Petit-spot lavas on the western Pacific Plate: contribution of carbonatite and recycled oceanic crust

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The manuscript is going to be submitted to \textit{Solid Earth}.

Keywords: Petit-spot volcano, alkali basalt, carbonatite, asthenosphere

1
Abstract

Petit-spot volcanism, which occurs owing to the plate flexure, have been reported from around the world. As the petit-spot melts ascent from the asthenosphere, they provide the essential information of the lithosphere–asthenosphere boundary (LAB). Here, we observed the lava outcrops of six monogenetic volcanoes formed by petit-spot volcanism in the western Pacific. Thereafter, we determined the \(^{40}\text{Ar}/^{39}\text{Ar}\) ages, major and trace element compositions, and Sr, Nd, and Pb isotopic ratios of the petit-spot basalts. The \(^{40}\text{Ar}/^{39}\text{Ar}\) ages of two monogenetic volcanoes were ca. 2.6 Ma (million years ago) and ca. 0 Ma, respectively. The isotopic compositions of the western Pacific petit-spot basalts suggest geochemically similar melting sources. They were likely derived from a mixture of high-\(\mu\) (HIMU) mantle-like and enriched mantle (EM)-1-like components related to carbonatite/carbonated materials and recycled crustal components. A mass balance-based melting model implied that the characteristic trace element composition (i.e., Zr, Hf, and Ti depletions) of the western Pacific petit-spot magmas could be explained by the partial melting of \(\sim5\%\) crust-bearing garnet lherzolite with 10\% carbonatite flux to a given mass of the source. This result confirms the involvement of carbonatite melt and recycled crust in the source of petit-spot melts and provides an implication for the genesis of tectonic-induced volcanoes including Hawaiian North Arch volcanics and Samoan petit-spot like rejuvenated volcanoes having similar trace element composition to petit-spot basalts.

Short Summary

Plate tectonics theory is understood as the motion of rocky plate (lithosphere) over ductile zone (asthenosphere). The causes of lithosphere–asthenosphere boundary (LAB) is controversial, but petit-spot volcanism supports the presence of melt at the LAB. We conducted geochemistry, geochronology, and geochemical modeling for petit-spot volcanoes on the western Pacific Plate, and the results suggested that carbonatite melt and recycled oceanic crust have induced the partial melting at the LAB.

1 Introduction

Among the upper mantle-derived alkali basaltic lavas in oceanic settings, those on thicker plates away from the mid-ocean ridge, could be divided into plume-related and non-plume-related volcanoes. For example, plume-related North Arch and post-erosional (rejuvenated-stage) volcanoes have been reported in Hawaii and Samoa (Bianco et al., 2005; Bizimis et al., 2013; Clague and Frey, 1982; Clague
and Moore, 2002; Dixon et al., 2008; Frey et al., 2000; Garcia et al., 2016; Hart et al., 2004; Konter and Jackson, 2012; Koppers et al., 2008; Reinhard et al., 2019; Yang et al., 2003). Non-plume related intraoceanic alkali volcanoes, called as petit-spot volcanoes, probably originate where nearby plate subduction causes plate flexures and upwelling of asthenospheric magma (Hirano et al., 2006; Hirano and Machida, 2022; Machida et al., 2015, 2017; Yamamoto et al., 2014, 2018, 2020). Therefore, the occurrence of petit-spot volcanisms supports the presence of melt at lithosphere-asthenosphere boundary (LAB) below the area at least.

The presence of melt in the uppermost asthenosphere could be due to small-scale convection, heating, or the presence of hydrous or carbonatitic components (Hua et al., 2023; Korenaga, 2020). In particular, the presence of CO$_2$ and carbonated/carbonatitic materials is key in the formation of alkaline, silica-undersaturated melt in the upper mantle (Dasgupta and Hirschmann, 2006; Dasgupta et al., 2007, 2013; Kiseeva et al., 2013; Novella et al., 2014). Experimental studies have shown that the solidus of carbonate-bearing peridotite is lower than that of CO$_2$-free peridotite (Falloon and Green, 1989; 1990; Foley et al., 2009; Ghosh et al., 2009). In addition, carbonatites and Si-undersaturated melts are generated through the partial melting of CO$_2$-bearing or carbonated peridotite. The produced melts could exhibit continuous chemical variations depending on pressure (i.e., depth). Namely, carbonatitic melts are produced in the deep asthenosphere (300 km to 110 km), while carbonated or alkali silicate melts are generated in the shallower upper mantle (~110 km to ~75 or 60 km) (Keshav and Gudfinnsson, 2013; Massuyeau et al., 2015, 2021). Indeed, primary carbonated silicate magma and evolved alkali basalts have been simultaneously observed at the post-spreading ridge in the South China Sea (Zhang et al., 2017; Zhong et al., 2021). Hawaiian rejuvenated volcanoes were also attributed to be explained by a carbonatite-metasomatized source with or without silicate metasomatism (Borisova and Tilhac, 2021; Dixon et al., 2008; Zhang et al., 2022).

Submarine petit-spot volcanoes on the subducting northwestern (NW) Pacific Plate may originate from carbonate-bearing materials and crustal components (pyroxenite/eclogite) based on the characteristic trace element, enriched mantle (EM)-1-like Sr, Nd, and Pb isotopic, and relatively low Mg isotopic compositions (Liu et al., 2020; Machida et al., 2009, 2015). In particular, the depletion of specific high-field-strength elements (HFSEs) (i.e., Zr, Hf, and Ti) and the abundant CO$_2$ of petit-spot basalts imply that their melting sources are related to carbonated materials (Hirano and Machida, 2022; Okumura and Hirano, 2013). Here, the nature of the uppermost part of the asthenosphere beneath the oldest Pacific Plate aged 160 Ma, was characterized using the eruptive ages and geochemical properties of six newly observed petit-spot volcanoes and lava outcrops. We verified the contribution of carbonatitic components and crustal materials to the melting source of petit-spot volcanoes to understand the nature of the underlying lithosphere-asthenosphere system and model the geodynamic evolution of the region.
In the last 20 years, the increasing knowledge of petit-spot volcanic settings has provided useful insights on the nature of the lithosphere-asthenosphere system, especially in the NW Pacific region (Hirano et al., 2006; Hirano and Machida, 2022). As other implications, subducted petit-spot volcanic fields with geological disturbances on the seafloor play a role in controlling the hypocentral regions of megathrust earthquakes (Fujiwara et al., 2007; Fujie et al., 2020; Akizawa et al., 2022), and the vestige of hydrothermal activity owing to petit-spot magmatism were recently reported (Azami et al., 2023).

Petit-spot melts, which originated from the asthenosphere unrelated to mantle plume, could be a key to elucidating the nature of the LAB (Hirano and Machida, 2022). Their asthenospheric origin was supported by MORB-like noble gas isotopic ratios, multi-phase saturation experiment, and geochemistry (Hirano et al., 2006; Hirano and Machida, 2022; Machida et al., 2015, 2017; Yamamoto et al., 2018). The LAB is identified as a discontinuous transition in seismic velocities at the base of the lithosphere, and its causes are attributed to hydration, melting, and mineral anisotropy with considerations for the unique characteristics in each tectonic setting (e.g., Rychert and Shearer, 2009). The occurrence of petit-spot volcanism substantiates the existence of melt at the LAB below the area at least (Hirano et al., 2006). Recently, similar volcanic activities have been observed worldwide including Java (Sunda) Trench, Tonga Trench, Chile Trench, Mariana Trench, Costa Rica, North American Basin and Range, and southern offshore of Greenland, implying the universal occurrence of petit-spot and similar magmatisms (Axen et al., 2018; Buchs et al., 2013; Falloon et al., 2022; Hirano et al., 2013, 2016, 2019; Reinhard et al., 2019; Taneja et al., 2016; Uenzelmann-Neben et al., 2012; Yamamoto et al., 2018, 2020; Zhang et al., 2019). Although there is still an open question of whether the LAB discontinuity is due to the differences in the physical properties of minerals (e.g., Hirth and Kohlstedt, 1996; Kang and Karato, 2023; Karato and Jung, 1998; Katsura and Fei, 2021; Stixrude and Lithgow-Bertelloni, 2005; Wang et al., 2006) or the presence of partial melts (e.g., Audhkhasi and Singh, 2022; Chantel et al., 2016; Conrad et al., 2011; Debayle et al., 2020; Herath et al., 2022; Hua et al., 2023; Kawakatsu et al., 2009; Mierdel et al., 2007; Sakamaki et al., 2013; Yoshino et al., 2006), the occurrence of petit-spot volcanism reveals the partial melting of the asthenospheric mantle of the region because they erupted on the seafloor without hotspot and ridge activities (Hirano et al., 2006; Hirano and Machida, 2022; Machida et al., 2015, 2017; Yamamoto et al., 2014, 2018, 2020).

