# 1 Revision 3

2	Thermal state of the upper mantle and the origin of the Cambrian-Ordovician ophiolite
3	pulse: constraints from ultramafic dikes of the Hayachine-Miyamori Ophiolite
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#### ABSTRACT

Ophiolite pulses, which are periods of enhanced ophiolite generation and emplacement, 20 are thought to have a relevance to highly active superplumes (superplume model). However, the 21 Cambrian-Ordovician pulse has two critical geological features that cannot be explained by such 22 a superplume model: predominance of subduction-related ophiolites and scarcity of plume-23 related magma activities. We addressed this issue by estimating mechanism and condition of 24 magma generation, including mantle potential temperature (MPT), from a ~500 Ma subduction-25 26 related ophiolite, the Hayachine-Miyamori ophiolite. We developed a novel method to overcome 27 difficulties in global MPT estimation from an arc environment by using porphyritic ultramafic dikes showing flow differentiation, which have records of the chemical composition of the 28 29 primitive magma, including its water content, because of their high pressure ( $\sim 0.6$  GPa) intrusion 30 and rapid solidification. The solidus conditions for the primary magmas are estimated to be 31  $\sim$ 1450 °C,  $\sim$ 5.3 GPa. Geochemical data of the dikes show passive upwelling of a depleted mantle 32 source in the garnet stability field without a strong influence of slab-derived fluids. These results 33 combined with the extensive fluxed melting of the mantle wedge prior to the dike formation 34 indicate sudden changes of the melting environment, its mechanism, and the mantle source from extensive fluxed melting of the mantle wedge to decompressional melting of the sub-slab mantle 35 36 which has been most plausibly triggered by a slab breakoff. The estimated MPT of the sub-slab mantle is ~1350 °C, which is very close to that of the current upper mantle, and may reflect the 37 global value of the upper mantle at ~500 Ma if small-scale convection maintained the shallow 38 39 sub-slab mantle at a steady thermal state. We therefore conclude that the Cambrian-Ordovician ophiolite pulse is not attributable to high-temperature of the upper mantle. Frequent occurrence of 40 slab breakoff, which is suggested by our geochemical compilation of Cambrian-Ordovician 41

ophiolites, and subduction termination, which is probably related to the assembly of theGondwana supercontinent, may be responsible for the ophiolite pulse.

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### **INTRODUCTION**

The Earth has been releasing heat into the space, which drives internal and surface processes, such as magmatism, plate tectonics, and evolution of the surface environment (Davies, 1999). In this regard, the thermal history of the Earth is an essential part of its evolution. The major volume (~80 %) of the Earth is occupied by the mantle, which is transferring heat to the Earth's surface by thermal convection. It is, therefore, important to reveal the thermal state of the mantle and temporal changes in its thermal state in order to better understand the past and future evolution of the Earth.

A quantitative proxy of a thermal state of the mantle is the mantle potential temperature 53 (MPT: McKenzie and Bickle, 1988). It has been estimated from the chemical compositions of 54 magmas erupted on the Earth's surface at various tectonic settings, such as mid-ocean ridges and 55 intraplate settings (Putirka, 2005; Lee at al., 2009; Herzberg et al., 2010) as well as volcanic arcs 56 (Lee at al., 2009; Sakuyama et al., 2009, 2014). However, a secular change of the global thermal 57 58 state of the mantle has been estimated from MPTs at restricted tectonic settings (non-arc environments, such as spreading centers; Herzberg et al., 2010). This restriction of applicable 59 tectonic environments guarantees that the global thermal state of the Earth is sampled by 60 decompressional melting of either upper mantle that passively upwells along divergent margins, 61 or deep mantle that actively upwells from a deep thermal boundary layer. Arc environments were 62 avoided for this purpose because of: (1) involvement of water input from subducting slab into the 63

64	magma generation environment and magma degassing during transportation to the surface and
65	volcanic events which leads to technical difficulties in the estimation of MPTs, and (2) a thermal
66	structure and flow pattern of the mantle wedge strongly affected by subduction of a slab, which
67	could be in various thermal states depending on its age. These difficulties in application of the
68	conventional methods to estimate a secular change of the global MPTs of the mantle from
69	magma-related materials occurring in an arc environment must be overcome since there are
70	several periods in the Earth's history when material records of primitive magmas are dominantly
71	available from arc environments (Dilek and Furnes, 2011)
72	Ophiolites, which are defined in this paper as suites of temporally and spatially associated
73	ultramafic, mafic, and felsic rocks representing remnants of ancient oceanic crust and upper
74	mantle (Dilek and Furnes, 2011, 2014), were formed and exhumed at least since the
75	Paleoproterozoic to the very recent past (Stern, 2005). Their age distributions show several
76	confined periods with high rates of occurrence, which are called "ophiolite pulses" (Abbate et al.,
77	1985; Ishiwatari, 1994; Yakubchuk et al., 1994). These pulses occur in late Neoproterozoic,
78	Cambrian-Ordovician, and Jurassic-Cretaceous times (Ishiwatari, 1994; Dilek, 2003). They have
79	been explained by connection to episodic activities of superplumes generated at the core-mantle
80	boundary region in 300 million years cycles (Ishiwatari, 1994; Dilek, 2003; Vaughan and
81	Scarrow, 2003). We will refer to this hypothesis for the ophiolite pulse origin as the "superplume
82	model". It does not, however, explain two critical features of the Cambrian-Ordovician ophiolite
83	pulse. First, the Cambrian-Ordovician ophiolite pulse is dominated by subduction-related
84	ophiolites thought to have formed mostly in a back-arc environment, which is in contrast to the
85	minor occurrence of subduction-related ophiolites for the Jurassic-Cretaceous pulse (Dilek and
86	Furnes, 2011; Furnes et al., 2014). This peculiarity of the pulse in ophiolite types requires some

mechanisms to explain the high rate of subduction-related ophiolite formation during the 87 Cambrian-Ordovician time. Second, there is almost no report of large igneous provinces (LIPs), 88 which are believed to be closely related to mantle plume activities, during the Cambrian-89 Ordovician time. Ernst and Buchan (2003) compiled world LIPs since 4000Ma, which shows that 90 LIP generation was significantly suppressed during the Cambrian-Ordovician compared to that 91 during the Jurassic-Cretaceous. Because there are such fundamental inconsistencies between the 92 93 essential features of Cambrian-Ordovician geological records and the superplume model, the origin of the Cambrian-Ordovician ophiolite pulse should be reappraised. 94 In order to construct a model for the Cambrian-Ordovician ophiolite pulse and to seek a 95 general model of ophiolite pulses through the Earth's history by considering the secular change 96 97 of thermal state of the mantle, it is imperative to carefully evaluate dynamics involved in the 98 formation of each ophiolite by examining the mechanisms of magma generation in the mantle, its 99 MPT, and the controlling tectonics. It is, however, not easy to conduct this task for the Cambrian-100 Ordovician ophiolite pulse because of the aforementioned difficulties in the estimation of global 101 MPT from arc volcanic rocks. The estimated MPTs may represent the thermal state of either the 102 global upper mantle, deep thermal boundary layers, or even arc mantle wedges. In this study, we extend the current estimation method of global MPTs, whose application has been restricted to 103 104 non-arc environments, to one applicable to arc environments. The novel approach is to use intrusive rocks, from which a proper melt composition, including its water content, can be 105 106 estimated.

We have developed a new procedure for the estimation of magma generation
 conditions and the MPTs by examining ultramafic intrusive rocks in the mantle section of an
 ~500 Ma arc ophiolite, the Hayachine-Miyamori Ophiolite (Ozawa et al., 2015), northern Japan.

110	We argue for a slab breakoff tectonic origin on the basis of the obtained magma generation
111	conditions and the geological and petrologic data from the ophiolite. The MPTs are shown to
112	represent the thermal state of the Cambrian-Ordovician upper mantle, which are comparable to
113	the present day MPTs of the upper mantle. The implications of the MPTs and slab-breakoff
114	tectonics for the origin of the Cambrian-Ordovician ophiolite pulse are discussed through
115	compilation of geochemical data from ophiolites of this age.

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#### **PREVIOUS STUDIES**

The Hayachine-Miyamori Ophiolite, located in the South Kitakami Massif, northeast 118 Japan (Fig. 1a), is composed of two complexes separated by the Hizume-Kesennuma fault. They 119 are the Miyamori complex distributed to the west and the Hayachine complex to the east of the 120 fault (Fig. 1a). The two complexes consist mostly of peridotites, with lesser amounts of 121 hornblende-rich mafic-ultramafic rocks and minor shallow intrusions and volcanic rocks. The 122 ophiolite is divided into the Tectonite Member characterized by penetrative plastic deformation 123 textures, and the Cumulate Member featuring cumulus textures without evidence of solid-state 124 deformation (Fig. 1b). Their boundary is delineated within the Miyamori Complex and is an 125 intrusive contact formed after the cessation of the main deformation of the Tectonite Member 126 (Ozawa, 1983; 1984). The boundary strikes nearly parallel to the elongate distribution and 127 layering of the Cumulate Member, ~N 40° W (Ozawa, 1984) (Fig. 1b). The Tectonite Member is 128 further divided by Cr/(Cr+Al) ratio (Cr#) of spinel into the Chromite-bearing Ultramafic Suite 129 (CRUS; Cr# > 0.4) and the Aluminous Spinel Ultramafic Suite (ASUS; Cr# < 0.4). The 130 Hayachine complex is mostly composed of the ASUS. The Miyamori complex is mostly 131

composed of the CRUS, and the rocks of ASUS occur as patches in the CRUS with sizes of 132 several hundred meters to a kilometer (Ozawa, 1987b; Ozawa, 1988; Ozawa et al., 2015). 133 Ozawa (1984) showed that the Miyamori complex is a thrust sheet tectonically 134 transported from the northwest to the current site over Paleozoic sedimentary rocks and 135 metamorphic rocks distributed to the southwest of the Miyamori complex. The metamorphic 136 rocks also occur in a tectonic window (1 km x 600 m) in the central part of the Miyamori 137 complex, where pelitic schist, greenschist, epidote amphibolite, and amphibolite are distributed 138 139 from the west to the east in this order. These rocks could have been a metamorphic sole of the 140 ophiolite as reported from beneath many ophiolites (Williams and Smyth, 1973; Jamieson, 1986) because of the short-distance transition of their metamorphic grade from the greenschist to 141 142 amphibolite facies over the spatial scale of a few hundred meters. Their relationship with the 143 ophiolite is, however, unclear due to the later tectonic disturbance.

