°N and Kane F.Z.	Kong et al., 1988
	SeaBeam	MAR axis 24°–22°30′N (gaps)	Karson et al., 1987
	Simrad	MAR and seafloor to 10 Ma 24°–20°N	Gente et al., 1991a
Magnetics	Surface ship	MAR and seafloor to 10 Ma 24°–20°N	Gente et al., 1991a
	Surface ship	MAR axis 22°–24°N and Kane F.Z.	Schulz et al., 1988
Gravity	Surface ship	Kane Transform	Louden and Forsyth, 1982
	Surface ship	MAR axis 22°–24°N and Kane F.Z.	Morris and Detrick, 1991
	Surface ship	MAR and crust to 10 Ma 24°–20°N	Deplus et al., 1992
	Near-bottom (from *Nautile*)	MAR axis at 23°20′N	Dubois et al., unpubl. data; Dubois et al., 1992
Seismics	Refraction	Kane F.Z.	Detrick and Purdy, 1980
	Refraction	MAR axis 22°–24°N (Kane F.Z.)	Purdy and Detrick, 1986
	Delay-time studies	Nodal basin at E RTI of Kane F.Z.	Cormier et al., 1984
	MCS reflection	MAR median valley 22°30′–24°N	Detrick et al., 1990; Mutter and Karson, 1992
	Microseismic network	MAR median valley at 22°40′N	Toomey et al., 1985, 1988
Side-scan sonar	SeaMARC I	MAR axis 22°15′–23°30′N	Kong et al., 1988
	TOBIE	W. Kane Transform and W RTI	Searle et al., 1992
	Deep Tow: Up-/downlooking sonar	E. Kane Transform and E RTI	Karson et. al., 1992b
Near-bottom photogeology	ANGUS	Kane F.Z. and MAR at E and W RTI's	Karson and Dick, 1983
	ANGUS	MAR 22°50′–23°N	Karson et al., 1987; Brown and Karson, 1988
	Deep tow	E. Kane F.Z. and MAR at E RTI	Karson et al., 1992b
	SCAMPI	MAR 22°–22°30′N	Cannat et al., unpubl. data
Submersible dives	*Alvin*	W. Wall, E RTI	Karson and Dick, 1983
	Alvin	MAR median valley 23°15′–22°55′N	Karson et al., 1987; Brown and Karson, 1988
	Nautile	MAR median valley 23°40′–23°15′N	Mével et al., 1991; Gente et al., 1991b
	Nautile	Kane F.Z. at E RTI	Auzende et al., 1993
	Nautile	MAR median valley 23°40′–23°15′N	Dubois et al., unpubl. data
Ocean Drilling Program	*JOIDES Resolution*, Leg 106	MAR Axis at 23°15′N (Snake Pit Vent site)	Detrick, Honnorez, Bryan, Juteau, et al., 1988
	JOIDES Resolution, Leg 109	MAR axis at 22°55′N and 23°15′N	Detrick, Honnorez, Bryan, Juteau, et al., 1988
	JOIDES Resolution, Leg 153	MAR median valley wall at 23°31′N and 23°05′N	Unpubl. data
Dredging	Dredging	MAR median valley at 22°–23°N	Melson et al., 1968
	Dredging	Kane Transform	Miyashiro et al., 1969, 1971
	Dredging	MAR median valley at 22°–24°N	Bryan et al., 1981
	Dredging and *Alvin*	MAR at 24°N	Karson and Dick, 1983
	Dredging and *Alvin*	MAR median valley at 22°–24°N	Karson et al., 1987; Bryan et al., 1994
	Dredging	MAR median valley at 22°–24°N	Zonenshain et al., unpubl. data; Gente et al., 1989
	Dredging	MAR and seafloor to 4 Ma, 21°–24°N	Cannat et al., 1993

rocks on the western median valley wall of the rift valley (Dick et al., 1981; Karson and Dick, 1983).

