Geodynamic implications of crustal lithologies from the southeast Mariana forearc

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ABSTRACT

The deep submergence research vehicle Shinkai 6500, diving on the Challenger segment of the Mariana forearc, encountered a superstructure of nascent arc crust atop a younger mantle with entrained fragments of metamorphosed crust. A plutonic block from this crust collected at 4900 m depth has a crystallization age of 46.1 Ma and mixed boninitic-arc tholeiitic geochemical signatures. A hornblende garnetite and two epidote amphibolites were retrieved from depths between 5938 m and 6277 m in an area dominated by peridotite. The garnetite appears to represent a crystal cumulate after melting of deep arc crust, whereas the amphibolites are compositionally similar to enriched mid-ocean ridge basalt (MORB). The initial isotopic compositions of these crustal fragments are akin to those of Eocene to Cretaceous terranes along the periphery of the Philippine plate. The garnetite achieved pressures of 1.2 GPa or higher and temperatures above 850 °C and thus could represent a fragment of the delaminated root of one of these terranes. This sample has coeval Sm-Nd, Lu-Hf, and 40Ar-39Ar ages indicating rapid ascent and cooling at 25 Ma, perhaps in association with rifting of the Kyushu-Palau arc. Peak P-T conditions were lower for the amphibolites, and their presence on the ocean floor near the garnetite might have resulted from mass wasting or normal faulting. The presence of relatively fusible crustal blocks in the circulating mantle could have contributed to the isotopic similarity of Mariana arc and backarc lavas with Indian Ocean MORB.

INTRODUCTION

Close association of amphibolites, garnet-rich crustal lithologies, and serpentinitized peridotites is found in mélanges and the deep roots of overthrust oceanic terranes. In mélanges, such as the Catalina schist and Franciscan Formation in California, variably metamorphosed lower-plated rocks become interspersed within matrices derived from ultramafic and sedimentary sources during burial and exhumation processes in subduction zones (e.g., Bebout and Barton, 1989; Sorensen and Grossman, 1989; Wakabayashi, 2012; Pennistom-Dorland et al., 2014). The crust-mantle transition zone of the obducted oceanic arc crust in the Jilal complex in Pakistan (e.g., Dhuime et al., 2007) has a basal crustal sequence of igneous and metamorphic rocks that includes hornblende garnetite. Here, we report an association of epidote amphibolite, hornblende garnetite, and partially serpentinitized mantle peridotites from the inner trench slope of the southern Mariana forearc west of the West Santa Rosa Banks fault (the Challenger segment; Ohara et al., 2012). This “naked” forearc (Stern et al., 2012) lacks accretionary sediments but has a suprasubduction zone stratigraphy reminiscent of ophiolites including volcanic rocks as young as Pliocene as Piocene (Ribeiro et al., 2013), gabbros, and melt-depleted peridotites at its greatest depths (e.g., Ohara et al., 2012). This segment of the forearc contrasts with that to the north and east of West Santa Rosa Banks fault (Fig. 1), which has basement lithologies thought to have formed during subduction initiation and early Izu-Bonin-Mariana (IBM) arc development in the Eocene (e.g., Stern and Bloomer, 1992; Reagan et al., 2010). We focus on key lithologies from three Shinkai 6500 dives that transected the trench slope of the Challenger segment. The dives recovered associated volcanic and plutonic rocks above partially serpentinitized dunites and harzburgites that host epidote amphibolite and hornblende garnetite. In this study, we explore the petrology, geochemistry, and geochronology of the amphibolites and garnetite to constrain their origin and path to the collection site, which bears on the tectonic development of the IBM system and the fate of crust in oceanic terranes. We also determine the age and composition of one complex granitoid to determine an age for magmatic activity associated with construction of the forearc crust in this area.

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Figure 1. (A, B) Location maps for geographic features in the Izu-Bonin-Mariana subduction system and Philippine plate mentioned in the text. (C) Locations for the 2010 Shinkai 6500 dive sites. WSRBF—West Santa Rosa boundary fault; SSF—Shinkai Seep Field (Ohara et al., 2012).
**GEOLOGICAL BACKGROUND**

The IBM arc-trench system is a complex suite of arc, backarc, and oceanic terranes formed along the eastern plate boundary of Eurasia. The oldest terranes include the Cretaceous to Paleocene remnant ocean islands and island arcs that make up the Oki-Daito and Daito ridges and the Amami Plateau in the northern West Philippine Basin (Hickey-Vargas, 1998; Hickey-Vargas, 2005; Ishizuka et al., 2011a; Fig. 1). The Huatung Basin abutting Taiwan on the west side of the West Philippine Basin is Cretaceous oceanic crust with an enriched mid-ocean ridge basalt (E-MORB) affinity (Hickey-Vargas et al., 2008). The southern end of the Philippine plate west of Palau, the Palau Basin, is thought to be Mesozoic oceanic crust (Taylor and Goodliffe, 2004).

The West Philippine Basin opened between ca. 54 and 30 Ma (Deschamps and Lallemand, 2002) and has a basement of MORB-like basaltic lavas (Savov et al., 2006). Subduction of the Pacific plate beneath the West Philippine Basin and older terranes began at ca. 52 Ma. Subduction initiation caused near-trench seafloor spreading and volcanism along the IBM forearc (e.g., Stern and Bloomer, 1992), which resulted in an ophiolitic stratigraphy of depleted peridotites, gabbroic rocks, dolerites, basalts, boninites, and early-arc rocks from deep to shallow and from the trench toward the arc (Reagan et al., 2010). The oldest lavas are highly deformed basalts with ages of 51–52 Ma (Ishizuka et al., 2011a; Reagan et al., 2013). Lavas with boninitic compositions erupted between 49 and 44 Ma (Cosca et al., 1998; Ishizuka et al., 2006). Normal arc volcanism began between 44 and 45 Ma along the length of the nascent IBM arc (Ishizuka et al., 2006; Reagan et al., 2008).

About the same time as subduction began, enriched ocean-island basalt (OIB)–like magmas were generated from a mantle plume (Ishizuka et al., 2013) near the West Philippine Basin spreading center. The oldest of these lavas (45–50 Ma) are preserved in the Minami Daito Basin (Hickey-Vargas, 1998), and progressively younger lavas are found on the western Oki-Daito Ridge (41–45 Ma) and the Urdenata Plateau (40–36 Ma; Ishizuka et al., 2013). Equivalent lavas south of the West Philippine Basin spreading center are found on the Benham Rise (ca. 36 Ma, Hickey-Vargas, 1998). Recent ages obtained on lavas from the Kyushu-Palau Ridge (Ishizuka et al., 2011b) suggest that its subduction-related volcanism occurred largely after the West Philippine basin opened.

