The uppermost mantle section below a remnant proto-Philippine Sea island arc: Insights from the peridotite fragments from the Daito Ridge

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ABSTRACT

The Amami Plateau, Daito Ridge, and Oki-Daito Ridge of the northwestern Philippine Sea Plate are remnants of Mesozoic island arcs. We have newly recovered samples of peridotite and peridotite-derived minerals from the Daito Ridge. The peridotite samples are composed of serpentinized/altered olivines, orthopyroxene porphyroclasts, small clinopyroxenes, and spinels, indicating a harzburgitic origin. Chondrite- and primitive mantle-normalized trace-element patterns for clinopyroxenes are characterized by a steep positive slope from middle rare earth elements to heavy rare earth elements (HREEs) plus yttrium. The light rare earth elements (LREEs) and Sr and Zr contents of clinopyroxenes vary in abundance, and some crystals have high LREE/HREE ratios coupled with positive Sr and Zr anomalies. These petrological and geochemical characteristics are not consistent with the Daito peridotites being the residue of a single partial melting event including melt extraction expected for mid-ocean ridge mantle. Instead, the peridotite source must have been enriched with slab-derived components, which are associated with arc-related magma. Thus, it is concluded that the studied peridotite fragments belong to an exhumed mantle section of a remnant proto-Philippine Sea island arc.

Keywords: Philippine Sea Plate, Daito Ridge, remnant arc, peridotite, slab-derived components, melting model; New Advances in Subduction Zone Magma Genesis

INTRODUCTION

The Philippine Sea Oceanic Plate has a complex architecture comprising several geotectonic domains, including three backarc basins: the West Philippine Basin, the Parece Vela Basin, and the Shikoku Basin (Lallemand 2016; Wu et al. 2016 and references therein). These back-arc basins, together with the Izu-Bonin-Mariana (IBM) Arc system, were formed following the initiation of subduction of the Pacific Plate under the Philippine Sea Plate at ~52 Ma (Ishizuka et al. 2011). Although the IBM Arc is one of the most studied intra-oceanic arcs, little is known about the proto-Philippine Sea Plate that existed along the Pacific Plate margin.

Cretaceous rocks have been reported in WNW-ESE trending sub-parallel ridges (the Amami Plateau, the Daito Ridge, and the Oki-Daito Ridge and Plateau; herein referred to collectively as the Daito Ridges) that line up from the north to south across the northwestern Philippine Sea Plate (Hickey-Vargas 2005; Ishizuka et al. 2011) (Fig. 1). Geophysical data indicate a total crustal thickness of 15–25 km in the Daito Ridge area (Nishizawa et al. 2014), making it an ideal location to reconstruct the tectonic evolution of the proto-Philippine Sea Plate.

Serpentinites and their minerals, such as chromian spinel and clinopyroxene, have been recovered from the Daito Ridge in the past (Site 445 in Fig. 1) (Yuasa and Watanabe 1977; Tokuyama et al. 1980), but their origin has remained unclear. To further explore the architecture of the proto-Philippine Sea Plate, we conducted sampling and direct observations of the Daito Ridges region during two expeditions operated by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC). The first expedition (YK10-04) of the R/V YOKOSUKA used a manned-submersible (SHINKAI 6500) coupled with a deep-tow camera survey, and the second expedition (KR13-15) used a remotely operated vehicle (KAIKO 66 7000II). Gabbroic, granitic, and metamorphic rocks were discovered in an exposed deep-crustal section, along with volcanic rocks ranging from basaltic to andesitic in composition. These findings are consistent with the proposed arc origin of the Daito Ridge region (Tokuyama 1995; Hickey-Vargas 2005; Ishizuka et al. 2011). Geochemical and age data for igneous and metamorphic rock samples collected from the ridge will be reported in a separate communication. Samples of serpentinite were recovered from the Daito Ridge (location 7K611 in Fig. 1) during the KR13-15 cruise. In this paper, we focus on these newly recovered samples and discuss their origin in the context of the tectonic evolution of the Philippine Sea Plate.

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[†] Special collection papers can be found online at http://www.minsocam.org/MSA/ AmMin/special-collections.html.



FIGURE 1. (a) Bathymetric map of the Daito Ridge, prepared using GeoMapApp (Ryan et al. 2009). (b) Enlargement of the area outlined in **a**, showing the sampling location (7 K#611) of this study. The location of Site 445 of DSDP Leg 58 is also shown (Klein and Kobayashi 1980). Ayu T. = Ayu Trough, C.B.F. rift = Central Basin Fault rift, L.O. FZ = Luzon-Okinawa Fracture Zone, Nankai T. = Nankai Trough.