The Hayachine-Miyamori Ophiolite is one of the ophiolites belonging to the Cambrian-144 Ordovician ophiolite pulse. The exhumation and cooling of the ophiolite took place at ca. 450 Ma 145 estimated by K-Ar ages of hornblende in hornblende-bearing mafic intrusive rocks (Ozawa et al., 146 1988; Shibata and Ozawa, 1992), and the magmatic events took place at  $499 \pm 65$  Ma estimated 147 by Sm-Nd isochron ages of clinopyroxene separated from lherzolite of the ASUS from the 148 Hayachine complex (Yoshikawa and Ozawa, 2007; Ozawa et al., 2015). The Hizume-Kesennuma 149 fault was active at least since the Jurassic time (Ehiro and Suzuki, 2003). The ophiolite is 150 characterized by several features suggesting its formation in an arc environment, such as the 151 presence of abundant amphibole and minor phlogopite in the mantle section with arc 152 geochemical signatures, and is classified into the supra-subduction type according to Dilek and 153 154 Furnes (2014). The geology, petrology, and geochemistry of the ophiolite have been extensively

studied, and its tectonic history is well constrained in spite of intensive late-stage secondary 155 processes, such as serpentinization, contact metamorphism by granite intrusions, and alteration 156 since the Cambrian-Ordovician time (Ozawa, 1987a, 1988; Ozawa and Shimizu, 1995; 157 Yoshikawa and Ozawa, 2007; Ozawa et al., 2015). The depth of formation of a magma body 158 crystallizing the Cumulate Member in the already deformed Tectonite Member is estimated to be 159  $18 \pm 3$  km (approx.  $0.6 \pm 0.1$  GPa) according to the crystallization of plagioclase in equilibrium 160 161 with olivine in the Cumulate Member at a high magmatic temperature and breakdown of plagioclase into orthopyroxene and spinel-pargasite symplectite along with the contact with 162 olivine at a low temperature (< ~800 °C; Ozawa, 1986). The pressure is too high for the depth of 163 ultramafic cumulate sequences commonly reported from ophiolites, and suggests a thickened 164 crust, though the crustal sections are mostly missing in the Hayachine-Miyamori ophiolites 165 owing to tectonic processes during or after the emplacement. According to the geochemical 166 features, such as TiO<sub>2</sub>/K<sub>2</sub>O in amphibole in ultramafic rocks, and Rb-Sr and Sm-Nd isotope 167 systems for clinopyroxene in mafic and ultramafic rocks, it is inferred that the ASUS of the 168 Hayachine complex was formed by decompressional melting of a depleted mantle in a back-arc 169 basin environment, the CRUS by flux melting of the ASUS under strong influx of fluids from the 170 subducting slab, and the Cumulate Member by decompressional melting of a depleted mantle 171 (Ozawa, 1988; Ozawa and Shimizu, 1995; Yoshikawa and Ozawa, 2007). 172

From these extensive data and geological relationships with the surrounding Silurian-Devonian geological units, such as the Motai metamorphic rocks and the Hikami granitic rocks, Ozawa et al. (2015) proposed a tectonic model for the Hayachine-Miyamori Ophiolite. The model consists of a series of events (1)-(4); (1) formation of the ASUS of the Hayachine complex by back-arc spreading in the Cambrian-Ordovician time (~500 Ma), (2) formation of the CRUS

of the Miyamori complex by extensive flux melting accompanying slab rollback, (3) formation of
the Cumulate Member by decompressional melting of the mantle triggered by slab breakoff, and
(4) exhumation of the ophiolite at ~450 Ma.

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# GEOLOGY OF ULTRAMAFIC INTRUSIVE ROCKS

The ultramafic intrusive rocks, which have a picritic composition, occur in the Cumulate 183 Member and never crosscut younger intrusive rocks such as the hornblende-rich mafic-ultramafic 184 rocks and the andesite-rhyolite porphyries in the Miyamori complex (Fig. 1b). The ultramafic 185 intrusive rocks are planar in morphology, sharply defined with a thickness of 0.5 to 30 cm, and 186 strike from N47°W to N10°E, which is mostly oblique to the layered structure of the Cumulate 187 Member. They crosscut the layering and weak foliation of the host rock and have many branches 188 (Figs. 2a and 2c). They are coarse-grained and porphyritic (Figs. 2b and 3a) and do not have 189 chilled margins in terms of grain size. Phenocrysts, mostly olivine with a minor amount of 190 clinopyroxene, are either heterogeneously or homogeneously distributed in the dikes. In the 191 192 former case, phenocrysts are symmetrically concentrated in the center, and their abundance gradually decreases towards the margin (Table 1, Figs. 2a, 2b, and 2c). The phenocrysts show 193 194 shape-preferred orientation exclusively in the marginal zone with elongate axes lying nearly 195 parallel to the dike wall (Fig. 2c). In the latter case, the homogeneously distributed phenocrysts 196 show less distinct shape-preferred orientation in the dikes.

197 These facts clearly show that the magmas intruded into the solid host rocks of the 198 Cumulate Member have solidified as dikes or veins. The Cumulate Member is the latest main 199 lithology in the evolution of the Hayachine-Miyamori ophiolite before the formation of

200	hornblende-bearing mafic-ultramafic intrusive rocks (Ozawa et al., 2015). The formation of the
201	ultramafic intrusive rocks, "ultramafic dikes" hereafter, is one of the latest events in the ophiolite
202	evolution. We found five outcrops of this type of intrusive rocks (Fig. 1b) and collected samples
203	from four thick dikes at four outcrops, two of which are covered with river gravels of a recent
204	flooding and only rootless blocks are available.
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206	PETROGRAPHY
207	The absence of chilled margins near the contact with their hosts suggests that the intrusion
208	took place at a fairly high temperature and possibly at high pressure, which must be clarified.
209	Moreover, the dikes contain secondary minerals locally forming veins, which sharply cut the
210	primary igneous textures. This indicates that the dikes may have undergone secondary processes
211	such as metamorphism and/or alteration, which could have modified the igneous chemical and
212	textural features. In this section, we describe petrographic features related to igneous processes
213	first to clarify the intrusion environment, and then describe secondary features in order to
214	evaluate the effect of secondary processes.

# 215 **Igneous textures**

Phenocrysts in the dikes are composed mainly of euhedral - subhedral olivine with a subordinate amount of euhedral clinopyroxene, which is absent in some dikes (Table 1). The size of phenocrysts is up to a few millimeters (Figs. 2b and 3a). Olivine is partially preserved everywhere in the dikes which show a homogeneous phenocryst distribution, though the extent of replacement with serpentine is variable (Table 1). Olivine is preserved at the margin and in the center but not at the edge of the dikes which show a central phenocryst concentration (Table 1).

There is a tendency that the matrix-dominant lithology is more susceptible to secondary 222 processes and form hydrous phases such as serpentine and chlorite. The olivine phenocrysts 223 locally include chrome spinel and are partially rimmed by orthopyroxene (Figs. 3a and 3b). 224 The matrix filling interstitial part of phenocrysts consists of anhedral amphibole and 225 clinopyroxene with minor chrome spinel and dusty fine-grained aggregate after plagioclase 226 (Table 1). Amphibole is the most dominant matrix phase (Table 1; Figs. 3a, 3b, 230 and 3c) and 227 its size ranges from several tens of microns to 1 mm. It forms a polycrystalline aggregate of a few 228 229 millimeters across and is in direct contact with olivine phenocrysts partially intervening thin

orthopyroxene rim. The amphibole locally includes euhedral chrome spinel. The clinopyroxene
forms equigranular aggregates, which show a patchy distribution in amphibole or plagioclase
pseudomorph. The grain size of clinopyroxene is from several tens to hundreds of microns and
tends to be smaller than that of amphibole.

All plagioclase crystals are replaced by aggregates of secondary minerals, but the 234 aggregates can be identified as plagioclase pseudomorphs as explained below. The plagioclase 235 236 pseudomorphs, up to a few millimeters across, are present only in the center of dikes (Table 1). The pseudomorphs show an anhedral outline, occur in the middle of polycrystalline amphibole, 237 and are locally in contact with clinopyroxene, but are never in direct contact with olivine (Figs. 238 3a, 3b, and 3c). Such features of plagioclase occurrence are the same as those observed in 239 plagioclase-bearing peridotites of the Cumulate Member (Ozawa, 1986). Spinel shows two 240 different modes of occurrence, which is intimately related to the color difference under an optical 241 microscope which reflects chemical compositions. Brownish chromite spinel is euhedral to 242 subhedral with grain sizes up to 1 mm. It is either included in olivine or occurs in amphibole 243 244 aggregates. Greenish aluminous spinel showing anhedral and locally vermicular morphology

occurs in amphibole surrounding plagioclase (Fig. S1). The texture is essentially the same as that of a reaction zone consisting of an inner zone of aluminous spinel-pargasite symplectite and an outer zone of orthopyroxene around plagioclase in plagioclase-bearing wehrlite of the Cumulate member (Ozawa, 1986). The texture suggests a reaction between olivine phenocrysts and fractionated interstitial melt after or during the crystallization of plagioclase from the interstitial melt, which may have taken place in the last stage of solidification of the dikes at the same pressure estimated for the Cumulate Member formation, ~0.6 ± 0.1 GPa (Ozawa et al., 2015).