The Kyushu Palau arc began rifting at 25 Ma, eventually forming the Parece-Vela–Shikoku Basin by seafloor spreading until ca. 6 Ma (Ishizuka et al., 2011a; Fig. 1). The shallower two dives (6K-1234 and 6K-1236) recovered an ophiolitic suite of samples including deformed peridotites, gabbroic to tonalitic intrusive rocks, and basalt to andesitic lavas. The deepest dive, 6K-1232, largely recovered variably serpentinized dunites and harzburgites. Dive 6K-1232 also recovered three loose fragments of metamorphic crustal rocks scattered amongst the serpentinized peridotites on the sedimented seafloor. These samples comprise a hornblende garnetite (6K1232R07) and two epidote amphibolites (6K1232R13 and 6K1232R16) collected at depths of 6276 m (11°37′21″N; 143°0.6132″E), 6046 m (11°37′33″N; 143°0.4764″E), and 5934 m (11°37′62″N; 143°0.4246″E), respectively. Although no direct contact relationships were observed, the close seafloor association of the amphibolites, garnetite, and peridotites suggests that these diverse lithologies were structurally interleaved.

An image analysis of the surface cut through the hornblende garnetite (Figs. 2B, 3A, and 3B) shows that it has ~69% garnet and 31% hornblende. Less than 1% of the sample consists of accessory minerals of titanite, magnetite, ilmenite, with trace pyrite and chalcopyrite. Garnets are anhedral, essentially unzoned, and average ~1 cm in diameter. Some grains are crosscut by amphibole veins (Fig. 4), consistent with retrograde hydration. Under plane-polarized light (PPL), amphiboles are normally pleochroic green to brownish green to tan but can have blue to tan pleochroism near grain boundaries (Fig. 3A).

The epidote amphibolites (Figs. 2C, 2D, 3C, and 3D) have two generations of amphibole fully grown 50%–70%. The first generation of amphiboles is millimeter scale, often rounded, and brownish-green hornblende. The second generation is bluish-green hornblende intergrown with fine-grained epidote (20%–35%) and 5%–10% accessory phases including titanite, ilmenite, and hematite after pyrite. These samples have ~5% plagioclase predominantly in veins a chlorite.

An intrusive rock (6K1232R14, Figs. 2A, 3E, and 3F) recovered from a depth of 4889 m (11°40′6405″N; 143°2.5850″E) as a loose block atop sediment during dive 6K-1236 consists of mingled fine-grained hornblende quartz diorite and
coarser tonalite and contains zircons. The tonalite has subhedral zoned plagioclase with a matrix dominated by graphically intergrown quartz and feldspar. The finer-grained quartz diorite has abundant round inclusions with rims outlined by fine opaque minerals. Intergrown euhedral plagioclase crystals compose the outer zone (10%–20%) of these inclusions. The core of the inclusions consists of anhedral quartz. We speculate that these round inclusions are segregation vesicles where gas-filled bubbles were filled by interstitial late-stage siliceous fluids. The entire sample is heavily altered, with considerable replacement of sodic plagioclase by fine clays and rare epidote. Secondary chlorite is present in much of the sample.

**METHODS**

The epidote amphibolite and hornblende garnetite whole rocks were analyzed for major-and trace-element compositions, as well as Nd, Hf, and Pb isotopic compositions. In addition, minerals separated from the garnetite were dated using Sm-Nd, Lu-Hf, and 40Ar/39Ar techniques. Minerals in this sample also were analyzed for major-element compositions by electron probe microanalysis (EPMA) and for trace-element compositions by laser ablation-inductively coupled plasma mass spectrometry (LA-ICPMS). Part of the granitoid sample (6K1236R14) was divided into tonalite and quartz diorite fractions.
Figure 3. Photomicrographs of samples used in this study. (A) 6K1223R7 garnetite plane polarized light (ppl); (B) 6K1232R7 garnetite crossed polars (xpl); (C) 6K1232R16 epidote amphibolite ppl; (D) 6K1232R13 epidote amphibolite ppl; (E) 6K1236R14 tonalite xpl; and (F) 6K1236R14 quartz diorite xpl. Scale bars = 1 mm. The principal mineral phases in each thin section are illustrated. Opaque grains in (A) and (B) are primarily ilmenite with inclusions of magnetite and titanite based on scanning electron microscopy-energy dispersive X-ray spectroscopy (SEM-EDS) analysis.
using a rock saw for whole-rock major- and trace-element analyses. The remainder was kept whole for zircon separation.

Weathered rinds were sawed off whole rocks before washing in distilled water and crushing. Randomized portions of the fresh rock chips were powdered for each sample using a ceramic mill. In preparation for Sm-Nd and Lu-Hf geochronology, the hornblende garnetite sample (6K1232R07) was ground and sieved to ~0.1 mm so that monocrystalline grains of garnet and hornblende could be separated using heavy liquids and handpicking. This process obtained five aliquots: hornblende, whole-rock powder, garnet with visible inclusions, “optically clean” garnet without visible inclusions, and “optically clean” garnet that underwent partial dissolution procedures. Mineral separates were hand crushed to a powder in a ceramic mortar and pestle.

**Major- and Trace-Element Analysis**

Whole-rock powders of the amphibolites and garnetite were analyzed by X-ray fluorescence (XRF) at Washington State University using methods described in Johnson et al. (1999). The granitoid was analyzed by inductively coupled plasma–optical emission spectroscopy (ICP-OES) techniques at the University of Iowa. In preparation for major-element analysis by ICP-OES, ~0.2 g of research grade LiBO₂ flux was added to the bottom of a heat-cleaned graphite crucible, followed by 0.1 g of rock powder and then a layer of 0.2 g LiBO₂ on top of the sample. The crucibles were placed in the furnace at 1050 °C for ~16 min, long enough for the sample to become molten. The molten sample was quickly poured into prepared bottles of 50 ml 5% HNO₃ and sonicated for ~10 min to assure dissolution. Approximately 10 ml aliquot of sample solution was placed into a 125 ml bottle and diluted with 70 ml HNO₃. Major-element data were collected from this solution using a Varian 720-ES ICP-OES with settings and conditions discussed in Reagan et al. (2013). Based on six replicate analyses of standard JA-1, errors on precision and accuracy for all major-element oxides are less than 2%. Exceptions are accuracy errors for P₂O₅ (10%) and CaO (3.5%).

For trace-element analysis, ~0.1 g of sample powders was weighed and added to Teflon beakers for dissolution using a combination of concentrated HF acid and concentrated HNO₃ acid. After drying, 5 ml of MilliQ water and 5 ml of concentrated HNO₃ were added to re-dissolve the samples. Solutions were heated at 90 °C for over 12 h and then transferred to 60 ml high-density polyethylene (HDPE) bottles and diluted with MilliQ water. An aliquot of 10 ml of this solution was put into a 250 ml HDPE bottle along with 80 ml of dilute HNO₃ and was spiked using an internal Rh-In spike. These samples were then run using the Thermal X-series ICP-MS according to methods described in Peate et al. (2010). Based on replicate analyses of standard BRP-1 (Peate et al., 2010), precision and accuracy errors for most elements are between 1% and 3%. Elements with higher precision errors are Sc (7%) and Th (8%). Elements with values that diverged from accepted values for BRP-1 by more than 3% are Ni (14%), Nb (7%), Pr (4%), Ta (5%), and Th (5%).

**Electron Probe Micro-Analysis**

The chemical compositions of garnet and hornblende were analyzed using a JEOL electron microscope (JXA733) housed at the Center for Instrumental Analysis, Shizuoka University, Japan. Analytical conditions were 15 kV accelerating voltage, 12 nA probe current, and a beam diameter of 20 μm.