The petit-spot volcanic province on the abyssal plain of the western Pacific is surrounded by Cretaceous seamounts and oceanic islands of the Western Pacific Seamount Province (Koppers et al., 2003) and located approximately 100 km southeast of the Minamitorishima (Marcus) Island (Fig. 1a). The study area corresponds to the oldest portion of the Pacific Plate aged at 160 Ma and the foot of
the outer-rise bulge related to the Mariana subduction system (Hirano et al., 2019; Fig. 1b). Such a subduction-related fore-bulge in front of the Mariana Trench has been numerically modeled and detected in satellite gravity maps despite crosscutting by several seamounts (Bellas et al., 2022; Hirano et al., 2019; Zhang et al., 2014, 2020). The petrography, geochemistry, and geochronology of petit-spot basalts and zircons in peperites, which were collected from a knoll, suggested that petit-spot magmas in this region ascend from the asthenosphere along the concavely flexed plate in response to subduction into the Mariana Trench at younger than ~3 Ma (Yamamoto et al., 2018; Hirano et al., 2019). Below the study area, low seismic velocity zone is observed under the lithosphere (Li et al., 2019; Fig. 1c). Notwithstanding the low velocity anomalies crosscutting the lower mantle (Fig. 1c), no active hotspots (i.e., heat supplies) have been reported around the western Pacific petit-spot province surrounded by Cretaceous Wake seamount chains including Minamitorishima Island and Paleogene intraplate volcanoes (Koppers et al., 2003; Aftabuzzaman et al., 2021; Hirano et al., 2021). The other petit-spot lava outcrops were observed in a volcanic cluster during three research cruises using the research vessel (RV) Yokosuka (YK16-01, YK18-08, and YK19-05S) with five dives using the submersible, Shinkai 6500 (6K#1466, 6K#1521, 6K#1522, 6K#1542, and 6K#1544; Fig. 2), and fresh basalts were collected. The information of sampling point, depth, thickness of palagonite rind and manganese-crust, and age of the western Pacific petit-spot basalts are provided in Table 1.
Fig. 1. Geological and geophysical information of the study area. (a) Bathymetry of the western Pacific near the Mariana Trench. The red box shows the study area to the southeast of Minamitorishima (Marcus) Island (Fig. 2). The bathymetric data are adopted from ETOPO1 (NOAA National Geophysical Data Center; http://www.ngdc.noaa.gov/). (b) Seafloor age map of the same area as (a). This study area is on a 160–170 Ma Pacific Plate, called the Jurassic Quiet Zone (JQZ) (Tivey et al. 2006). The present absolute motion of the Pacific Plate and the seafloor age are derived from studies by Gripp and Gordon (1990) and Müller et al. (2008), respectively. (c) The cross-section P-wave tomography beneath the thick yellow line including the study area on the ETOPO1 bathymetry map (left). The tomographic image (right) was drawn using the SubMachine (Hosseini et al., 2018; http://www.earth.ox.ac.uk/~smachine/cgi/index.php) on applying the data of Lu et al. (2019).
Fig. 2. Detailed bathymetry of the study area. The onboard multibeam data were surveyed during the YK10-05 and YK18-08 cruises by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC). The petit-spot knolls and outcrops were investigated during several dives as 6K#1466, 6K#1521, 6K#1522, 6K#1542, and 6K#1544. The pink-colored stars represent the sampling points. The age information was obtained in the present study and Hirano et al. (2019).

### 3 Field observations, sample locations, and petrography

<table>
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<tr>
<th>Cruise</th>
<th>Dive</th>
<th>Sample name</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
<th>Depth, m</th>
<th>Palagonite rind, mm *1</th>
<th>Manganese crust, mm *1</th>
<th>Ar-Ar age, Ma</th>
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*1: The samples which have no data of palagonite and/or Mn-crust thickness are due to the lack of them or crumbled.

*2: This is a reference value due to the lack of radiogenic 40Ar in this sample.
Here, the eruptive sites of monogenetic volcanoes or lava outcrops are approximately along each
dive site numbered 6K#1466, #1521, #1522, #1542, and #1544 conducted using the Shinkai 6500.
Only the 6K#1466 dive was conducted at two types of monogenetic volcanoes, divided into the glassy
(R3) and crystalline, vesicular (R6 and R7) types based on the geochemical and petrographic
descriptions and occurrence of basaltic samples.

3.1 YK16-01 cruise and 6K#1466 dive

During the YK16-01 cruise, a small conical knoll (ca. 0.04 km$^3$) was investigated by a
submersible dive, 6K#1466 (Figs. 2 and 3a). The lava flows, observed hollow lava tube resulting in
sediment-rolling/disturbing eruption, were located approximately 600 m south of the top of the knoll
(extremely fresh and glassy samples; 6K#1466R3-001 and R3-004 basalts) (Fig. 3a). Vesicular pillow
basalts were collected on the western slope of the knoll (samples 6K#1466R6-001, R7-001, and R7-
003; Fig. 3a). Although only the strong acoustic reflection could not completely distinguish the petit-
spot lava fields in ferromanganese nodule fields, this dive revealed lava outcrops using a sub bottom
profiler (SBP) and a multi narrow beam echo sounder (MBES). In detail, the petit-spot lava field, as
an acoustically opaque layer, was identified by a vigorous backscattering intensity in the MBES with
the distributions of the basement and sediment layers in the SBP.

The 6K#1466R3-001 and R3-004 samples were extremely fresh glassy basalts. The R3-001 and
R3-004 basalts exhibited similar petrographic features (Fig. 3a). These basalts were covered by 3.0–
4.5 mm-thick palagonite (hydrated quenched glass), and their outermost parts were surrounded by
5.8–7.2 mm-thick ferromanganese crust (Fig. 3a). They were less vesicular (<3 vol.%) and dominantly
basaltic glass with euhedral–subhedral olivine microphenocrysts (~100–500 μm in size), ferrotitanium
oxide (<50 μm in size), and minor plagioclase (~500 μm in size) (Fig. 3a). Secondary phases (e.g.,
clay minerals) were not observed.

The 6K#1466R6-001, R7-001, and R7-003 basalts, covered with 4.3–5.2 mm-thick
ferromanganese crust over 5.5–6.6 mm-thick palagonite rinds, exhibited high vesicularity (20–40
vol.%) (Fig. 3a). Certain pyroxene-dominated xenocrysts and peridotite xenoliths have been reported
by Mikuni et al. (2022). The basaltic groundmass comprised needle-shaped clinopyroxene (50–400
μm in size), subhedral olivine partly with aureoles of iddingsite (up to 100 μm in size), ferrotitanium
oxide, minor spinel (up to 10 μm in size), glass, and crystallite, notably without remarkable
phenocrysts (Fig. 3a). The photomicrograph of R6-001 is shown in Fig. 3a.

3.2 YK18-08 cruise and 6K#1521 and #1522 dives
Two submersible dives (6K#1521 and 1522) were conducted during the YK18-08 cruise to investigate petit-spot volcanoes. During the 6K#1521 dive, a small lava outcrop was discovered in the abyssal plain by tracing the strong acoustic reflection, which was expectedly derived from intrusive rock bodies, in the sedimentary layer detected by deep-sea SBP equipped on the *Shinkai 6500*. We observed that the strong reflective surface gradually became shallow during the navigation, revealing the small lava outcrop (Figs. 2 and 3b). Fresh and massive (nonvesicular) basalts were collected from this outcrop (samples 6K#1521R04 and R05; Fig. 3b). The samples from the 6K#1522 dive at a seamount exhibited highly irregular shapes, and massive lava flows, pillows, and lava breccia were observed (Fig. 3c). All the samples were fresh vesicular basalts (6K#1522R01, R02, R05, R12, R13, R16, and R17; Fig. 3c).

The fresh, massive, and nonvesicular basalts were obtained by 6K#1521 dive (R04 and R05) and comprised euhedral olivine microphenocrysts (150–400 μm in size), two types of ferrotitanium oxide (50–150 μm in size), and crystallite (Fig. 2b). Secondary phases were not observed as well. They were covered with 5.6–5.9 mm-thick ferromanganese crust and ~1.0 mm-thick palagonite rinds (Fig. 3b), but R05 did not have palagonite rinds. The photomicrograph of R04 is shown in Fig. 3b.

The seven fresh basalts collected during the 6K#1522 dive (6K#1522R01, R02, R05, R12, R13, R16, and R17), exhibited high vesicularity (20–40 vol.%) with 2.9–6.0 mm-thick palagonite rinds covered with 2.7–5.9 mm-thick ferromanganese crusts (Fig. 3c). Euhedral–subhedral olivine microphenocrysts (glomeroporphyritic, 30–200 μm in size), radial–needle-shaped clinopyroxene, iddingsite (<200 μm in size), spinel, and glass with minor xenocrystic olivines were observed (Fig. 3c). The photomicrograph of R01 is shown in Fig. 3c.

### 3.3 YK19-05S cruise and 6K#1542 and #1544 dives

A petit-spot knoll and related lava flows were surveyed by the 6K#1542 and #1544 dives, respectively, during the YK19-05S cruise (Fig. 2). During the 6K#1542 dive, geological survey and rock sampling were conducted from two points on the eastern slope of the knoll (Figs. 2 and 3d). Here, the 6K#1542R03 and R05 basalts were collected from the lava-breccia field covered with thin ferromanganese crust (Fig. 3d). Samples R06 and R09 were obtained from the lobate-surface lava between tubular lavas closer to the summit than R03 and R05 (Fig. 3d).

High-resolution (one-meter scale) bathymetric mapping was successfully conducted during the 6K#1544 dive, and this can contribute to future oceanographic investigations using the Human Occupied Vehicle (Kaneko et al., 2022). During this acoustic survey, several mounds, 10–20 m in height and a few hundred meters in diameter, were recognized (Fig. 3d). We observed these mounds and collected samples from outcrops during the second half of the dive. Pillow lavas, tumuli, and lava breccias were observed, and basaltic samples (6K#1544R04, R05, and R06) were collected (Fig. 3d).
Four vesicular basalts (10–30 vol.% vesicularity; 6K#1542R03, R05, R06, and R09) were covered with 4.3–4.4 mm-thick ferromanganese crust. The outer palagonitic rinds were 3.2–3.4 mm-thick (Fig. 3d). A few to 300-μm-sized euhedral–subhedral olivine microlites and microphenocrysts were glomeroporphyritic (Fig. 3d). The groundmass was dominated by needled dendritic clinopyroxenes (~100 μm in size). The others were olivine, spinel, glass, and xenocrystic olivine megacrysts. The photomicrograph of R06 is shown in Fig. 3d.

The basaltic samples from the 6K#1544 dive (6K#1544R04, R05, and R06) were covered with ferromanganese crust (5.0–5.8 mm-thick) over palagonitic rinds (3.4–4.4 mm-thick). All the samples exhibited high vesicularity in the range of 20–35 vol.% (Fig. 3d). They comprised olivine microphenocrysts (30–250 μm in size, euhedral–subhedral or columnar), clinopyroxene (<100 μm, needled, columnar, radial or dendritic shape), spinel, and glass without secondary phases (Fig. 3d).