Prior to ODP Leg 106, a major site survey resulted in a complete SeaBeam bathymetric map of the Kane Fracture Zone and adjacent ridge segments to the north and south (Detrick et al., 1985). This included approximately 100 km of the median valley south of the Kane Transform. The site survey included the collection of gravity (Morris and Detrick, 1991) and magnetics (Schulz et al., 1988) data, which are summarized below. In addition, a SeaMARC I side-scan sonar study of the area provided a comprehensive view of the morphology and large-scale structure of the entire rift valley for nearly 100 km south of the Kane Transform (Kong et al., 1988). ODP Legs 106 and 109 were conducted in the framework of these investigations. During Leg 109, an *Alvin* and *ANGUS* study of the rift valley and adjacent walls in the MARK area provided detailed sampling, structural data, the first detailed look at the Snake Pit hydrothermal vents, and geological evidence of different spreading styles along the ridge axis in the MARK area (Karson et al., 1987; Brown and Karson, 1988; Karson and Winters, 1992).

Since 1986, substantial new results have been obtained from studies of the MARK area. Many of these are the result of French-American joint study of the Mid-Atlantic Ridge, which is referred to as FARA. These explorations have included a series of *Nautile* dives on the neovolcanic zone, the Snake Pit, and the western median valley wall (Mével et al., 1991). A second *Nautile* study examined variations in the crustal structure exposed along the southern wall of the Kane Transform out to crustal ages of about 5 Ma (Auzende et al., 1993).

More recently, an on-bottom gravity survey of the Snake Pit and adjacent axial valley walls was conducted with *Nautile* (Dubois et al., unpubl. data). This survey extended off-axis to the west, into 3 Ma lithosphere. Other recent efforts to characterize the past history of spreading in the MARK region include a survey of bathymetry (Fig. 2), gravity, and magnetic properties of the lithosphere on both sides of the ridge axis out to ages of 10 Ma (Gente et al., 1991a). This survey also extended the along-axis coverage of the MAR as far south as 20°N (Fig. 2). Off-axis dredging in the context of these data was conducted during a later cruise (Cannat et al., 1993).

A side-scan sonar study of the western Kane Transform was carried out by a British team (Searle et al., 1992). In 1992, a side-scan sonar survey of the entire eastern ridge-transform intersection and southern wall of the Kane Transform was completed using an up- and down-looking sonar configuration (Karson et al., 1992b). Multichannel seismic data collected from along and across the rift valley contributed to a three-dimensional view of the crustal architecture (Detrick et al., 1990; Mutter and Karson, 1992).

The above and related studies (Table 1) have provided a huge amount of data with which to constrain the geology and crustal structure of the ridge-transform intersection area and adjacent parts of the transform and rift valley. Drilling operations during Leg 153 sought to investigate details of major processes related to seafloor spreading in the rift valley just south of the Kane Transform in the context of this extensively documented segment of the Mid-Ocean Ridge.

Regional Setting

A SeaBeam bathymetric map of the MARK area was produced during the site survey for Leg 106 (Detrick et al., 1984). This map has served as the base map for all subsequent studies and has been the basis for morphological and large-scale structural investigations (Kong et al., 1988; Pockalny et al., 1988). Magnetic data also collected during the Leg 106 site survey were interpreted by Schulz et al. (1988). The off-axis distribution of magnetic anomaly lineations indicates asymmetrical spreading with about 14.1 mm/yr to the west and 11.3 mm/yr to the east. Possible duplication of anomalies to the west suggest that discrete eastward ridge jumps may also have occurred.