**Ar²⁶/Ar Geochronology**

Hornblende from sample 6K1232R07 was incrementally heated with a 25W CO₂ laser, and the gas was analyzed using a MAP 215-50 mass spectrometer at the University of Wisconsin–Madison following methods described in Jicha et al. (2006). Data are calculated relative to a Fish Canyon sanidine standard age of 28.201 Ma (Kuiper et al., 2008) using the decay constants of Min et al. (2000).

**U-Pb Zircon Ages**

A heavy-mineral separate from a portion of granitoid sample (6K1236R14) was obtained by standard pulverizing, magnetic, and heavy-liquid methods. Individual zircon grains were handpicked under alcohol, mounted in epoxy resin with natural zircon standards, and polished to expose the grain centers.
for analysis by secondary ion microprobe spectrometry (SIMS) at the U.S. Geological Survey (USGS)–Stanford Microanalytical Center sensitive high-resolution ion microprobe–reverse geometry (SHRIMP-RG) facility. Zircon grains were imaged by cathodoluminescence (CL) to expose intra-grain zoning or complexity and aid in placing SIMS spots. The U-Pb and trace-element analyses were performed simultaneously following routines outlined in Barth and Wooden (2006) and Mazdab and Wooden (2006). Additional trace-element data were collected from analytical spots placed adjacent to the initial U-Pb analysis site within the same CL domain. Fractionation corrections were calibrated by replicate analysis of the zircon standard R33 (421 Ma; Black et al., 2004; Mattinson, 2010) with a 2σ calibration error for the R33 206Pb/238U ratio of 0.76% for the analytical session. Ages were calculated from 206Pb/238U ratios corrected for common Pb using the 206Pb method and 207Pb/206Pb ratios corrected for common Pb using the 204Pb method (see Williams, 1998). Initial common Pb isotopic composition was approximated from Stacey and Kramers (1975). The U concentration was calibrated with Madagascar (MAD) green (4196 ppm U, Barth and Wooden, 2010). Data reduction and plotting were performed using the programs Squid 1.13b (Ludwig, 2001) and Isoplot 3.00 (Ludwig, 2003). The trace-element routine collected 7Li, 9Be, 11B, 19F, 23Na, 30Si, 31P, 39K, 40Ca, 45Sc, 48Ti, 49Ti, 56Fe, 89Y, 93Nb, 94Zr, 1H, 169Tm, 172Yb, 175Lu, 90Zr, 139La, 140Ce, 146Nd, 147Sm, 153Eu, 155Gd, 157Gd, 163Dy, 165Ho, 157Gd, 165Ho, 159Tb, 163Dy, 166Er, 168Yb, 176Hf, 177Hf, 180Hf, 206Pb, 232Th, and 238U. Concentrations were calibrated against zircon standard MAD (Mazdab and Wooden, 2006). The estimated errors based on repeated analysis of MAD are 3%–10% for P, Y, Hf, Th, and U and the rare-earth elements (REEs) except for La (20%). The 49Ti data, with an estimated error of 4% based on analysis of MAD, were used to determine the Ti content to avoid interference of 92Zr 4+ with the 49Ti peak (Watson and Harrison, 2005). Ti-in-zircon temperatures were calculated using Ferry and Watson (2007), assuming the activity of SiO 2 is equal to one (and activity of TiO 2 is ~0.7 for rutile-absent siliceous melts; Hayden and Watson, 2007).

Pb Isotopes

Lead isotopic compositions were measured at the University of Wyoming High-Precision Isotope Laboratory using a ThermoFinnigan™ NeptunePlus multicollector–inductively coupled plasma mass spectrometer (MC-ICPMS). Rock powders were dissolved using a mixture of concentrated HF-HNO3 followed by 6M HCl. Lead was purified using standard anion exchange chromatography in HBr after Strelow and Toerien (1966). Samples were analyzed in 5% HNO3 solutions and introduced into the plasma using an Aridus II desolvating nebulizer. Thallium (National Bureau of Standards [NBS]-997) was added to each sample to correct for mass fractionation with Pb/Tl ratio target of ~3/1. Lead and thallium isotopes were analyzed in static mode using six Faraday collectors with ratios normalized to 205Tl/204Pb = 0.418922 to account for instrumental mass bias. A seventh Faraday collector was used to monitor mercury.

Lead-isotope ratios are reported relative to NBS-981 values of Thirlwall (2002). Internal precisions for 206Pb/204Pb, 207Pb/204Pb, and 208Pb/204Pb were ~30 ppm. Analyses of USGS standard BCR-2 gave 206Pb/204Pb = 18.7631, 207Pb/204Pb = 16.6275, 208Pb/204Pb = 38.7254, in agreement with values reported by Weis et al. (2006). External reproducibility for BCR-2 and other rock standard analyses ranges from 200 to 300 ppm. The procedural blank for this analysis was 51 picograms (pg), which is insignificant relative to the sample size.

Lu-Hf and Sm-Nd Ages; Initial Nd and Hf Isotopic Compositions

Neodymium and Hf isotopic compositions of whole-rock powders and mineral separates were performed at the University of Houston using a Nu Plasma II multiple collector–inductively coupled plasma mass spectrometer. Whole-rock, garnet with inclusions, and hornblende samples were powdered, spiked with mixed 176Lu-177Hf and 149Sm-150Nd isotope tracers, and then digested in 8:1 mixtures of distilled HF and HNO3 on a hotplate. One of the picked garnet fractions was washed in HCl to remove any labile and REE-rich components such as phosphate leaving a garnet leached residue fraction. All of the garnet fractions were spiked with mixed 176Lu-177Hf and 149Sm-150Nd isotope tracers and then digested in 8:1 mixtures of distilled HF and HNO3 on a hotplate. After dissolution of all samples, chemical separation of Hf, Nd, Lu, and Sm was conducted following methods outlined in Lapen et al. (2004, 2017). Mass spectrometry, instrumental mass fractionation correction, isotope-dilution calculations, and spike subtraction follow methods outlined in Lapen et al. (2004). Measured Hf and Nd isotopic compositions of standard BCR-2 analyzed during the course of this study are 0.282870 ± 6 and 0.512631 ± 13 (2sd), respectively. External precision of the 176Hf/177Hf and 143Nd/144Nd isotope ratio measurements based on replicate analyses of rock and solution standards during the course of the study are ± 0.006 and 0.0035%, respectively. Blanks of Hf, Nd, Lu, and Sm are <50 pg, <30 pg, <5 pg, and <40 pg, respectively, and are negligible.

RESULTS

Tonalite–Quartz Diorite

The mingled tonalite–quartz diorite sample (6K1236R14) yielded a small population of euhedral, oscillatory zoned zircon (Fig. 5A). Sixteen analyses give a weighted-mean 206Pb/238U age of 46.1 ± 0.7 Ma (mean square of weighted deviates [MSWD] = 1.6; Table 1; Fig. 5B) that is interpreted as the formation age of the crustal lithologies from the southeast Mariana forearc.