GEOLOGICAL BACKGROUND AND SAMPLE DESCRIPTIONS

The Philippine Sea Plate has a lozenge shape, with dimensions of ~3400 km from north to south and ~2600 km from east to west. Previous studies have shown that it has experienced a complex tectonic evolution (e.g., Uyeda and Ben Avraham 1972; Hilde and Lee 1984; Seno and Maruyama 1984; Hickey-Vargas 1991; Hall et al. 1995; Okino et al. 1999; Lallemand 2016; Wu et al. 2016). To the northwest, the Philippine Sea Plate is subducting below the Eurasian Plate along the Ryukyu Trench, up to the Nankai Trough. To the east, the Pacific Plate is subducting below the Philippine Sea Plate, resulting in the formation of the IBM Arc. In the southern part of the Philippine Sea Plate, a small divergent segment, the Ayu Trough, is present. In addition, the plate hosts two back arc basins: the Eastern Shikoku and Parece Vela basins, which opened during the Neogene (Karig 1971; Uyeda and Ben Avraham 1972). The Daito Ridges itself has been interpreted as a remnant Cretaceous-age segment of an island arc (Tokuyama 1995; Hickey-Vargas 2005; Ishizuka et al. 2011).

Shiki et al. (1977) summarized the results of early 1970s cruise expeditions to the northwestern Philippine Sea, and documented arc-type igneous rocks and metamorphic schists (up to epidote-amphibolite facies) in the Daito Ridge region, which they attributed to pre-Cenozoic island-arc activitiy. Yuasa and Watanabe (1977) reported serpentinites, hornblende schist, and tremolite schist from the Daito Ridge area, with the serpentinites consisting of mesh-structure serpentine, green spinel, and magnetite, with small amounts of clinopyroxene. Tokuyama et al. (1980) reported detrital minerals such as chromian spinel, diopside, and enstatite at DSDP Site 445 (Daito Ridge), which were likely derived from serpentinized peridotites.

The samples of the present study are characterized by peridotitic clasts and the peridotite-derived minerals, such as orthopyroxene, spinel, and clinopyroxene, cemented in a matrix consisting predominantly of carbonate minerals (Fig. 2; Supplemental¹ Fig. A1). Although the peridotitic fragments are typically highly serpentinized/altered to dark brown-colored minerals, they still contain fresh crystals of orthopyroxene, clinopyroxene, and spinel (Fig. 2). Tremolite is locally associated with the orthopyroxene.

MINERAL CHEMISTRY: ANALYTICAL METHODS

Major-element compositions of minerals (Tables 1–3 and Supplemental¹ Tables S1–S3 in the Supplementary material) were determined using an electron probe micro-analyzer (EPMA, IEOL JXA-8800 Superprobe) at Kanazawa University, Japan. The analyses were performed with an accelerating voltage of 15 kV and a beam current of 20 nA, using a 3 µm beam diameter. Natural and synthetic mineral standards were used for calibration, and JEOL software using ZAF corrections was employed for data reduction. In-house mineral standards (olivine, chromian spinel, diopside, and K-feldspar) were measured repeatedly to monitor data quality. The measured concentrations in these minerals were consistent with the averaged values from long-term analyses, being within 1 St.dev. for every element. Data precision, which was established through multiple analyses of one point in the house-prepared standard minerals, was better than 5 and 10% relative standard deviation from the averaged values for elements with abundances of >0.5 and <0.5 wt%, respectively. Details of the EPMA analyses can be found in Morishita et al. (2003).



FIGURE 2. Typical examples of serpentinized clasts (brown-colored areas in the image) plus peridotite-derived minerals, hosted in a carbonate matrix (transparent areas). (a) Thin section of sample R08-1. (b) Photomicrograph (plane-polarized light) of a lithic clast within sample R08-1, containing spinel, orhotpyroxene, and clinopyroxene. (c) Same as in b, but in cross-polarized light. (Color online.)

TABLE 1. Averaged spinel compositions

<u> </u>	0.04	6.0	Daa	6.0	Dea	6.0	D00.4	6.0	
Sample	R01	S.D.	R02	S.D.	K03	S.D.	R08-1	S.D.	
wt%	spinel	(n = 7)	spinel	(n = 3)	spinel	(n = 6)	spinel	(<i>n</i> = 11)	
TiO ₂	< 0.05		< 0.05		< 0.05		< 0.05	0.01	
AI_2O_3	33.73	2.7	30.32	2	32.14	1.4	29.96	1.7	
Cr_2O_3	35.29	2.7	38.84	3.1	36.39	1.7	39.32	1.9	
$Fe_2O_3^*$	2.19	0.3	1.14	1.7	1.71	0.8	0.78	0.3	
FeO	13.78	0.6	15.68	0.5	16.36	0.3	16.27	0.7	
MnO	0.21	0.03	0.23	0.04	0.24	0.05	0.27	0.04	
MgO	15.53	0.3	13.81	0.6	13.61	0.4	13.35	0.5	
CaO	< 0.03		< 0.03		< 0.03		< 0.03		
NiO	0.11	0.02	0.11	0.04	0.1	0.03	0.08	0.03	
Total	100.8		100.1		100.6		100		
Mg#	0.668	0.01	0.611	0.02	0.597	0.01	0.594	0.02	
Cr#	0.413	0.04	0.462	0.04	0.432	0.02	0.468	0.03	
YFe ³⁺	0.024	0.003	0.013	0.02	0.019	0.009	0.009	0.003	
Notes: S.D. = 1σ standard deviation. Fe ₂ O ₃ and FeO were calculated based on stoichiometry. Mg# = Mg/(Mg+Fe ²⁺); Cr# = Cr/(Cr+Al); YFe ³⁺ = Fe ³⁺ /(Cr+Al+Fe ³⁺).									