Off-axis swath mapping (Fig. 2), gravity, and magnetic survey coverage (Gente et al., 1991a; Deplus et al., 1992) shows that the lithosphere created along this region of the MAR during the last 10 Ma is segmented into rhombohedral domains of relatively shallow depth and negative residual mantle Bouguer anomalies (rmBa); these domains are inferred to be underlain by relatively thick crust, and they are bounded by areas of relatively greater depth and positive rmBa (Gente et al., 1991a; Deplus et al., 1992), inferred to be areas of thinner crust. The dimensions of the rhombohedral domains are a few tens of kilometers parallel to both isochrons and flow lines (Fig. 2), suggesting a periodicity of 2 to 6 m.y. for the creation of these domains. Each rhombohedral domain is interpreted as having been created along a ridge segment, with maximum magma supply and consequently thicker magmatic crust at the segment center, and minimum magma supply, resulting in thinner crust, at the segment ends. The serpentinized peridotite outcrops of the MARK area and DSDP Site 395 (where serpentinite-bearing breccias were drilled close to the surface) are located in domains of greater than average depth and positive rmBa interpreted as the off-axis traces of magma-poor ridge segments ends. Serpentinized peridotites have also been dredged off-axis of the MAR in these inferred thin-crustal domains (Cannat et al., 1993).

The MAR Axis in the MARK Area

For nearly 200 km south of the Kane Transform, the MAR is relatively linear and has no major transform offsets (Figs. 1 and 2). The floor of the rift valley (below 3500 m depth) varies in width from 14 to 20 km, with relief ranging from less than 1 to more than 4 km from the valley floor to the adjacent crestal mountains. In general, the inner floor deepens and broadens toward the north from <3500 mbsl near the saddle at 22°55′N to >6100 mbsl in the nodal basin, marking the intersection with the Kane Fracture Zone (Fig. 3). Superimposed on this regional pattern are a series of discrete spreading segments or spreading cells that are defined on the basis of their morphology, geological character, and geophysical characteristics. These segments are several tens of kilometers in length and appear to represent portions of the MAR that are distinguished by somewhat different spreading processes.

Leg 153 drill sites are located along the western wall of the Mid-Atlantic Ridge axial valley in the MARK area (Fig. 3), where extensive low-angle normal faults, or detachment faults, as well as gabbro and peridotite outcrops have been documented. The drill sites are located in a discrete spreading segment or cell located immediately south of the eastern intersection of the Kane Transform and the MAR. This particular ridge segment has experienced a history of extreme mechanical extension over the past 1 m.y. that has resulted in the widespread exposure of rock types generally considered to be derived from the middle to lower crust and upper mantle.

A broad dome-like uplift on the order of 20 km across occupies the lithosphere plate corner bounded to the north by the Kane Transform and to the east by the median valley of the MAR. This is referred to as an inside-corner high or ridge-transform intersection massif. The opposing median valley wall on the east side of this ridge segment is substantially deeper (~3500 m), less steep, and has exposures of only block-faulted basaltic pillow lavas. This type of highly asymmetrical morphology and distribution of volcanic and plutonic rock types is typical of ridge-transform intersections along slow-spreading ridges (Tucholke and Lin, 1994) and may also characterize ridge segments far from transform faults where extreme mechanical extension and uplift have occurred (Karson, 1991).

A median neovolcanic ridge extends southward from the nodal basin for the entire length of this northern spreading cell. It is approximately 40 km long, up to 4 km wide, and as much as 400 m high. This volcanic edifice is estimated to be approximately 5000 yr old based on its sparse pelagic sediment cover. The Snake Pit hydrothermal field (ODP Leg 106 Scientific Party, 1986; Fig. 3) is located on the summit of this ridge.

A wide transition zone (23°18′N to 23°05′N) separates the two cells (Fig. 3). This transition zone corresponds with both a probable offset of magnetic anomalies (Schulz et al., 1988) and a marked change of rift valley morphology (Karson et al., 1987; Kong et al., 1988). It has been interpreted as a zero-offset transform fault (Purdy and Detrick, 1986) or as an accommodation zone linking two regions of different spreading styles (Brown and Karson, 1988; Karson and Winters, 1992).

The southern spreading cell is located at a local high point along the ridge axis. It has a simple, *U*-shaped form and is underlain by fissured and block-faulted basaltic pillow lavas and numerous small conical volcanic structures (Kong et al., 1988; Brown and Karson, 1988; Humphris et al., 1990). Near the center of this segment lies Serocki Volcano (ODP Leg 106 Scientific Party, 1986; Site 648).

