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Research Article

The composition of back-arc basin lower crust and upper mantle in the Mariana Trough: A first report

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Abstract The Mariana Trough is an active back-arc basin, with the rift propagating northward ahead of spreading. The northern part of the Trough is now rifting, with extension accommodated by combined stretching and igneous intrusion. Deep structural graben are found in a region of low heat flow, and we interpret these to manifest a low-angle normal fault system that defines the extension axis between 19°45′ and 21°10′N. A single dredge haul from the deepest (~5.5 km deep) of these graben recovered a heterogeneous suite of volcanic and plutonic crustal rocks and upper mantle peridotites, providing the first report of the deeper levels of back-arc basin lithosphere. Several lines of evidence indicate that these rocks are similar to typical back-arc basin lithosphere and are not fragments of rifted older arc lithosphere. Hornblende yielded an $^{40}Ar/^{39}Ar$ age of 1.8 ± 0.6 Ma, which is interpreted to approximate the time of crust formation. Harzburgite spinels have moderate Cr# (<40) and coexisting compositions of clinopyroxene (CPX) and plagioclase (PLAG) fall in the field of mid-ocean ridge basalt (MORB) gabbros. Crustal rocks include felsic rocks (70–80% SiO_2) and plutonic rocks that are rich in amphibole. Chemical compositions of crustal rocks show little evidence for a ‘subduction component’, and radiogenic isotopic compositions correspond to that expected for back-arc basin crust of the Mariana Trough. These data indicate that mechanical extension in this part of the Mariana Trough involves lithosphere that originally formed magmatically. These unique exposures of back-arc basin lithosphere call for careful study using ROVs and manned submersibles, and consideration as an ocean drilling program (ODP) drilling site.

Key words: back-arc basin crust, back-arc basin rifting, ophiolite.

INTRODUCTION

Back-arc basins form where extension occurs behind magmatic arcs. One endmember is characterized by seafloor spreading, and with lower extension rates this grades into amagmatic lithospheric stretching. Actively spreading back-arc basins offer valuable perspectives into several important problems that lie at the disciplinary juncture between seafloor spreading, arc magmagenesis, and convergent margin tectonics. Because back-arc basin basalts (BABB) form by adiabatic decompression of asthenospheric mantle that has been modified by the ‘subduction component’, they provide insights into controls of decompression melting and the composition of mantle that has been metasomatized at convergent margins. More commonly than not, ophiolites are interpreted to have formed in back-arc basins. By analogy with mid-ocean ridges, slow back-arc extension is expected to be accomplished amagmatically. At present, little is known about the importance of this process for BABB evolution. This evolution in our thinking about how extension occurs in back-arc basins follows our

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evolving understanding of mid-ocean ridge (MOR) processes, with seafloor spreading being recognized earlier than, and often to the exclusion of, mechanical extension. Following the establishment of plate tectonic principles, the nature of magmatic extension was the focus of research throughout the 1970s and most of the 1980s before the importance of mechanical extension for slow-spreading ridges was acknowledged in the mid- to late 1980s. Our understanding of the nature of extension in back-arc basins is lagging by perhaps a decade behind that of MOR extension. It is perhaps timely to document amagmatic extension in slow-spreading back-arc basins.

The Mariana Trough is a slow-spreading back-arc basin (Fryer 1995). At 18°N, where the Trough is widest, full spreading rates range between 3 and 4.4 cm/y (Hussong & Uyeda 1981), similar to that of slow-spreading mid-ocean ridges (Fox & Gallo 1984). Slow-spreading mid-ocean ridges are dominated by segmented, asymmetric, block-faulted depressions that are structurally similar to continental rifts. Major normal faults and exposures of lower crustal and upper mantle rocks are common only along slow-spreading ridges, indicating that mechanical deformation is far more important in crustal formation at slow-spreading ridges than it is at fast-spreading ridges (Mutter & Karson 1992). Mechanical extension and exposures of lower crustal rocks are poorly known from back-arc basins, although exhumation of deep crustal metamorphic rocks due to orogenic collapse in Mediterranean Sea back-arc basins has recently been summarized (Jolivet *et al.* 1994). The lack of samples of back-arc basin oceanic crust has limited our understanding of back-arc basin lithospheric evolution as well as our ability to confidently infer the tectonic setting of ophiolites from their lower crustal and upper mantle components.

In this paper, we present first results from our study of an exposure of lower crustal and upper mantle rocks exposed along the extension axis of a back-arc basin. The samples come from a single dredge haul collected during the 1991 Tunes expedition aboard the Research Vessel *Thomas Washington*. While this small sample suite is an inadequate base for developing grand models of back-arc basin lithospheric evolution, it is a good place to start. We report our results and integrate them with other data from the region in an attempt to stimulate efforts to understand this important aspect of the oceanic lithosphere, and to provide a basis for planning future studies of this unique tectonic window on back-arc basin lower crust and upper mantle.

REGIONAL SETTING

The Mariana Trough is a slowly extending back-arc basin that is propagating northward through the Mariana Arc system. This preserves evolutionary stages along strike from north to south that mimic the evolution through time of any single section of the basin. Early studies focused on the southern and central Trough and inferred that the entire basin formed by seafloor spreading (Hussong & Uyeda 1981). Recent studies indicate that stretching of pre-existing arc crust dominates early rift development in the northern Mariana Trough, although details of the location and nature of the transition from rifting in the north to spreading in the south await resolution (cf. Martinez *et al.* 1995; Yamazaki *et al.* 1993). In spite of these uncertainties, the style of extension in the Mariana Trough varies from seafloor spreading from at least the latitude of Guam as far north as 19°45′N (Fig. 1). North of this latitude, spreading is supplanted by amagmatic extension resulting in the formation of deep graben; these extend from 19°45′ to 21°10′N and are the focus of the present study. The extension axis shifts eastward, close to the line of the magmatic arc, where the rift axis has captured the arc magmatic budget north of Fukujin (~22°N). Extension in this region is magmatic yet distinct from seafloor spreading in being dominated by central volcanoes, a poorly developed ridge or rift, and volcanism that has strong arc affinities and is often felsic. This stage of extension is analogous to that observed in the Sumisu Rift in the Bonin Arc (Taylor *et al.* 1991). This portion of the extension axis is referred to as the ‘Volcano–Tectonic Zone’ or VTZ (Martinez *et al.* 1995; Baker *et al.* 1996). The VTZ intersects the arc at Nikko. The region north of Nikko has not yet rifted and is dominated by active shoshonitic volcanoes of the Hiyoshi Volcanic Complex, Fukutoku-oka-no-ba (Sin Iwo Jima) and Iwo Jima.

The focus of this study is the deep, amagmatic graben that define the rift axis for over 150 km between 19°45′ and 21°10′N. These are called the Central Graben by Martinez *et al.* (1995). The axis of extension here is over 4 km deep, and two regions, near the north and south ends, are over 5 km deep. The southern chasm is the deepest point in the Mariana Trough, and is remarkably deep compared with the anomalously deep 3.6 km mean depth of the Mariana Trough spreading axis (Park *et al.* 1990). The lower crustal and upper mantle samples were collected from the southern deep (Location 1 in Fig. 1); other sampling locali-

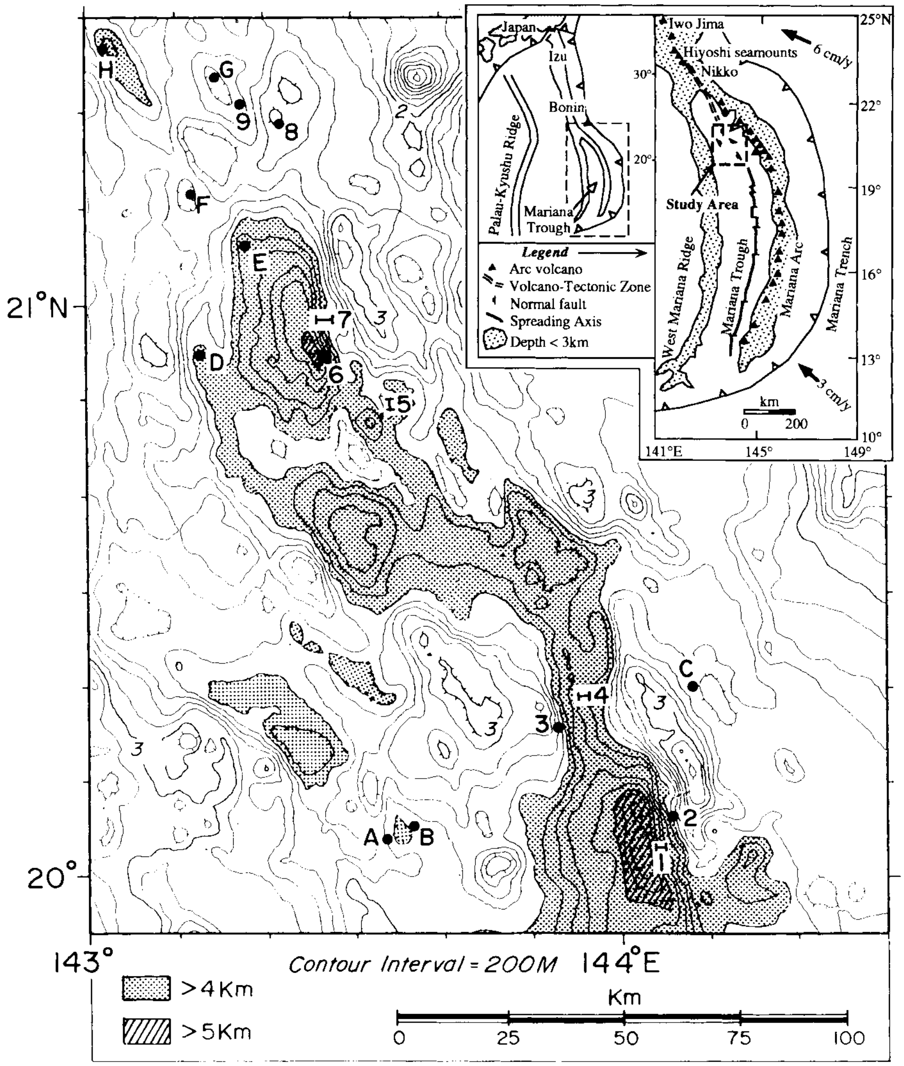

Fig. 1 Map of the central graben in the northern Mariana Trough. Insets show the location of the Mariana arc system and the Mariana Trough, and the location of the study area. Arrows indicate relative motion between the Pacific plate and Mariana arc (Eguchi 1984). Numbered locations refer to sampling locations, and lettered locations denote positions of heat flow measurements; further details are listed in Tables 1 and 2. Sources of bathymetric data are presented in Martinez *et al.* (1995).

ties recovered only altered volcanics (Figure 1, Table 1).