Rare earth element (REE) and trace-element (Li, Ti, Sr, Y, Zr, and Nb) compositions of orthopyroxene and clinopyroxene were determined using 193 nm ArF excimer laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) at Kanazawa University (Agilent 7500S equipped with MicroLas GeoLas Q-plus, Ishida et al. 2004). Clinopyroxenes and orthopyroxenes were analyzed by ablating 50–90 and >90 μm spot diameters, respectively, at 8 Hz. The NIST SRM 612 standard was used as the primary calibration standard and was analyzed at the beginning of each batch consisting of $n \le 8$ unknowns, with a linear drift correction applied between each calibration. The element concentrations of NIST SRM 612 for the calibration were selected from the preferred values of Pearce et al. (1997). Data reduction was facilitated using ²⁹Si as an internal standard, based on Si contents obtained by EPMA following the protocol essentially identical to that outlined by Longerich et al. (1996). NIST SRM 614 was analyzed for quality control during measurement. The measured concentrations in NIST SRM 614 glass, which is a synthetic silicate glass with a nominal concentration of 1 ppm for 61 elements including REEs, were consistent with

TABLE 2 Representative orthopyroxene compositions

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Sample no.	R01	R01	R02	R03	R03	R03	R08-1	R08-1
An. no.	1-67	1-68	2-135	2-101	3-102	3-125	1-44	1-10
SiO ₂	57.27	57.27	57.02	57.24	56.81	56.24	57.41	56.28
TiO ₂	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05
AI_2O_3	2.46	2.43	2.51	2.17	2.40	2.55	2.21	2.34
Cr ₂ O ₃	0.68	0.64	0.75	0.46	0.61	0.74	0.65	0.70
FeO	5.51	5.28	5.62	5.45	5.45	5.46	5.44	5.18
MnO	0.14	0.18	0.08	0.14	0.13	0.13	0.16	0.14
MgO	34.47	34.10	34.38	34.42	34.19	33.36	34.18	33.25
CaO	0.75	0.84	0.54	0.45	0.52	1.40	0.80	2.36
Na₂O	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	<0.02
K ₂ O	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	< 0.02	<0.02
NiO	< 0.05	0.08	< 0.05	0.11	0.06	0.11	0.08	0.09
Total	101.3	100.8	100.9	100.4	100.2	100.0	100.9	100.4
Mg#	0.918	0.920	0.916	0.918	0.918	0.916	0.918	0.920
Cr#	0.155	0.149	0.166	0.125	0.145	0.163	0.165	0.168
LA#	#201	#203	#101			#Y111	#112	#111

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			Parts p	er milli	on					
Ti	70.1	31.9	47.2	n.a.	n.a.	62.0	51.0	52.7		
Sr	0.005	0.039	0.149	n.a.	n.a.	0.478	0.311	0.004		
Υ	0.115	0.09	0.092	n.a.	n.a.	0.124	0.089	0.082		
Zr	0.034	0.079	0.062	n.a.	n.a.	0.526	0.271	0.031		
Nb	0.040	0.041	0.040	n.a.	n.a.	0.073	0.027	0.029		
Dy	0.01	0.005	0.005	n.a.	n.a.	0.009	0.006	0.004		
Ho	0.003	0.003	0.002	n.a.	n.a.	0.006	0.003	0.003		
Er	0.024	0.019	0.017	n.a.	n.a.	0.016	0.017	0.016		
Tm	0.007	0.005	0.005	n.a.	n.a.	0.005	0.004	0.004		
Yb	0.066	0.05	0.055	n.a.	n.a.	0.058	0.050	0.048		
Lu	0.017	0.009	0.014	n.a.	n.a.	0.009	0.010	0.011		
Notes: An. no. = analytical point number, Mg# = Mg/(Mg+Fe) atomic ratio, Cr# = Cr/(Cr+Al) atomic ratio. n.a. = not analyzed.										