The photomicrograph of R04 is shown in Fig. 3d. During macroscopic observations, practically all the basalts from the 6K#1542 and 6K#1544 dives exhibited similar vesicularity and freshness. Their geochemical features were also similar to each other and are described in Sect. 5-1 and 5-2.
Fig. 3. Bathymetric map with photos of the outcrop, the collected samples, and their photomicrographs with detailed bathymetry of the sampling points. (a) The 6K#1466, (b) 6K#1521, (c) 6K#1522, and (d) 6K#1542 and 6K#1544 dives using the Shinkai 6500 by JAMSTEC. The 1-m gridded bathymetry of the 6K#1544 dive is shown in (d), obtained using an MBES equipped with the Shinkai 6500 over a 100-m resolution map.
obtained using the surface ship, R/V Yokosuka (Kaneko et al., 2022). The photomicrographs of representative samples are shown for plane-polarized light (PPL), cross-polarized light (XPL), and backscatter electron (BSE). Ol, olivine; Cpx, clinopyroxene; Mgt, magnetite; Spl, spinel.

4. Analytical methods

4.1 Major and trace element analysis of volcanic glass, mineral, and whole-rock

Major element compositions of glasses and minerals were determined using an electron probe micro analyzer (EPMA), JXA-8900R at Atmosphere and Ocean Research Institute (AORI), the University of Tokyo was used for glass analysis and JXA-iHP200F at GSJ, AIST was used for mineral analysis. The analyses were performed using an accelerating voltage of 15 kV, a beam current of 12 nA, and a beam diameter of 10 µm for glass and 2 µm for mineral. A peak counting time of 20 s and a background counting time of 10 s were used, except for Ni, for which a peak counting time of 30 s and a background counting time of 15 s. For Na analysis of glass, the peak counting time was 5 s and the background counting time was 2 s. Natural and synthetic minerals were used as standards, and data were corrected using a ZAF online correction program (Akizawa et al., 2021). Major element composition of glass was determined by the mean value of 10 analytical points.

Trace element compositions of minerals were determined using a laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS; New Wave Research UP-213 and Agilent 7500s) at Kanazawa University. The Nd: YAG deep UV (ultraviolet) laser’s wavelength is 213 nm. The analyses were conducted with 100 µm spot size. A repetition frequency of 6 Hz and a laser energy density of 8 J cm⁻² were used. NIST612 glass (distributed by National Institute of Standards and Technology) was employed for calibration, using the preferred values of Pearce et al. (1997). Data reduction was undertaken with ²⁹Si as the initial standard, and SiO₂ concentrations were obtained by an electron microprobe analysis (Longerich et al., 1996). BCR-2G (distributed by the United States Geological Survey) was used as a secondary standard to assess the precision of each analytical session (Jochum and Nohl, 2008).

Whole-rock major and trace element compositions of rock samples were analyzed by Activation Laboratories Ltd., Canada, using Code 4Lithoresearch Lithogeochemistry and ultratrace5 Exploration Geochemistry Package. The former package uses lithium metaborate/tetraborate fusion with inductively coupled plasma optical emission spectrometry (FUS-ICP-OES) and inductively coupled plasma mass spectroscopy (FUS-ICP-MS) for the major and trace element analyses, respectively. The latter package uses inductively coupled plasma optical emission spectrometry (ICP-OES) and inductively coupled plasma mass spectroscopy (ICP-MS) for the major and trace element analyses,
4.2 Sr, Nd, and Pb isotope analysis

4.2.1 Acid leaching

Acid leaching was conducted for the selected basaltic samples on the basis of the procedure of Weis and Frey (1991, 1996) as follows: [1] About 0.3–0.4 or 0.6 g of rock powder is weighed into an acid-washed 15 mL Teflon vial (Savilex®). [2] 10 or 12 mL of 6N (N: normality) HCl were added, and then heated at 80°C for 20–30 min. [3] After heating, the suspension is ultra-sonicated in 60°C water for 20 min. [4] The supernatant is decanted. Steps [2] to [4] were repeated more than 4 times (up to 6 times) until the supernatant become clear or pale yellow to colorless. [5] TAMAPURE-AA Ultrapure water (Tama Chemicals; Co., Ltd.), which includes a lower Pb blank than milli-Q H2O, were added instead of 6N HCl, and the suspension is ultra-sonicated for 20 min. This step is conducted twice. [6] The leached rock powder is dried on a hot plate at 120°C. [7] After cooling, the powder is weighed.

4.2.2 Extraction of Pb, Sr, and Nd

The extraction of Pb, Sr, and Nd was performed following the procedures of Tanimizu and Ishikawa (2006) and Machida et al. (2009). First, from ~50 to ~100 mg of rock powder was weighted in a 7 mL Teflon vial (designated as “vial A”), and digested using mixed acid composed of HF and HBr. The separation was conducted by cation exchange resin (AG-1X8; Bio-Rad Laboratories Inc.) on the basis of procedures described in Tanimizu and ishikawa (2006). All fractions from the first and second supernatant loading (0.5 M HBr) to the elution of other elements (mixed acid composed of 0.25 M HBr and 0.5 M HNO3) were collected in another 7 mL Teflon vial (designated as “vial B”) for Sr and Nd separation. Finally, Pb was extracted by 1 mL of 1M HNO3 in another 7 mL Teflon vial (designated as “vial C”). The procedural blanks for Pb totaled less than 23 pg.

The Sr and Nd-bearing solution in the vial B was transferred into the vial A containing residues of digested samples. 2 mL of HClO4 and 2 mL HNO3 was further added to the vial A, and the residue was dissolved at 110 °C. Both Sr and Nd were separated by column with a cation exchange resin (AG50W-8X; Bio-Rad Laboratories Inc.) and a Ln resin (Eichrom Tech-nologies Inc.) on the basis of procedures described in Machida et al. (2009). The separated Sr and Nd were further purified by column separation with a cation exchange resin. The total procedural blanks for Sr and Nd were less than 100 pg.

4.2.3 Analytical procedure
Pb isotopic ratios were obtained using the multi-collector ICP-MS (MC-ICP-MS; Neptune plus, Thermo Fisher Scientific), with nine Faraday collectors, at Chiba Institute of Technology (CIT), Japan. The NIST SRM-981 Pb standard was also analyzed and yielded the average values of $^{206}\text{Pb}/^{204}\text{Pb} = 16.9303 \pm 0.0005$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4828 \pm 0.0006$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.6710 \pm 0.0016$. These correspond to previous values determined using MC-ICP-MS with Tl normalization, but they were slightly lower than values determined by TIMS in Tanimizu and Ishikawa (2006) from the $^{207}\text{Pb}/^{204}\text{Pb}$ double-spike. Reproducibility was monitored by an analyses of the JB-2 GSJ standard, and the obtained values were $^{206}\text{Pb}/^{204}\text{Pb} = 18.3326 \pm 0.0005$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.5453 \pm 0.0006$, and $^{208}\text{Pb}/^{204}\text{Pb} = 38.2240 \pm 0.0017$.

Sr and Nd isotopic analyses for powdered rocks and glasses were conducted using the thermal ionization mass spectrometry (TIMS; Triton XT, Thermo Fisher Scientific) with nine Faraday collectors, at CIT. 1.5 μL of 2.5M HCl and 0.5M HNO$_3$ was used for loading of separated Sr and Nd of sample on the single and double Re-filament, respectively. The measured isotopic ratios were corrected for instrumental fractionation by adopting the $^{86}\text{Sr}/^{85}\text{Sr}$ value to be 0.1194 and that of $^{146}\text{Nd}/^{144}\text{Nd}$ to be 0.7219. The average value for the NIST SRM-987 Sr standard was 0.710239 ±0.000005 (2σ, n =2), and that for the GSJ JNd-1 Nd standard was 0.512103 ±0.000005 (2σ, n =2). They agree well with values from the literature for the NIST SRM-987 $^{87}\text{Sr}/^{86}\text{Sr} = 0.710252–0.710256$; Weis et al., 2006) and JNd-1 ($^{143}\text{Nd}/^{144}\text{Nd} = 0.512101$; Wakaki et al., 2007). Consequently, we did not correct the values of the unknowns for offsets between the measurements and the values for the Sr and Nd standards.

### 4.3 $^{40}\text{Ar}/^{39}\text{Ar}$ dating

Samples for $^{40}\text{Ar}/^{39}\text{Ar}$ dating were prepared by separating crystalline groundmass after crushing them to sizes between 100 and 500 μm. The separated groundmass samples were leached by HNO$_3$ (1 mol/L) for one hour to remove clays and altered materials. All samples were wrapped in aluminum foil along with JG-1 biotite (Iwata, 1998), K$_2$SO$_4$, and CaF$_2$ flux monitors. Any amorphous (e.g., quenched glass) was removed because $^{39}\text{Ar}$ may move from one phase to another in a process known as “recoil.” This can create a disturbed age spectrum when $^{39}\text{Kr}$ is produced from $^{39}\text{Ar}$ in amorphous material through interaction with fast neutrons during irradiation of the sample. Samples were irradiated for 6.6 days in the Kyoto University Research Reactor (KUR), Kyoto University. Argon extraction and isotopic analyses were undertaken at the Graduate School of Arts and Sciences, the University of Tokyo. The sample gases were extracted by incremental heating of 10 or 11 steps between 600°C and 1500°C. The analytical methods used are the same as those used by Ebisawa et al. (2004) and Kobayashi et al. (2021).
5 Results

To describe the geochemical and chronological results, each sample group was denoted by its dive number, e.g., the sample group obtained from the 6K#1521 dive was labeled “1521 samples”. The basalts from the 6K#1466 dive were divided into two groups for R3 (collected from the seafloor south of the knoll) and R6–R7 (sampled on the knoll) based on their geographical, petrological, and compositional differences. The mineral compositions of each petit-spot basalt are shown in Fig. S1 and Table S1, S2 and S3.