Lu-Hf and Sm-Nd Ages; Initial Nd and Hf Isotopic Compositions

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elements that are commonly enriched in subduction-related lavas, such as K, Rb, and Ba (Fig. 6), although these enrichments might have been affected by alteration. The quartz diorite portion of this sample has higher REE concentrations, a flatter REE pattern, and a more pronounced depletion in Nb compared with the tonalite. Overall, the compositions are consistent with mingling between a differentiated boninitic magma and a more mafic but still differentiated arc tholeiite.

Amphibolites and Garnetite

The two epidote amphibolite samples have major-element compositions resembling those of enriched mid-ocean ridge basalts (Table 4), with flat REE patterns, a low Ba/La and no Nb anomaly (Fig. 6B). Uranium is somewhat enriched in both samples, and Ba is relatively enriched in sample 6K1232R16, but these isolated enrichments most likely occurred during metamorphism and hydration. In contrast, the hornblende garnetite has relatively low concentrations of SiO₂, Na₂O, and K₂O compared to basaltic lavas and has relatively high concentrations of Al₂O₃, Fe₂O₃, and CaO (Table 4). The enrichment in the middle rare-earth elements (REEs) compared to heavy and light REE and low Zr/Sr in this sample would be highly unusual for a basalt from any tectonic setting. All of these compositional traits suggest that this sample consists of accumulated garnet and amphibole or pyroxene after extraction of melt either during melting of mafic crust at high pressure (e.g., Garrido et al., 2006) or crystal fractionation of a mafic lava in the deep crust (e.g., Jagoutz et al., 2009). The relatively low Mg # suggests that the associated melt and bulk initial rock or bulk initial magma were relatively differentiated. The relatively high concentrations of Rb, Ba, Sr, and Pb in this sample are consistent with the addition of a hydrous fluid from a subducting plate to its parental magma source.

The amphiboles in the hornblende garnetite are pargasites (Table 5) using the classification scheme of Hawthorne and Oberti (2007). The garnets are dominated by pyrope, almandine, and grossular components and classify as G3 (Grütter et al., 2004) and eclogitic (Schulze, 2003). Al-in-hornblende geobarometry (Schmidt, 1992) gives a pressure of ~1.2 GPa, which we interpret as a pressure along the retrograde path because of the textural evidence that hornblende grew after garnet. A pseudosection was drawn using a whole-rock composition constructed from mineral compositions and proportions in this sample but without Mn, Ti, and P using Theriak/Domino (de Capitani, and Petrakakis, 2010). At 1.2 GPa, an assemblage lacking chlorite and spinel and dominated by garnet, hornblende, and melt is produced at temperatures between 850 and 1000 °C (Fig. 7). Approximately 4% clinopyroxene appears in the pseudosection under these conditions, although it is not present in the garnetite sample. Doubling the Fe³⁺/Fe₂⁺ used in the modeling halves the clinopyroxene proportion but does not remove it entirely. Either the appearance of clinopyroxene is an artifact of errors in mineral solution models, or clinopyroxene once existed in this sample but was replaced by hornblende during retro-
Mineral separates from hornblende garnetite 6K1232R07 produced a Sm-Nd isochron age of 24.8 Ma ± 2.5 Ma and a Lu-Hf isochron age of 25.21±25 Ma, their initial εNd values are significantly more radiogenic (8.3–8.5; Fig. 10A). The 206Pb/204Pb values for the epidote amphibolites are moderate for IBM crustal rocks (18.748 and 18.832) with small positive Δ8/4 values (Table 1; Fig. 10B), whereas the hornblende garnetite has a lower 206Pb/204Pb value (18.345) and a higher Δ8/4 value. This range of isotopic compositions is shared by early arc boninites and pre-arc terranes such as the Amami Plateau and the Oki-Daito Ridge. However, they differ from later plume-related lavas from the Benham Rise, the Oki-Daito Rise, the Urdaneta Plateau, and the Minami-Daito Basin (cf. Ishizuka et al., 2013). All of these comparisons remain valid after adjusting the Pb isotopic composition for the 25 Ma age of the hornblende garnetite.

### DISCUSSION

#### Tonalite-Quartz Diorite

Sample 6K1236R14 consists of mingled boninitic tonalite and arc tholeiitic quartz diorite, illustrating that it is a magma generated in association with subduction. Its age of 46.1 Ma is during the time when the nascent IBM arc
TABLE 2. MAJOR- (WT%) AND TRACE-ELEMENT (PPM) DATA FOR 6K1236R14

|         | Diorite | Tonalite |
|---------|---------|----------|
| SiO₂    | 63.16   | 73.87    |
| TiO₂    | 0.52    | 0.32     |
| Al₂O₃   | 15.86   | 13.21    |
| Fe₂O₃*  | 7.08    | 3.08     |
| MnO     | 0.13    | 0.06     |
| MgO     | 2.15    | 0.82     |
| CaO     | 5.30    | 3.41     |
| Na₂O    | 4.60    | 4.03     |
| K₂O     | 1.12    | 1.14     |
| P₂O₅    | 0.08    | 0.05     |
| SUM     | 101.40  | 100.81   |

XRF trace elements

|         |      |        |
|---------|------|--------|
| Sc      | 21.0 | 6.2    |
| V       | 180  | 57     |
| Cr      | 7    | 2      |
| Ni      | b.d. | b.d.   |
| Cu      | 17   | 9      |
| Zn      | 40   | 21     |
| Sr      | 189  | 165    |
| Y       | 26.5 | 9.2    |
| Zr      | 50   | 128    |
| Ba      | 127  | 145    |

| ICP-MS (measured) |      |        |        |
|-------------------|------|--------|--------|
| Cs                 | 0.16 | 0.106  | 0      | 0.16  | 0.11  |
| Rb                 | 18.1 | 14.5   | 0      | 18.10 | 14.50 |
| K                  | 1.51 | 1.62   | 0.9    | 9463  | 9546  |
| Ba                 | 131.2| 152.3  | 0      | 131   | 152   |
| Sr                 | 200  | 171    | 0      | 200   | 171   |
| Pb                 | 1.42 | 1.24   | 0.9    | 1.42  | 1.24  |
| Th                 | 0.429| 0.437  | 34.8   | 0.43  | 0.45  |
| U                  | 0.153| 0.085  | 96     | 0.16  | 0.11  |
| Nb                 | 1.51 | 1.62   | 2.1    | 1.51  | 1.62  |
| Ta                 | 0.097| 0.083  | 0.084  | 0.083 | 0.083 |
| La                 | 5.62 | 4.07   | 0.01   | 5.62  | 4.07  |
| Ce                 | 14.91| 9.2    | 8.42   | 14.91 | 9.20  |
| Pr                 | 2.19 | 1.19   |        |       |       |
| Nd                 | 10.55| 5.09   | 0.64   | 10.55 | 5.09  |
| Zr                 | 11.6 | 3.2    | 430000 | 50    | 128   |
| Hf                 | 0.67 | 0.19   | 10033  | 1.57  | 3.10  |
| Sm                 | 3.09 | 1.24   | 2.00   | 3.09  | 1.24  |
| Eu                 | 0.74 | 0.55   | 0.56   | 0.74  | 0.55  |
| Gd                 | 3.62 | 1.39   | 23.5   | 3.62  | 1.40  |
| Ti                 | 3175 | 1890   | 9.31   | 3175  | 1889  |
| Tb                 | 0.63 | 0.23   | 9.50   | 0.63  | 0.23  |
| Dy                 | 4.13 | 1.44   | 122.8  | 4.14  | 1.48  |
| Ho                 | 0.9  | 0.3    | 54.83  | 0.90  | 0.32  |
| Er                 | 2.6  | 0.91   | 281.3  | 2.62  | 0.99  |
| Yb                 | 2.53 | 0.92   | 603.2  | 2.58  | 1.09  |
| Lu                 | 0.38 | 0.14   | 121.8  | 0.39  | 0.18  |
| Y                  | 25.7 | 9      | 1509   | 25.8  | 9.4   |

Correction for zircon loss

|         |        |        |        |
|---------|--------|--------|--------|
| Zircon  | 0.01%§ | 0.03%§ |        |

Note: Major-element data are normalized. Abbreviations: bd—below detection limit; ICP-MS—inductively coupled plasma–mass spectrometry; XRF—X-ray fluorescence.