Sample no.	R01	R01	R01	R01	R01	R01	R01	R01	R01	R02	R02	R03	R03	R03	R08	R08	R08
An. no.	2-69	1-75	2-76	2-78	1-79	2-81	1-86	2-91	1-95	4-137	3-142	2-99	2-105	1-115	1-23	1-25	1-51
								wt%									
SiO ₂	53.10	53.91	53.39	53.95	53.84	53.37	54.05	53.40	53.48	53.86	53.97	54.06	53.61	53.90	54.33	53.45	53.28
TiO ₂	< 0.05	0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	<0.05	<0.05	< 0.05	<0.05	<0.05	< 0.05	<0.05
Al ₂ O ₃	2.69	1.70	2.89	1.62	2.77	2.10	1.70	2.50	2.55	2.10	2.21	2.33	2.12	2.21	2.06	1.98	2.24
Cr_2O_3	0.96	0.70	1.04	0.38	0.96	0.56	0.62	0.78	0.93	0.62	0.74	0.83	0.88	0.73	0.85	0.69	0.89
FeO	1.85	1.82	2.24	1.80	2.21	1.93	1.80	2.10	1./1	2.00	1.89	1.07	1.08	2.02	1.50	1.05	1.85
MnO	0.08	< 0.05	17.07	0.06	19.08	0.13	10.08	17/2	0.07	1756	1749	1722	0.05	0.05	0.05	17.46	1747
CaO	23 73	24 31	22 75	23 57	22.36	23 57	21.89	23 51	24 35	23 75	23.89	24.42	24.42	23 75	24 33	24.01	23.46
Na ₂ O	0.08	0.08	0.07	0.06	0.08	0.07	0.18	0.06	0.06	0.04	0.07	0.04	0.05	0.04	0.11	0.10	0.09
K ₂ O	< 0.02	< 0.02	< 0.07	< 0.02	< 0.02	< 0.07	< 0.02	< 0.02	< 0.02	< 0.02	<0.07	< 0.02	< 0.02	< 0.02	< 0.02	<0.02	<0.02
NiO	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	< 0.05	0.05	0.05	0.05	< 0.05	0.05	0.05	0.06	< 0.05	0.05	< 0.05
Total	99.7	100.0	100.4	99.4	100.3	99.6	98.6	99.9	100.4	100.1	100.3	100.8	100.3	100.3	100.6	99.5	99.4
Mg#	0.943	0.944	0.934	0.945	0.936	0.943	0.946	0.937	0.947	0.940	0.943	0.949	0.949	0.939	0.952	0.950	0.944
Cr#	0.193	0.215	0.195	0.135	0.189	0.152	0.198	0.174	0.197	0.166	0.183	0.193	0.218	0.181	0.216	0.189	0.210
LA dia (µm)	50	60	90	50	90	60	90	50	50			50	50	50	60	50	60
							Pa	rts per n	nillion								
Li	5.2	6.9	7.5	5.6	5.4	5.7	6.7	6.0	4.0	n.a.	n.a.	2.8	2.4	3.1	4.2	5.3	5.7
Sc	45	54	52	51	51	53	56	54	57	n.a.	n.a.	60	61	50	50	55	48
Ti	58	66	61	130	82	78	79	68	103	n.a.	n.a.	57	64	47	121	139	90
V	139	143	150	155	153	133	126	148	166	n.a.	n.a.	164	176	159	178	173	165
Cr	6697	7277	7827	5902	7200	4737	6118	6661	8079	n.a.	n.a.	5354	6605	5330	6568	7229	7037
CO NI:	22	19	23	25	23	21	1/	23	20	n.a.	n.a.	18	18	22	20	19	20
INI Cr	0.21	55/	309 1.40	0 1 7	0 1 0	0 2 2	070	0.20	0.76	n.d.	n.d.	55Z	524	0 27	012	0.21	570
51 V	0.21	0.55	0.71	0.17	0.10	0.55	0.70	0.20	0.70	n.d.	n.d.	0.10	0.02	0.57	0.15	0.51	0.24
7r	0.04	0.09	146	0.00	0.00	0.03	1 10	0.80	1.14	na.	n a	0.58	0.09	113	0.00	0.75	0.33
Nb	0.041	0.039	0.053	0.036	0.047	0.45	0.044	0.043	0.048	n a	n a	0.035	0.048	0.038	0.10	0.043	0.034
Ba	0.047	0.106	0.382	<	< 0.017	0.092	0.300	< 0.015	0.276	n.a.	n.a.	<	< 0.010	0.220	< 0.012	0.064	0.079
La	0.006	0.009	0.027	<	0.003	0.006	0.010	<	0.014	n.a.	n.a.	<	<	0.020	<	0.006	0.005
Ce	0.016	<	0.026	0.008	0.010	0.007	0.011	0.012	0.017	n.a.	n.a.	<	<	<	<	<	<
Pr	<	<	0.005	<	<	0.001	0.002	<	<	n.a.	n.a.	<	<	<	<	<	<
Nd	<	<	0.019	<	<	<	0.009	<	<	n.a.	n.a.	<	<	<	<	<	<
Sm	<	<	<	<	<	<	<	<	<	n.a.	n.a.	<	<	<	<	<	<
Eu	<	<	0.003	<	<	<	<	<	<	n.a.	n.a.	<	<	<	<	<	<
Gd	<	<	<	<	<	<	<	<	<	n.a.	n.a.	<	<	<	<	<	<
Tb	<	<	0.004	<	0.004	0.003	0.004	<	0.004	n.a.	n.a.	<	<	<	<	<	<
Dy	0.051	0.064	0.067	0.074	0.072	0.065	0.069	0.059	0.102	n.a.	n.a.	0.047	0.039	0.060	0.052	0.055	0.044
Но	0.024	0.026	0.022	0.034	0.033	0.029	0.030	0.023	0.036	n.a.	n.a.	0.017	0.025	0.020	0.021	0.023	0.023
Er T	0.11	0.14	0.12	0.13	0.15	0.15	0.14	0.13	0.20	n.a.	n.a.	0.10	0.14	0.10	0.12	0.12	0.09
Im	0.019	0.026	0.021	0.032	0.030	0.030	0.027	0.026	0.039	n.a.	n.a.	0.023	0.031	0.022	0.022	0.018	0.019
YD	0.21	0.21	0.20	0.22	0.26	0.25	0.25	0.23	0.37	n.a.	n.a.	0.20	0.24	0.18	0.20	0.20	0.17
Lu	0.031	0.036	0.036	0.039	0.047	0.039	0.041	0.037	0.062	n.a.	n.a.	0.036	0.042	0.031	0.029	0.035	0.031
Notes: An. no	. = analyt	ical poin	t numbe	r, Mg# = l	Mg/(Mg+	-Fe _{total}) at	omic rati	10, Cr# = (_r/(Cr + A	AI) atomio	c ratio, LA	dia = la	ser diam	eter, < =	lower tha	in detect	ion limit,