5.1 Major and trace element compositions

The major and trace element compositions for the whole rock and glass of the petit-spot basalts are listed in Table 2 and 3, respectively. The basalt compositions for a petit-spot knoll were reported by Hirano et al. (2019) (expressed as “1203, 1206” in each figure). The data are discussed along with the reported NW Pacific petit-spots (Hirano and Machida, 2022). Using a total alkali vs. silica (TAS) diagram, virtually all the samples were classified as alkalic rocks, but the 1542 and 1544 basalts were plotted near the boundary between alkalic and non-alkalic (Fig. 4a). Two petit-spot basalts (1466R7-001 and R7-003) from the petit-spot knoll were notably silica-undersaturated (i.e., SiO$_2$ = 39.3–39.4 wt%) and classified as foidite (Mikuni et al., 2022). All the western Pacific petit-spot basalts, except for the 6K#1466R7 basalts, were sodic (K$_2$O/Na$_2$O = 0.24–0.58) and were notably discriminated to the potassic NW Pacific petit-spots (Fig. 4b).

Selected major element oxides and trace element ratios vs. MgO plots for the petit-spot basalts are shown in Figs. 5 and 6, respectively. The MgO concentrations of the 1466R3 and 1521 samples each exhibiting similar petrographic features (i.e., nonvesicular, and glassy) were characterized by values (4.0–4.4 wt%) lower than those of other vesicular samples (6.6–9.3 wt%). The K$_2$O, Na$_2$O, Al$_2$O$_3$, and SiO$_2$ contents negatively correlated with MgO (Figs. 5a–d). The CaO, FeO$_T$, and CaO/Al$_2$O$_3$ abundances exhibited positive correlations with MgO (Figs. 5e–g). The TiO$_2$ concentrations exhibited no correlations with MgO (Fig. 5h), as well as the selected trace element ratios (Figs. 6a–g) except for the Sm/Hf ratio with positive correlations (Fig. 6h). The Sm/Hf ratio also negatively correlated with SiO$_2$ (Fig. S2). The study samples exhibited whole-rock loss on ignition (LOI) in the range of 0.67–1.72 wt%, excluding two relatively altered samples, 6K#1466R7-001 (LOI = 2.68 wt%) and R7-003 (LOI = 6.29 wt%).

The PM-normalized (Sun and McDonough, 1989) trace element patterns for the petit-spot basalts, including those reported by a previous study (Hirano et al., 2019), were shown for each dive compared to the representative ocean island basalt (OIB) in Figs. 7a–f. The petit-spot basalts generally
showed high light rare earth element (LREE)/heavy REE (HREE) ratios. Negative Zr, Hf, Ti, and Y anomalies were commonly observed in these western Pacific petit-spots as well as those of the NW Pacific petit-spots (Fig. 7g). The 1466 basalts collected on the seafloor south of the knoll (6K#1466R3-001 and R3-004) were compositionally different from those obtained on the knoll (6K#1466R7-001, R7-003). The basalts from the 6K#1542 and 1544 dives, collected from nearby locations, had the same compositions in major and trace element ratios in both whole rock and glass, respectively (Figs. 4, 5, 6, 7e, and f). These samples in the Ba/Nb and Sm/Hf diagrams were plotted in the range of “Group 3” in the discrimination of the NW Pacific petit-spot basalts (Machida et al., 2015), indicating their negative Zr and Hf anomalies without notable U, Th, Nb, and Ta anomalies in the PM-normalized trace element patterns (Fig. 7h). The Sm/Hf ratio of the differentiated 1466R3 samples was lower than that of other samples. A positive correlation between fluid mobile and immobile elements, Ba vs. Nb (Fig. 8a) and U vs. Th (Fig. 8b), respectively, was observed, excluding the Ba of the 1466R7 samples (Fig. 8a).
Table 3

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**Analysis:** by ActLab

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The data are plotted as the total 100 wt%. The triangles and circles show the whole-rock and quenched-glass compositions, respectively. The compositions of the NW Pacific petal spots are represented by gray triangles (Hirano and Machida, 2022). The data of the 6K#1203 and 1206 basalts are from Hirano et al. (2019), and those of the 6K#1466R7 basalts are from Mikuni et al. (2022). (b) K$_2$O vs. Na$_2$O diagram.

The maximum K$_2$O/Na$_2$O value of kimberlite is from PetDB database (https://search.earthchem.org/). The data of OIB and MORB are compiled from Stracke et al. (2022) as “Expert datasets” in GEOROC database (https://georoc.eu/georoc/new-start.asp).

Fig. 4. Relationships between the SiO$_2$ and alkali contents. (a) Total alkali vs. silica diagram using the platform of Le Bas et al. (1986). The dividing line of alkaline and sub-alkaline is from Irvine and Baragar (1971). The data are plotted as the total 100 wt%. The triangles and circles show the whole-rock and quenched-glass compositions, respectively. The compositions of the NW Pacific petal spots are represented by gray triangles (Hirano and Machida, 2022). The data of the 6K#1203 and 1206 basalts are from Hirano et al. (2019), and those of the 6K#1466R7 basalts are from Mikuni et al. (2022). (b) K$_2$O vs. Na$_2$O diagram.

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18
Fig. 5. Selected major-element oxides against MgO. The symbols and compiled data correspond to those in Fig. 3.
Fig. 6. Selected trace-element ratios against MgO. The symbols and compiled data correspond to those in Fig. 3.
Fig. 7. Primitive mantle (PM, Sun and McDonough, 1989)-normalized trace-element patterns (a)–(g) and element ratios (h). (g) The compositional range of the study samples and NW Pacific petit-spots (Hirano and Machida, 2022). (h) The Ba/Nb and Sm/Hf ratios of the petit-spot basalts to discriminate the three groups after Machida et al. (2015). The data of 6K#1203, 1206 basalts and 6K#1466R7 basalts are from Hirano et al. (2019) and Mikuni et al. (2022), respectively. The symbols and compiled data in the (h) correspond to those in Fig. 3.

Fig. 8. Alteration sensitive elements (Ba and U) vs. insensitive elements (Nb and Th). The symbols and compiled data correspond to those in Fig. 3.

5.2 Sr–Nd–Pb isotopic composition

The Sr, Nd, and Pb isotopic compositions of the leached, unleached whole rock, and fresh glasses
in this study (presented in Table 4) were in practically identical ranges of \(^{87}\text{Sr}/^{86}\text{Sr}\) (0.703412–0.704424), \(^{143}\text{Nd}/^{144}\text{Nd}\) (0.512694–0.512890), \(^{206}\text{Pb}/^{204}\text{Pb}\) (18.6582–18.7778), \(^{207}\text{Pb}/^{204}\text{Pb}\) (15.5086–15.5749), and \(^{208}\text{Pb}/^{204}\text{Pb}\) (38.6506–38.8041) despite their different locations (Figs. 9a–d, Table 4). The isotopic compositions of the quenched glass and whole rock were identical, indicating that the characteristics of the melting source could be obtained through the geochemistry of the young and fresh volcanic quenched glass. The leached and unleached materials of the same sample also had similar isotopic ratios, except for the 1466R7-003 basalt, which had a relatively high LOI (6.29 wt%) (Figs. 9a–d). The Sr–Nd–Pb isotopic three-dimensional (3D) plot is shown in Fig. 9e.

![Fig. 9. Sr–Nd–Pb isotopic variations of the petit-spot basalts. The mantle endmembers are derived from a study by Zindler and Hart (1986). The open triangles in (a)–(d) represent the acid-leached samples. Carbonatite data were compiled from GEOROC (https://georoc.eu/georoc/new-start.asp) with Bizimis et al. (2003). Carbonatite data with \(^{87}\text{Sr}/^{86}\text{Sr} > 0.706\) by GEOROC were eliminated. The northwestern (NW) Pacific petit-spots and petit-spots off the Tonga Trench are from Hirano and Machida (2022) and Reinhard et al.](image-url)
Table 4
Sr, Nd, and Pb isotopic compositions of western Pacific petit-spot basalts and measured standards.  