The structure of the Central Graben is poorly understood. There is a relatively low rate of sedimentation in this part of the Mariana Trough, so that bathymetry expresses structure. The larger structure is a complex graben that is composed of four asymmetric deeps (Fig. 1). The southernmost (~20°N) is a down-to-the east half-graben, as is the northernmost (~21°N) and the easternmost of the central pair (~20°35′N, 144°E). The westernmost of the central pair (~20°35′N, 143°30′) is a down-to-the west half graben, as is the narrow corridor that connects the southern and east-central graben. The structure is reminiscent of complex half-grabens that characterize continental rifts, such as the East African Rift Valley (Rosendahl 1987). Similar structures have been documented for the earliest stages in the evolution of some back-arc basins such as the Sumisu Rift in the Bonin arc (Taylor *et al.* 1991). Mutter & Karson (1992) argued that the exposure of lower crust and upper mantle results from movement on low-angle (~30°) normal faults. This interpretation is consistent with the asymmetry observed for half-graben in the graben complex. The consistent location of down-to-the-east half graben in along the eastern margin of the deep is consistent with the presence of a west-dipping, low-angle normal fault. This is the general interpretation that we prefer (Fig. 2), although we caution that many structural details await resolution.

The region around the Central Graben is characterized by anomalously low heat flow (Table 2). Two different expeditions in the Mariana Arc and Trough yielded 8 heatflow sites within the area encompassed by Fig. 1. Most of these locations were measured multiple times, and with one exception where high heat flow was measured (location G, north of the northernmost basin), means for the

Table 1 Sampling localities in the Mariana Trough Central Graben.

Sample locality	Grid reference	Depth (m)
1. Tunes 7 D45 (dredge) 300 kg serpentinites, felsic plutonics, gabbros, basalts — no significant Mn encrustation.	20°2.5′N, 144°04′E	5175–4800
2. KH 84-1-23 (dredge) basalts	20°07′N, 144°06′E	3960–4300
3. KH 84-1-24 (dredge) basalts	20°16′N, 143°52′E	4160–4250
4. Tunes 7 D71 (dredge) 200 kg of pillowed basalt. Somewhat altered, some Mn encrustation; not zero age volcanics.	20°18.5′N, 143°55–57′E	4300–4400
5. Tunes 7 D46 (dredge) 200 kg of brown-gray altered basalt	20°48–50′N, 144°33′E	3700
6. Shinkai 6500 dive #142 basalt, dolerite, dacite	20°54′N, 143°26′E	5195–4650
7. Tunes 7 D47 (dredge) 400 kg altered and fresh basalts, volcaniclastics, some diabase; abundant Mn-encrustation	20°58′N, 143°25–27′E	5100–3700
8. Tunes 7 D48 (dredge) 1000 kg weathered and hydrothermally altered basalt	21°19′N, 143°22′E	2900
9. Tunes 7 D68 (dredge) 500 kg fresh, glassy pillow basalt	21°21′N, 143°16′E	3900–3800

Numbered locations correspond to locations in Fig. 1.
Tunes 7: R/V *Thomas Washington*, December 1991.
KH-84-1 data are from Shibata and Segawa (1985).
Shinkai 6500 dive 142 data are from Yamazaki and Yuasa (1993).

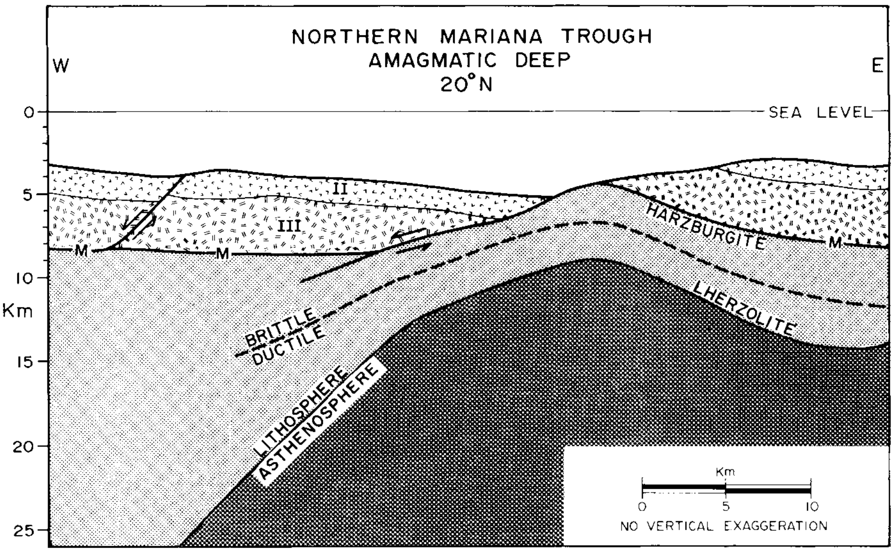

Fig. 2 True scale cross-section and structural interpretation consistent with the bathymetric expression of structure and occurrence of lower crustal and upper mantle rocks exposed in the southernmost sub-basin of the central graben. Modified after a model presented by Karson (1990) for part of the Mid-Atlantic Ridge. Note that the asthenosphere-lithosphere boundary is elevated beneath the graben, and this would lead to high heat flow. This interpretation is only consistent with the observed low heat flow if lithospheric thinning is extremely recent.

7 localities range from 10 to 49 mW/m^2, well below the global heatflow value of about 65 mW/m^2. The heatflows measured around the deep graben are not anomalously low compared to measurements made elsewhere in the Mariana Trough, even adjacent to the extension axis, where fluxes of less than 40 mW/m^2 are commonly measured (Anderson 1975; Uyeda & Horai 1981; Yamazaki 1992). However, many of the low heatflow values around other parts of the Mariana Trough spreading axis may reflect local sites of cold water recharge for hydrothermal convective cells, but this cannot explain the low heat flow around the Central Graben, which lacks recent igneous and hydrothermal activity. The low heat flow in this region must reflect instead cold crust and mantle. This is consistent with sampling in the region that indicate a lack of fresh volcanic rocks (Table 1) and with subdued seafloor magnetization over the Central Graben, in spite of the fact that the positive

Table 2 Central Graben heat flow.

Location*	Station	N	Mean heat flow (mW/m^2)
A	KH84-1-HF6A-C[1]	3	13
B	KH84-1-HF3A to D[1]	4	42
C	KH84-1-HF4A&B[1]	2	27
D	H339 RC564[2]	1	10
E	H341 RC566[2]**		
	H342-1 to 4[2]		
	H377-1 to 5[2]	9	26
F	H343-1 to 4[2]	4	49
G	H366RC569[2]		
	H262RC502[2]	2	205
H	H345-1 to 5[2]		
	H262RC501[2]	6	12

[1] Yamano 1985.

[2] Yamazaki 1992.

*Letters correspond to locations in Fig. 1.

**H341 has a significant non-conductive component of heat flow.

magnetization band associated with the VTZ can be traced into the Central Graben (Martinez *et al.* 1995). The only identifiable volcano in the region shown in Fig. 1 is the southwesternmost of the Kasuga cross-chains at 21°23′N, 143°40′E. The Central Grabens are defined by free-air gravity lows between 0 and 25 mGal which suggests that low-density material underlies the Central Graben at depth (Martinez *et al.* 1995).

There are several possibilities to explain the presence and location of the Central Graben. First, these could simply represent the natural occurrence of mechanical extension along slow-spreading ridges (Mutter & Karson 1992). Second, the Central Graben could be a zone of mechanical extension present in front of a propagating spreading ridge. This interpretation is similar to that inferred for the location of the Hess Deep in advance of the westward-propagating Cocos–Nazca Ridge (Francheteau *et al.* 1990). Finally, the Central Graben may reflect the localization of the extension axis away from the arc and towards a more central position within the basin. As described by Martinez *et al.* (1995), the localization process is the first step in establishing a well-defined plate boundary within the backarc basin. This moves the extension axis away from the arc magma supplied to the VTZ, and extension may become amagmatic until true seafloor spreading is established. The composition of samples recovered from the grabens provides an opportunity to test these hypotheses. For example, if the first hypothesis is correct, we should recover samples of crust that formed by spreading, whereas if either of the other two hypotheses are correct, we should recover VTZ igneous rocks or rifted arc basement.

SAMPLE LOCATION AND DESCRIPTIONS

Dredge 45 was the first dredge station of the December 1991 TUNES 7 expedition. We targeted the lower part of a west-facing scarp on the east side of the southernmost of the Central Graben (Location 1, Fig. 1). From beginning to end of the dredge, the ship moved ESE, from 20°02.73′N, 144°03.46′E to 20°02.28′N, 144°04.63′E; water depth decreased from 5175 m to 4800 m during the 2.5 hours that the dredge was on the bottom. The form and weathering of most samples indicate that they came from a talus deposit at the base of the scarp. The dredge haul was remarkably heterogeneous, with about 300 kg of serpentinized ultramafic, felsic plutonic, gabbroic, and altered volcanic rocks recovered. There was no significant Mn-encrustation on any of the samples. Serpentinized ultramafics comprise the largest group, making up about one-third of the haul. The second most abundant lithology are variously altered basalts. The third most abundant group is made up of felsic to mafic plutonic rocks. A group of anorthositic plutonic rocks made up the final group. These groups were sampled for the analyses reported in this paper.