*Before normalization.
§Percent zircon added.
¶Stoichiometric concentration.
### Table 3. Trace-element data for zircon from 6K1236R14 collected by secondary ion mass spectrometry (SIMS) during U-Pb analysis

| Spot | Y (ppm) | La (ppm) | Ce (ppm) | Nd (ppm) | Sm (ppm) | Eu (ppm) | Dy (ppm) | Er (ppm) | Yb (ppm) | Hf (ppm) |
|------|---------|----------|----------|----------|----------|----------|----------|----------|----------|----------|
| 1.1  | 389     | 0.07     | 3        | 0.2      | 0.7      | 0.25     | 8        | 43       | 113      | 263      |
| 2.1  | 789     | 0.22     | 4        | 0.4      | 1.1      | 0.35     | 13       | 66       | 153      | 331      |
| 3.1  | 1184    | 0.06     | 5        | 0.5      | 1.6      | 0.44     | 19       | 96       | 229      | 523      |
| 4.1  | 2991    | 0.05     | 16       | 1.7      | 4.6      | 1.26     | 52       | 243      | 552      | 1120     |
| 5.1  | 1002    | 0.02     | 4        | 0.5      | 1.6      | 0.52     | 17       | 84       | 191      | 407      |
| 6.1  | 498     | 0.03     | 3        | 0.2      | 0.7      | 0.25     | 7        | 39       | 95       | 219      |
| 7.1  | 5910    | 0.14     | 71       | 4.6      | 13.8     | 3.08     | 131      | 530      | 1026     | 1913     |
| 8.1  | 2654    | 0.02     | 11       | 1.8      | 4.9      | 1.38     | 49       | 232      | 487      | 1010     |
| 9.1  | 523     | 0.01     | 2        | 0.2      | 0.6      | 0.22     | 7        | 41       | 100      | 239      |
| 10.1 | 1890    | 0.01     | 14       | 0.8      | 2.6      | 0.60     | 31       | 150      | 341      | 705      |
| 11.1 | 1199    | 0.01     | 6        | 0.5      | 1.8      | 0.57     | 20       | 98       | 229      | 504      |
| 12.1 | 2257    | 0.01     | 10       | 1.2      | 3.7      | 1.08     | 40       | 192      | 428      | 915      |
| 13.1 | 1723    | 0.01     | 9        | 0.7      | 2.3      | 0.59     | 26       | 137      | 320      | 696      |
| 14.1 | 1176    | 0.01     | 5        | 0.4      | 1.3      | 0.52     | 17       | 93       | 225      | 504      |
| 15.1 | 4208    | 0.03     | 22       | 3.4      | 9.2      | 2.43     | 86       | 368      | 746      | 1496     |

**Note:** All analyses were performed on the sensitive high-resolution ion microprobe–reverse geometry (SHRIMP-RG) at the U.S. Geological Survey–Stanford Microanalytical Center at Stanford University, following procedure outlined in Mazdab and Wooden (2006). Temperature estimates calculated from Ferry and Watson (2007).

*aSpot labeled as grain number.spot number; t—trace-element analysis only.*
transitioned from boninite genesis during mantle upwelling and shallow hydrous mantle melting to normal arc volcanism melting related to mantle counterflow (see Ishizuka et al., 2006; Reagan et al., 2010; Ishuzika et al., 2011a). Plutonic rocks collected by dredging between 8150 m and 7200 m depth in the Mariana forearc southeast of Guam and east of the Santa Rosa boundary fault have similar mixed boninite-arc tholeiite affinities (Johnson et al., 2014). This portion of the Challenger segment with its elevated seafloor, therefore, likely represents a fragment of the Mariana early-arc terrane that is present along much of the length of the IBM forearc (Reagan et al., 2013) and is subaerially exposed nearby on the forearc island of Guam (Reagan and Meijer, 1984).
### TABLE 4. MAJOR- (WT%), TRACE-ELEMENT (PPM), AND LEAD-ISOTOPE DATA

|          | Hornblende garnetite | Epidote amphibolite | Epidote amphibolite |
|----------|-----------------------|---------------------|---------------------|
| SiO₂     | 39.72                 | 42.58               | 44.77               |
| TiO₂     | 0.77                  | 1.87                | 1.66                |
| Al₂O₃    | 19.18                 | 14.67               | 14.57               |
| Fe₂O₅    | 17.46                 | 12.88               | 13.03               |
| MnO      | 0.50                  | 0.20                | 0.23                |
| MgO      | 9.63                  | 8.94                | 8.94                |
| CaO      | 11.22                 | 13.57               | 11.57               |
| Na₂O     | 0.96                  | 2.12                | 2.66                |
| K₂O      | 0.34                  | 0.29                | 0.41                |
| P₂O₅     | 0.02                  | 0.17                | 0.14                |
| Sum      | 99.80                 | 97.29               | 97.97               |
| Mg#      | 0.50                  | 0.55                | 0.55                |
| Li       | 1.2                   | 13.8                | 8.5                 |
| Be       | n.a.                  | 0.50                | 0.84                |
| Sc       | 26.8                  | 54.3                | 58.2                |
| V        | 270                   | 392                 | 393                 |
| Cr       | 21                    | 321                 | 199                 |
| Co       | n.a.                  | 54                  | 62                  |
| Ni       | 19                    | 102                 | 81                  |
| Cu       | 35                    | 45                  | 178                 |
| Zn       | 59                    | 120                 | 196                 |
| Ga       | 7                     | 21                  | 22                  |
| Mo       | n.a.                  | 2.37                | 0.93                |
| Sn       | n.a.                  | 1.02                | 1.08                |
| Cs       | n.a.                  | 0.014               | 0.040               |
| Rb       | 1.10                  | 0.99                | 1.48                |
| Ba       | 59.30                 | 9.56                | 96.55               |
| Sr       | 113                   | 124                 | 144                 |
| Pb       | 0.51                  | 0.39                | 0.90                |
| Th       | b.d.                  | 0.264               | 0.300               |
| U        | b.d.                  | 0.178               | 0.208               |
| Nb       | 0.8                   | 4.7                 | 4.5                 |
| La       | 1.30                  | 5.04                | 5.27                |
| Ce       | 4.70                  | 14.45               | 12.94               |
| Pr       | 0.82                  | 2.39                | 2.01                |
| Nd       | 5.20                  | 12.68               | 10.57               |
| Zr       | 19                    | 91                  | 100                 |
| Sm       | 2.27                  | 4.27                | 3.58                |
| Eu       | 1.12                  | 1.61                | 1.34                |
| Gd       | 4.86                  | 5.58                | 4.80                |
| Tb       | 1.16                  | 0.99                | 0.87                |
| Dy       | 9.41                  | 6.60                | 5.91                |
| Ho       | 1.84                  | 1.41                | 1.31                |
| Er       | 4.99                  | 4.06                | 3.86                |
| Tm       | n.a.                  | 0.59                | 0.62                |
| Yb       | 3.91                  | 3.99                | 3.84                |
| Lu       | 0.53                  | 0.58                | 0.58                |
| Y        | 54                    | 37                  | 34                  |
| Hf       | 0.29                  | 2.50                | 2.66                |
| Ta       | 0.01                  | 0.34                | 0.35                |
| ²⁰⁶Pb/²³⁵Pb | 18.3538  | 18.8323           | 18.7478           |
| ±        | 0.0006                | 0.0007              | 0.0006             |
| ²⁰⁶Pb/²³⁵Pb | 15.4962  | 15.5385           | 15.5566           |
| ±        | 0.0005                | 0.0005              | 0.0004             |
| ²⁰⁶Pb/²³⁴Pb | 38.2734  | 38.4302           | 38.4496           |
| ±        | 0.0013                | 0.0014              | 0.0013             |