<table>
<thead>
<tr>
<th>Cruise</th>
<th>Sample name</th>
<th>Sample type</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>$^{143}$Nd/$^{144}$Nd</th>
<th>$^{206}$Pb/$^{204}$Pb</th>
<th>$^{207}$Pb/$^{204}$Pb</th>
<th>$^{208}$Pb/$^{204}$Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>YK16-01</td>
<td>6K#1466 R3-004</td>
<td>Glass</td>
<td>0.703568 (06)</td>
<td>0.512842 (05)</td>
<td>18.6582 (07)</td>
<td>15.5086 (06)</td>
<td>38.6506 (19)</td>
</tr>
<tr>
<td>YK16-01</td>
<td>6K#1466 R7-001</td>
<td>Whole rock leached</td>
<td>0.703790 (05)</td>
<td>0.512817 (07)</td>
<td>18.7054 (20)</td>
<td>15.3377 (20)</td>
<td>38.8041 (50)</td>
</tr>
<tr>
<td>YK16-01</td>
<td>6K#1466 R7-005</td>
<td>Whole rock leached</td>
<td>0.703989 (05)</td>
<td>0.512790 (06)</td>
<td>18.7788 (09)</td>
<td>16.5093 (08)</td>
<td>38.7991 (22)</td>
</tr>
<tr>
<td>YK16-01</td>
<td>6K#1468 R7-003</td>
<td>Whole rock leached</td>
<td>0.703933 (11)</td>
<td>0.512815 (05)</td>
<td>18.7107 (06)</td>
<td>15.5749 (06)</td>
<td>38.7618 (17)</td>
</tr>
<tr>
<td>YK18-08</td>
<td>6K#1521 R04</td>
<td>Glass</td>
<td>0.703605 (05)</td>
<td>0.512832 (04)</td>
<td>18.6954 (06)</td>
<td>15.5428 (06)</td>
<td>38.7005 (19)</td>
</tr>
<tr>
<td>YK18-08</td>
<td>6K#1522 R01</td>
<td>Whole rock leached</td>
<td>0.703544 (05)</td>
<td>0.512861 (06)</td>
<td>18.7778 (09)</td>
<td>15.5209 (08)</td>
<td>38.7991 (22)</td>
</tr>
<tr>
<td>YK18-08</td>
<td>6K#1522 R01</td>
<td>Whole rock leached</td>
<td>0.703590 (05)</td>
<td>0.512866 (06)</td>
<td>18.7705 (07)</td>
<td>15.5244 (07)</td>
<td>38.7905 (22)</td>
</tr>
<tr>
<td>YK18-08</td>
<td>6K#1522 R01</td>
<td>Whole rock leached</td>
<td>0.703656 (06)</td>
<td>0.512872 (04)</td>
<td>18.7773 (09)</td>
<td>15.5178 (07)</td>
<td>38.7904 (21)</td>
</tr>
<tr>
<td>YK19-05S</td>
<td>6K#1542 R03</td>
<td>Whole rock leached</td>
<td>0.703412 (07)</td>
<td>0.512890 (06)</td>
<td>18.7559 (10)</td>
<td>15.5244 (11)</td>
<td>38.7574 (36)</td>
</tr>
<tr>
<td>YK19-05S</td>
<td>6K#1542 R05</td>
<td>Glass</td>
<td>0.703517 (06)</td>
<td>0.512847 (04)</td>
<td>18.7663 (08)</td>
<td>16.5519 (07)</td>
<td>38.7893 (19)</td>
</tr>
<tr>
<td>YK19-05S</td>
<td>6K#1544 R04</td>
<td>Whole rock leached</td>
<td>0.703480 (04)</td>
<td>0.512883 (05)</td>
<td>18.7413 (14)</td>
<td>15.5262 (14)</td>
<td>38.745 (41)</td>
</tr>
<tr>
<td>YK19-05S</td>
<td>6K#1544 R04</td>
<td>Whole rock leached</td>
<td>0.703568 (05)</td>
<td>0.512863 (04)</td>
<td>18.7400 (08)</td>
<td>15.5253 (09)</td>
<td>38.7347 (22)</td>
</tr>
<tr>
<td>YK10-05</td>
<td>6K#1206 R04</td>
<td>Glass</td>
<td>0.703492 (05)</td>
<td>0.512890 (04)</td>
<td>18.7074 (06)</td>
<td>15.5109 (07)</td>
<td>38.6970 (19)</td>
</tr>
<tr>
<td>YK10-05</td>
<td>6K#1206 R04 duplicate</td>
<td>Glass</td>
<td>0.703707 (06)</td>
<td>0.512872 (04)</td>
<td>18.7071 (06)</td>
<td>15.5119 (17)</td>
<td>38.6950 (18)</td>
</tr>
</tbody>
</table>

Errors shown in parentheses represent 2σ and apply to the last two digits.

5.3 Age determination and estimation

The $^{40}$Ar/$^{39}$Ar ages were determined for two samples (1466R6-001 and 1522R01) (Fig. 10a, Table S4). The secondary material (e.g., alteration products) plausibly causes the recoil loss and redistribution of Ar during irradiation of samples, particularly fine-grained groundmass separates of submarine basalt (Koppers et al., 2000). This effect is negligible for $^{40}$Ar/$^{39}$Ar dating samples in this study because the total K/Ca ratios estimated using the irradiated $^{39}$Ar/$^{37}$Ar ratio (0.089 for 6K#1466R6, 0.080 for 6K#1522R01; Table S4) are mostly correspond to the bulk K/Ca ratios calculated using the major element compositions of Table 2 (0.088 for 6K#1466R7-001, 0.076 for 6K#1522R01). This is supported by the rock descriptions recognized no secondary materials of...
crystalline $^{40}$Ar/$^{39}$Ar specimens. Sample 1466R6-001 had a plateau age of 3.03 ± 0.18 Ma in seven fractions comprising 94.1% released $^{39}$Ar. However, the plateau age was recognized as apparently old, owing to excess $^{40}$Ar, as indicated by the initial $^{40}$Ar/$^{36}$Ar ratio of 325 ± 15, which exceeded the atmospheric ratio (296.0; Nier, 1950) in the inverse isochron. The inverse isochron age of 2.56 ± 0.34 Ma showed the best age estimate for the 1466R6-001 basalt (Fig. 10a). The 1522R01 sample released almost no radiogenic daughter nuclide of $^{40}$Ar in the K–Ar age system (Fig. 10a).

The ranges of eruption age were estimated for all the samples using the average thickness ($n = 20$) of ferromanganese crust and palagonite rind (hydrated quenched glass) with their deposition/formation rates on the seafloor (ferromanganese crust, 1–10 mm/Myr; Hein et al., 1999; palagonite, 0.03–0.3 mm/Myr; Moore et al., 1985) (Fig. 10b). Using this approach, the western Pacific petit-spots were expected to have erupted later than ca. 9 Ma. The ranges of eruption age estimated from palagonite rind did not overlap with those from ferromanganese crust showing older durations, although they had general correlations (Fig. 10b). The $^{40}$Ar/$^{39}$Ar ages of two samples and the U-Pb age of zircon in the 1203 and 1206 peperites (Hirano et al., 2019) were overlaid within these ranges.

Fig. 10. Geochronological data. (a) The $^{40}$Ar/$^{39}$Ar ages of the 6K#1466R6-001 and 6K#1522R01 basalts. The errors show a 2-sigma confidence level. (b) Estimated relative ages using the thickness of ferromanganese crust (green bands) and palagonite (hydrated quenched-glass rind; red bands) covered with petit-spot basalts. These values were estimated using the average for each sample ($n = 20$). The U-Pb age of zircon in the 6K#1203 and 1206 peperites are from Hirano et al. (2019).

6 Discussion

6.1 Eruptive setting of western Pacific petit-spots

Here, two crystalline petit-spot basalts were subjected to $^{40}$Ar/$^{39}$Ar dating. A previously reported petit-spot knoll in this region (examined during the 6K#1203 and 1206 dives) aged “younger than 3 Ma” was investigated using the U–Pb dating of eight zircons in peperites (Fig. 10b) (Hirano et al.,
2019). The results showed that the silica-undersaturated vesicular basalt of 6K#1466R6-001, as a host of ultramafic xenoliths (Mikuni et al., 2022), exhibited a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 2.56 ± 0.34 Ma (Fig. 10).

Oppositely, the fresh vesicular basalt of 6K#1522R01, which erupted at the foot of the 100-Ma Takuyo-Daigo seamount (Fig. 2) (Nozaki et al., 2016), did not exhibit radiogenic $^{40}\text{Ar}$ highlighting that this sample is quite young (approximately 0 Ma) (Fig. 10). The ranges of eruption ages were estimated using the average thickness of ferromanganese crust and palagonite rind (seawater-hydrated quenched glass) with their deposition/formation rates on the seafloor. The $^{40}\text{Ar}/^{39}\text{Ar}$ and zircon U–Pb ages were within these ranges (Fig. 10). Here, the petit-spot volcanic field is surrounded by Cretaceous seamounts (Koppers et al., 2003) and irregular Paleogene volcanoes (Aftabuzzaman et al., 2021; Hirano et al., 2021). However, no zero-aged hotspots were observed in this region, and the P-wave tomographic image of the surface to the core–mantle boundary of the study area did not exhibit a plume-like low-velocity zone (Fig. 1c; Lu et al., 2019). Furthermore, the MORB-like to more depleted noble-gas isotopic compositions of the petit-spot knoll (6K#1203, 1206) suggested its upper mantle origin (Yamamoto et al., 2018). Along with the outer-rise bulge in front of the Mariana Trench detected through a positive gravitational anomaly (Hirano et al., 2019), these data suggest that the western Pacific petit-spot volcanoes could have erupted at ~0–3 Ma owing to the flexure of the subducting Pacific Plate into the Mariana and Ogasawara Trenches.

The petit-spot basalts from the 6K#1542 and 1544 dives could have originated from the same eruptive source based on their similar petrographic and geochemical features despite a distance of approximately 6.8 km between both (Figs. 3d, 4, 5, 6, 7, 8, and 9). Contrarily, in terms of their petrography and geochemistry, the basalts from the 6K#1466 dive are discriminated between the samples from the lava flows on the abyssal plain (R3-001 and R3-004) and the samples from the knoll site (R6-001, R7-001, and R7-003). The R3 basalts were collected at a lava outcrop 600 m south of the knoll, and the R6 and R7 samples were collected on the western slope of the knoll (Fig. 3a). The 6K#1466R3 series are glassy with a high SiO$_2$ content (50.6–51.6 wt%), including minor plagioclase and less vesicles (Figs. 3a and 4a). However, the 6K#1466R6 and R7 series exhibited silica-undersaturated compositions (SiO$_2$ = 39.3–39.4 wt%) and high vesicularities (20–40 vol.%) (Figs. 3b and 4a). Combining these observations with the differences in MgO contents and trace element compositions, the R3 and R6–R7 basalts are implied to have different parental magmas (Figs. 6 and 7b). Generally, vesicular samples (6K#1203, 1206, 1466R7, 1522, 1542, and 1544 basalts) are relatively primary (i.e., MgO > 6.63 wt%), whereas nonvesicular samples (6K#1466R3 and 1521 basalts) are evolved (i.e., MgO < 4.43 wt%). This correlates with the compositions of olivine microphenocrysts in the low forsterite content (Fo# = 100 × Mg/([Mg+Fe$^{2+}$]$_{cation}$) of olivine in evolved basalts and the high Fo# of olivine in the relatively primary basalts (Figs. S1a–c).

The CI chondrite-normalized REE ratios of these samples are within those of OIBs, and the REE patterns exhibit HREE-depleted patterns (Fig. S3). However, among the western Pacific petit-
spots, the REE and trace element ratios differ for each volcano (i.e., parental magmas) (Figs. 6 and S3). Given the lack of correlation between MgO and the trace element ratios, each volcano could have originated from isolated sources (i.e., melt ponds) with different chemical compositions and degrees of melting (Fig. 6). Oppositely, the radiogenic Sr, Nd, and Pb isotopic ratios of the samples are nearly identical, and the components in the source are probably equivalent (Fig. 9).