TABLE 3. Representative clinopyroxene compositions

n.a. = not analyzed.

previously reported values to within 10% relative standard deviation, and the data precision was better than 10% relative standard deviation for all the analyzed elements. Details of the analytical method and data quality control are given by Morishita et al. (2005a, 2005b). Representative trace-element compositions of clinopyroxenes and orthopyroxenes in the studied samples are listed in Table 3.

RESULTS

The Cr# [= Cr/(Cr+Al) atomic ratio], Mg# [= Mg/(Mg+Fe²⁺) atomic ratio], and YFe³⁺ [= 100 Fe³⁺/(Cr+Al+Fe³⁺) atomic ratio] values of spinel in the studied samples are 0.37–0.52, 0.56–0.68, and <2.2, respectively (Fig. 3). The TiO₂ content of spinel is commonly below the detection limit (<0.05 wt%).

The Mg# [= Mg/(Mg+Fe^{total}) atomic ratio], Al₂O₃, and TiO₂ content of orthopyroxene are 0.91–0.92, 2.0–2.6, and <0.05 wt%, respectively. Chondrite-normalized REE patterns for orthopyroxenes show an abrupt decrease from heavy REEs (HREEs) to middle REEs (MREEs) (Fig. 4a). Middle and light REEs (LREEs) are generally below the detection limit. Some of the clinopyroxenes exhibit positive Zr and Sr anomalies (Fig. 4b).

The Mg# and Al_2O_3 content of clinopyroxene are 0.94–0.95 and 1.0–2.6 wt%, respectively. The TiO₂ and Na₂O contents of

clinopyroxene are low, at <0.05 and <0.1 wt%, respectively. Chondrite-normalized REE patterns for clinopyroxenes show an abrupt decrease from HREEs to MREEs, and MREEs are generally below the detection limit (Fig. 4a). Clinopyroxene shows variations in LREEs, Sr, and Zr concentrations. Some of the clinopyroxenes exhibit high LREE/HREE ratios coupled with low abundances of MREEs (lower than the detection limit), resulting in roughly V-shaped REE patterns and are Zr and Sr positive anomalies (Fig. 4b). The partition coefficients of HREEs and Y between clinopyroxene and orthopyroxene are consistent with those in earlier studies (Supplemental¹ Table S4) (Ozawa and Shimizu 1995; Kelemen et al. 2003; McDade et al. 2003a, 2003b), indicating chemical equilibrium with regard to these elements.