Abbreviations: n.a.—not analyzed; b.d.—below detection limit.
Amphibolites and Garnetite

The Pb, Nd, and Hf isotopic compositions (Fig. 10) of the epidote amphibolites overlap with those of pre-arc and early-arc lavas in the IBM system and are near the boundary between mid-ocean ridge basalt (MORB) domains from the Pacific and Indian oceans as defined by Pearce et al. (1999). The hornblende garnetite has a significantly lower initial Nd isotopic composition, placing it firmly in the Indian MORB field like most arc and backarc lavas on the Philippine plate (see Pearce et al., 1999; Hickey-Vargas, et al., 2008; Reagan et al., 2008). Similarly, the $\Delta$8/4 Pb value for the hornblende garnetite is high, and its Pb isotopic compositions plot in the Indian Ocean MORB field. Siliciclastic sediments atop the Pacific plate (Meijer, 1976; Plank and Langmuir, 1998)

| Phase | Garnet | Garnet | Garnet | Garnet | Garnet | Garnet | Garnet | Garnet | Garnet | Amphibole | Amphibole | Amphibole | Amphibole | Amphibole | Amphibole | Amphibole | Amphibole |
|-------|--------|--------|--------|--------|--------|--------|--------|--------|--------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|-----------|
| SiO₂  | 38.31  | 38.71  | 39.09  | 38.64  | 38.49  | 38.92  | 39.22  | 39.32  | 39.44  | 39.58  | 39.86 |
| TiO₂  | 0.14   | 0.16   | 0.12   | 0.10   | 0.12   | 0.10   | 0.08   | 0.08   | 0.13   | 1.44   | 1.40 |
| Al₂O₃ | 21.27  | 21.29  | 21.79  | 21.57  | 21.59  | 21.46  | 21.51  | 21.64  | 16.18  | 16.27  | 16.43 |
| FeO⁺  | 19.59  | 18.99  | 19.54  | 19.50  | 19.24  | 19.61  | 19.15  | 12.97  | 13.25  | 13.22  | 13.18 |
| MnO   | 0.93   | 0.86   | 0.93   | 0.81   | 0.82   | 0.80   | 0.89   | 0.15   | 0.16   | 0.11   | 0.14 |
| MgO   | 7.36   | 7.58   | 7.55   | 7.62   | 7.26   | 7.61   | 7.72   | 11.16  | 11.11  | 11.30  | 11.33 |
| CaO   | 11.02  | 11.15  | 11.15  | 11.13  | 11.08  | 11.02  | 11.34  | 12.17  | 11.68  | 11.75  | 11.86 |
| Na₂O  | 0.00   | 0.01   | 0.01   | 0.00   | 0.00   | 0.00   | 0.00   | 0.73   | 0.68   | 0.74   | 0.72 |
| K₂O   | 0.00   | 0.04   | 0.03   | 0.03   | 0.01   | 0.01   | 0.02   | 0.00   | 0.00   | 0.00   | 0.00 |
| Total | 98.64  | 98.78  | 100.21 | 99.40  | 98.60  | 99.53  | 99.93  | 96.16  | 95.95  | 96.56  | 97.00 |

| Phase | Core | Rim | Core | Rim | Core | Rim | Core | Rim | Core | Rim | Core | Rim |
|-------|------|-----|------|-----|------|-----|------|-----|------|-----|------|-----|
| SiO₂  | 39.13| 39.54| 39.72| 39.27| 39.66| 39.52| 39.28| 39.30| 38.98| 38.68| 1.11| 1.06|
| TiO₂  | 1.44 | 1.23 | 1.49 | 1.34 | 1.43 | 1.33 | 1.29 | 1.22 | 1.11 | 1.06 |
| Al₂O₃ | 16.28| 16.48| 16.30| 16.09| 16.40| 16.92| 16.42| 16.77| 17.69| 17.97| 10.93| 11.04|
| FeO⁺ | 13.36| 13.12| 13.33| 13.94| 13.27| 13.29| 13.01| 13.38| 13.19| 13.14| 13.17| 13.14|
| MnO | 0.15 | 0.23 | 0.20 | 0.21 | 0.15 | 0.19 | 0.13 | 0.14 | 0.17 | 0.14 |
| MgO | 11.81| 11.82| 11.59| 11.69| 11.76| 11.91| 11.74| 11.78| 11.74| 12.03| 12.03|
| CaO | 11.02| 11.15| 11.15| 11.13| 11.08| 11.02| 11.34| 12.17| 11.68| 11.75| 11.86|
| Na₂O | 0.00 | 0.01 | 0.01 | 0.00 | 0.00 | 0.00 | 0.00 | 0.73 | 0.68 | 0.74 | 0.72 |
| K₂O | 0.00 | 0.04 | 0.03 | 0.03 | 0.01 | 0.01 | 0.02 | 0.00 | 0.00 | 0.00 | 0.00 |
| Total | 96.01| 96.35| 96.54| 96.15| 96.63| 96.60| 95.62| 96.30| 96.69| 96.87|