Because of the long time that the dredge was on the bottom, we have few constraints on the original igneous stratigraphy beyond the generalized ophiolite/oceanic crust paradigm. However, dredging by Japanese scientists (KH-84-1-23; Location 2, Fig. 1) sampled the upper parts of the same scarp between 3960 and 4300 m depth and recovered fresh pillow basalt and dolerite. We infer from this that exposures of lower crustal and upper mantle rocks lies at a depth of >4300 m. Hand specimens show two generations of felsic melt intruding metamorphosed mafic rock, with an earlier generation of darker andesitic/dacitic melt showing diffuse margins with the mafic host and the younger tonalitic melt intruding along brittle fractures (Fig. 3). We infer from this that emplacement of the felsic igneous rocks postdates consolidation of the mafic igneous crust.

We selected seven samples to study. These include six intrusive rocks and a basalt. Intrusive rocks define a heterogeneous suite that includes a hornblende diorite (D45-2-22), an anorthosite (D45-3-8), a diabase (D45-4-1) and three felsic samples (D45-3-2, 3-11, and 3-29). Hornblende

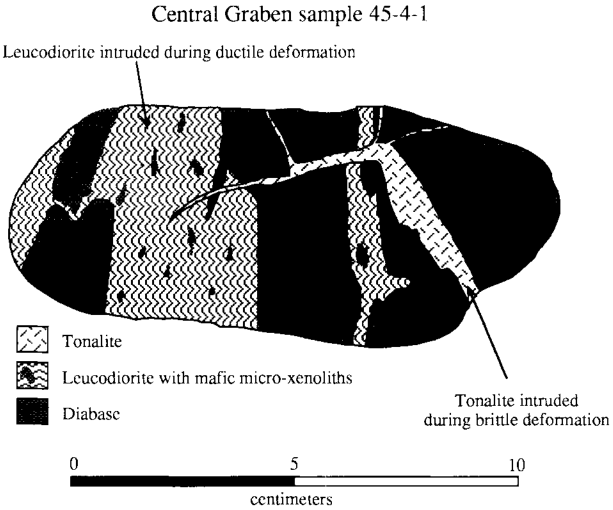

Fig. 3 Sketch of relations displayed in sawed sample D45-4-1. Mafic rock was first intruded by andesitic to dacitic material. This intrusion occurred under conditions of ductile deformation, as shown by diffuse margins, partial digestion of mafic micro-xenoliths, and the presence of a fabric in the intrusive rock. Tonalitic intrusion occurred under brittle conditions, with tonalite veins following nearly orthogonal fracture or fault sets.

diorite D45-2-22 contains large cumulate crystals of plagioclase (5–20 mm; An$_{40-44}$) and large (10–20 mm) black edenitic hornblende (on a basis of 24 O, tetrahedral Al = 1.4 to 1.5 and Na + K = 0.9 to 1.1) that contains about 3.4% TiO$_2$ and is poor in potassium (0.18 to 0.21% K$_2$O). Many amphibole-bearing gabbros result from hydrothermal interaction (Mével & Cannat 1991), and some of the D45 gabbros show intermediate stages in the alteration of clinopyroxene to brown amphibole. However the amphiboles in D45-2-22 are coarse (up to 2 cm long), euhedral, and contain no relict cores of clinopyroxene, and we interpret these as igneous in origin. Anorthosite D45-3-8 is composed of large (>1 cm), euhedral plagioclase (~95%), with the remainder being smaller disseminated crystals of hornblende, clinopyroxene, and biotite. Diabase D45-4-1 is a fine-grained (~0.1–0.3 mm) rock, composed of 45% pleochroic green amphibole, probably actinolite, 55% plagioclase, and 5% opaques. Tonalite sample D45-3-2 is fine grained (<1–2 mm) and composed almost entirely of quartz and plagioclase, with a small amount of chlorite and opaques replacing unknown mafic minerals. Granitic sample D45-3-11 is fine grained (0.3–0.6 mm) and composed of 40% subhedral quartz, 30% euhedral plagioclase, and 30% perthitic feldspar as rims around plagioclase. Granitic sample D45-3-29 is similar to D45-3-11 in grain size and the abundance of quartz and the relationship between plagioclase cores and perthite rims; this sample also contains ~3% hornblende and biotite. Basalt D45-5-2 is sparcely phyric, with 5% skeletal olivine (~0.5 mm) and 1% plagioclase (~0.5 mm) set in a dark brown, devitrified, variolitic groundmass.

ANALYTICAL TECHNIQUES

Samples were ground in alumina in order to preclude contamination by Nb and Ta. Major elements for five samples and some trace elements (Sr, Nb, Zr, Y, Zn, Cu, and Ni) on all seven samples were analyzed by X-ray fluorescence techniques at the University of Oklahoma using techniques outlined by Weaver (1990). Two samples (45-3-8 and 3-11) were analyzed for major elements using XRF by X-Ray Assay Laboratories. Concentrations of REE, Th, Ta, Hf, Co, Cr, and Sc were determined by instrumental neutron activation at the University of Oklahoma using procedures outlined by Weaver (1990). Concentrations of Rb, Ba, and Pb were determined by standard isotope dilution techniques at the University of Texas at Dallas (UTD). Isotopic compositions of Sr, Nd, and Pb were determined at UTD using procedures outlined by Gribble *et al.* (1996). Sr isotopic compositions are reported relative to Eimer and Amend (E&A) SrCO$_3$ ^{87}Sr/^{86}Sr = 0.70800. Nd isotopic compositions were fractionation-corrected to ^{146}Nd/^{144}Nd = 0.7219 and are reported relative to UCSD Nd ^{143}Nd/^{144}Nd = 0.511847 and BCR ^{143}Nd/^{144}Nd = 0.512618. External reproducibility is better than ± 0.00002 for ^{143}Nd/^{144}Nd. In-run precision is quoted in Table 4 when this is worse than the external reproducibility. ε_{Nd} was calculated from a value for the bulk earth appropriate for the UTD lab using the ε_{Nd} values reported by (Pier *et al.* 1989) for BCR-1 and UCSD standards. Pb isotopic compositions are corrected for fractionation: 0.15%/amu. Sixteen analyses of NBS 981 standard yield ^{206}Pb/^{204}Pb = 16.943 ± 26, ^{207}Pb/^{204}Pb = 15.495 ± 33, and ^{208}Pb/^{204}Pb = 36.736 ± 99. All samples analyzed for isotopic composition were whole-rock powders that were leached in warm 6N HCl prior to chemical processing. Sr isotopic compositions in several instances showed significant change before and after leaching; this is interpreted to reflect the result of alteration. It is believed that the Sr isotopic compositions determined after leaching approximates the isotopic composition of Sr in many but not all of the samples that we have analyzed, and samples with low concentrations of Sr are most vulnerable to isotopic shifts accompanying alteration. Feldspars were separated from two samples with low Sr contents (45-3-2: 62 ppm and 45-3-11: 9 ppm), and these were analyzed before and after leaching.

These isotopic compositions are slightly higher than for the corresponding whole rock powders. The authors therefore accept that the Sr isotopic compositions of the unaltered rocks are best approximated by the leached whole-rock powders, although even these probably have slightly elevated $^{87}Sr/^{86}Sr$ due to alteration. $^{40}Ar/^{39}Ar$ procedures at University of California, Los Angeles (UCLA) are outlined in Harrison *et al.* (1991).

RESULTS

When the dredge suite was first examined aboard the ship, we thought that we had sampled rifted Mariana arc crust. As work progressed on the suite, this hypothesis became increasingly untenable, and it is now certain that rifted back-arc basin crust was sampled. Comparable suites have not been reported from other back-arc basins, so before geochemical and isotopic results are reported from the suite that we have studied, we prove that it is a remarkably diverse sampling of young back-arc basin crust. This is accomplished by examining peridotite petrology and mineral compositions, gabbroic mineral composition, and one radiometric age.

The serpentinized peridotites from D45 were originally harzburgites with an average of 2% modal clinopyroxene. The mineralogy and mineral composition of these rocks will be reported elsewhere, but important results are summarized here. Serpentinization of the peridotites was moderate to complete (50–100% serpentinized), and many peridotites preserve abundant magnesian olivine (Fo90.5), orthopyroxene (Mg/Mg + Fe = 0.90), and clinopyroxene (Mg/Mg + Fe = 0.90). Peridotite spinels have Cr/Cr + Al = 0.25 to 0.40, well within the field typical of depleted oceanic peridotites, and distinct from spinels in boninites and peridotites recovered from forearcs (Fig. 4a). These peridotites show mineralogical features indicating moderate depletion, similar to that expected from slow-spreading ridges globally. There is no evidence that the peridotites represent attenuated arc lithosphere. Insofar as they are indistinguishable from peridotites recovered from spreading ridges, they are most readily interpreted as residues after partial melting to form back-arc basin basalts.

No olivine gabbros were found, only coarse-grained clinopyroxene gabbros which contain pleochroic-brown hornblende as discrete crystals, optically continuous rims around clinopyroxene, and partially replacing clinopyroxene interiors. These hornblendes may have crystallized late in the evolution of a hydrous, fractionated mafic magma or may reflect operation of a deep crustal hydrothermal system. D45 gabbros contain clinopyroxene and plagioclase that are distinct from Mariana arc gabbros (Fig. 4b). At a given clinopyroxene composition, Mariana and other arc gabbros have more calcic plagioclase than do MORB gabbros. D45 gabbros have coexisting clinopyroxene and plagioclase compositions that are indistinguishable from MORB, and so are best interpreted as forming in association with a spreading ridge. This is again consistent with an origin in a back-arc basin magmatic system.