252 Orthopyroxene exclusively occurs rimming olivine phenocrysts with thickness of a few 253 hundred µm (Figs. 3a and 3b) and is less abundant near the edge of dikes than in the dike center (Table 1). The dikes have a selvage of orthopyroxene aggregates with a thickness of  $\sim 0.1$  mm 254 255 along the contact with the host peridotite. The same relationship between pyroxenite dikes and 256 harzburgite host was reported from the Tectonite Member by Ozawa (1994), who attributed the texture to a reaction of hydrous melt transported through fractures with the wall rock under heat 257 258 loss. The same textural relationship between the olivine phenocrysts with orthopyroxene rim suggests a reaction between the olivine crystals and a magma flowing through fractures before its 259 complete solidification filling the fractures as dikes. The systematic spatial variation of the 260 thickness of orthopyroxene rim on olivine phenocrysts suggests a more extensive reaction in the 261 dike center than in the margin. There are no vesicles or miarolitic cavities, suggesting that no 262 vesiculation occurs during the solidification of the dikes (Figs. 2a, 2b, and 3a). 263

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# 265 Textures related to secondary low-temperature processes

Igneous textures were modified in various degrees by low-temperature processes, 266 which are defined in this paper as low-temperature ( $< \sim 700$  °C) metamorphism and/or alteration 267 with introduction of fluids to form hydrous phases such as serpentine, chlorite, and tremolite 268 associated with crystallization of magnetite and spindle-shaped diopside. We hereafter use 269 "alteration" to denote this process. The orthopyroxene is partially (or totally) replaced by fine-270 grained aggregates of tremolite and diopside while keeping the primary grain outline 271 272 (orthopyroxene pseudomorph; Fig. 3a) in samples taken from outcrop (DKSH1 and DKSH2). Some rootless rocks from the dikes (RYO1 and RYO2), however, have fresh orthopyroxene 273 (Figs. 3b and 3c). The aggregates are partly replaced by chlorite. Amphibole is less susceptible to 274 275 the alteration, but is replaced by tremolite, diopside, chlorite, and serpentine particularly at the dike margin. Plagioclase pseudomorph consists of fine-grained grossular, but the original outline 276 of plagioclase is preserved. In spite of the complete replacement, the origin of the grossular 277 aggregate as plagioclase is supported by its trace element patterns, which are shown below. 278 Olivine phenocrysts are replaced partially or completely by serpentine and magnetite. Such 279 replacement is limited in fresh samples to the formation of serpentine and magnetite along minor 280 fractures in olivine (Figs. 3a and 3b). These textures suggest multistage open-system alteration 281 reactions after the complete solidification of the dike. 282

The extent of the replacement of each primary mineral by secondary phases was qualitatively evaluated under an optical microscope, and three levels of alteration were distinguished: weak, intermediate, and extensive (Table1). This information was used to classify alteration degrees of the samples into weak, intermediate, and strong (Table 1). The one sample with strong alteration is from the edge of dike DKSH1. Weakly altered samples are from the marginal and central parts of dike DKSH2, which was used to examine alteration processes in detail. A more quantitative parameter for alteration is the mass loss on ignition, which isdescribed below.

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### **ANALYTICAL METHODS**

Several slabs were obtained sequentially from the center to edge of two dikes DKSH2, 293 which has a homogeneous phenocryst distribution, and DKSH1, which shows a high 294 concentration of phenocrysts in the center (Table 1; Figs. 2a and 2b). Two samples of rootless 295 blocks from locality #3 (RYO1) with a homogeneous phenocryst distribution and locality #4 296 (RYO2) with a zoned phenocryst distribution at sample scale ( $\sim 15$  cm) were also examined 297 (Table 1). Each slab sample (100 - 150 g) was rinsed for 15 minutes in deionized water and 298 methanol using an ultrasonic vibrator and was dried in an oven at 80 °C for 12 hours. The dried 299 slab samples were crushed to chips of  $\sim 5$  mm diameter by a tungsten mortar. The chips were 300 crushed to powder by the rod mill: Retsch Vibratory Disc Mill RS 200 with an agate rod for 301 several minutes. The powder samples of 1.5 - 2.0 g were put in the oven and heated at 120 °C 302 303 for 12 hours to outgas the absorbed water. Following this procedure, the powder samples were ignited in a programable multistage heater in a Muffle Furnace of Yamato FO510 to outgas the 304 305 water and other volatile components in the crystal structures at several temperatures and for the 306 respective durations: 100 °C for 10 minutes, 500 °C for 30 minutes, 800 °C for 30 minutes, and 307 1000 °C for 6 hours. The mass loss including water in samples is regarded as a loss on ignition 308 (LOI in wt%). Thin section observation of the examined samples shows that only a minor amount of carbonates and sulfides are present ( $< \sim 2 \text{ vol}\%$  in total). Therefore, the contribution 309 310 of these minerals to the LOI is negligible.

311	Major and trace element concentrations of the whole-rock samples were measured with an
312	X-ray Fluorescence Spectrometer (XRF): PANalytical Axio at the Department of Earth and
313	Planetary Science, the University of Tokyo by the glass bead method for major elements and the
314	pressed powder method for trace elements. The analytical procedures have been described by
315	Yoshida and Takahashi (1997). The glass beads were prepared as follows. The powder samples
316	and the flux of anhydrous lithium tetraborate were weighed 0.4 g (< 0.1% in error) and 4 g (<
317	0.1% in error), respectively and were mixed. The mixed powder and a drop of lithium iodide
318	were put in a platinum crucible and heated with a fuse sampler, Tokyo Kagaku TK-4100, to
319	obtain glass bead.
320	Major element concentrations in minerals were measured with an electron probe
321	microanalyzer (EPMA; JEOL JXA-8900L), at the Department of Earth and Planetary Science,
322	the University of Tokyo. The adopted analytical conditions for olivine, pyroxene, amphibole, and
323	spinel are: a focused beam, an acceleration voltage of 15 kV, a beam current of 12 nA on a
324	Faraday cup, and 30 s counting time for peak and 15 s for background on both sides of each peak.
325	The beam diameter of 50 $\mu$ m is adopted for analyses of the core of plagioclase pseudomorph
326	replaced by fine-grained mineral aggregates to obtain the bulk composition. We adopted an
327	acceleration voltage of 25 kV and a beam current of 50 nA on a Faraday cup and 100 s counting
328	time for peak and 50 s for background on both sides of each peak for trace elements in olivine
329	(NiO and MnO). The correction method to obtain wt% from peak intensity after correcting
330	background is ZAF.

Trace elements in the whole-rock samples were measured on the glass beads prepared for the XRF analyses with a Thermo Fisher Scientific iCAP Q inductively coupled plasma mass spectrometer (ICP-MS) coupled with a CETAC LSX-213 G2+ Nd:YAG laser ablation (LA)

334	system at the Department of Earth and Planetary Science, the University of Tokyo. The analytical
335	procedures described by Itano and Iizuka (2017) were adopted. The glass beads were cut
336	vertically to the disks and the cut surface was polished with diamond paste (1 $\mu$ m size). The laser
337	beam with a spot size of 50 $\mu m,$ a shot frequency of 10 Hz, and a scan rate of 5 $\mu m/s$ in line
338	analysis were used. Analyzed elements are Ca, Cr, Mn, Ni, Sr, Y, Zr, Nb, Ba, La, Ce, Pr, Nd, Sm,
339	Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, Pb, and U. The obtained signal intensities of trace
340	element data were corrected for gas blank, and calibrated against NIST SRM 613 (Hollocher and
341	Ruiz, 1995). Trace element concentrations on each sample were normalized using the Ca
342	concentrations of the glass beads measured with XRF as an internal standard.
343	Trace element concentrations of amphibole and clinopyroxene in thin sections samples
344	polished with diamond paste (1 $\mu$ m size) were measured with the LA-ICP-MS. Focused-laser
345	with a spot size of 40 $\mu$ m and a shot frequency of 5 Hz were used. The integration time is 60
346	seconds, including 20 seconds gas blank measurement. Analyzed elements are Ca, Sr, Y, Zr, Nb,
347	Ba, La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb, Lu, Hf, Ta, Pb, and U. The correction
348	procedures of obtained signal intensities are basically the same as those adopted for the whole-
349	rock analyses described above. The Ca concentration measured by EPMA on each analysis spot
350	was used as an internal standard.
351	Nd and Sr isotope ratios in the whole-rock samples and clinopyroxenes were measured
352	with a Thermo Fisher Scientific Neptune plus multiple collector ICP-MS at the Graduate School

of Science, Hokkaido University. Clinopyroxene grains were carefully handpicked from coarsely

crushed samples of the ultramafic dikes under a binocular microscope. They were powdered by

an agate mortar. The analytical procedures for chemical separation followed the methods used in

Pin et al. (1994) and Noguchi et al. (2011) for Sr, Pin et al. (1994) and Pin and Zalduegui (1997)

357	for Nd. Mass fractionation factors for Sr and Nd were calculated using ${}^{86}$ Sr/ ${}^{88}$ Sr = 0.1194 and
358	$^{146}$ Nd/ $^{144}$ Nd = 0.7219, respectively. Additional corrections were performed by applying a
359	standard bracketing method using NIST987 and JNdi-1 for Sr and Nd isotopic analyses,
360	respectively. The data were finally normalized to ${}^{87}\text{Sr}/{}^{86}\text{Sr} = 0.710214$ for NIST 987, and
361	$^{143}$ Nd/ $^{144}$ Nd = 0.512117 for JNdi-1. The isotopic ratios of the GSJ standard JB-3, measured
362	during the course of this study, were ${}^{87}$ Sr/ ${}^{86}$ Sr = 0.703384 ± 0.000024 (n = 32, 2 $\sigma$ ), ${}^{143}$ Nd/ ${}^{144}$ Nd =
363	$0.513065\pm0.000011$ (n = 23, 25). The Rb/Sr and Sm/Nd ratios of whole-rock samples were
364	measured by ICP-MS using a Thermo Fisher X-series at the Graduate School of Science,
365	Hokkaido University. The analytical uncertainties are typically 0.7 % for the Rb/Sr ratios and 0.5
366	% for the Sm/Nd ratios.
367	
368	WHOLF DOCK CHEMISTRY
	WHOLE-ROCK CHEMISTRY
369	Major elements
369 370	Major elements The whole-rock major element oxide contents including the loss on ignition (LOI) of dike
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376

# 377 Rare earth and other trace elements

378	The whole-rock rare earth element (REE) and other trace element of dike samples are
379	listed in Table 1. The CI-chondrite normalized trace element and REE patterns of the samples are
380	plotted in Figs. 5a, 5b, 5c, and 5d. The dikes have essentially the same trace element patterns,
381	which are characterized by strong positive anomalies of Ba and Sr, a strong negative anomaly of
382	Pb, and weak negative anomalies of Hf and Zr relative to the REEs, and U, Nb, Ta lying on a
383	smooth extension of the REE variations (Fig. 5a). The REE patterns are essentially the same for
384	all samples and are characterized by smooth overall variations: a depletion of light rare earth
385	elements (LREE), and a weak depletion of heavy rare earth elements (HREE) (Figs. 5c and 5d).
386	The abundances of REE in the dikes having homogeneous phenocryst distributions are nearly
387	constant (Fig. 5c), but those of the dikes having a central enrichment of phenocrysts are high in
388	the edge and margin and low in the center while keeping the same variation pattern (Fig. 5d).