| Phase | Core | Rim | Core | Rim | Core | Rim | Core | Rim | Core | Rim | Core | Rim |
|-------|------|-----|------|-----|------|-----|------|-----|------|-----|------|-----|
| SiO₂  | 39.31| 39.01| 39.97| 39.90| 39.74| 39.45| 39.74| 39.45| 39.74| 39.45| 39.74| 39.45|
| TiO₂  | 1.12 | 1.03 | 1.14 | 1.20 | 1.19 | 1.14 | 1.19 | 1.14 | 1.19 | 1.14 |
| Al₂O₃ | 17.33| 17.64| 17.07| 17.20| 17.76| 18.02| 17.76| 18.02| 17.76| 18.02| 17.76| 18.02|
| FeO⁺ | 13.01| 13.37| 12.85| 13.35| 12.92| 13.02| 12.92| 13.02| 12.92| 13.02| 12.92| 13.02|
| MnO | 0.16 | 0.13 | 0.14 | 0.11 | 0.13 | 0.11 | 0.16 | 0.11 | 0.13 | 0.11 |
| MgO | 10.93| 10.78| 11.36| 11.10| 10.74| 10.74| 10.74 | 10.74| 10.74 | 10.74 |
| CaO | 12.09| 12.07| 12.05| 12.16| 12.18| 12.18| 12.18 | 12.18| 12.18 | 12.18 |
| Na₂O | 2.04 | 1.78 | 1.71 | 1.79 | 1.81 | 1.80 | 1.81 | 1.80 | 1.81 | 1.80 |
| K₂O | 0.57 | 0.58 | 0.60 | 0.56 | 0.56 | 0.56 | 0.56 | 0.56 | 0.56 | 0.56 |
| V₂O₅ | 0.09 | 0.05 | 0.04 | 0.04 | 0.06 | 0.06 | 0.06 | 0.06 | 0.06 | 0.06 |
| Total | 96.66| 96.42| 96.92| 97.42| 97.09| 97.06| 97.09| 97.06| 97.09 | 97.06 |

Note: FeO⁺—total Fe as FeO.
also are characterized by high $\Delta$B/4 Pb, but these sediments have much higher $^{208}$Pb/$^{204}$Pb values compared with the hornblende garnette. These data suggest that the amphibolites and garnette are more likely affiliated with the Philippine plate rather than the subducting Pacific plate.

The enriched (E)–MORB-like trace-element compositions of the epidote amphibolites are most consistent with basaltic protoliths that erupted at a spreading center or in an intraplate setting. The tectonic setting for genesis of the hornblende garnette is more ambiguous. Its major- and trace-element composition indicates that it is a crystal cumulate either produced by anatexis and melt removal (Garrido et al., 2006) or crystal cumulate of high-pressure basalt fractionation (Alonso-Perez et al., 2009; Jagoutz et al., 2009). Its enrichment in fluid-soluble elements is consistent with this basaltic crust being the roots of an island arc. The Eocene forearc basalts found on the nearby Mariana trench-slope are too depleted in incompatible trace elements, and most have $^{208}$Pb/$^{206}$Pb values (Reagan et al., 2010) that are too radiogenic for these lavas to be the precursors to any of the crustal samples analyzed here. Similarly, West Philippine basin basalts have Pb isotopic compositions that differ from the amphibolites and garnette and are unlikely protoliths.

More likely equivalents of the metamorphic lithologies from dive sites are the pre-IBM arc terranes of the Philippine plate. These terranes are currently situated in the far backarc, to the west of the Kyushu-Palau Ridge (Fig. 1). Prior to the opening of the Parece-Vela Basin and the Mariana Trough at 25 Ma, the present forearc would have been situated much closer to these terranes (Fig. 1). The best matches in terms of spread in Pb and Nd isotopic compositions (Fig. 10) are the lavas from the eastern Oki-Daito Ridge and Amami Plateau, which have affinities to both ocean-island and arc settings (Hickey-Vargas et al., 2008; Ishizuka et al., 2013). The unusually high Hf isotopic composition for the garnette sample with respect to its Nd isotopic composition is consistent with a pre-arc history for this sample.

The chemical association of the hornblende garnette with older Philippine plate terranes suggests that it was a portion of deep and dense crustal root of a similar terrane that delaminated (e.g., Tatsumi, 2000), melted and disaggregated, and became entrained in asthenospheric mantle. Currently, we can only speculate at the trigger for deep crust delamination. One possibility is that mantle upwelling associated with subduction initiation at 52 Ma (Stern and Bloomer, 1992; Ishizuka et al., 2011a) heated the deep roots of one or more pre-arc terranes, causing them to melt and shed their dense residues (Fig. 11). Thermobarometric constraints from the hornblende garnette suggest it achieved pressures of at least 1.2 GPa (>40 km) and temperatures exceeding 850 °C (Fig. 7). The lack of zonation for its cm-scale garnets suggests these high-grade conditions lasted previous.
The consistent 25 Ma age of the hornblende garnetite determined by multiple techniques (i.e., Lu-Hf, Sm-Nd, and 40Ar-39Ar) with differing closure temperatures suggests rapid cooling at that time. This age is coeval with the initial opening of the Parece-Vela Basin (Ishizuka et al., 2011b). We posit that this rifting at 25 Ma (Fig. 11) caused mantle with entrained hornblende garnetite and related metamorphic blocks to upwell into the rifted Kyushu-Palau arc. This assemblage of rocks then cooled and eventually became part of the mantle beneath the crust sampled by dive 6K-1236. The epidote amphibolites could also have been entrained in the mantle associated with rifting at 25 Ma, but the close association of rocks with different P-T histories and compositions needs additional explanation. Possibilities include: (1) faulting associated with the collision of the Caroline Ridge with the Mariana Trench in the Late Miocene tectonically shuffled rocks in the southern Mariana forearc as has been postulated for interleaved metamorphic rocks and peridotites on Yap Island (see Hawkins and Batiza, 1977; Ohara et al., 2002); (2) rocks with differing histories were chaotically mixed during serpentine diapirism (e.g., Maekawa et al., 1993); (3) epidote amphibolites were mass wasted from crustal outcrops farther upslope onto garnetite-bearing mantle; or (4) normal faulting associated with subduction erosion and exposure of the mantle in the forearc (Reagan et al., 2017) juxtaposed deep crustal amphibolites and garnetites embedded in the upper mantle. Direct evidence for (1) and (2) in the dive area is lacking, whereas evidence for normal faulting and mass wasting in the Mariana forearc is widespread (Martínez et al., 2000; Chapp et al., 2008), suggesting that some combination of (3) and (4) moved rocks from the crust section or uppermost mantle downslope to where they were collected during Shinkai 6500 dive 6K-1232.

Dispersed fragments of crust in the mantle would be relatively fusible, and their presence could affect Nd and Hf isotope values for modern-day lavas from Mariana Arc and Mariana Trough. However, these crustal materials are rare, with only one garnetite collected during all of the recent Shinkai 6500 diving in the Mariana and Bonin forearcs (e.g., Reagan et al., 2010; Ishizuka et al., 2011a; Ohara et al., 2012). Thus, it is unclear both how widespread these fragments are and what their bulk isotope contributions to mantle sources of lavas would be. Nevertheless, the relatively unradiogenic Nd isotopic compositions compared with Hf for the garnetite indicate that the presence of such materials in the mantle might contribute to its overall Indian Ocean–like isotopic nature (cf. Hickey-Vargas, 1998; Pearce et al., 1999; Woodhead et al., 2012).