Finally, the age of the plutonic suite is most consistent with a back-arc basin origin. One sample

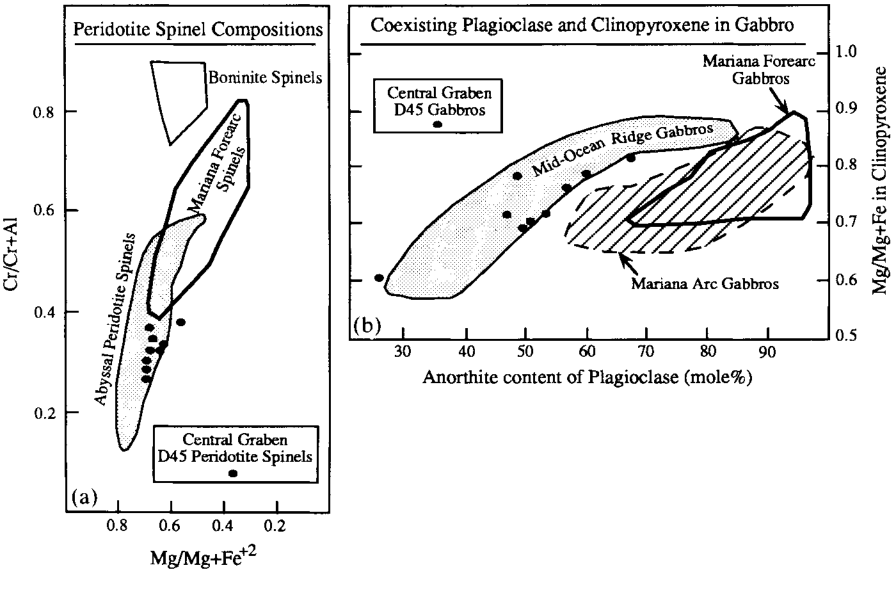

Fig. 4 Mineral chemistry of Central Graben peridotites and gabbros recovered in D45. (a) Plot of Cr vs. Mg composition for spinels in harzburgite. Fields for abyssal peridotites and boninites are from Dick and Bullen (1984); fields for peridotites of Eocene age from the Mariana forearc are from Bloomer and Hawkins (1983). (b) Compositions of co-existing plagioclase and clinopyroxene in gabbros. Fields are modified after Bloomer *et al.* (1995). Note that the D45 samples plot in the field of normal oceanic crust and outside fields for arc rocks in both fields.

of hornblende gabbro-diorite (45-2-22) was dated using the $^{40}Ar/^{39}Ar$ step-heating technique. Hornblende was separated by hand-picking and analyzed at UCLA (Table 5). Fourteen heating steps were analyzed, with >85% of the Ar released during two heating steps (1120° and 1180°C). The data yield an inverse isochron age of 1.8 ± 0.6 Ma [mean square of the weighted deviates (MSWD) = 63], which we tentatively assign to the entire igneous section that we have sampled. The large error on this age is largely due to the low K content (0.18–0.21% K_2O) of this young hornblende. Apatite was also separated from this sample in an attempt to obtain a fission track age, but the combination of very low U contents and very recent cooling (younger than 1.8 Ma) prevented the derivation of a meaningful cooling age. It is not known how long back-arc basin extension has occurred in this part of the Mariana Trough; estimates range from spreading since about 3.5 Ma (Yamazaki *et al.* 1993) to rifting since at least 2.5 Ma (Martinez *et al.* 1995). In either case, the $^{40}Ar/^{39}Ar$ age reported indicates that the crust formed after backarc basin extension was well underway, and precludes formation in an arc setting. It is not known how long after formation this crust was rifted to establish the present morphology, but the absence of Mn encrustation on lower crustal and upper mantle samples suggests that these surfaces have not been exposed for very long.

Chemical results are reported in Table 3 and isotopic data are reported in Table 4. The data are presented in a K_2O-SiO_2 plot (Fig. 5a) and define a low-K suite with the exception of the very felsic (>77% SiO_2) samples D45-3-11 and 29, which fall in the medium-K field. Attention should be given to the fact that felsic lavas are not found along the Mariana Trough spreading ridge (ruled field in Fig. 5a). These samples are also plotted on a conventional alkalies-iron-magnesium (AFM) plot, which also shows the large compositional range among these samples (Fig. 5b). The mafic samples plot within or near the tholeiitic field; sample 45-4-1 plots in the calc-alkaline field, but this may be the result of seawater alteration, as this sample has unusually high Na_2O contents (Table 3). In contrast, the felsic samples (and anorthositic sample 45-3-8) plot within the calc-alkaline field.

Trace element data demonstrate further that the crustal rocks comprise a compositionally diverse suite. K/Rb ratios are high, greater than 600 for the suite and generally >1000. There is little evidence for a 'subduction component' in the form of elevated Ba/La and Sr/Nd or low Ce/Pb. With the exception of anorthositic sample 45-3-8 which has very high (>700 ppm) Ba, Ba/La is less than 9, much more like typical MORB (Ba/La = 3.6; Hofmann 1988) than typical Mariana arc lavas (Ba/La = 20 to 75; Lin *et al.* 1989). The Ce/Pb ratio shows tremendous variability, from as great as 87 to as low as 3.1. The very high value is due to accumulative hornblende in sample D45-2-22. The diabase and basalt samples have Ce/Pb of 25 and 26, respectively, very close to that expected for MORB (24.5; Hofmann 1988)). The felsic samples show a range of Ce/Pb, from values typical of MORB (Ce/Pb = 26 for 45-3-2) to very low, arc-like values (Ce/Pb <4; 45-3-11 and 3-29). The Sr/Nd ratio is 13 or less, even in the anorthositic sample; this is as low as or lower than Sr/Nd for MORB (~10; Hofmann 1988), and unlike Sr/Nd for arc basalts, which range up to 80 or more (DePaolo & Johnson 1979). Finally, Y/Nb in the D45 suite is also low, 12 or less, and similar to MORB (Y/Nb ~10; Hofmann 1988). However, it should be noted that arc basalts can also have moderately low Y/Nb (average arc basalt Y/Nb = 13; Hawkesworth *et al.* 1991).

The Rare Earth Element patterns for the suite further demonstrate a compositional variability that is not approached anywhere along the Mariana Trough spreading axis to the south (Fig. 6). Whereas the olivine basalt (45-5-2) has a REE pattern that is indistinguishable from that of typical Mariana Trough basalt, the diabase (45-4-1) is very enriched in REE. The hornblende diorite (45-2-22) is very enriched in REE but relatively depleted in the LREE and Eu. The anorthosite (45-3-8) has the steepest REE pattern of any of the samples and has a positive Eu anomaly. The felsic samples show LREE-enrichments with strongly negative Eu-anomalies superimposed on an overall U-shaped REE pattern.

Data for three felsic samples are plotted on diagrams designed to test for tectonic setting of granitic rocks (Fig. 7). It should be noted that there are no fields defined for back-arc basin felsic rocks. The data plot outside of fields defined by volcanic arc granitic associations, and instead lie within and between fields defined for Within-Plate and Ocean-Ridge granites. As is the case for all rocks from this dredge, the felsic rocks are not similar to felsic rocks from intra-oceanic arcs.

The tectonic affinities of the crustal rocks can be further examined using the compatibility or 'spider' diagram (Fig. 8). The patterns are relatively flat, at about 2 to 10 times N-MORB for both felsic

Table 3 Whole rock geochemistry for the Mariana Trough Central Graben.

	45-2-22 Hb Diorite	45-3-2 Hb Tonalite	45-3-8 Anorthosite	45-3-11 Granodiorite	45-3-29 Granodiorite	45-4-1 Diabase	45-5-2 Ol. Basalt	KH-84-1-23 Mean Basalt[1]
SiO_2	47.99	70.50	61.1	79.7	77.99	51.59	51.32	51.2
TiO_2	2.22	0.46	0.06	0.08	0.10	1.72	1.22	1.68
Al_2O_3	14.55	15.62	21.1	11.8	13.11	17.94	17.10	15.59
$Fe_2O_3{}^T$	13.33	4.38	1.17	0.81	1.13	10.41	9.33	9.91
MgO	6.56	1.95	2.16	0.43	0.40	4.98	6.74	6.09
CaO	8.82	1.05	3.87	0.32	0.25	8.35	11.36	10.20
Na_2O	3.94	6.36	8.24	5.06	5.17	4.80	2.89	3.61
K_2O	0.11	0.67	0.55	2.06	2.92	0.09	0.40	0.32
P_2O_5	0.73	0.12	0.24	< 0.01	0.01	0.35	0.15	0.23
Total	98.25	101.11	98.5	100.0	101.08	100.23	100.51	99.13
Trace Elements (ppm)								
Rb	0.57	4.8	3.1	16.8	39.7	0.19	5.4	—
Sr	222	62	299	9	9	331	204	—
Ba	39	98	707	119	183	72	51	—
Pb	0.51	2.0	2.2	12.2	13.5	1.4	1.1	—
Nb	12.4	13.8	2.7	11.6	21.6	7.8	3.7	—
Ta	0.62	1.10	0.31	1.18	2.50	0.53	0.29	0.29
Zr	111	443	183	137	175	131	98	—
Hf	3.68	10.6	4.55	5.33	7.95	3.37	2.16	8.1
Th	0.26	2.72	0.57	3.18	4.59	0.36	0.50	0.25
Sc	61	7.6	0.9	1.8	2.1	27	35	39
Cr	16.4	2.3	4.4	1.3	1.4	40	219	223
Co	26.8	3.5	4.2	0.9	0.6	26	34	33
Ni	104	13	37	1	4	51	93	—
Cu	3	28	< 1	5	6	45	62	—
Zn	92	49	24	31	42	93	65	—
Y	149	84	35	81	82	60	30	—
La	12.2	21.3	13.6	15.1	21.3	13.4	6.0	6.0
Ce	44.4	51.5	35.2	37.5	52.0	34.6	17.3	15.0
Nd	45.1	32.1	24.3	23.4	25.0	25.7	12.5	—
Sm	16.9	8.78	5.58	6.66	6.58	7.14	3.28	4.3
Eu	4.46	2.03	2.40	0.45	0.53	2.32	1.21	1.58
Tb	3.76	1.76	0.91	1.53	1.48	1.45	0.71	0.91
Yb	12.4	8.82	3.25	9.09	10.1	5.85	2.92	3.2
Lu	1.54	1.21	0.39	1.29	1.53	0.75	0.37	0.54
K/Rb	1600	1160	1470	1020	610	3900	614	—
K/Ba	23.4	57.4	6.47	144	132	10.4	65	—
Ba/La	3.2	4.6	52	7.9	8.6	5.4	8.5	—
Ce/Pb	87	26	16	3.1	3.9	25	16	—
Sr/Nd	4.92	1.93	12.3	0.39	0.36	12.9	16.3	—
Y/Nb	12	6.1	13	7	3.8	7.7	8.1	—

Rb, Ba, and Pb by isotope dilution; REE, Th, Ta, Hf, Co, Cr, and Sc by INAA; all others by XRF.