389

### **390** Isotope ratios

The whole-rock <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd ratios of samples from four dikes are listed in 391 Table 2 and are plotted in Fig. 6a. The <sup>143</sup>Nd/<sup>144</sup>Nd ratios are similar irrespective of localities and 392 positions within each dike and range from 0.512975 to 0.513079. There is a systematic difference 393 394 of Sr isotopes depending on the degree of alteration in samples from dike DKSH2, as shown in Fig. 6a. The strongly and intermediately altered samples with high LOI have higher <sup>87</sup>Sr / <sup>86</sup>Sr 395 396 ratios than the less altered samples with low LOI. The lines that join samples with different degrees of alteration in the <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd diagram (Fig. 6a) are nearly horizontal, 397 which indicates that secondary processes shifted <sup>87</sup>Sr/<sup>86</sup>Sr to higher values but did not affect the 398 399 Nd isotope ratios. The least altered samples of the dikes with homogeneous phenocryst distribution (DKSH2MW and DKSH2CW) have low <sup>87</sup>Sr / <sup>86</sup>Sr ratios, which are thought to 400

401	represent the primary values before the alteration. The range of the initial ENd of four dikes and
402	the values of the initial ${}^{87}$ Sr/ ${}^{86}$ Sr of DKSH2MW and DKSH2CW are 6.0 - 7.1 and ~0.7032,
403	respectively and plot just at the lower end of the Cambrian-Ordovician MORB field (Yoshikawa
404	and Nakamura, 2000; ɛNd, 7 - 12; <sup>87</sup> Sr / <sup>86</sup> Sr, 0.7018 - 0.7027).
405	
100	
406	MINEKAL CHEMISIKY
407	Cation ratios of minerals are abbreviated as follows: $Mg# = Mg/(Mg + Fe^{2+})$ , $Cr# =$
408	$Cr/(Cr + Al)$ . The total iron is assumed to be $Fe^{2+}$ , except for spinel and amphibole, for which
409	ferric iron contents were calculated from stoichiometry constraints with charge balance
410	considerations. The procedure for amphibole is from Leake et al. (1997).
411	Major and trace element compositions of minerals
412	Olivine. Olivine in the dikes and their host peridotites is homogeneous in Mg# on a grain
413	scale (a few millimeters). A variation in Mg# on the scale of 10 cm is, however, observed not
414	only in the dikes but also in the host peridotites. This is well documented in dikes with a
415	homogeneous phenocryst distribution as shown in Fig. 7 for dike DKSH2 because of the limited
416	alteration near the contact. The Mg# decreases from 0.90 to 0.87 over the distance of 54 mm
417	from the dunite host to the contact and from 0.87 to 0.83 over the distance of 60 mm from the
418	contact to the dike center (Fig. 7a). Olivines in the dikes are heterogeneous in NiO content on a
419	grain scale, which is manifested by the larger scatter in composition at a given distance than that
420	of Mg# in Fig. 7a. It shows variation across the contact with the host dunite as in the case of Mg#
421	(Fig. 7b), but the spatial extent of variation is narrower than that of Mg# (cf. Figs. 7a and 7b).

422	The average NiO content in olivine at the center of the dike is 0.21 wt% with a standard deviation
423	of 0.01 wt% in 1 $\sigma$ . It is 0.35 wt% with a standard deviation of 0.01 wt% in 1 $\sigma$ in the host dunite.
424	Spinel. Spinel is present in all samples and shows diverse chemical composition,
425	particularly in terms of tetravalent cations (Table 1; Fig. 8). It shows zoning in each grain.
426	Euhedral spinel grains show concentric zoning with decreases in Cr# and Fe <sup>3+</sup> /(Cr+Al+Fe <sup>3+</sup> )
427	from the core to the rim (see Figs. S2a and S2b). The Cr# and Fe <sup><math>3+</math></sup> /(Cr+Al+Fe <sup><math>3+</math></sup> ) vary from 0.37
428	and 0.14 at the grain center to 0.13 and 0.05 at the rim, respectively. The largest euhedral spinel
429	grain registers the highest $Cr#$ and $Fe^{3+}/(Cr+Al+Fe^{3+})$ at the core. Subhedral and anhedral spinel
430	grains occurring in adjacent to plagioclase pseudomorphs are rich in Al and poor in Cr and $Fe^{3+}$ .
431	Their Cr# range 0.00 - 0.03 and Fe <sup>3+</sup> /(Cr+Al+Fe <sup>3+</sup> ) 0.01 - 0.02. The Cr# in large grains of such
432	spinel shows an asymmetric zoning pattern, which increases from the rim closer to plagioclase to
433	the opposite rim (see Figs. S2c and S2d). Small green spinel, locally vermicular in morphology,
434	is particularly aluminous, and it is almost Cr-free in an extreme case (Figs. 8 and S1).
435	Amphibole. Amphibole is present in all the dikes but is completely replaced by
436	secondary minerals near the contact with the host peridotites (Table 1). All the amphiboles are
437	classified as pargasite according to Leake et al. (1997) (see Fig. S3). Amphibole in the dike
438	center is homogeneous in Mg# on a grain scale (~1 mm). The Mg# of amphibole varies from 0.82
439	to 0.92 depending on the local modal abundance of olivine phenocrysts: the Mg# tends to be
440	higher near olivine phenocrysts. A variation of Mg# across the contact with the host peridotite
441	cannot be clarified because of the complete alteration of amphibole near the contact (Fig. 7a).
442	The $TiO_2$ / $K_2O$ weight ratios of amphibole, shown in Fig. 9, are within the range for normal mid-
443	ocean ridge basalts (NMORB) and are similar to the values of ASUS of the Hayachine complex
444	(Ozawa, 1988) and the Cumulate Member of the Miyamori complex (Ozawa et al., 2015). The

445	ratio is clearly distinct from those of CRUS in the Miyamori complex, which is comparable to the
446	ratio of high magnesian andesite (Ozawa, 1988).

447	Chondrite-normalized trace element patterns of 7 amphibole grains in weakly altered
448	samples of dike DKSH2 are plotted in Fig. 10a. The patterns have peculiar features of a
449	substantial decline from middle rare earth elements (MREEs) to light rare earth elements
450	(LREEs), a slight decline from MREEs to heavy rare earth elements (HREEs), positive anomalies
451	of some of highly incompatible elements (Ba and Sr), a strong negative anomaly of Pb and weak
452	negative anomalies of Hf and Zr relative to rare earth elements (REEs).
453	Orthopyroxene. Orthopyroxene survived alteration in the least altered dikes (Table1;
454	RYO1 and RYO2) is homogeneous on a few hundred-micrometer scale and has Mg#, Cr#, Al <sub>2</sub> O <sub>3</sub>
455	and CaO wt% ranging from 0.82 to 0.84, 0.01 to 0.03, 2.1 to 3.1 wt%, 0.47 to 0.67 wt%,
456	respectively. The spatial chemical variations of orthopyroxene on the scale of $\sim 10$ cm in the dikes
457	and their host peridotite cannot be clarified because of the extensive alteration.
458	Clinopyroxene. Clinopyroxene phenocrysts, that occur in some of the dikes (Table1;
459	DKSH1, RYO1, and RYO2), have chemical zoning in terms of Mg#, Cr#, Al <sub>2</sub> O <sub>3</sub> , and CaO. The
460	core tends to have higher Mg#, Cr#, and lower Al2O3, CaO wt% than the rim. The Mg#, Cr#,
461	Al <sub>2</sub> O <sub>3</sub> , and CaO wt% vary from 0.86 to 0.88, 0.03 to 0.10, 2.9 to 5.1 wt%, 21.4 to 23.5 wt%,
462	respectively. The <sup>87</sup> Sr/ <sup>86</sup> Sr and <sup>143</sup> Nd/ <sup>144</sup> Nd ratios of separated clinopyroxene from dike RYO1
463	are 0.513055 and 0.703701, respectively, which are similar to those of the whole rock (Fig. 6a).
464	The Mg#, Cr#, Al <sub>2</sub> O <sub>3</sub> and CaO wt% of matrix clinopyroxenes show variations on a few
465	hundred micrometers scale, ranging from 0.83 to 0.89, 0.03 to 0.08, 2.7 to 4.9 wt%, 21.1 to 23.5