Summarily, (1) the western Pacific petit-spot volcanoes erupted at ~0–3 Ma owing to the plate flexure related to the subduction of the Pacific Plate into the Mariana Trench (Figs. 1 and 2). (2) The 6K#1542 and 1544 samples originated during the same magmatic event (Fig. 3d). However, the basalts from the 6K#1466 dive are subdivided into two parental magmas (R3 and R6–R7 basalts) (Fig. 3a). (3) Each volcano originated from isolated source and/or ascending processes based on the independent trace element ratios. The geochemical components involved in the source, however, were similar among the western Pacific petit-spot volcanoes because of the nearly identical Sr, Nd, and Pb isotopic compositions (Figs. 6 and 9). A variation in the trace element compositions among the volcanoes is plausibly due to the degree of contribution of carbonatite flux and/or the recycled crustal component to the source, as discussed below.

6.2 Petit-spot magma composition and its evaluation

Post-eruption seawater alteration might have affected the chemical composition of oceanic basalts. Thus, various approaches, including petrographic observation, geochemical investigation, and acid leaching, have been employed to evaluate the primary features and the removal of this effect for isotopic analysis (Hanano et al., 2009; Melson et al., 1968; Miyashiro et al., 1971; Nobre Silva et al., 2009; Resing and Sansone, 1999; Staudigel and Hart, 1983; Zakharov et al., 2021). The study samples exhibit whole-rock LOI <1.72 wt%, excluding two relatively altered samples, 6K#1466R7-001 (LOI = 2.68 wt%) and R7-003 (LOI = 6.29 wt%). Pristine quenched glasses are preserved in most of the samples, excluding three exceptional samples (the 6K#1466R6-001, R7-001, and R7-003 basalts). Positive correlations are observed between the alteration-insensitive (e.g., Nb, Th) and -sensitive (e.g., Ba, U) incompatible elements. This indicates that the effect of seawater alteration was not extensive, excluding the 6K#1466R7-001 and R7-003 basalts (Fig. 8). Although each sample was derived from different volcanic edifices, the positive correlation of all the study samples is due to the chemical similarity of the source compositions for certain elements (i.e., Ba/Nb and U/Th ratios are nearly constant among the samples), as well as the Sr, Nd, and Pb isotopic compositions (Fig. 9). These observations showed that most of the petit-spot basalts were unaffected by seawater alteration with a few exceptions (i.e., 1466R7-001 and R7-003 basalts).

The variable MgO (4–9 wt%), Ni (<263 ppm), and Cr (<350 ppm) contents in the samples are lower than the expected values of primary mantle-derived melt (MgO >10 wt%, Ni >400 ppm, Cr
Similarly, the Mg# (100 × Mg/[Fe$^{2+}$ + Mg]) values are differentiated in the range of 41–57 (Table 2) against the primary basaltic melt, which is equilibrated with the upper mantle (Mg# = 66–75; Irving and Green, 1976). No phenocrysts were observed (only microphenocryst), despite such differentiated compositions as well as most of the NW Pacific petit-spot basalts. This suggests that the western Pacific petit-spots experienced crystal fractionation in the lithosphere as well as the case of NW Pacific petit-spot (Machida et al., 2017; Valentine and Hirano, 2010; Hirano, 2011; Yamamoto et al., 2014). Therefore, the calculation of primary composition of the petit-spot basalts using the mineral modal composition on the thin section could not be performed. However, the trends of the major elements of the samples imply the crystal fractionation of the same phases. The negative trends of the Al$_2$O$_3$ content and the positive trends of the CaO and CaO/Al$_2$O$_3$ content with a decrease in MgO indicate the occurrence of olivine, spinel, and clinopyroxene fractionation (Figs. 5c, e, and g). The absence of visible correlations of the K$_2$O, Na$_2$O, SiO$_2$, and TiO$_2$ contents against MgO suggests that the fractionation of plagioclase and the Fe–Ti oxides was insignificant. The Fe–Ti oxides as minor phases in the groundmasses and plagioclases were only observed in the most differentiated 1466R3-001 and R3-004 basalts (Figs. 3, 5a, b, d, and h). However, these major elemental trends should be interpreted as apparent trends because each petit-spot volcano originated from an isolated parental magma with different chemical composition or degree of partial melting as discussed above.

The trace element composition of alkali basalts can be used to determine the melting source rather than major elements (Hofmann, 2003; Machida et al., 2014, 2015). Trace element composition of magma, however, could be modified by crustal and/or mantle assimilation and fractionation of certain minerals. The relatively primitive basalts (6K#1203, 1206, 1466R6, R7, 1522, 1542, and 1544) included xenocrystic olivines and partly ultramafic xenoliths, indicating a rapid magma ascent (Hirano et al., 2019; Mikuni et al., 2022; Fig. S4). However, since the stagnation of ascending petit-spot magma could occur to create fertile peridotite and pyroxene-rich veins from the middle to lower depths of the lithosphere (Mikuni et al., 2022; Pilet et al., 2016), the chemical composition of the petit-spot magma could be modified because of assimilation with the ambient lithospheric peridotite. According to Hirano and Machida (2022), ascending silica-undersaturated melt would mainly consume orthopyroxene (±spinel) and become a more silicic composition with Zr and Hf depletion. This is because of the relatively higher Zr–Hf partition of orthopyroxene than those of other trace elements (Pilet et al., 2008; Shaw, 1999; Tamura et al., 2019). The orthopyroxenes of fertile pyroxenites and lherzolite xenoliths metasomatized by petit-spot melts exhibit Zr and Hf enrichment (Mikuni et al., 2022; Fig. S5). If this silica-enrichment (i.e., melt–rock interaction) was significant, a positive correlation between SiO$_2$ and Sm/Hf is expected as a mantle assimilation trend. However, the samples exhibit a negative correlation, similar to those of the NW Pacific petit-spots (Hirano and Machida, 2022) (Fig. S2). Considering the relationship between the Sm and Hf partition coefficients of
clinopyroxene (i.e., $D^{\text{Hf}} < D^{\text{Sm}}$; McKenzie and O’Nions, 1991; Kelemen et al., 2003), we suggest that
the negative correlation between the Sm/Hf and SiO$_2$ of the petit-spot basalts probably reflects the
crystal fractionation of clinopyroxene rather than mantle assimilation. The Ba/Nb ratios of the samples
are nearly constant and do not correlate with the MgO and SiO$_2$ contents (Figs. 6g and S2g). The lack
of correlation between the other trace element ratios, excluding Sm/Hf and Ba/Nb (i.e., La/Y, La/Lu,
Sm/Yb, La/Sm, Nb/Ta, Zr/Hf), and the MgO concentration imply that crystal fractionation may not
have been involved with those of the incipient melt (Fig. 6). However, it is difficult to independently
follow the evolution of the trace element composition for each volcano since each volcano originated
from isolated sources. Thus, considering the observations above, the fresh and zero-aged 6K#1522
basalts (the highest Sm/Hf ratios and lowest SiO$_2$ contents among the fresh samples and higher MgO
contents) were selected for further analysis with geochemical modeling. Considering that the 6K#1522
samples had MgO in the range of 6.63–7.36 wt%, olivine was expectedly the dominant phase of crystal
fractionation (Asimow and Langmuir, 2003; Helz and Thornber, 1987; Herzberg, 2006). When the
olivine maximum fractionation model (Takahashi et al., 1986; Tatsumi et al., 1983) was applied to test
two samples, the calculated primary trace element contents did not significantly differ from those of
the analytical compositions (Table S5 and Fig. S6). Thus, the 6K#1522 basalts were assumed to be the
most primary petit-spot basalts and were used to evaluate the geochemical modeling results.

6.3 Melting source of western Pacific petit-spots

The depletions of specific elements (e.g., Ta, Zr, Hf, and Ti) of petit-spot basalts potentially
demonstrate the involvement of carbonatitic materials in conjunction with a large amount of CO$_2$ and
lower Mg isotopic ratio than that of the normal mantle (Bizimis et al., 2003; Dasgupta et al., 2009;
Hirano and Machida, 2022; Hoernle et al., 2002; Liu et al., 2020; Okumura and Hirano, 2013). Other
oceanic lavas originating from the asthenosphere (e.g., Hawaiian rejuvenated lavas and North Arch
volcanoes) exhibited characteristic trace element signatures (i.e., Zr and Hf depletion) similar to those
of petit-spot lavas. This implies that their melting sources were involved with carbonatitic materials
with or without plume-derived components (Fig. S7; Borisova and Tilhac, 2021; Clague and Frey,
1982; Clague et al., 1990; Dixon et al., 2008; Yang et al., 2003). In addition, the involvement of
recycled crustal components was inferred from the geochemical features of the petit-spot basalts, and
the upper mantle was revealed to be heterogeneous (Liu et al., 2020; Machida et al., 2009, 2015). Such
scenario of the source on petit-spot magma is consistent with the previously suggested petrogenesis
of alkaline rocks explained by the addition of CO$_2$-rich components and/or recycled crustal materials
with or without sediment to the mantle (e.g., Dasgupta et al. 2007; Hofmann, 1997). Conversely, the
melting of an amphibole-rich metasomatic vein explains the major and trace element composition of
alkali basalts (Pilet et al., 2008; Pilet, 2015). However, the experimentally produced melts exhibit Pb
depletion and a positive Nb-Ti anomaly in the PM-normalized trace element patterns (Fig. S8) inconsistent with the petit-spot basalts (Fig. 7). In addition, Juriček and Keppler (2023) demonstrated that amphibole dehydration is not the cause for the oceanic LAB by high-pressure experiment on the realistic condition. The fertile pyroxenitic xenoliths and pyroxene xenocrysts occurring in the 1466R6 and R7 basalts, which originated from the metasomatic vein related to prior petit-spot magmatism, had neither amphiboles nor other hydrous minerals (Mikuni et al., 2022).