[1] Major elements are averages of 47 probe analyses of glasses (Shibata & Segawa 1985). Trace element data are for sample KH-84-1-23-204 (Nakamura & Wakita 1985).

and mafic suites. Strong depletions in Sr and Ti seen in the felsic rocks result from crystal-liquid fractionation during magmatic evolution or anatexis. The strong enrichment in Ba and depletion in Ti seen for 'mafic' sample 45-3-8 is probably due to a cumulate origin. The strong decoupling between large ion lithophile elements (LILE) such as K, Rb, and Ba and high field strength cations (HFSC), especially Nb and Ta that is so characteristic of arc rocks (e.g. Pearce & Parkinson 1993) is not seen in these patterns. The patterns for the mafic rocks do not show the strong depletions in Nb and Ta characteristic of arc basalts, or even the more modest depletions displayed by the MTB18 suite. The data for the Central Graben crustal suite is clearly more similar to MORB or back-arc basin crust than it is to arc crust.

Isotopic data are listed in Table 4 and presented in Fig. 9. Nd and Pb isotopic compositions of these samples closely approximate magmatic composi-

Table 4 Isotopic data for the Mariana Trough Central Graben.

Sample	$^{87}Sr/^{86}Sr^1$ Unleached	Leached*	$^{143}Nd/^{144}Nd^{*2}$	ε-Nd	$^{206}Pb/^{204}Pb^{*3}$	$^{207}Pb/^{204}Pb^{*3}$	$^{207}Pb/^{204}Pb^{*3}$
45-2-22	—	0.70303	0.51311	+ 9.3	18.513	15.509	38.138
45-3-2	0.70308	0.70294	0.51310	+ 9.2	18.232	15.480	37.962
fspr	0.70315	0.70305	—	—	—	—	—
45-3-8	—	0.70317	0.51311 ± 3	+ 9.3	18.339	15.516	38.080
45-3-11	0.70386	0.70343	0.51313	+ 9.7	18.255	15.505	38.006
fspr	0.70391	0.70353	—	—	—	—	—
45-3-29	0.70399	0.70294	0.51312	+ 9.5	18.239	15.493	37.962
45-4-1	—	0.70331	0.51310	+ 9.2	18.566	15.511	38.180
45-5-2	—	0.70307	—	—	18.487	15.498	38.117

*Leached overnight in warm 6N HCl.

[1] Relative to E&A $SrCO_3$ $^{87}Sr/^{86}Sr = 0.70800$.

[2] Relative to UCSD Nd $^{143}Nd/^{144}Nd = 0.511847$ and BCR $^{143}Nd/^{144}Nd = 0.512618$. External reproducibility is better than ± 0.00002 for $^{143}Nd/^{144}Nd$. In-run precision is given when this is worse than the external reproducibility.

[3] Corrected for fractionation: 0.15%/amu.

Table 5 Argon isotopic and age data for D45-2-22 Hornblende.

Step, Temp (°C)		$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	mol^{39}Ar ($\times 10^{-16}$)	Cumulative ^{39}Ar released	%Radiogenic	$^{40}Ar^{*}/^{39}Ar_K$	Age (Ma)
1	700	757.1	2.016	2.591	3.76	0.005	- 1.1	- 8.5	- 12.5 ± 11.9
2	750	654.7	2.223	2.254	3.55	0.009	- 1.7	- 11.1	- 16.3 ± 13.2
3	800	413.1	2.439	1.423	6.29	0.017	- 1.8	- 7.3	- 10.7 ± 4.3
4	850	367.7	5.266	1.274	4.66	0.023	- 2.3	- 8.5	- 12.4 ± 7.1
5	900	399.5	19.95	1.386	5.94	0.030	- 2.1	- 8.7	- 12.7 ± 5.9
6	950	457.6	20.69	1.612	4.63	0.036	- 3.7	- 17.3	- 25.4 ± 6.5
7	980	387.5	16.72	1.349	4.03	0.041	- 2.6	- 10.0	- 14.7 ± 6.9
8	1010	193.0	16.20	0.6701	7.72	0.051	- 2.0	- 3.8	- 5.6 ± 3.0
9	1030	108.5	18.58	0.3719	7.88	0.061	0.1	0.1	0.1 ± 1.4
10	1050	58.11	21.31	0.2036	28.6	0.097	- 0.6	- 0.4	- 0.5 ± 0.5
11	1080	319.3	19.61	1.055	18.2	0.120	2.9	9.2	13.4 ± 3.3
12	*1120*	*16.96*	*24.22*	*0.0611*	*175*	*0.340*	*4.8*	0.8	*1.2 ± 0.1*
13	*1180*	*13.97*	*24.57*	*0.0488*	*512*	*0.985*	*10.7*	*1.5*	*2.2 ± 0.1*
14	1250	857.6	25.41	2.936	9.11	0.996	- 0.9	- 8.0	- 11.8 ± 16.2
15	1400	3144	21.44	10.926	3.17	1.000	- 2.6	- 84.5	- 128 ± 186

Data in italics (1120° to 1180°C) contain 86% of Ar.

tions. This is not the always the case for $^{87}Sr/^{86}Sr$, where significant differences are observed between leached and unleached whole-rock samples. With the exception of the Sr-isotopic data, for which we suspect that magmatic values have been disturbed by alteration, the isotopic data for D45 crustal samples are indistinguishable from basalts from portions of the Mariana Trough undergoing seafloor spreading and distinct from those of the Mariana Arc. The crustal samples are also distinct from the more arc-like compositions found for lavas from the Mariana Trough farther north where rifting occurs (Fig. 1).

DISCUSSION

The new geochemical and isotopic data for crustal samples from the Central Graben are used to address two questions concerning the early stages in back-arc basin evolution. First we examine how early back-arc basin extension is accomplished and what magma sources are tapped. We then address the question of the origin and significance of felsic back-arc basin magmas. Finally, we discuss the paradox presented by the various estimates of crustal age inferred from geochronology, magnetics, and heat flow.

EXTENSION IN THE NORTHERN MARIANA TROUGH: MAGMATIC OR MECHANICAL?

Models for the evolution of the northern Mariana Trough are rapidly evolving. A consensus exists that the tectonics of the region reflects a coupled rifting and spreading regime that is propagating northward, with the 'nose' of rifting now lying