The ridge axis to the south has an asymmetry that is the opposite of the northern cell's, with a relatively high and steep eastern median valley wall and a more subdued western wall. Low-angle normal faults and basaltic pillow lavas with extensive diabase dikes crop out on the eastern wall, while only block faulted pillow basalts are found on the western wall. Relatively young basaltic cones are dispersed across the old, sediment covered and fissured median valley floor (Lawrence et al., unpubl. data). Seismic studies in the area demonstrate active faulting beneath the eastern median valley wall that extends downward into the axial upper mantle (Toomey et al., 1985, 1988). This seismically active spreading cell is interpreted as a major half-graben structure bounded by a juvenile detachment fault (Lawrence et al., unpubl. data). The cell boundary zones to the north and south of this segment are not yet defined.

Based on a seismic survey, Purdy and Detrick (1986) suggested that crustal thickness in the MARK area decreases along-axis towards the transform intersection. These authors defined two spreading cells along the MARK area axis, with contrasting crustal thicknesses. The northern cell extends from the nodal basin down to 23°18′N and has a seismic crust 4–5 km thick; the southern cell has a seismic crust 6–7 km thick (Purdy and Detrick, 1986). Residual gravity anomalies calculated for the MARK area by Morris and Detrick (1991) agree with the seismic data and suggest thicker crust or higher mantle temperatures south of the 23°18′N to 23°05′N axial discontinuity. They also show positive residual mantle Bouguer anomalies, suggesting thin crust beneath this axial discontinuity. Similar relations have been inferred beneath many other nontransform segment boundaries along the MAR (Lin et al., 1990; Kuo and Forsyth, 1988; Patriat et al., 1991; Deplus et al., 1992) and have been attributed to a focusing of the magma supply beneath ridge segment centers and attenuation of magmatism near segment boundaries.

Multichannel seismic reflection surveys in the central North Atlantic have demonstrated that the Atlantic oceanic crust is characterized by major middle to lower crustal reflectors that generally dip toward younger crust (i.e., toward the spreading center where the crust was formed) and also toward nearby fracture zones (Mutter et al., 1985; White et al., 1990; Morris et al., 1993). Mutter and Karson (1992) interpreted these reflectors as major detachment faults that penetrate the entire thickness of the crust and correlated them with the major detachment fault structures of the median valley walls of the MARK area. Multichannel seismic studies of the axial region of the MARK area are hampered by high reflectivity of the basalt-water interface and rough topography. However, gently inclined reflectors also have been recognized in this young axial lithosphere. Dipping reflectors, observed in an axial seismic line, suggest major faults interpreted as lateral ramps linking detachments in a complex three-dimensional extensional system (Mutter and Karson, 1992). Reflectors in seismic lines crossing the median valley have not been correlated with surface geological features. However, subtle reflectors dipping toward the axis are present beneath Site 920, suggesting shear zones or high-impedance lithologic units in the serpentinized peridotite (Mutter and Karson, unpubl. data). Detrick et al. (1990) used these same data to demonstrate that no axial magma chamber similar to that of the East Pacific Rise exists beneath the neovolcanic zone of the northern cell.

Geochemically, the basaltic rocks recovered from the MARK region appear to indicate relatively low degrees of mantle melting or depletion. As suggested by Klein and Langmuir (1987), the $Na_{(8.0)}$ (the "corrected" value of Na_2O to 8.0 wt% MgO) and $CaO_{(8.0)}/Al_2O_{3(8.0)}$ value (similarly recast to 8.0 wt% MgO) in basalt glasses can be used as a measure of the extent of mantle melting or depletion in the mid-ocean ridge basalt (MORB) source. The value of $Na_{(8.0)}$ in the MARK area glasses varies between 2.6 and 3.1, and with the exception of several high-alumina basalts, the $CaO_{(8.0)}/Al_2O_{3(8.0)}$ value of basalts varies between 6.8 and 7.3 (Humphris et al., 1990; Langmuir et al., 1992; Bryan et al., 1994; Casey et al., unpubl. data). These values have been shown to be among the highest and lowest along the Mid-Atlantic Ridge between 0° and 50°N (Casey et al., unpubl. data). The MARK area would appear to represent a minimum in melting or mantle depletion along this segment of the ridge axis.