To discuss the involvement of carbonatitic and crustal components in petit-spot melts, a partial melting model of the heterogeneous mantle is provided. The involvement of carbonatitic fluids and recycled materials in the genesis of petit-spot melts has been suggested, and the open-system model with carbonatite influx from the outer system was employed using “OSM-4” of Ozawa (2001), referring the parameters of Borisova and Tilhac (2021). This model is based on the mass conservation equations of one-dimensional steady-state melting. In this present study, the model uses a critical melt fraction ($\alpha_c$; mass fraction of melt when melt separation begins = melt connectivity threshold) at 0.005 or 0.01. The system is opened to fluxing at a constant melt-separation rate ($\gamma$) when the system reaches the $\alpha_c$. The final trapped melt fraction ($\alpha_f$; mass fraction of melt trapped in the residue) was fixed at ~0 (it was calculated as 10^{-6} owing to mass balance). We calculated the trace element composition of partial melts at various degree of melting ($F$), a few rates of influx ($\beta$) and melt separation ($\gamma$). We assumed a primitive mantle (PM) source as a lherzolite with or without a normal (N)-MORB source as the recycled oceanic crust (Sun and McDonough, 1989), such as pyroxenite and eclogite. The recycled crust (N-MORB component) was mixed in the source as compositional heterogeneity calculated as “0.05N-MORB + 0.95PM” for the trace element concentration, and the considered mineral phases and their proportions were derived only from garnet lherzolite (i.e., olivine, orthopyroxene, clinopyroxene, and garnet). The mineral mode of garnet lherzolite (olivine 55%, orthopyroxene 20%, clinopyroxene 15%, and garnet 10%) and the melting reaction mode (olivine 8%, orthopyroxene −19%, clinopyroxene 81%, and garnet 30%) are based on studies by Johnson et al. (1990) and Walter (1998), respectively. The proportion of olivine and garnet was also changed to evaluate the effect of garnet modal ratio to the produced melt composition. In this situation, the clinopyroxene is consumed at an $F$ (degree of partial melting) of ~ 19%; therefore, the system was calculated up to 18% partial melting. The carbonatite melt, as an influx, in this model is “average carbonatite” from a study by Bizimis et al. (2003). The partition coefficient of trace elements is generally based on a study by McKenzie and O’Nions (1991, 1995) excluding Ti for clinopyroxene and garnet (Kelemen et al., 2003). The variables of $\beta$ (influx rate) and $\gamma$ (melt-separation rate) were changed during the modeling within the mass balance ($\gamma \leq \beta + 1$). The modeled melts were outputted as “total melt,” considering the instantaneous and accumulated melts. For the carbonatite composition, the value of “average carbonatite” of Bizimis et al. (2003) is applied because the chemical composition of carbonatite is largely diverse, and this value is recommended for geochemical modeling (Bizimis
et al., 2003). The parameters are listed in Table S6. As a result, partial melting of garnet lherzolite with 10% carbonatite influx to a given mass of source (i.e., garnet lherzolite) can roughly explain the trace element pattern of petit spot basalts (Figs. 11a–e), and the presence of 5% crustal component in the source is the most plausible model of petit-spot magma generation (Figs. 11b and d). In addition, slightly less garnet in the lherzolite source than the modal ratio of Johnson et al. (1990) fits the petit-spot better (Fig. 11b). In both cases, the presence of a crustal component in the source yields more plausible results (Figs. 11a–d). The higher carbonatite influx ($\beta = 1.0$) could not explain the trace element composition of the petit-spot basalts (Fig. 11f). The melt connectivity threshold ($\alpha_c$) of 0.01 is plausible because higher connectivity of melt (i.e., lower $\alpha_c$ value) leads to enrichment of LILEs and LREEs (Fig. 11g). The results also showed that the melt-separation ratio is insignificant to the trace element composition of the calculated melts (Figs. 11d and e). Thereafter, we concluded that the partial melting of ~5% crustal component-bearing garnet lherzolite with ~10% carbonatite flux to a given mass of the source plausibly explains the melting source of petit-spot volcanoes (Figs. 11b and d).

Assuming that the trace element composition of 6K#1203, 1206, 1542, and 1544 basalts are also primitive, they may be explained by a partial melting of garnet lherzolite with 5% crustal component and lower carbonatite influx rate ($\beta = 0.03$) (Fig. S9). Actually, the 6K#1203, 1206, 1542, and 1544 basalts exhibited the similar MgO contents and Mg# to those of 6K#1522 basalts (Fig. 4 and Table 2). These results provide quantitative evidence on the petrogenesis of petit-spots, i.e., the contribution of carbonatite melt and recycled oceanic crust.

Although the melting source contained small proportions of carbonatite melt and crustal components, these components could have contributed to the isotopic composition because of their abundant incompatible elements rather than the ambient mantle. The determination of the Sr, Nd, and Pb isotopic compositions revealed that they had geochemically identical prevalent mantle (PREMA)-like sources (Fig. 9). They do not belong to any mantle isotopic endmembers (i.e., depleted MORB mantle (DMM); EM-1, -2; and HIMU; Fig. 9) contrary to those of NW Pacific petit-spots toward the EM-1 isotopic composition (Machida et al., 2009; Liu et al., 2020). In the Pb isotopic space, the present samples do not correlate with those of the neighboring HIMU-like Cretaceous seamounts (Fig. 9a) (N-Wake, S-Wake seamounts; Konter et al., 2008; Koppers et al., 2003; Natland, 1976; Smith et al., 1989; Staudigel et al., 1991). For the melting source of the NW Pacific petit-spot basalts, the contributions of the eclogite/pyroxenite endmember as recycled oceanic crust and the carbonated endmember were suggested based on the major and trace elements and the Mg, Sr, Nd, and Pb isotopic compositions with the Mg diffusion modeling (Liu et al., 2020). The higher FeO/MnO ratios of the present melts (65.9–78.0), compared with those of partial melts originating from peridotite (50–60), are attributed to the presence of recycled pyroxenite (Herzberg, 2011). This could have contributed to the crustal components in the melting source. However, the western Pacific petit-spots in this study identically exhibited a PREMA-like isotopic signature without extreme endmember contributions as described...
above (Fig. 9). Such isotopic compositions with the world’s petit-spots can be possibly explained by
the diverse mixing proportion of HIMU and EM-1 components (Fig. 9e). The isotopic compositions
of the NW Pacific petit-spots (off the Japan Trench), Samoan petit-spots (off the Tonga Trench), petit-
spot dikes in Christmas Island (off the Java trench), and western Pacific petit-spots (off the Mariana
Trench in this study) are roughly along the HIMU–EM-1 mixing line (Fig. 9e). Furthermore, the
isotopic compositions of global carbonatites can be generally explained by the mixing of HIMU and
EM-1 (Bell and Tilton, 2002; Hoernle et al., 2002; Hulett et al., 2016). The contributions of the
carbonated material/carbonatite and crustal components to the melting source were suggested in terms
of the origin of HIMU and EM-1 (Collerson et al., 2010; Hanyu et al., 2011; Wang et al., 2018; Weiss
et al., 2016; Workman et al., 2004; Zindler and Hart, 1986). However, the EM-1 and HIMU
components could not be determined to be carbonated component and recycled crust, respectively,
owing to the various views on each tectonic setting for the mantle endmember. The variability of global
carbonatite isotopic compositions also makes it difficult to determine their representative isotope ratios
(Fig. 9). Although such issues make a quantitative isotopic mixing model challenging, the HIMU-EM-
1 like trend of the global petit-spot volcanoes may reflect the involvement of carbonatitic and recycled
crustal materials. Conclusively, the mass balance models on the trace elements and the isotopic
variations in the petit-spot volcanoes confirmed the contribution of carbonatite melt and the recycled
oceanic crust to the melting source of the western Pacific petit-spots (Fig. 12). Experimental studies
have revealed the various petrogenesis of carbonatite and carbonatitic alkali-rich magma under high
pressures (Dasgupta et al., 2006; Ghosh et al., 2009). The geochemistry of petit-spot basalts including
Mg isotopes suggested that the conceivable origin of carbonatite related to the petit-spot melt is
subducted “carbonated” pelite, pyroxenite/eclogite, or peridotite stored as diamond or metal carbide
in the reduced lower portion of the upper mantle (Liu et al., 2020; Rohrbach et al., 2007). Subducted
carbonated pelite, for example, would melt under high pressure (>8 GPa) through the oxidation at the
redox boundary where the iron-wüstite (IW) buffer changes to the quartz–fayalite–magnetite
(QFM) buffer (i.e., redox melting; Grassi and Schmidt, 2011). Chen et al. (2022) demonstrated that
the alkali-rich carbonatite melt could occur under a pressure higher than 6 GPa, particularly exhibiting
K-rich and Na-rich carbonatitites under 6–12 and >12 GPa, respectively. This pressure-dependent
alkalinity of the produced carbonatite melts might explain the variation between potassic NW Pacific
petit-spot lavas and present sodic petit-spot lavas (Fig. 4b). On the other hand, an experimental study
pointed out the existence of carbonate-rich layer in the LAB owing to the horizontally spread carbonate
from around the wedge mantle rather than upwelling from the deep mantle (Hammouda et al., 2020).
Several high pressure–temperature experiments and modeling revealed that the chemical composition
of intraplate magmas from upper mantle depends on their original depth; carbonatitic melt can be
generated beneath thick cratonic lithosphere (~250–200 km), kimberlitic melt would be produced at
>120 km in depth, and alkali basalt would occur at 100–60 km in depth by partial melting of “original”
CO₂ and H₂O-bearing mantle (Massuyeau et al., 2021). These depth-dependent compositional variation, that is, K-rich kimberlite to alkali basalt may also explain the geochemical gap between K-rich NW Pacific petit-spots and K-poor western Pacific petit-spots (Fig. 4b). Although the multiple origins of carbonatite are merely suggested and remain unclear, carbon-rich components play a key role in the partial melting of mantle at the LAB (Sifré et al., 2014), that is, the source of petit-spot magma.