**CONCLUSIONS**

Shinkai 6500 diving on the Challenger segment of the Mariana forearc during the YK10-12 expedition of the R/V Yokosuka collected a wide variety of mantle and crustal lithologies. A plutonic rock collected at relatively shallow...
Atmospheric argon ratios:
\[ ^{40}\text{Ar}/^{36}\text{Ar} = 298.56 \pm 0.31 \quad \text{Lee et al. (2006)} \]
\[ ^{38}\text{Ar}/^{36}\text{Ar} = 0.1885 \pm 0.0003 \quad \text{Lee et al. (2006)} \]

Interfering isotope production ratios:
\[ ^{40}\text{Ar}/^{36}\text{Ar}_{\text{ex}} = 0.00005 \pm 0.00014 \]
\[ ^{38}\text{Ar}/^{36}\text{Ar}_{\text{ex}} = 0.01210 \pm 0.00002 \]
\[ ^{40}\text{Ar}/^{36}\text{Ar}_{\text{ca}} = 0.000695 \pm 0.00001 \]
\[ ^{38}\text{Ar}/^{36}\text{Ar}_{\text{ca}} = 0.000198 \pm 0.000001 \]
\[ ^{40}\text{Ar}/^{36}\text{Ar}_{\text{ca}} = 0.000265 \pm 0.000002 \]

Decay constants:
\[ \lambda_{^36\text{Ar}} = (0.580 \pm 0.014) \times 10^{-10} \text{ a}^{-1} \quad \text{Min et al. (2000)} \]
\[ \lambda_{^38\text{Ar}} = (4.884 \pm 0.099) \times 10^{-10} \text{ a}^{-1} \quad \text{Min et al. (2000)} \]
\[ \lambda_{^40\text{Ar}} = (2.58 \pm 0.03) \times 10^{-9} \text{ a}^{-1} \]
\[ \lambda_{^38\text{Cl}} = (8.23 \pm 0.042) \times 10^{-10} \text{ h}^{-1} \]
\[ \lambda_{^40\text{K}} = (2.303 \pm 0.046) \times 10^{-4} \text{ a}^{-1} \]

depths was derived from a proto-arc terrane with boninite and arc tholeiite affinities and an age of 48.1 Ma, illustrating that shallow portions of the Challenger segment are fragments of the IBM nascent arc terrane that has been dissected by subsequent rifting. Epidote amphibolites and a hornblende garnetite collected at greater depth from an area dominated by serpentinized dunite and harzburgite have isotopic affinities with the Oki-Daito Ridge and similar pre-IBM arc terranes. Coincident Hf, Nd, and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 25 Ma indicate that the garnetite was uplifted rapidly and cooled at this time. We conclude that the garnetite is a delaminated fragment of the deep arc crust of a Late Cretaceous to Early Eocene terrane in the West Philippine basin. The epidote amphibolites are fragments of shallower crust with an E-MORB affinity. We speculate that mantle flow brought the garnetite to the trench, where it was uplifted and cooled during rifting of the Kyushu-Palau Ridge. The lower-grade epidote amphibolites also might have been transported in the mantle or could represent fragments of the basement of the overlying crust. In either case, the amphibolites were likely juxtaposed with the garnetite by normal faulting and mass wasting related to forearc extension and erosion. Thus, the forearc south of the Mariana Trough has Eocene arc crust underlain by younger mantle with em-
Figure 9. $^{36}$Ar/$^{40}$Ar age spectrum and inverse isochron diagram for hornblende separated from sample 6K1232R7. Data are from Table 7.

Figure 10. (A) Plot of initial $\epsilon$Hf versus $\epsilon$Nd values for the epidote amphibolites and garnetite calculated at 25 Ma. Data are from Table 6. (B) Plot of $\Delta$8/4 versus $^{206}$Pb/$^{204}$Pb for the epidote amphibolites and hornblende garnetite. Data are from Table 4. Delta values were calculated using equations from Hart (1984). The red line is the location of the boundary between the Indian Ocean mid-ocean ridge basalt (MORB) field with high $\Delta$8/4 and $\Delta$7/4 and Pacific MORB field with lower $\Delta$8/4 and $\Delta$7/4 as defined in Pearce et al. (1999). The dashed line labeled “NHRL” represents the Northern hemisphere regression line of Hart (1984). The green fields represent lavas from pre-arc terranes including the Oki-Daito Ridge, Amami Plateau, and Huatung Basin. The yellow fields labeled “ODMP” are lavas associated with the Oki-Daito Mantle Plume. The orange fields labeled “WPB” are seafloor basalts from the West Philippine Basin. The blue fields labeled “KPR” represent lavas from the Kyushu-Palau Ridge. The gray fields labeled “FAB” represent forearc basalts, and the patterned fields represent boninites from forearc dive sites, Deep Sea Drilling Project sites 458 and 459, and Chichijima (data sources: Hickey-Vargas, 1998; Pearce et al., 1999; Hickey-Vargas 2005; Savov et al., 2006; Hickey-Vargas et al., 2008; Reagan et al., 2010; Ishizuka et al., 2011a, 2011b; Ishizuka et al., 2012; Hickey-Vargas et al., 2013).
Figure 11. Schematic diagrams illustrating the potential origin of the crustal rocks discussed in this paper. (A), (C), and (E) represent cross sections along lines illustrated on schematic maps (B), (D), and (F). Brown areas illustrate arc crust; basaltic crust originating in a subduction initiation, backarc, or mid-ocean ridge setting is gray; mantle is illustrated in light green. The dashed line in (F) is the West Santa Rosa boundary fault. AP—Amami Plateau, DR—Daito Ridge, ODR—Oki-Daito Ridge, WPB—West Philippine Basin, MOC—Mesozoic oceanic crust, KPR—Kyushu Palau Ridge, WMR—West Mariana Ridge, PVSB—Parece-Vela and Shikoku basins, Bon—early-arc boninitic crust. The red star illustrates the location of the Shinkai 6500 dive sites 6K1232 and 6K1236, which were on a lithospheric block consisting of Eocene crust underlain by more recently emplaced mantle. The AP-DR-ODR area consists of Mesozoic arc terranes, plume-related ocean islands, and basins of late Mesozoic and Eocene age (see Ishizuka et al., 2013; Arculus et al., 2015). MOC probably consisted of oceanic lithosphere with minor arc and ocean-island terranes. Izu-Bonin-Mariana (IBM) plate rotations over time are not illustrated. (A, B) Early-arc phase shortly after initiation of subduction when upwelling mantle might have triggered delamination of deep segments of a crustal terrace, such as the Oki-Daito Ridge. (C, D) Rifting at 25 Ma to create the Parece-Vela and Shikoku basins, causing disaggregated fragments of delaminated crust to rise to shallow levels in the mantle. The dots with varied colors represent fragments of the crust, which could have varying metamorphic grade. (E, F) The situation today, where mantle with embedded crustal blocks crop out near the Mariana Trench.
ACKNOWLEDGMENTS

Rasmus Andreasen, David Peate, and Guillaume Girard are thanked for analytical assistance. We thank the Japan Agency for Marine-Earth Science and Technology for funding the cruises of the RV Yokosuka and the Shinkai 6500 diving. We also thank the Shinkai 6500 and RV Yokosuka crews for their outstanding work. MR was funded by National Science Foundation grant 0804862; LH was funded by an Iowa Center for Research by Undergraduates grant from the University of Iowa; KM was funded by the Japan Society for the Promotion of Science (Kiban-S 16H06347). Constructive reviews by Bob Stern and two anonymous reviewers helped us significantly improve the manuscript.

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