Machado et al. (1983) have also shown that basalts from the region are isotopically typical of normal-MORB (N-MORB), being among the most depleted in $^{87}Sr/^{86}Sr$ and enriched in $^{143}Nd/^{144}Nd$ along the northern MAR axis. Basalts from this region also have relatively unradiogenic Pb-isotope compositions (Kempton et al., 1993). Trace element data suggest that most MARK basalts are light rare earth element (REE) depleted, typical of N-MORB (Bryan et al., 1981; Humphris et al., 1990) with a few exceptions that tend to exhibit flat REE patterns and can be classified as transition-MORB (T-MORB) (Casey and Bryan, unpubl. data). These data suggest that basalts from the MARK area lie near the nonplume end-member of the compositional spectrum of MORBs.

These geochemical data are consistent with mineral chemistry studies of residual mantle rocks in the region. The Al_2O_3 content of clinopyroxene and orthopyroxene and the Cr/Cr + Al (Cr#) for spinel can, to a first approximation, be used as an index of the extent of mantle melting and depletion (e.g., Dick, 1989). The Cr# of spinel increases with increasing degrees of melting and values from the MARK region generally lie between 0.22 and 0.44 (e.g., Komor et al., 1990; Juteau et al., 1990; Casey, unpubl. data). The Al_2O_3 content of pyroxene will decrease as melting increases and can be used as a measure of mantle depletion. The Al_2O_3 content lies between 3.2 and 5.0 wt% for mantle orthopyroxene and between 4.4 and 6.0 wt% for mantle clinopyroxene in the MARK region (Komor et al., 1990; Juteau et al., 1990; Casey, unpubl. data; Tartarotti et al., unpubl. data). These are again among the lowest values of Cr# for spinel and the highest Al_2O_3 values for mantle pyroxenes along the northern MAR and have similarly been interpreted to represent a minimum in the extent of mantle melting or depletion along the N. MAR between 0° and 50°N (Casey et al., unpubl. data).

The Ridge/Transform Intersection Massif

Dredges (Miyashiro et al., 1969, 1971) as well as *Alvin* (Dick et al., 1981; Karson and Dick, 1983) and *Nautile* (Mével et al., 1991; Auzende et al., 1993) dives along the east- and north-facing slopes of the ridge/transform intersection massif (Fig. 3) have documented extensive outcrops of gabbro, metagabbro, metabasalt, metadiabase, and lesser serpentinized peridotite. Samples from these outcrops commonly display slickensides, shear zones, or schistosity. The outcrops are marked by moderately dipping faults and shear zones defining a nearly continuous dip-slope. Some fault zones can be followed for several hundred meters upslope and may correspond to major detachment faults. They dip to the east-northeast along the eastern face of the massif and to the north and northeast along the northern transform-parallel wall. These gently dipping fault zones commonly are cut by higher angle faults that impart a stepped pattern to the edges of the massif. Faults on the eastern side are normal faults, whereas faults with moderately to gently plunging striae occur along the transform valley wall. The summit of the massif, however, is covered with highly fractured and degraded basaltic rocks.

The gabbroic samples collected from the intersection massif include variably deformed and metamorphosed olivine gabbro through gabbronorite and ferrogabbro to trondhjemite (Karson and Dick, 1983; Mével et al., 1991; Marion et al., 1991; Marion, 1992). Deformation structures range from brittle to ductile, and metamorphic assemblages indicate seawater penetration and hydrothermal alteration (Kelley and Delaney, 1987; Marion et al., 1991; Marion, 1992; Kelley et al., 1993; Gillis et al., 1993). Along the eastern and northern walls of the massif, mafic to ultramafic plutonic rocks crop out between 6000 and 2200 mbsl.