Fig. 11. Geochemical modeling for the primitive mantle (PM)-normalized trace-element pattern. The calculated hypothetical melts are a production of carbonatite influx melting of garnet lherzolite with or without 5% crustal component. Detailed information of the parameters is described in Section 6-3 and Table S6. $F$ is the degree of melting (%). The trace-element composition of the western Pacific petit-spot basalts from the 6K#1522 dive is shown as black lines for comparison. The PM composition of lherzolite and the N-MORB composition of recycled crust were based on a study by Sun and McDonough (1989). The influx carbonatite is the “average carbonatite” of a study by Bizimis et al. (2003). The parameters used in the open-system melting models were as follows: $a_c$ is a critical melt fraction, $a_f$ is a final trapped melt fraction, $\beta$ is a melt influx rate, and $\gamma$ is a melt-separation rate. Model results are compared by varying each parameter, i.e., garnet modal ratio and presence of crustal material ($a$–$d$), melt-separation rate ($d$ and $e$), carbonatite influx rate ($d$ and $f$), and critical melt fraction ($d$ and $g$). Each figure is expressed based on the difference from the condition in (d).
Carbonatitic melt and recycled oceanic crust potentially induce partial melting of asthenospheric mantle beneath the western Pacific region. Carbonatitic melt might have originated from a carbon-rich component horizontally migrated from a subduction zone (Hammouda et al., 2020), or a redox melting of reduced carbon in the deep mantle (Chen et al., 2022; Grassi and Schmidt, 2011; Rohrbach et al., 2007). Petit-spot magma stagnated in the lithosphere with fractional crystallization and melt-rock interaction (Mikuni et al., 2022), and they have erupted at ~0–3 Ma.

7 Conclusion

The occurrence of petit-spot volcanism supports partial melting at the LAB, providing crucial implications for the nature of this geophysical discontinuity. Multiple petit-spot magmatisms on the western Pacific Plate occurred at ~0–3 Ma, originating from similar PREMA-like melting sources based on 40Ar/39Ar dating and the Sr, Nd, and Pb isotopic compositions. The mass balance-based open-system modeling for trace elements revealed that the western Pacific petit-spot magma was generated by the partial melting of a small amount (5%) of oceanic crust-bearing garnet lherzolite with 3–10% carbonatite influx to a given mass of the source. The Sr, Nd, and Pb isotopic compositions of this study samples, with those of the NW Pacific petit-spots, off the Tonga and Java Trenches, could be explained by mixing the EM-1-like and HIMU-like components, which contribute to subducted carbonated/crustal materials. The tectonic-induced magmatism, like a petit-spot, may have the same melting mechanism.

Authorship contributions

K. Mikuni and N. Hirano conceived the project and performed all experiments. S. Machida and Y. Kato contributed the Sr, Nd, and Pb isotopic analysis using TIMS and MC-ICP-MS. H. Sumino contributed the 40Ar/39Ar dating. N. Akizawa, A. Tamura, and T. Morishita helped and performed
EPMA and LA-ICP-MS analyses. S. Machida and N. Hirano conducted the research cruises to gain the rock samples. All authors interpreted the data and wrote the manuscript with comments and improvements.

Competing Interest

The authors declare that they have no conflict of interest.

Data availability

The data newly analyzed in this study and results of geochemical modeling are included in digital format in the online data repository of this paper (Tables 1, 2, 3 and 4, and Supplementary Tables S1 to S6) and the EarthChem online database (DOI will be obtained when it is accepted).

Acknowledgement

We would like to thank the captains, crews, and shipboard scientific parties of the R/V Yokosuka and the operating team of the submersible Shinkai 6500 for their great work during the YK16-01, YK18-08, and YK19-05S cruises. The Kyoto University Research Reactor Institute is gratefully acknowledged in their assistance of undertaking the radiometric dating. We would like to express our great appreciation to Prof. T. Tsujimori (ORCiD: 0000-0001-9202-7312) for his effort in management of the laboratory at Tohoku University. We also thank R. Fukushima (ORCiD: 0000-0003-2683-6757) for improving the wording in the manuscript. We are really grateful Y. Matamura, Y. Shimbo, and Y. Jindo for their help and discussion on scientific matters. The authors would like to thank Enago (www.enago.jp) for the English language review. This research was supported by the Cooperative Program (No. 106, 202) of Atmosphere and Ocean Research Institute, The University of Tokyo. The Japan Society for the Promotion of Science (Grant Numbers 17K05715, 18H03733, 20K04098) also supported this research.

References


Akizawa, N., Ozawa, K., Tamura, A., Michibayashi, K. and Arai, S.: Three-dimensional evolution of melting, heat and melt transfer in ascending mantle beneath a fast-spreading ridge segment


Bizimis, M., Salters, V.J.M., Garcia, M.O. and Norman, M.D.: The composition and distribution of


Dasgupta, R. and Hirschmann, M.M.: Melting in the Earth's deep upper mantle caused by carbon


Jochum, K.P. and Nohl, U.: Reference materials in geochemistry and environmental research and the


https://doi.org/10.1109/OCEANSChennai45887.2022.9775366, 2022.


Moore, J.G., Fornari, D.J. and Clague, D.A.: Basalts from the 1877 Submarine Eruption of Mauna Loa, Hawaii; New Data on the Variation of Palagonitization Rate with Temperature. United

Müller, R.D., Sdrolias, M., Gaina, C. and Roest, W.R.: Age, spreading rates, and spreading


Nier, A.: A redetermination of the relative abundances of the isotopes of carbon, nitrogen, oxygen,
argon, and potassium, Phys. Rev., 77, 789-793, https://doi.org/10.1103/PhysRev.77.789,
1950.

Nobre Silva, I.G., Weis, D., Barling, J. and Scoates, J.S.: Leaching systematics and matrix
elimination for the determination of high-precision Pb isotope compositions of ocean island
2009.

Novella, D., Keshav, S., Gudfinnsson, G.H. and Ghosh, S.: Melting phase relations of model
carbonated peridotite from 2 to 3 GPa in the system CaO-MgO-Al$_2$O$_3$-SiO$_2$-CO$_2$ and further
indication of possible unmixing between carbonatite and silicate liquids, J. Geophys. Res.,

compositions and resource potential of ferromanganese crust at Takuyo Daigo Seamount,

Okumura, S. and Hirano, N.: Carbon dioxide emission to earth's surface by deep-sea volcanism,

Orihashi, Y., Maeda, J., Tanaka, R., Zeniya, R. and Niida, K.: Sr and Nd isotopic data for the seven
GSJ rock reference samples; JA-1, JB-1a, JB-2, JB-3, JG-1a, JGb-1 and JR-1, Geochem. J.,

Ozawa, K.: Mass balance equations for open magmatic systems: Trace element behavior and its
application to open system melting in the upper mantle. J. Geophys. Res., 106, 13407–

S.P.: A compilation of new and published major and trace element data for NIST SRM 610
and NIST SRM 612 glass reference materials, Geostand. Newsl., 21, 115–144,


Pilet, S.: Generation of low-silica alkaline lavas: Petrological constrains, models, and thermal
implications, The Interdisciplinary Earth: A Volume in Honor of Don L. Anderson, Gillian
Baumgarther, P.O. and Müntener, O.: Pre-subduction metasomatic enrichment of the oceanic
lithosphere induced by plate flexure, Nat. Geosci., 9, 898–903,
https://doi.org/10.1038/NGEO2825, 2016.
Regelous, M., Weinzierl, C.G. and Haase, K.M.: Controls on melting at spreading ridges from
correlated abyssal peridotite – mid-ocean ridge basalt compositions, Earth Planet.
Rohrbach, A., Ballhaus, C., Golla-Schindler, U., Ulmer, P., Kamenetsky, V.S. and Kuzmin, D.V.:
Metal saturation in the upper mantle, Nature, 449, 456–458,
https://doi.org/10.1038/nature06183, 2007.
Rychert, C. A. and Shearer, P. M.: A global view of the lithosphere–asthenosphere boundary,
Sakamaki, T., Suzuki, A., Ohtani, E., Terasaki, H., urakawa, S., Katayama, Y., Funakoshi, K.-I.,
Wang, Y., Hernlund, J.H. and Ballmer, M.D.: Ponded melt at the boundary between the
lithosphere and asthenosphere, Nat. Geosci., 6, 1041–1044,
Shaw, C.S.J.: Dissolution of orthopyroxene in basanitic magma between 0.4 and 2 GPa: Further
implications for the origin of Si–rich alkaline glass inclusions in mantle xenoliths, Contrib.
conductivity during incipient melting in the oceanic low-velocity zone, Nature, 509, 81–85,
https://doi.org/10.1038/nature13245, 2014.
Cretaceous record of the South Pacific isotopic and thermal anomaly, J. Geophys. Res., 94,


xenoliths from northwestern Pacific lithosphere, Chem. Geol., 268, 313–323,

Yamamoto, J., Korenaga, J., Hirano, N. and Kagi, H.: Melt-rich lithosphere-asthenosphere boundary inferred from petit-spot volcanoes, Geology, 42, 967–970,


Yamamoto, J., Hirano, N. and Kurz, M.D.: Noble gas isotopic compositions of seamount lavas from the central Chile trench: Implications for petit-spot volcanism and the lithosphere asthenosphere boundary, Earth Planet. Sci. Lett., 552, 116611,


Yoshino, T., Matsuzaki, T., Yamashita, S. and Katsura T.: Hydrous olivine unable to account for conductivity anomaly at the top of the asthenosphere, Nature, 443, 973–976,
https://doi.org/10.1038/nature05223, 2006.


Zhang, W., Johnston, S. and Currie, C.A., Kimberlite magmatism induced by west-dipping subduction of the North American plate, Geology, 47, 395–398,