DISCUSSION

Protolith of the studied peridotite fragments

The protolith of the studied peridotite fragments prior to serpentinization/alteration was harzburgitic, as evident from the presence of coarse-grained orthopyroxene porphyroclasts with minor amounts of clinopyroxene and spinel, together with the



FIGURE 3. Relationship between the Mg# and Cr# of spinel. (a) Comparison with the Philippine Sea back-arc peridotites. Data are from Loocke et al. (2013) and the Dunite-type (D-type) peridotites of Ohara et al. (2003). (b) Comparison with typical mid-ocean ridge and fore-arc peridotites. The compositional ranges for mid-ocean ridge peridotites and fore-arc peridotites are compiled by Morishita et al. (2011), and those for mid-ocean ridge lherzolites and harzburgites are from Warren (2016).



FIGURE 4. (a) Chondrite-normalized REE patterns and (b) primitivemantle-normalized trace-element patterns of the Daito clinopyroxenes and orthopyroxenes. Chondrite and primitive mantle values are from McDonough and Sun (1995). The typical detection limit of elemental analyses is also shown for 50 μ m (+) and 90 μ m (X) diameter laser beams.



FIGURE 5. Relationship between the Al₂O₃ content (wt%) in the cores of orthopyroxene porphyroclasts and the Cr# in coexisting spinel. The compositional ranges for mid-ocean ridge and fore-arc peridotites are from Morishita et al. (2015).

serpentinized/altered olivines, as the dominant minerals. Another line of evidence supporting a pre-serpentinization harzburgite lithology is the reasonably high Cr# of the spinel in each sample (Fig. 3b). Major element compositions of the Daito peridotite minerals overlap with those from abyssal peridotites recovered from the mid-ocean ridges and fore-arc settings (Figs. 3 and 5). Therefore, it is not possible to use major element compositions of minerals to constrain the tectonic setting from which the studied samples originated. However, the trace-element compositions of clinopyroxene does yield information regarding the tectonic setting and/or magmatic processes experienced by the studied samples, as explained below.

Comparison with other peridotites

Figure 6 compares the trace-element compositions of clinopyroxenes in the studied samples with that of other spinel peridotite samples analyzed in previous studies. Geochemical data from peridotites with plagioclase and/or gabbroic veins in previous works are not included for further discussions because plagioclase-bearing (and/or gabbroic vein-bearing) peridotites undergo significantly chemical alteration when they interact with melts during the formation of plagioclases/gabbroic veins (e.g., Ohara 2006).

The Daito clinopyroxenes are characterized by higher LREE/ HREE ratios and Sr contents than those of back-arc peridotites from the Philippine Sea Plate (Ohara et al. 2002, 2003). The Zr and Sr are enriched relative to elements with similar compatibility in some Daito clinopyroxenes (Fig. 4b). Figure 7 compares the primitive-mantle-normalized Sr/Yb and Zr/Yb ratios in the Daito clinopyroxenes with those of peridotite samples analyzed in other studies. The Yb content of clinopyroxene in residual peridotites is a good indicator of the degree of partial melting (e.g., Hellebrand et al. 2001). The Sr/Yb and Zr/Yb ratios decrease as the degree of simple partial melting (e.g., fractional melting) increases. Clinopyroxenes from the Izu-Bonin-Mariana fore-arc peridotites (Parkinson and Pearce 1998) and the Hess Deep (Dick and Natland 1996) do not exhibit Zr enrichment, with the exception



FIGURE 6. Comparisons between primitive-mantle-normalized trace-element patterns for clinopyroxenes in the studied samples (gray) and those from previous studies (black). The primitive-mantle values are from McDonough and Sun (1995). (**a**) Comparison with Philippine Sea backarc peridotites, which have been modified to a lesser extent by melt impregnation. Data are from Ohara et al. (2002), and the fertile peridotites (F-type) studied by Ohara et al. (2003). (**b**) Comparison with Izu-Bonin-Mariana fore-arc peridotites. Data are from Parkinson and Pearce (1998). (**c**) Comparison with the mantle section of the Josephine Ophiolite, Oregon, U.S.A. Data are from Le Roux et al. (2014). The stippled lines represent samples that show anomalous trends. (**d**) Comparison with peridotites samples from the Hess Deep of the East Pacific Rise. One sample, depicted by the dotted line, displays a similar trend to the Daito Ridge peridotites. Data are from Dick and Natland (1996).



FIGURE 7. Correlations between primitive-mantle-normalized Sr/Yb and Zr/Yb ratios in clinopyroxenes from the Daito Ridge (this study), Izu-Bonin-Mariana (IBM) fore-arc peridotites (Parkinson and Pearce 1998), the Josephine Ophiolite (Le Roux et al. 2014), and the Hess Deep (Dick and Natland 1996). The primitive mantle values are from McDonough and Sun (1995). The fractional melting trend is also shown.

of one sample from the Hess Deep that has geochemical characteristics similar to those of the Daito clinopyroxenes (Figs. 6 and 7). Elevated Sr and Zr concentrations are also observed in some Josephine Ophiolite peridotite samples, which are thought to represent residual mantle after flux melting in a subduction zone (e.g., Le Roux et al. 2014; Figs. 6 and 7).