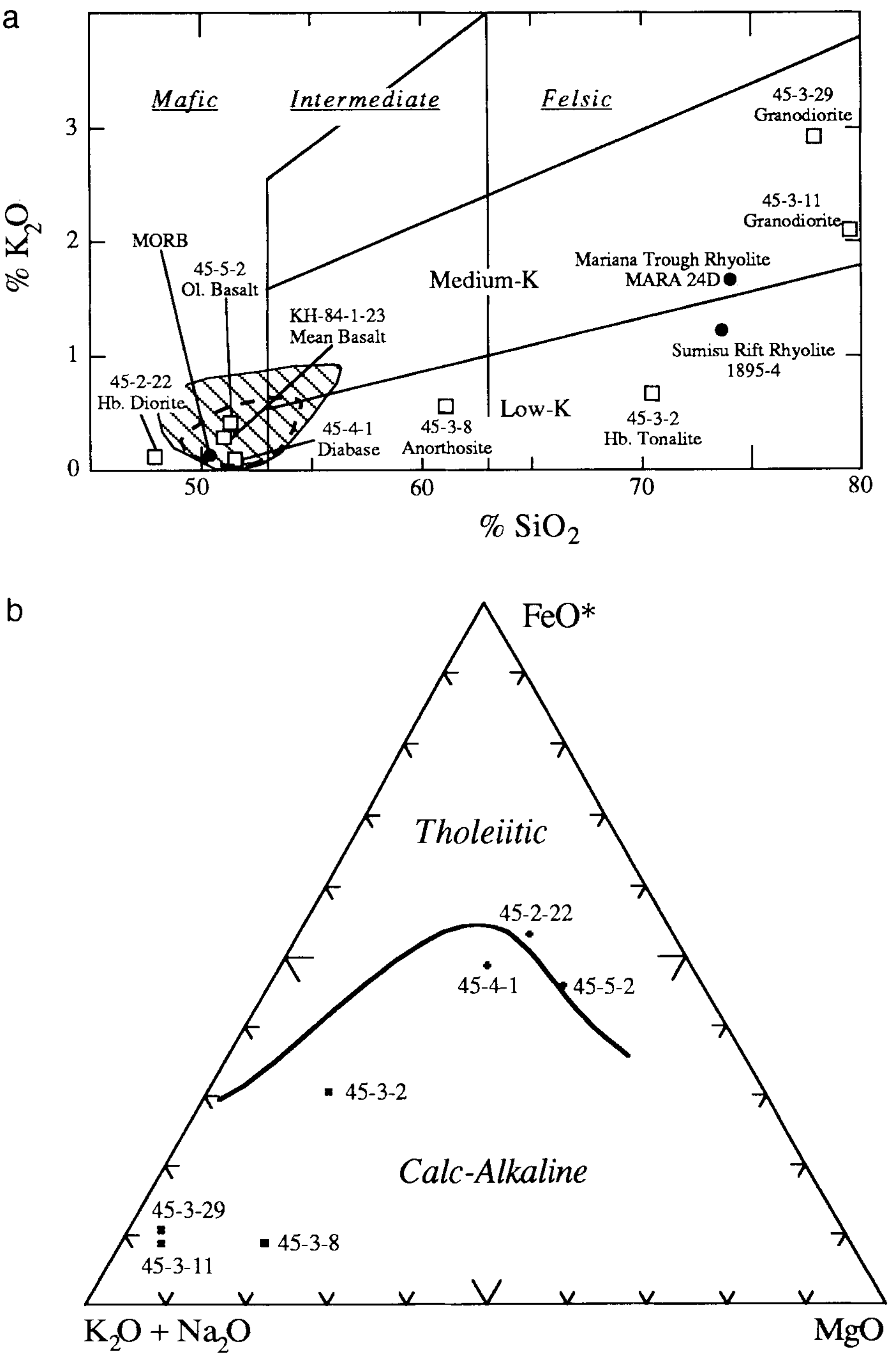

Fig. 5 (a) Potassium-silica diagram for samples from the D45 suite (□) and related rocks (●). KH-84-1-23 is the mean composition of basalts recovered from locality 2 in Fig. 1. MORB is composition of N-MORB average reported by Hofmann (1988). MARA 24D is the composition of pumice recovered from rhyodacite dome at 18°01′N, 144°18′E (Lonsdale & Hawkins 1985). 1895-4 is a typical rhyolite from the Sumisu Rift in the Bonin arc (Gill *et al.* 1992). Basaltic glasses from that part of the Mariana Trough where seafloor spreading is now occurring (i.e. south of 19°45′N) define a low-K suite, with SiO_2 <55%. Ruled field encompasses all data from the Mariana Trough spreading axis (south of 19°45′N), interior dashed field encompasses >95% of this data. (b) AFM diagram for samples analyzed in Table 3. Note the bimodal character of the dredge 45 suite, with a group of more primitive mafic rocks (45-2-22,4-1, 5-2) and a group of felsic rocks (45-3-2, 11, 29). Note that anorthosite sample 45-3-8 is very poor in ferromagnesian minerals and thus plots with the felsic rocks. The boundary between tholeiitic and calc-alkaline fields is that of Irvine and Baragar (1971).

north of 23°N while the location of the spreading 'nose' to the south is more controversial. Gravity data indicate that Mariana Trough crust becomes much thinner just south of 22°N, where it approaches normal oceanic crustal thickness (Ishihara & Yamazaki 1991). This result is consistent with the interpretation of Yamazaki *et al.* (1993), based on magnetic data, that spreading occurs as far north as 22°N. This conclusion was challenged by Martinez *et al.* (1995), who argued that true seafloor spreading exists only south of 20°N. Martinez *et al.* (1995) nevertheless concluded that the magnetization bands in the Northern Mariana Trough volcano tectonic zone (NMT VTZ) required that the locus of igneous activity at any one time must be: (i) relatively narrow and (ii) migrate laterally. A similar interpretation was applied to linear magnetic anomalies in the Havre Trough, where the basin for over 1600 km along strike, from the Valu Fa ridge to New Zealand, is undergoing rifting similar to that in the NMT (Wright 1993). The difference between the style of extension now occurring in the NMT VTZ and the Havre Trough and that of true seafloor spreading is subtle but largely concerns the abundance of sheeted dykes in the true spreading and the inferred abundance of stretched arc lithosphere in the VTZ.

The Central Graben suite offers a valuable perspective on the controversy. Two estimates of rift propagation rates exist: an estimate of 16 cm/yr is based on palinspatic reconstruction of the basin (Stern *et al.* 1984), and a range of 10–40 cm/yr which is based on the geometry of magnetic bands (Martinez *et al.* 1995). These estimates are similar and indicate that the crustal section exposed in the Central Graben formed in a tectonic position analogous to the present VTZ, between 21°30′ and 23°N. This is a zone where extension is believed to be accomplished by a combination of magmatic intrusion and stretching of pre-existing arc lithosphere (Martinez *et al.* 1995), very similar to the 'Basin-and-Range' style of extension envisioned for the first 3 million years of extension in the Lau Basin (Parson & Hawkins 1994). A fundamental question is how much igneous activity occurs dur-

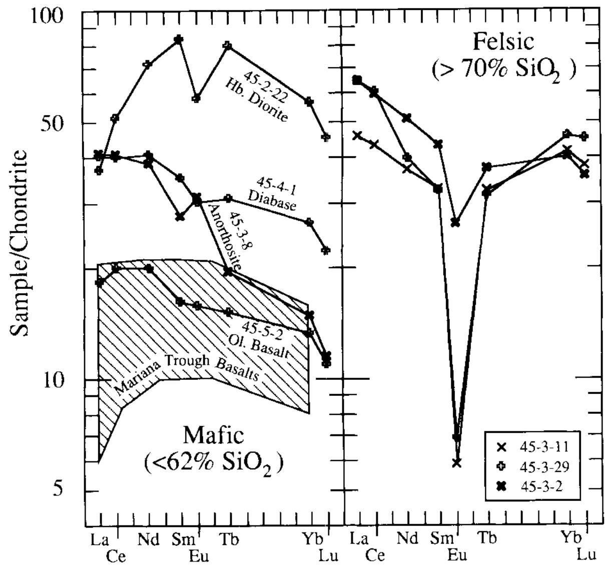

Fig. 6 Chondrite-normalized plots of Rare Earth Element abundances for mafic (left) and felsic (right) representatives of the D45 suite. The field defined by basalts recovered from along the Mariana Trough spreading axis is outlined and shown in the diagonally ruled pattern on the left (Gribble *et al.* 1996). Note that with the exception of olivine basalt sample 45-5-2, all samples are greatly enriched in REE relative to typical Mariana Trough back-arc basin basalt.

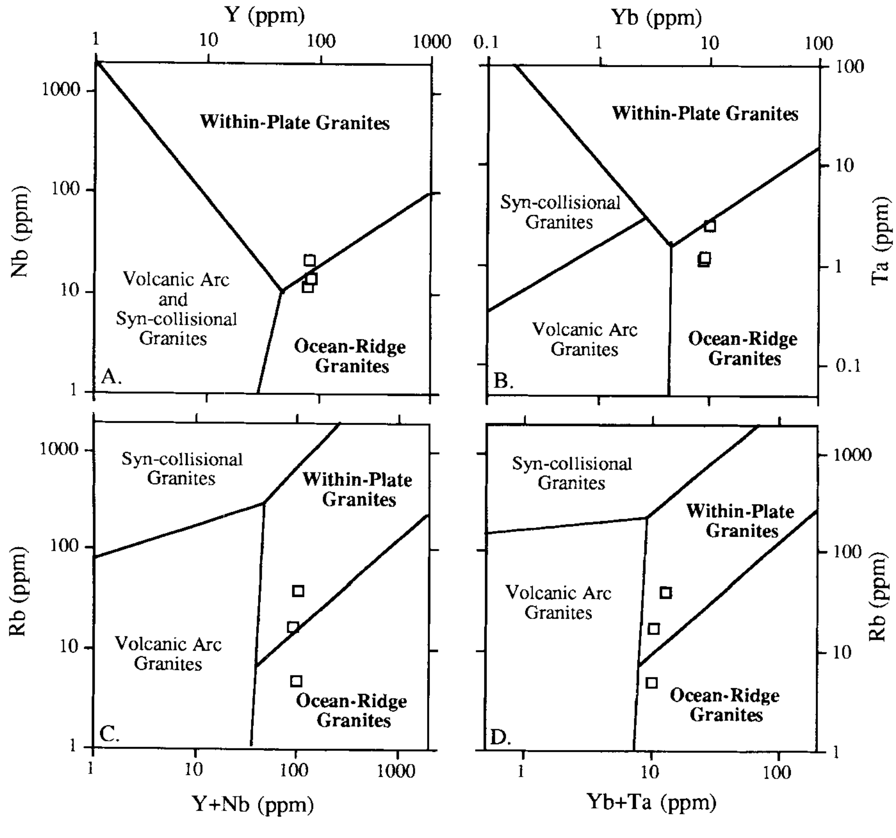

Fig. 7 Tectonic discriminant diagrams after Pearce *et al.* (1984), with data for three felsic samples plotted (45-3-2, 45-3-11, 45-3-29). Most appropriate fields are labeled in bold. Note that the data plot in the field of Within-Plate Granites and Ocean-Ridge Granites and plot outside of the field of arc granites on all four diagrams (Nb-Y, Ta-Yb, Rb-Y + Nb, and Rb-Yb + Ta).