467	7a), though the variation is not so distinct as exhibited by olivine. This is partly due to the
468	variability of the Mg# depending on the local modal abundance of olivine phenocrysts. Spindle-
469	shaped clinopyroxene, which is inferred to have crystallized during alteration according to its
470	occurrence and morphology, has the highest CaO (~24.4 wt%) and lowest Al <sub>2</sub> O <sub>3</sub> (~0.4 wt%)
471	contents and classified as diopside.
472	Chondrite-normalized trace element patterns of matrix clinopyroxene from weakly altered
473	dike DKSH2 are plotted in Fig. 10a. All of 5 analyses are essentially the same except for Ba,
474	which shows a large variation. The trace element patterns are characterized by a depletion of Ta,
475	strong negative anomalies of Pb and Nb. The REE abundance of matrix clinopyroxene is lower
476	than that of the matrix amphibole. Their patterns are distinct in that clinopyroxene has a more
477	sloped CI-chondrite normalized variations of LREEs and HREEs than amphibole.
478	<b>Plagioclase (nseudomorph).</b> Trace element patterns of fine-grained aggregate which
170	Tingtoone (population ph), Theo common particular of the granne aggregate when
479	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct
479 480	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This
479 480 481	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph).
479 480 481 482	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph).
<ul> <li>479</li> <li>480</li> <li>481</li> <li>482</li> <li>483</li> </ul>	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph). MAGMA GENERATION CONDITION AND MECHANISM
479 480 481 482 483 484	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph). MAGMA GENERATION CONDITION AND MECHANISM In this section, after showing that melt compositions are restorable from the dikes, we will
<ul> <li>479</li> <li>480</li> <li>481</li> <li>482</li> <li>483</li> <li>484</li> <li>485</li> </ul>	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph). <b>MAGMA GENERATION CONDITION AND MECHANISM</b> In this section, after showing that melt compositions are restorable from the dikes, we will take the following steps to estimate magma generation conditions, melting mechanisms, and
<ul> <li>479</li> <li>480</li> <li>481</li> <li>482</li> <li>483</li> <li>484</li> <li>485</li> <li>486</li> </ul>	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph). MAGMA GENERATION CONDITION AND MECHANISM In this section, after showing that melt compositions are restorable from the dikes, we will take the following steps to estimate magma generation conditions, melting mechanisms, and mantle potential temperatures on the basis of the data presented above. (1) We first restore the
<ul> <li>479</li> <li>480</li> <li>481</li> <li>482</li> <li>483</li> <li>484</li> <li>485</li> <li>486</li> <li>487</li> </ul>	consist mostly of grossular in a weakly altered sample DKSH2CW are characterized by distinct Eu positive anomaly and depletion of HREEs, which are typical for plagioclase (Fig. 10b). This clearly supports that the aggregate is an alteration product of plagioclase (pseudomorph). <b>MAGMA GENERATION CONDITION AND MECHANISM</b> In this section, after showing that melt compositions are restorable from the dikes, we will take the following steps to estimate magma generation conditions, melting mechanisms, and mantle potential temperatures on the basis of the data presented above. (1) We first restore the original whole-rock compositions for major and trace elements by correcting for alteration

489	modifications and olivine accumulation. (3) We estimate primary melt compositions in
490	equilibrium with the mantle by adding olivine to the intruded melts. (4) We use the primary melt
491	compositions to estimate melting mechanism, mantle source composition, melting/melt
492	segregation conditions, and solidus conditions. (5) We finally estimate mantle potential
493	temperatures. (6) Additionally, we evaluate the openness of the melting system.

494

495 Melt compositions preserved in the dikes

The examined dikes are similar in their field occurrences, petrography, whole-rock 496 chemical compositions, mineral chemical compositions, and isotopic compositions, as described 497 above, irrespective of the phenocryst distribution in each dike. This indicates that they formed in 498 a series of intrusion events from a common magma and that the phenocryst distribution was 499 controlled by intrusion and solidification processes. The high concentration of phenocrysts in the 500 dike center and shape-preferred orientation of the olivine phenocrysts in the marginal zone 501 indicate the operation of flow differentiation owing to a velocity gradient in flowing dikes, which 502 503 is known as "Bagnold effect" (Bagnold, 1954; Komar, 1972). The homogeneous distribution and less distinct shape-preferred orientation of phenocrysts in some dikes can be explained by a 504 505 turbulent flow associated with a high Reynolds number. The distributions of phenocrysts in the 506 dikes suggest that the magmas were injected as crystal-laden magmas which solidified in a closed system on a short time scale to maintain the distribution of phenocrysts under the gravitational 507 508 field irrespective of the zoned or homogeneous distribution.

Examination of the variation of Mg# of olivine across the contacts between the host
peridotite and the dike (Fig. 7a) also supports such rapid solidification. The area ratio of the Mg-

Fe profile of olivine in the dike and the host sides (Fig. 7a) is 0.69 : 0.31, which is clearly 511 different from 1:1. The host peridotite is mainly composed of olivine (~99 wt%), whereas the 512 dike side is composed of olivine phenocrysts (~37.0 wt%) and matrix minerals (amphibole, ~58.3 513 wt%; orthopyroxene, ~3.9 wt%; and clinopyroxene, ~0.1 wt%; Table 3) which are calculated 514 from the igneous modal composition of the DKSH2EI in vol% (Table 1). Taking the igneous 515 modal abundance of the dike side into consideration, it is shown that the mass balance of MgO is 516 517 maintained in a closed system, within a uncertainty smaller than ~15% (Table 3). Therefore, the Mg-Fe profile in the homogeneous-type dike is due to subsolidus diffusional Mg-Fe exchange 518 after the solidification of the magma on a time scale much longer than Fe-Mg exchange involving 519 520 magma over 10 cm. The Mg# profile across the contact is asymmetric and characterized by a shorter tail on the host dunite side and a longer tale within the dike (Fig. 7a). This feature is 521 522 attributable to Mg-Fe interdiffusion with composition dependence and the difference in igneous modal abundance. This is consistent with the much narrower variation of NiO content in olivine 523 than that of Mg# because the Ni diffusion is slower than the Fe-Mg interdiffusion (Fig. 7b; Petry 524 et al., 2004; Dohmen and Chakraborty, 2007). It is highly plausible that the melt composition in 525 the dikes is preserved with limited interaction with the wall rocks by rapid solidification after the 526 magma intrusion. 527

528

# 529 Restoration of original whole-rock compositions for major elements

Subsolidus modifications due to alteration. The presence of secondary hydrous
minerals, such as serpentine, diopside, and chlorite, as described in the petrography section,
indicates that the dikes suffered from alteration. In order to clarify if the formation of secondary
hydrous phases modified the whole-rock compositions, we compared two pairs of samples from

dike DKSH2, which show a different degree of alteration (Table 1) while keeping the same 534 igneous modal abundances (Table 1). Loss on ignition (LOI) was used as a quantitative measure 535 of the degree of alteration, and the effects of alteration were examined by plotting oxide wt% 536 data against the LOI (Fig. 11). A vector defined by a pair of samples with a difference in LOI and 537 oxide contents is regarded as a trend of alteration. The obtained mean alteration vectors from the 538 two sample pairs are listed in Table S1. The initial whole-rock water contents for each sample 539 540 can be estimated from the water content of pargasite ( $\sim 2.5$  wt%) and its modal abundances (Tables 1 and 4). We obtained whole-rock compositions prior to alteration for nine samples by 541 542 extrapolating the whole-rock data to the initial whole-rock water contents. The corrected compositions are listed in Table 4. 543

Subsolidus high-temperature modifications due to elemental exchange with the host. 544 545 The variation of the Mg# of olivine observed near the contact of dike DKSH2 and the host (Fig. 546 7) shows that the whole-rock MgO and FeO concentrations in the dike were modified by 547 subsolidus diffusion at high temperatures as discussed above. Figure 7 shows that the 548 modification is limited to within 4 cm from the contact with the host. Most of the samples 549 including the rootless blocks with more than 10 cm size were, therefore, not affected by the diffusional exchange. However, the whole-rock chemical compositions of the samples near the 550 551 contact (DKSH1ES and DKSH2EI; Table 1) were shifted from the values that these rocks had soon after the dike solidification. The whole-rock composition of sample DKSH2EI before the 552 553 subsolidus high-temperature modification was restored by using the initial olivine Mg# in the 554 center (~0.83, Fig. 7a), which was not modified by this modification, and the igneous modal abundance (Table 4). The same procedure cannot be applied to DKSH1ES because of its 555 extensive alteration. The restored composition for DKSH2EI is lower in MgO and higher in FeO 556

(Table 4). The whole-rock major element compositions corrected for the alteration and
diffusional elemental exchange with the host are plotted in Fig. 12 excepting sample DKSH1ES.

559

# 560 Estimation of intruded melt compositions for major elements and igneous olivine

561 compositions.

The liquid composition of magmas intruded into fractures and frozen as ultramafic dikes 562 is not simply the whole-rock compositions after the corrections made above. This is because the 563 dikes contain a large number of phenocrysts. The liquid parts of the magmas including 564 phenocrysts at the time of intrusions are called "intruded melt", hereafter. The MgO 565 concentrations of whole rocks corrected for alteration and elemental exchange with the host vary 566 from 14.6 to 27.7 wt% and are negatively and linearly correlated with Al<sub>2</sub>O<sub>3</sub> at nearly constant 567 FeO concentrations (Fig. 12; Table 4) as in the case of Hawaiian basalts (e.g., Rhodes and 568 Vollinger, 2004; Putirka, 2005). The correlation of the whole-rock composition and phenocryst 569 570 abundance suggests that the linear relationships among oxide wt% are explained by a binary 571 mixing of the phenocrysts and intruded melt for each dike. We need to know the accurate abundance and composition of the phenocrysts to properly estimate the intruded melt 572 573 compositions. Although we can obtain the average amount of crystals loaded in the intruded 574 magma by integrating the heterogeneously distributed phenocrysts over the dike width if 575 necessary, we cannot easily estimate the chemical composition of the phenocrysts. The measured 576 Mg# of olivine (~0.83, Fig 7a), the lowest value in the dikes, does not represent an igneous composition. This is because the current compositions of olivine phenocrysts are not those in 577 578 equilibrium with the intruded melt owing to the modification through elemental exchange with 579 the interstitial melt and further with the other constituent minerals at subsolidus by diffusion over

the grain scale. The scale of diffusional exchange between olivine phenocrysts and the surrounding melt in the dikes after the cessation of magma transportation is smaller than the sample scale. On the basis of this limitation of olivine-melt interaction, we can estimate the olivine and host melt composition at high temperature from the measured whole-rock composition listed in Table 4 and the olivine phenocryst-matrix ratio listed in Table 1 by taking the following steps.