Magmatic processes deduced from geochemical modeling

Here we discuss the magmatic processes that the samples are inferred to have undergone, based on geochemical modeling using the trace-element compositions of clinopyroxenes. The trace-element compositions of clinopyroxenes, in terms of LREE/ HREE ratios and Sr content, are distinct from the residue after a single episode of fractional melting and melt extraction (Figs. 7 and 8), when compared with typical residual peridotites that formed by decompression melting (e.g., at a mid-ocean ridge). In addition, the positive Zr spike in clinopyroxene is a unique characteristic of the studied samples (Figs. 5b and 7).

We modeled the clinopyroxene trace-element patterns using (1) an open-system melting model in which the input and output of fluids/melts are associated with partial melting and melt



FIGURE 8. Observed and modeled primitive-mantle-normalized trace-element patterns for clinopyroxene (Cpx). Details of how the data were obtained are outlined in the text. (a) Modeled clinopyroxene trace-element compositions after fractional melting under conditions of the spinel (spl) peridotite stability field. (b) Modeled clinopyroxene trace-element compositions after fractional melting under conditions of the spinel (spl) peridotite stability field, following previous fractional melting under conditions of the garnet (Grt) peridotite stability field. The patterns shown are for residue after 5, 10, and 15%, and 5, 10, and 13% fractional melting from depleted MORB mantle (DMM) of Workman and Hart (2005). (c) Modeled influx and clinopyroxene compositions after open-system melting. (d) Modeled clinopyroxene compositions after influx melting under conditions of the spinel peridotite stability field. The influx composition, taken from Li et al. (2013), is estimated from slab-derived boninite melts from the Izu-Bonin-Mariana fore-arc. (e and f) Modeled clinopyroxene are the same as those used in the open-system melting model (d). The percolating melt mass ratio (α_j) decreases gradually from $\alpha_1 = 1$ to $\alpha_j = 2 \times (j-1)$ (j > 1) away from the location where the melt–rock reaction was initiated. The patterns shown are for compositions after $\alpha_1-\alpha_7$ plate models.

Open system m	nelting		Mineral/melt partition coefficients						
Starting compositions	5% fractional melting from a DMM at			Dol/lq	Dsp/lq	Dopx/lq	Dcpx/lq		
(first stage melting)	garnet peridotite condition (see Fig. 8a)								
			La	0.000007	0.0008	0.0025	0.035		
Influx melting	13% from residue after first melting		Ce	0.00001	0.0008	0.005	0.055		
(second stage melting)	at spinel peridotite condition		Sr	0.00019	0.0008	0.007	0.065		
degree of melting			Nd	0.00007	0.0008	0.01	0.15		
	Estimation of influx melting	Using influx composiitons	Zr	0.007	0.07	0.07	0.15		
		of Li et al. (2013)	Sm	0.0007	0.0008	0.02	0.25		
Result	Figure 8c	Figure 8d	Eu	0.00095	0.0008	0.03	0.32		
critical melt fraction: α	0.05	0.02	Ti	0.015	0.15	0.15	0.23		
dimensionless influx rate: β	1	0.02	Dy	0.004	0.0015	0.05	0.35		
Others	The effect of trapped melt crysta	Ý	0.0065	0.00225	0.06	0.37			
			Er	0.009	0.003	0.07	0.37		
Simplified pl	late model		Yb	0.009	0.0045	0.08	0.37		
Starting compositions	5% fractional melting at g	5% fractional melting at garnet peridotite then							
(see Fig. 8b)	13% fractional melting	13% fractional melting at spinel peridotite							
percolating melt/fluid ratio	Gradual decreasing awasy								
	clinopyroxene/melt mass ratio is char	nged from $\alpha_1 = 1$, then $\alpha_i = 2(j-1)$							
Result	Figures 8e	and 8f							

TABLE 4. Melting conditions and parameters used for geochemical models

extraction of the peridotite itself (Ozawa and Shimizu 1995; Ozawa 2001) and (2) a simplified plate model in which porous fluid/melt flow through mantle rocks is simulated by fluid/melt transfer from one cell to another [e.g., Akizawa et al. (2016) after Vernières et al. (1997)]. The melting conditions and parameters used in the models are summarized in Table 4.