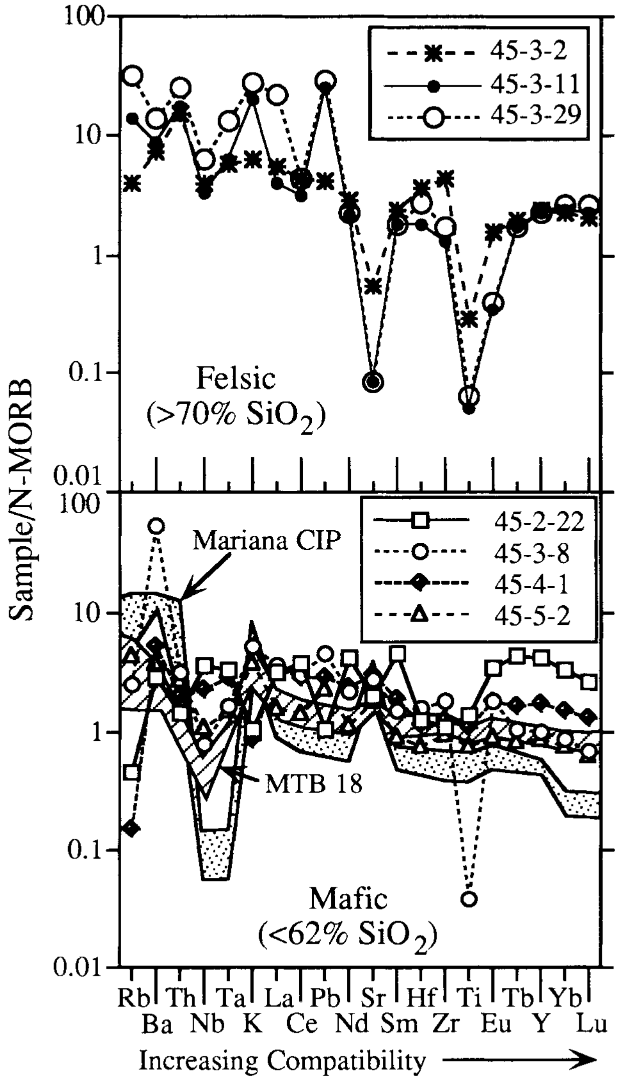

Fig. 8 Element compatibility or 'spider' diagrams for felsic (upper) and 'mafic' (lower) crustal rocks dredged from the Central Graben. Sequence of elements and concentrations in N-MORB are from Hofmann (1988). Note that the felsic rocks are strongly depleted in Sr and Ti and may be moderately enriched in Pb and K relative to adjacent elements; otherwise, the elemental patterns are relatively flat. With the exception of strong enrichment in Ba and depletion of Ti in anorthositic sample 45-3-8, the 'mafic' samples also display a relatively flat pattern. For comparison, the diagonally ruled field MTB-18 is defined by three samples of Mariana Trough basalt recovered from along the spreading axis near 18°N (Hawkins *et al.* 1990). The stippled field Mariana CIP is defined by five basalts from the Central Island Province of the active Mariana arc (Woodhead 1988). Note that the data sets for Mariana CIP and MTB 18 used to construct these fields do not include Pb or Hf and in some instances Th or Ta.

ing the rift stage, which is directly related to the question of how much of the extension is taken up magmatically as opposed to mechanically. For example, Clift (1995) argues that the rift stage generally does not involve voluminous igneous activity in the basin itself. Magnetic bands in the Havre Trough are interpreted as 'pseudo-linear magnetic anomalies' resulting from the spatial and temporal emplacement of highly magnetized sheeted lava and intrusion of dykes between rifted block segments flanking axial rift graben (Wright 1993).

Two important points need to be made with regard to the hypothesized nature of rifting in the northern Mariana Trough. First, no samples of rifted arc crust have been recovered from any location. This does not dismiss the model of rifting because younger flows would be expected to cover such crust, however one might expect to recover older arc crust from some part of the highly faulted Mariana Trough. The only place where arc crust may have been recovered is from basins in the westernmost Mariana Trough at 18°N, where DSDP Site 453 recovered breccias that were clearly derived from an arc infrastructure exposed by faulting (Natland 1982; Party 1981). However, these breccias probably represent debris flows that sampled the West Mariana Ridge and do not indicate the nature of the subjacent crust at the western limit of the Mariana Trough. Within the study area shown in Fig. 1, there are a total of 9 sampling sites. Most of these sites are from steep scarps around the Central Graben, and one or more of these sites would be expected to expose rifted arc crust, if it was a significant component of the back-arc basement in this region. None of the sites listed in Table 1 include material that on the basis of preliminary geochemical data (R. F. Gribble and R. J. Stern, unpubl. data 1996) can be interpreted as rifted arc crust, albeit none of the sites other than the D45 site sampled lower crust and upper mantle.

The second important point is that the suite that we have studied is more similar to Mariana Trough basalts than to Mariana arc lavas. This is clear from every petrologic, geochemical and isotopic perspective that we have considered, including the nature of ultramafic and gabbroic rocks (Fig. 4), elemental compatibility diagrams (Fig. 8), and isotopic data (Fig. 9). This is a reconnaissance investigation, but it was our working hypothesis that we had sampled rifted arc crust, so if our sample selection of the D45 suite was biased, it is towards the identification of rifted arc crust. The felsic rocks are especially significant. Although their origin is not resolved, these may be crustal melts, as discussed in the following section. If they are crustal melts, these sample the crust and especially their isotopic compositions provide a mean isotopic composition of local crust. Nd and Pb isotopic compositions discriminate between Mariana arc and back-arc magmas. Mariana arc magmas

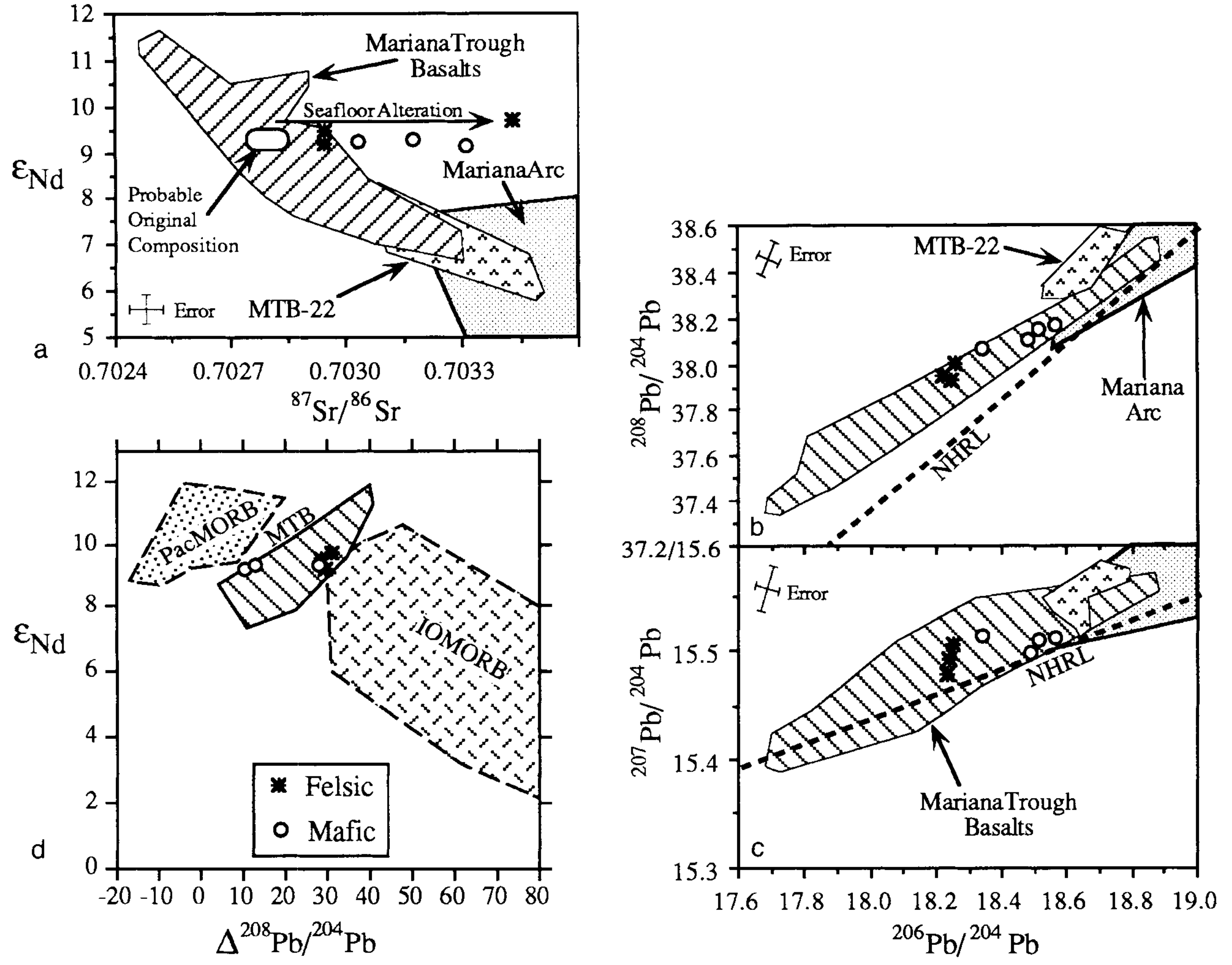

Fig. 9 Isotopic characteristics of Central Graben samples. Fields Mariana Trough Basalts MTB are for samples from the spreading ridge around 18°N and farther south (Volpe *et al.* 1987; Volpe *et al.* 1990; Gribble *et al.* 1996). Fields MTB-22 are for samples from the extension axis to the north of the Central Graben, just north of the study area in Fig. 1, between 21°35′N and 21°50′N (Stern *et al.* 1990). Fields Mariana Arc are defined from DePaolo and Wasserburg (1977); Lin *et al.* (1990); Meijer (1976); Woodhead (1989); Woodhead and Fraser (1985) and Morris *et al.* unpubl. data (1996). (a) Epsilon-Nd vs. $^{87}Sr/^{86}Sr$ of leached whole-rock powders. The samples analyzed here are affected to differing degrees by seafloor alteration. The oval Probable Original Composition is the inferred original composition of the suite. (b) and (c) Pb isotopic composition. NHRL is the northern hemisphere reference line (Hart 1984). (d) Isotopic composition of Nd vs. Pb. Fields for Pacific MORB (Pac MORB) and Indian Ocean MORB (IOMORB) are from the literature summarized in Gribble *et al.* (1996). $\Delta^{208}Pb/^{204}Pb$ measures the deviation of a sample's $^{208}Pb/^{204}Pb$ from that expected for a given $^{206}Pb/^{204}Pb$ on the NHRL, following the algorithm of Hart (1984).

throughout the history of the arc have been characterized by $^{87}Sr/^{86}Sr > 0.7032$, εNd $< +8$ and $\Delta^{208}Pb/^{204}Pb < +30$, whereas MTB basalts extend from arc-like values to $^{87}Sr/^{86}Sr = 0.7025$ to 0.7032, εNd $= +8$ to $+12$ and $\Delta^{208}Pb/^{204}Pb = +30$ to $+60$ (Fig. 9d). The isotopic compositions of Nd and Pb for all three felsic samples fall in the range of back-arc basin magmas, and however they formed, they cannot be melts of arc crust.