Step1: We assume a Mg#, NiO and MnO content for olivine in equilibrium with a 586 587 hypothetical intruded melt, starting with the measured Mg#x, NiOx, and MnOx for example. 588 Step2: We subtract the observed amount of all olivine phenocrysts with the assumed Mg#x, NiOx, and MnOx contents from a whole-rock composition (Tables 1 and 4). After subtracting 589 590 phenocrysts, we obtain a hypothetical intruded melt as a residue of the subtraction. This is called 591 meltz. Step3: We calculate a Mg#, NiO, and MnO content for an olivine in equilibrium with the 592 meltz using the partition coefficients of Mg, Fe, Ni, and Mn between olivine and melt (Beattie et 593 al., 1991; Kinzler et al., 1990). The calculated contents are called Mg#<sub>Y</sub>, NiO<sub>Y</sub>, and MnO<sub>Y</sub>. Step4: We check if the values of Mg#x, NiOx, and MnOx, and those of Mg#y, NiOy, and MnOy, are 594 595 respectively the same. If we assumed appropriate values of Mg#, NiO, and MnO of olivine in Step1, these values of X and Y should be the same. However, this is usually not the case. Step5: 596 597 When the values of Y are different from those of X, then we go back to Step 1 and assume slightly different values of X, and go to Step 2. By repeating Step1 to Step5 until the X and Y 598 sets of olivine compositions coincide, we obtain the intruded melt composition and the Mg#, 599 600 NiO, and MnO contents of olivine in equilibrium with the intruded melt. The ratios of  $Fe^{3+}$ / total Fe of the intruded melts are assumed to be  $\sim 0.13$ , which was estimated by applying the olivine-601 orthopyroxene-spinel oxygen geobarometry to our samples (Ballhaus et al., 1991; Sack et al., 602

603	1980; O'Neill, 1987). The results are listed in Table 5. The estimated Mg#, NiO, and MnO
604	contents of olivine vary from 0.85 to 0.89, 0.08 to 0.21, and 0.17 to 0.32, respectively.
605	The examined dikes show no features suggesting vesiculation during solidification, as
606	mentioned above. This observation and the estimated emplacement pressure of the host Cumulate
607	Member (~0.6 GPa; Ozawa, 1986) suggest that the intruded melts were undersaturated with
608	water, which was retained in hydrous phases without degassing. Since the initial hydrous mineral
609	in the dikes is pargasite with $\sim$ 2.5 wt% water (Table 1), we estimated the initial water content for
610	each dike by using the modal abundance of pargasite (Table 5). The estimated water contents in
611	the intruded melt vary from 0.6 to 2.3 wt%. Even for the maximum estimate (2.3 H <sub>2</sub> O wt%), the
612	saturation pressure of the intruded melt is $< \sim 0.1$ GPa according to VOLATILECALC (Newman
613	and Lowenstern, 2002), which is distinctively lower than the pressure of solidification of the
614	dikes.

615

# 616 Estimation of intruded melt compositions for trace elements

The rare earth element (REE) concentrations of the measured samples do not show any systematic relationships with the degree of alteration or LOI (Fig. 5c). Therefore, we infer that the REE in dike samples were not modified by secondary processes, and we calculated the REE of the intruded melts from the raw whole-rock data by subtracting the contribution of the phenocrysts (Table 5).

We also estimated most of the trace elements of the intruded melts in the same way as the REEs (Table 5). However, some trace elements, particularly fluid-mobile incompatible elements, which are prone to secondary open processes such as alteration, show variations controlled by the

625	degree of alteration or LOI. Our samples contain secondary minerals replacing primary minerals
626	and occurring in veinlets cross-cutting the primary textures. These indicate some corrections
627	must be made for these elements. We calculated the whole-rock compositions of intruded melt
628	from the trace element and the igneous modal abundances of matrix minerals for the least altered
629	dike, DKSH2, whose matrix is composed mostly of amphibole and clinopyroxene (> 97 %; Table
630	1). The results are listed in Table 6 and plotted in Fig. S4. The calculated trace element patterns
631	of the intruded melt are smooth, and several positive anomalies of, Ba, U, and Sr noted in the
632	whole-rock data are subdued.

633

#### 634 Estimation of primary melt compositions

It is very plausible that a magma frozen in the form of the ultramafic dike underwent 635 fractional crystallization during its ascent from the melting/melt segregation depth in the deeper 636 mantle to the intrusion site. We have to evaluate this process in order to estimate the composition 637 638 of a primary melt generated in the mantle. Because fractionation of olivine, the most plausible 639 fractionated mineral, strongly affects Mg#, MnO, and NiO of olivine in equilibrium with the melt, these parameters are used as proxies for the extent of fractionation. We have to know the 640 641 mantle composition in order to adopt the proxies. The ranges of Mg#, NiO and MnO of olivine in 642 the upper mantle are known as the olivine mantle array estimated from compiled data of mantle xenoliths (Takahashi et al., 1987) (see Fig. S5). The NiO contents of olivine for a given value of 643 644 Mg# varies by ~0.08 wt%, which corresponds to the width of the array. We adopt the lowest edge of the array data field (e.g., NiO wt% = 0.36 at Mg#, 0.90). This is because olivine in the mantle 645 646 xenoliths analyzed by Takahashi et al. (1987) underwent various degrees of subsolidus Ni 647 exchange with pyroxenes and spinel, which increases NiO in olivine depending on the subsolidus

648	thermal history. The amount of fractionated olivine and the primary melt compositions for each
649	dike are estimated by repeatedly adding olivine phenocrysts with a chemical composition in
650	equilibrium with the melt in 0.1 wt% step until the olivine composition reaches the olivine mantle
651	array. We used Beattie et al. (1991) and Kinzler et al. (1990) for olivine-melt partition
652	relationships for Mg-Fe, Ni, and Mn. Starting melt compositions are the above estimated intruded
653	melt compositions (Table 5). The estimated amounts of fractionated olivine for the four dikes are
654	listed in Table 7. The major element compositions of the estimated primary melts are also listed
655	in Table 7. Even if we adopt the half width of the olivine mantle array as the uncertainty of the
656	mantle NiO (~0.04 NiO wt%), the error in the MgO of the primary melt is $< 1.0$ wt%.
657	The water and rare earth and trace element concentrations in the primary melts are also
658	corrected for olivine addition. The results are listed in Table 7 and plotted in Fig. 13. The
659	chondrite-normalized ratios of Dy and Yb ((Dy/Yb)n) of the primary melt of the four dikes are
660	1.22 - 1.40 and larger than the range of the compiled literature data of the present-day NMORB
661	and normal back-arc basin basalts (NBAB) (0.95 - 1.20) (see below for discussion on (Dy/Yb)n).
662	
663	Estimation of melting mechanisms and mantle source compositions

melting/melt segregation conditions: pressure, temperature, degree of melting/melt segregation,
and mantle source compositions. Melt segregation processes could also affect the composition.
We constrained the mantle source composition and the segregation processes on the basis of the
REE compositions of the primary melts estimated from the four dikes (DKSH1, DKSH2, RYO1,
and RYO2). They are modeled with a one-dimensional steady-state decompressional melting

model (Ozawa, 2001) using published partition coefficients (Table S2) (Fig. 14). The modeling 670 conditions are summarized in Table 8, and their complete lists are available in Table S3 - S6. 671 Careful examination of the accumulated melt compositions in the near fractional melting models 672 without any fluid influx for the four dikes show the followings. (1) Melting must have taken 673 place in the garnet stability field irrespective of whether the mantle source is the primitive mantle 674 (PM) or MORB mantle source (MM) in order to reproduce the weak decline of heavy rare earth 675 676 elements (HREE). The extent of melting in the garnet stability field ranges from 75 to 100% of the total melting, if a near fractional melting model with complete melt accumulation is assumed 677 (Table 8). (2) A depleted mantle source, such as a MORB mantle source (MM), is required in 678 order to better fit the pattern of the middle rare earth elements. (3) There are two possible models 679 to better reproduce light rare earth elements (LREE): instantaneous melt formed by near 680 fractional melting of the MORB mantle source or accumulated melt formed by fractional melting 681 of a slightly more depleted mantle source, such as the depleted MORB mantle source (DMM). 682 The modeling results are essentially the same for the four examined dikes (Fig. 14 and Table 8). 683 684 The CRUS of the Tectonite Member of the Miyamori complex was inferred to have 685 formed by fluxed fractional melting of the mantle wedge under extensive influx of a slab-derived fluid from the subducting slab strongly enriched in LREE and with anomalies of several trace 686 687 elements (Ozawa and Shimizu, 1995; Yoshikawa and Ozawa, 2007). On the other hand, the REE and trace element patterns of the ultramafic dikes, as well as rocks of the Cumulate Member, do 688 not indicate any involvement of such fluids at all. The K2O/TiO2 ratios of pargasite (Fig. 6), REE 689 690 modeling results (Fig. 14), initial Nd and Sr isotope ratios of the whole-rock and separated clinopyroxene of the ultramafic dikes and the Cumulate Member (Yoshikawa and Ozawa, 2007) 691 consistently show that they were derived from a depleted mantle, such as the MORB mantle 692

source, with insignificant involvement of any types of H<sub>2</sub>O/CO<sub>2</sub>-rich fluids or exotic melts. The
most plausible melting mechanism is, therefore, adiabatic decompressional melting, if the wide
occurrence of such MORB-type mantle material in the uppermost mantle (Cipriani et al., 2004;
Workman and Hart, 2005) is taken into consideration.