First, we estimated the flux composition using the opensystem melting model formulated by Ozawa (2001). As shown below, several parameters in the melting model were adjusted to obtain the best possible results. To meet the requirement of a steep gradient in REE distribution from MREEs to HREEs plus Y (compare Figs. 8a and 8b), we assumed that the starting composition was a residue following 5% fractional melting of a depleted MORB mantle (DMM) (Workman and Hart 2005) under pressure-temperature conditions of the garnet peridotite stability field (first-stage melting). Melting proportions for the garnet peridotite are as follows: 0.045 olivine + 0.965 clinopyroxene + 0.14 garnet = 0.14 orthopyroxene + 1 melt (simplified after Walter 1998). Then, the mineral mode for the garnet stability conditions was converted to spinel peridotite based on the following reaction: 3 orthopyroxene + 1 clinopyroxene + 1 spinel = 4 garnet + 1 olivine (Takazawa et al. 1996). A total of 13% melting was used for the second-stage melting (flux-induced open system melting). Melting proportions for the second-stage open system melting are as follows: 1 melt = 0.05 spinel + 0.45orthopyroxene + 0.75 clinopyroxene - 0.25 olivine (simplified after Kinzler and Grove 1992). Influx composition with regard to La, Ce, Sr, and Zr concentrations was manipulated to match the analytically obtained values, assuming open-system melting under the following conditions: (1) the presence of a critical melt fraction in the system at which the system becomes open to melt separation at a constant rate of $\alpha = 0.05$; and (2) a dimensionless influx rate (influxed mass fraction of the initial solid, divided by the degree of melting) of $\beta = 1$. The effect of trapped melt crystallization was not considered. In contrast, influx compositions for HREEs plus Y and Ti were assumed to be 0.002 times (except for Sm = 0.005 times) the primitive mantle values. The calculated influx composition is characterized by high LREE/ HREE ratios, with positive Sr and Zr spikes (Fig. 8c). Of note, positive Zr spikes are associated with boninites (Cameron et al. 1983; Bloomer and Hawkins 1987; Taylor et al. 1994; Li et al.

2013). Li et al. (2013) inferred a slab origin (via flux melting) for boninites exhibiting a positive Zr spike from the Izu-Bonin-Mariana Arc.

We have also estimated the composition of clinopyroxene composition by modeling the second-stage open-system melting of the residual peridotites following first-stage melting under the conditions stated above. We used the flux composition of Li et al. (2013), which was estimated for slab-derived melts. The conditions used for influx melting were as follows: degree of partial melting = 13%, α = 0.02, and β = 0.02. Our calculation suggests that the observed positive Zr spike in clinopyroxenes could be qualitatively reproduced by interaction with a Zr-enriched slab-derived influx component under reasonable melting parameters. However, the steep MREE-HREE gradient could not be accurately reproduced (Fig. 8d).

The clinopyroxenes (and orthopyroxenes) observed in thin sections show a wide range of trace-element compositions. Clinopyroxenes in a single thin section might be derived from several peridotite samples because the studied samples are aggregates of small peridotite fragments (Fig. 2). Although the special distribution of the samples is not clear, this small-scale variation could be explained by the percolation of fluids/melts through the peridotites. We tested this theory by applying the plate model (Akizawa et al. 2016, simplified from Vernières et al. 1997) to simulate the chemical characteristics of the Daito clinopyroxenes. The chemical compositions of percolating melt changes with increasing cell numbers, j, of the percolation column, according to the following equation: $C_{melt j} = (C_{melt (j-1)} +$ $\alpha_i C_{Cpx 0} / (1 + D\alpha_i)$. The initial trace-element composition of the percolating melt (C_{melt0}) and initial clinopyroxene (C_{Cpx0}) are the same as those used in the open-system melting model (taken from Li et al. 2013). The residue after 13% fractional melting under conditions of the stability of spinel peridotite is assumed to be formed from a residual following 5% fractional melting under the stability conditions of garnet peridotite (Fig. 8b). We assume that the percolating melt/fluid mass ratio decreases gradually away from the location where the melt-rock reaction was initiated. The clinopyroxene/melt mass ratio (α_i) is assumed to change from $\alpha_1 = 1$ to $\alpha_i = 2 \times (j-1)$ (j > 1). Positive Sr and Zr anomalies in clinopyroxenes are observed in reacted columns that are located far from the first percolating column (Figs. 8e and 8f).

In conclusion, both open-system melting and plate models with Sr- and Zr-enriched fluids/melts, which considered representative of slab-derived melts, can qualitatively reproduce the chemical characteristics of the Daito clinopyroxenes, implying that the Daito peridotites were formed in an island arc environment. These results support the theory that the Daito Ridge region represents a section of the uppermost mantle to the lower crust from a remnant arc of the proto-Philippine Sea Plate.

IMPLICATIONS

Newly recovered peridotite samples from the Daito Ridge are of subarc origin. Our results, coupled with acquisition of granitic samples from the Daito Ridges, suggest that the Daito Ridge represents an exposed mantle section of the proto-Philippine Sea Plate. This raises the question of how the uppermost mantle section of island arc was brought to the ocean floor without having been thrust there during plate convergence, as occurs during ophiolite obduction. Further detailed investigations of the studied samples, plus similar samples from ophiolites, are needed to enhance our understanding of such mantle section and their exhumation.

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Endnote:

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