A series of *Nautile* dives (Auzende et al., 1993) and a side-scan sonar and photographic survey along the southern transform wall to the west of the intersection massif (Karson et al., 1992a) have documented outcrops of variably tectonized gabbro, metagabbro, serpentinized peridotite, metadiabase, and metabasalt. These studies suggest that the geology of the ridge/transform intersection massif, with variously deformed and metamorphosed plutonic rocks exposed in the footwall of large, gently dipping fault zones, has persisted over the past 5 m.y.

Outcrops of Serpentinized Peridotites on the Western Rift Valley Wall

Two serpentinite outcrops were discovered during *Alvin* dives on the western wall of the axial valley (Karson et al., 1987): one at 23°10′N and the other at 23°21′N. Site 670 at the southern outcrop was drilled during ODP Leg 109. These outcrops occur in rugged terrane of the western median valley wall of the northern spreading cell and extend into the cell boundary zone to the south. On the basis of a linear magnetic anomaly that might be produced by high-susceptibility serpentinite, Brown and Karson (1988) proposed that serpentinized peridotite occurs in a broad belt along the western median valley wall.

In 1988, three dredge hauls by the *Akademik Mstislav Keldysh* in the northernmost outcrop area recovered more serpentinized peridotite (Gente et al., 1989). This area was also explored during three *Nautile* dives (Mével et al., 1991). The summit of the western rift valley wall forms a hill that culminates at a depth of 2600 mbsl (Fig. 3). The relatively young, fissured pillow basalts of the axial valley floor abut the base of the rift valley wall at a depth of about 3700 m. The base of the slope is covered by talus of serpentinite blocks and basalt fragments.

Serpentinized peridotite crops out nearly continuously from about 3500 mbsl, to about 3100 mbsl. This portion of the median valley wall is a tectonically active slope, dominated by east-facing, presumably normal, faults. As in the gabbroic rocks of the intersection massif, faults with relatively steep dips (40° to 70°) cut a more pervasive, gently east-dipping (20°–40°) schistosity, which is thought to be related to earlier, low-angle faults and shear zones. Samples show serpentinite shear zones and veins and a high-temperature plastic fabric defined by elongated pyroxene grains (Mével et al., 1991).

The upper slopes of the rift valley wall are covered with pillow basalt and basaltic talus. The contact between the serpentinized peridotite and the pillow basalt is concealed by talus. It may mark the base of basaltic flows extruded over serpentinized peridotite that was once exposed on the seafloor. Alternatively, it may be a low-angle fault contact separating an allochthonous slice of basalt from underlying serpentinite.

Most samples of serpentinized peridotite recovered along the western rift valley wall of the MARK area appear to have been derived from a clinopyroxene-bearing harzburgite protolith (Juteau et al., 1990; Tartarotti et al., in press; Casey et al., unpubl. data), but there are also rare samples of serpentinized dunite and pyroxenite. Localized high-temperature shear zones in these ultramafic rocks contain synkinematic hornblende (Casey, 1986; Tartarotti et al., in press), and a few samples display dikelets and veins of altered, zircon-bearing, gabbroic to dioritic material (Mével et al., 1991; Tartarotti et al., in press).

SUMMARY

Investigations over the past 30 years have made the MARK area the most extensively studied portion of any slow-spreading ridge. Extensive geological and geophysical surveys, many of which result from ODP's efforts, have permitted a detailed picture of the composition and structure to be constructed. This information has also fostered the development of new and provocative models for seafloor spreading. Future drilling will test these hypotheses for the creation and evolution of oceanic crust in the slow-spreading environment of the MARK area.

Recent studies of the MAR demonstrate that the interplay of magmatic construction and mechanical extension varies significantly both regionally and even locally. Still, information derived from studies of plutonic rocks and major fault zones drilled in the northern spreading cell of the MARK area should be widely applicable to other segments of slow-spreading ridges.

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*Abbreviations for names of organizations and publications in ODP reference lists follow the style given in *Chemical Abstracts Service Source Index* (published by American Chemical Society).

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