697

### 698 Estimation of melt segregation and solidus conditions and mantle potential temperatures

We estimated the melting/melt segregation pressure by plotting the estimated primary 699 melt compositions in the Ne'-Ol'-Qtz' ternary norm diagram (Ne'= nepheline + 0.6 albite, Ol'= 700 701 olivine + 0.75 orthopyroxene and Qtz'= Quartz + 0.4 albite + 0.25 orthopyroxene) according to the method of Irvine and Barager (1971) (Fig.15). The major elements compositions are strongly 702 controlled by pressure and temperature conditions but relatively weakly dependent on the source 703 composition, which is close to the depleted mantle in our case as discussed above from the 704 geochemical data. We extended the pressure isopleths constructed by Sakuyama et al. (2009) for 705 706 pressures < 3 GPa to higher pressures > 3 GPa on the basis of melting experiments made at 3 and 707 6 GPa (Walter, 1998; see Fig. S6). The normative values of the primary melt compositions of the dikes are listed in Table 7, and the estimated melting pressures are listed in Table 9. 708

There are barometers based on Si-activity models (Putirka, 2008; Lee et al., 2009). We applied the model of Putirka (2008) to melt compositions produced in experiments at 6 GPa by Walter (1998) and found that the barometers give lower pressure (4.1 - 4.9 GPa). Moreover, there is a negative correlation between the difference of experimental and calculated pressures ( $\Delta P$ ) and temperatures (e.g.,  $\Delta P = +1.1$  GPa at 1800 °C and +1.9 GPa at 1710 °C). Therefore, these barometers may not be applicable to pressures close to 6.0 GPa particularly for temperatures

lower than ~1700 °C, the condition of which can be realized by the decline of solidus owing to
the presence of water. We, therefore, use Fig. S6 to estimate melt segregation pressures.
Melting/melt segregation temperature is firstly estimated for the dry case by applying

thermometer based on MgO content in the primary melt (Sakuyama et al., 2014):

719 
$$T = 1080.7 + 54.75(-0.27074P^2 + 2.21634P - 0.99731) + 15.08MgO$$
 (1)

where *T* is the temperature in °C, *P* is the pressure in GPa, and MgO is MgO content of the melt in wt%. The estimates for the dry case are listed in Table 9. The presence of amphibole in the dikes, however, indicates an involvement of water in the melting. The water contents of primary melts constrained above are used to correct the effect of water on melting pressures according to Sakuyama et al. (2009) by the following formula after Tatsumi et al. (1983):

725 
$$P^{wet} = P^{dry} + H_2 O^{primary} \times \frac{0.5}{3.0}$$
 (2),

where  $P^{wet}$  and  $P^{dry}$  are melting pressures in the wet and dry cases, respectively, in GPa and *H*<sub>2</sub>*O*<sup>*primary*</sup> is the H<sub>2</sub>O content in the melt in wt%, and the effect on temperature according to Medard and Grove (2008):

729 
$$T^{dry} - T^{wet} = 40.4C_{H20}^{melt} - 2.97(C_{H20}^{melt})^2 + 0.0761(C_{H20}^{melt})^3$$
 (3),

where  $T^{wet}$  and  $T^{dry}$  are melting temperatures in the wet and dry cases, respectively, in °C, and  $C^{melt}_{H2O}$  is the H<sub>2</sub>O content in the melt in wt%. The melt segregation temperatures and pressures corrected for the water effects are 1416 - 1424 °C and 4.9 - 5.0 GPa, respectively, and are listed in Table 9. Degrees of melting were estimated for the four dikes by two methods. One is based on the dependence of Ne' on melting degree, which is constrained from melting experiments on depleted mantle compositions (0.28 - 0.30 Na<sub>2</sub>O wt%) with reports of melt fractions (Hirose and

736	Kushiro (1993) with the correction made by Hirschmann (2000); Walter, 1998; Laporte et al.,
737	2004; Fig. 16). This method gives 7.1 - 8.7 $\pm$ 3.5 % of degree of melting for the four dikes
738	(DKSH1, DKSH2, RYO1, and RYO2) (Fig. 16 and Table 9). The other is based on the
739	relationships between (Nd/Sm)n and (Dy/Yb)n using the open-system fractional melting model
740	(Ozawa, 2001; Tables S3-S6) and the partition coefficients listed in Table S2, and yields a degree
741	of melting of $13.7 - 20.0 \pm 3.8$ % (Table 9). The water concentrations in the mantle source thus
742	range from 0.04 $\pm$ 0.02 to 0.18 $\pm$ 0.07 wt% for a 7.1 - 8.7 % degree of melting, and from 0.11 $\pm$
743	0.02 to $0.31 \pm 0.08$ wt% for a 13.7 - 20.0 % degree of melting. It is expected that such amounts of
744	water present in the mantle source at $\sim$ 5 GPa is mostly partitioned into the melt once the solidus
745	is reached because of the extremely high solubility of the melt at such high pressure (e.g., $\sim 27$
746	H <sub>2</sub> Owt% at 3 GPa, Hodges (1974)), and that the melting degree does not increase much with
747	further decompression. The plausible ranges of the degree of melting and water content in the
748	source will be constrained below by comparing the solidus conditions with the water-preset
749	solidi.

From the melting/melt segregation conditions, we further estimated solidus conditions by assuming adiabatic (decompressional) melting, which is the plausible melting mechanism, as discussed above. We extrapolate the melt segregation conditions to the solidus by the melting degrees obtained above by the two methods and the latent heat of melting according to the following equation after Asimow et al. (1997):

755 
$$\left(\frac{\partial T}{\partial P}\right)_{S} = \frac{T\{\alpha^{S}V^{S} + F(\alpha^{lq}V^{lq} - \alpha^{S}V^{S})\} + \Delta H\left(\frac{\partial F}{\partial T}\right)_{P}\left(\frac{\partial T}{\partial P}\right)_{F}}{C_{p}^{S} + F\Delta C_{p} + \Delta H\left(\frac{\partial F}{\partial T}\right)_{P}}$$
(4)

where *T* is the temperature in K, *P* is the pressure in GPa, *F* is the melting degree,  $\Delta H$  is the enthalpy change in J/Mol,  $\alpha^s$ , and  $\alpha^{lq}$  are the thermal expansions for solid and melt, respectively,

758	in 1/K, $V^s$ , and $V^{lq}$ are the partial volumes for solid and melt, respectively, in J/GPa, and $Cp^s$ and
759	$Cp^{lq}$ are the isobaric heat capacities for solid and melt, respectively, in J/Mol/K. We used
760	expressions for thermal expansion, molar volume, heat capacity, and enthalpy of fusion
761	according to Lange and Carmichael (1987), Fei (1995), Berman and Brown (1985), Berman
762	(1998), Stebbins et al. (1984), Kojitani and Akaogi (1997). The calculated adiabatic gradients and
763	parameters used in the estimation procedures are listed in Table S7 after King (1957), Krupka et
764	al. (1985a), Robie et al. (1982a, 1982b), Haselton and Westrum (1980).
765	All the solidus conditions calculated by using the degrees of melting estimated by the two

767 (2003) (~0.08 wt%; Fig. S7). This range of water content is, however, more consistent with that 768 based on the degree of melting estimated from the Ne' norm ( $0.04 \pm 0.02 - 0.18 \pm 0.07$  wt% H<sub>2</sub>O

766

methods are plotted between the water-present solidi for 0.05 and 0.1 H<sub>2</sub>O wt% after Katz et al.

in the source) than that estimated from the REE contents  $(0.11 \pm 0.02 - 0.31 \pm 0.08 \text{ wt}\% \text{ H}_2\text{O})$ .

The higher melting degree based on the REE modeling may be attributed to a slightly more

depleted source or more effective melt separation. We thus accepted the former estimates for the solidus conditions more plausible, which range from 1440 -1460  $\pm$  30 °C and 5.3 - 5.4  $\pm$  1.0 GPa

(Table 9). The errors are estimated by propagating the largest uncertainty in the estimation of the

melting pressures and temperatures. The solidus temperatures are higher than those of melt

segregation by 20 to 25 °C according to Takahashi et al. (1993) (Table S7). The estimated solidus

pressure indicates that the melting initiated deeper than  $\sim$ 170 km. These deep melting conditions

are consistent with the conditions of high-pressure melting inferred from the REE patterns of

primary melts (> 75 % of the total melting in the garnet stability field >  $\sim$ 2.5 GPa).

The estimated mantle potential temperatures (MPTs) range  $1340 - 1360 \pm 40$  °C (Table 9) by extrapolating the solidus condition to the surface along the adiabatic slope of the solid mantle, 18 °C/GPa (Katsura et al., 2009). The errors are estimated by propagating the uncertainty in the
estimation of the solidus temperature and pressure.

783

# 784 Evaluation of openness of the melting systems

We successfully reproduced the REE pattern of the estimated primary melt of dike 785 DKSH2 by the near fractional melting model with the DMM source and without material influx 786 (Fig. 14). However, highly incompatible trace elements cannot be fitted reasonably well. The 787 estimated water contents of the mantle source (0.04 - 0.18 wt%), which are notably higher than 788 the MORB mantle source ( $\sim 0.02$  wt%: Dixon et al., 1998), indicate the presence of water as 789 hydrous phase before melting or the influx of a H<sub>2</sub>O-bearing agent during melting. Therefore, it 790 is quite plausible that a water-bearing agent enriched in highly incompatible elements was added 791 before or during melting processes. We estimated the ratios of the highly incompatible elements 792 (Ba, Sr, Nb, and Ta) of the influxed agent required to reproduce those of the primary melt of dike 793 DKSH2 by an open-system near fractional melting model (Ozawa, 2001) with the same model 794 795 parameters adopted to reproduce less incompatible trace elements above except for assuming material influx before or during decompressional melting. The required chondrite-normalized 796 797 ratios ((Ba/Nb)n, (Ba/Ta)n, (Sr/Nb)n, and (Sr/Ta)n) of the agent fall within a narrow range (2.1 -798 4.5) irrespective of influx timing. The ratios of partition coefficients between solid mantle and water-rich fluids at low temperature (< 900 °C), for pairs of Ba-Nb, Ba-Ta, Sr-Nb, and Sr-Ta, are 799 800 12 - 110, 24 - 350, 5.0 - 6.1 and 12 - 19, respectively, and are much higher than for those of a hydrous silicate melt or a supercritical fluid (0.9 - 2.1, 1.3 - 2.9, 0.7 - 1.6 and 1.0 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3 - 2.9, 0.7 - 2.2, 1.3, 1801 802 respectively, at 1200 °C) (Kessel et al., 2005). This implies that the water-rich agent should be a
hydrous silicate melt or a supercritical fluid. The latter is stable at a pressure higher than  $\sim 4$  GPa (Mibe et al., 2007), which is consistent with the estimated melting depth >  $\sim 130$  km.