We conclude that extension in much of the Northern Mariana Trough was accomplished magmatically. This style of extension is nevertheless distinct from normal seafloor spreading in ways that are incompletely understood, however all of the data indicate that stretched arc crust makes up a negligible part of the substrate of the NMT south of 22°N. The rough 'Basin-and-Range' bathymetry that characterized the seafloor reflects faulting and stretching of this newly formed backarc basin crust and lithosphere.

ORIGIN AND SIGNIFICANCE OF FELSIC MAGMAS

One of the most interesting and enigmatic questions resulting from this study concerns the origin and significance of the felsic plutonic rocks. These are increasingly being recognized as an important component of backarc basin magma systems. They have been long recognized as important parts of back-arc basins developed on continental crust, such as the 'Green Tuff' region related to the Miocene opening of the Sea of Japan, or the felsic volcanism of the Taupo Volcanic Zone of New Zealand, related to the southward propagation of Lau Basin–Havre trough rifting. The importance of felsic igneous activity in intraoceanic back-arc basins has more recently been recognized, beginning with the recovery of dacites from Zephyr Shoal in the Lau Basin (Hawkins 1976). Felsic volcanics have been recovered from the Mariana Trough west of the spreading axis at 18°N (Lonsdale & Hawkins 1985), in the Sumisu Rift and Myojin Basin (Fryer *et al.* 1990; Hochstaedter *et al.* 1990; Ikeda & Yuasa 1989), and are inferred to be an important part of the early rift phase in the Lau Basin (Clift 1995). Our research in the Mariana Trough indicates that there are at least two important felsic volcanic centers along the VTZ north of the Central Graben, and the experience in recognizing the Mariana 'mounds' area as predominantly a zone of off-ridge felsic volcanism with subdued bathymetric expression (Lonsdale & Hawkins 1985) indicates that felsic igneous activity in the Mariana Trough is more important than is commonly acknowledged.

A first-order scientific controversy surrounds the question of how these felsic magmas are generated. Where such rhyolites are found on continental crust, isotopic contrasts between mantle and crust often result in a strong argument for the involvement of pre-existing crust or sediments. For example, Taupo rhyolites are interpreted as resulting from extensive fractionation of basalt that was contaminated by 15–25% of metasediments (McCulloch *et al.* 1994). The issue becomes more problematic for the generation of felsic magmas in oceanic crust, where isotopic differences between possible sources are not easily distinguished. An overview of the problem and plausible explanations for the origin of low-K silicic magmas in oceanic arcs is presented by Beard (1995). Beard's conclusions concern felsic magmagenesis in mafic and hydrous arc systems but are applicable to the problem of felsic magmagenesis in intra-oceanic back-arc basins. He concluded that melts with SiO_2 >62%, K_2O <3%, and Na_2O/K_2O >1 were readily produced by either extensive closed-system fractionation of basalt at $P < 100$ MPa or dehydration melting of low-K amphibolite at $P < 700$ MPa, but generally could not distinguish between these two possibilities. This echoes a larger problem for the origin of so-called 'plagiogranites' common in many ophiolites, with advocates for both fractionation of mafic melts (Coleman & Donato 1979) and anatexis of amphibolite crust (Gerlach *et al.* 1981). These range of possibilities are raised in discussions of the origin of felsic rocks in back-arc basin settings, with some arguing for fractionation (Gill *et al.* 1992) and others for anatexis (Ikeda & Yuasa 1989) for felsic rocks from the same back-arc basin.

Valuable perspectives on the problem of the generation of felsic melts in extensional tracts of oceanic crust come from Iceland, where rhyolitic rocks make up about 10% of the crust and comprise the subordinate component of a bimodal volcanic terrane (Sigurdsson 1977). It is also where careful petrologic research into the problem has been going on for several decades. These rhyolitic rocks are similar in many respects to those that we have studied from the Central Graben, including moderate to low potassium contents and flat to gently sloping REE patterns with negative europium anomalies. The similarity in isotopic compositions between Iceland felsic and mafic rocks has led to equivocal interpretations, with careful study leading some scientists to argue strongly for an origin by extensive fractionation of basalt (Furman *et al.* 1992; Macdonald *et al.* 1990) and others to argue just as strongly for anatexis of mafic crust (Sigmarsson *et al.* 1991). The fact that this controversy continues in a well studied area such as Iceland indicates that it will not be a simple matter to resolve the origin of felsic melts in the Mariana Trough and other back-arc basins.

The data for the felsic material from the Central Graben do not allow us to answer the question of how the Central Graben felsic rocks were generated, but they do offer some new perspectives and allow two hypotheses to be rejected. We can see that the felsic material intrudes the mafic material (Fig. 3), and this suggests that some time elapsed between mafic and felsic phases of activity. There are at least two phases of felsic magma generation, with the first phase being less evolved than the second. The felsic rocks are associated with a range of coarse-grained plutonic rocks and it is possible that this association results in the disruption of a deep crustal magma chamber. There is evidence for felsic rocks intruding amphibolite-facies metabasalts or metadiabases, but none available for felsic rocks intruding other plutonic rocks, suggesting that they are spatially restricted to the top or sides of a magma chamber. These observations — that felsic melts evolve with time and that they are associated with a magma chamber, perhaps with its upper portions — are consistent with a differentiation model. However, we reiterate that the Pb isotopic data for the felsic rocks are distinct from those of the mafic rocks that we have sampled (Fig. 9b,c,d), so if the felsic rocks are derived by magmatic differentiation, less evolved products of the parental magma remain to be identified. The same point must be made with regard to an origin by anatexis. If the felsic rocks are crustal melts, representatives of this crustal progenitor are not identified in the dredge suite, although the isotopic systematics are generally consistent with the isotopic characteristics of Mariana Trough basalts. The data do allow us to reject the possibility that Mariana Trough felsic magmas result from melting of sediments, as was suggested for the rhyodacites near 18°N (Lonsdale & Hawkins 1985). The isotopic data also preclude an origin for the felsic rocks by melting of arc crust. The origin of the felsic rocks suggests either melting of newly formed back-arc basin (BAB) crust or fractionation of BAB basaltic melts. Careful study and sampling of the scarp above the dredge site using a manned submersible may allow us to determine the method of origin.

AGE PARADOX

The time of formation of oceanic crust may be determined by several means, including interpretation of heat flow and magnetics, as well as isotopic determinations. A paradox results when these three approaches are used for Central Graben crust. The radiometric age of 1.8 ± 0.6 Ma is discussed above. Interpretation of seafloor magnetic anomalies in the northern Mariana Trough indicate that the crust beneath the Central Graben is normally magnetized, and this has been interpreted as forming during the Brunhes chron (<0.7 Ma; Yamazaki *et al.* 1993; Martinez *et al.* 1995). Heat flow data for the region around the Central Graben (Fig. 1, Table 2) are generally low (<50 mW/m^2), whereas conductive heatflow for 1.8 Ma old crust should be much higher. Heatflow similar to that measured around the Central Graben is characteristic of seafloor that is much older (>50 Ma; Stein and Stein 1992) than the radiometric age of 1.8 ± 0.6 Ma. Resolution of this paradox awaits a better understanding of the crustal structure, style and age of faulting, and extent of hydrothermal circulation of this region.

CONCLUSIONS

The preliminary examination of the rocks exposed in the Central Graben of the Mariana Trough leads to some valuable insights:

(1) While the importance of mechanical extension has been recognized for intra-arc basins such as the Sumisu Rift (Taylor *et al.* 1991; Klaus *et al.* 1992), we present primary evidence that mechanical extension in back-arc basins can take place in a manner similar to that of continental rifts and slow-spreading mid-ocean ridges. Extension to form the Central Graben took place along low-angle detachment faults, exposing lower crust and upper mantle. These results support the hypothesis based on studies of the Sumisu Rift and the Mid-Atlantic Ridge, that mechanical extension in slowly extending tracts of seafloor is, to a first order, indistinguishable from that of continental crust, with interesting and unexplored implications for the rheology of the lower crust and upper mantle.

(2) This paper documents for the first time the composition of parts of a back-arc basin crustal and upper mantle section. This crustal section is presently thin, no more than 3 km thick, although we cannot evaluate possible effects of tectonic thinning. The crustal suite is remarkably heterogeneous, with a variety of coarse-grained gabbroic, anorthositic, and felsic plutonic rocks making up the lower to middle crust. The upper mantle rocks are partially serpentinized, and the possibility exists that such low-velocity serpentinites make up the lowermost few kilometers of the seismically-defined crust.

(3) We have demonstrated that the mantle and crustal rocks of this sample of back-arc basin lithosphere are similar mineralogically and in terms of LIL/REE/HFSC fractionation to normal MORB lithosphere and dissimilar to typical arc lithosphere. On the other hand, the Central Graben crustal section is more similar to arc lithosphere in having an abundance of felsic rocks and hydrous minerals.

(4) No evidence has been found to support the theory that crustal extension in this part of the Mariana Trough involved mechanical stretching of arc lithosphere, neither in the composition of the rocks themselves nor in the composition of possible anatectic melts. Mechanical extension is important in the evolution of this part of the Mariana Trough, but this extension affects back-arc basin crust that formed 1.8 Ma ago.

(5) The exposure of lower crust and upper mantle in the Central Graben presents an unparalleled opportunity to examine the nature and evolution of back-arc basin lithosphere, the style of mechanical extension in back-arc basins, and the origin of felsic magmas in oceanic crust. The most

interesting exposures probably lie at a depth of 4300 m and greater, up to a depth of about 5500 m. This is beyond the depth range of most submersibles and requires the use of deep ROVs and manned submersibles like the Shinkai 6500. The exposures are of sufficient scientific importance to warrant a carefully planned and intensive program of mapping and sampling that could lead to scientific drilling in the near future.

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