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806

#### DISCUSSION

## 807 Tectonic model in the Cambrian-Ordovician northeast Japan

The tectonic environment of the decompressional melting of the depleted mantle from 808  $\sim$ 170 km depth involving a minor influx of supercritical fluid is not realized in a mantle wedge at 809 steady-state subduction because of the geochemical characteristics of the ultramafic dikes are 810 distinct from those of arc magmas. Neither back-arc basin spreading nor ridge subduction setting 811 are plausible because the their depths of melting are as shallow as 30 - 70 km (e.g., McKenzie 812 813 and Bickle, 1988; McKenzie and O'Nions, 1991; Kinzler and Grove, 1992; Forsyth, 1993; Hirschmann and Stolper, 1996; Yamashita and Tatsumi, 1994). The flux melting of the mantle 814 wedge forming the CRUS of the Tectonite Member had occurred in the Cambrian-Ordovician 815 northeast Japan just before decompressional melting of the depleted mantle forming the 816 Cumulate Member and the ultramafic dikes. Such sudden change of the melting regime to 817 adiabatic decompressional melting of the depleted mantle at the depth of ~170 km is best 818 explained by a slab breakoff model. 819

A slab breakoff model invokes a detachment of a subducting oceanic plate which induces upwelling of asthenosphere materials from above or beneath the subducting plate (Davies and von Blanckenburg, 1995). In the former case, distinct geochemical features of the mantle wedge must be inscribed in the primary melt, but it is not the case for the Cumulate Member and the ultramafic dikes as discussed above. Therefore, mantle upwelling must be dominated from the

bottom side of the plate. This is consistent with a result of geodynamic modeling of slab breakoff 825 (Gerya et al., 2004), where a substantial upwelling takes place from beneath the subducting plate. 826 Ozawa et al. (2015) proposed a model that the Cumulate Member was crystallized from 827 melts formed by decompressional melting of a sub-slab mantle triggered by a slab breakoff 828 during the Cambrian-Ordovician time in northeast Japan Arc. The slab breakoff model 829 constrained in our study is not in conflict with this model. Figure 17 shows a tectonic model 830 based on Ozawa et al. (2015) with our proposition of generation and transportation of the primary 831 832 magma frozen as the ultramafic dikes in the Cumulate Member.

833

# 834 Thermal state of the Cambrian-Ordovician upper mantle

The present-day MPT varies from  $1370 \pm 49$  °C for mid-ocean ridges to  $1620 \pm 55$  °C for 835 ocean island basalt (Putirka, 2005). The former magma generation is induced by passive 836 upwelling, and the latter by active upwelling. Thus, the wide variation of the present-day MPT 837 838 implies the diversity in the depths of mantle upwelling. It should be noted that the estimated MPTs are within the range of 1350 - 1450 °C if the magma generation mechanism is restricted to 839 passive upwelling regardless of whether it occurs beneath mid-ocean ridges or back-arc basins 840 (Wiens et al., 2006). This implies that the upper mantle is globally in a fairly homogeneous 841 thermal state, which might be maintained by small-scale convection underneath the plates 842 (Richter and Parsons, 1975; van Hunen and Čadek, 2009; Takeo et al., 2018). 843

We have estimated the Cambrian-Ordovician MPT of ~1350 ± 40 °C in the northeast
Japan arc from ultramafic dikes with the specification of melting mechanism as decompression
melting of MORB mantle source from the sub-slab mantle. Therefore, the estimated MPT reflects

847	the global thermal state of the upper mantle because the thermal state of sub-slab mantle is
848	similar to the mantle beneath mid-ocean ridges provided that small-scale convection is
849	maintained beneath the oceanic lithosphere. We conclude that the thermal state of the upper
850	mantle in Cambrian-Ordovician time was globally similar to that of the present-day upper mantle.
851	

## 852 IMPLICATIONS: ORIGIN OF THE CAMBRIAN-ORDOVICIAN OPHIOLITE PULSE

Cambrian-Ordovician ophiolites classified as subduction-related according to the Th/Yb-853 Nb/Yb discrimination diagram of Pearce (2008) have high Nb/Yb ratios above the MORB-OIB 854 855 array (Dilek and Furnes, 2011; Furnes et al., 2014). These ophiolites, however, plot mostly in the MORB field in the V-Ti discrimination diagram of Shervais (1982) (Dilek and Furnes, 2011; 856 Furnes et al., 2014). This implies that primary magmas for such ophiolites formed mostly by 857 decompressional melting of a MORB mantle source under negligible influence of slab-derived 858 fluids. The plausible mechanisms to reproduce such features have been thought to be either back-859 860 arc spreading (Dilek and Furnes, 2011; Furnes et al., 2014) or ridge subduction (Shervais, 2001). We compiled data of volcanic or shallow-intrusive basaltic rocks from the Cambrian-861 Ordovician subduction-related ophiolites and examined the relationship between the chondrite-862 normalized ratio of Dy and Yb ((Dy/Yb)n) and that of Nd and Sm ((Nd/Sm)n). They are 863 compared with primary melts formed by steady-state decompressional melting (Ozawa, 2001) of 864 a MORB mantle source that ascent from various depths with a variable extent of melting in the 865 garnet peridotite field, which is represented by percentages of melting in the garnet stability field 866 in the total melting degree (PMGF) (Figs. 18a and 18b). The data from the subduction-related 867 Cambrian-Ordovician ophiolites have a similar range of (Nd/Sm)n as that of the present-day 868

869	NMORB and NBABB, and a wide range of (Dy/Yb)n (0.97 - 1.39) with many data higher than
870	(Dy/Yb)n of the present-day NMORB and NBABB (1.22 - 1.40; Figs. 18a and 18b). The
871	predominance of magmas that have (Nd/Sm)n and (Dy/Yb)n ratios within the range of the
872	present-day NMORB and NBABB suggests that the thermal state of the Cambrian-Ordovician
873	upper mantle was not distinct from that of the present-day upper mantle. This is consistent with
874	our estimation of MPT from the ultramafic dikes. We further infer that these NMORB or
875	NBABB-like magmas were generated from the MORB mantle source at depths similar to the
876	present-day spreading center (30 - 70 km, Kinzler and Grove (1992)), though the generation
877	depths are not estimated for these ophiolites.
878	The (Dy/Yb)n from some of the ophiolites are as high as that of the magmas filling
879	ultramafic dikes in the Hayachine-Miyamori ophiolite (Dy/Yb)n (> 1.2) with $(Nd/Sm)n < 1.0$
880	(Fig. 18a). Such ophiolites are: Bymarka, Norway (Slagstad, 2003); Solund-Stavfijord, Norway
881	(Furnes et al., 2012); Star Lake, Canada (Lissenberg et al., 2005); Cabo Oltegal Complex, Spain
882	(Sánchez-Martínez et al., 2007); Kudi, China (Fang, 1998; Jiang et al., 1992; Wang et al., 2002).
883	The presence of magmas characterized by the high $(Dy/Yb)n (> 1.2)$ in these ophiolites indicates
884	partial melting of MORB mantle source at depth as deep as ~160 km. It is speculated that these
885	magmas were generated by adiabatic decompression of sub-slab mantle caused by slab breakoff,
886	though other magmas were generated at depths as shallow as those of MORB either through
887	back-arc spreading or ridge subduction. This might have an implication on the Cambrian-
888	Ordovician ophiolite pulse.

An appropriate geodynamic model for the ophiolite pulse must explain the following geological and petrological features in the Cambrian-Ordovician time. First, the MPT of the Cambrian-Ordovician upper mantle is similar to that of the present-day upper mantle. Second,

40

892	LIPs and subduction-unrelated ophiolites are scarce in contrast with the Jurassic-Cretaceous
893	ophiolite pulse (Ernst and Buchan, 2003). Third, U-Pb zircon ages that represent ages of acidic
894	magmatic activity have a peak at 490-530 Ma (Hawkesworth et al., 2009). The first feature
895	suggests that the thermal effect on the surface tectonics was the same as those of the present-day
896	plate tectonics. The second feature cannot be explained by superplume models (Ishiwatari, 1994;
897	Dilek, 2003; Vaughan and Scarrow, 2003). The third feature suggests that the Cambrian-
898	Ordovician ophiolite pulse corresponds to the period of the assembly of the Gondwana
899	supercontinent (Hawkesworth et al., 2009).

A subduction termination is accompanied by a slab breakoff and ophiolite exhumation 900 and is highly expected during arc-continent collision which induces passive upwelling of the sub-901 902 slab mantle from at least ~200 km depth (Davies and von Blanckenburg, 1995; Gerya et al., 903 2004). We propose a new model for the origin of the Cambrian-Ordovician ophiolite pulse, 904 which may provide one of the alternative mechanisms for termination of one cycle of subduction 905 as proposed by Shervais (2001) by invoking ridge subduction. The frequent occurrence of 906 subduction termination followed by slab breakoff was the principal cause of the ophiolite pulse, 907 which led to the formation of the Gondwana supercontinent. The slab breakoff induced the decompressional melting of the sub-slab mantle and subsequent obduction of the ophiolites. The 908 909 mantle potential temperatures of Cambrian-Ordovician ophiolites are similar to the present-day value, but magmas were generated under a similar degree of melting by passive upwelling at 910 various depths from similar to the present-day mid-ocean ridge and back-arc basin environments 911 912 to the depth ~170 km (Figs. 18a and 18b). This implies that the water content of the shallow Cambrian-Ordovician upper mantle was similar to the present-day value, but that the water 913 content at depth  $\sim 170$  km or deeper beneath the subducting slab was as high as 0.04 - 0.18 wt%. 914

915 916 ACKNOWLEDGEMENTS We are grateful to H. Yoshida for the analytical support and S. Sasaki for kind help in the 917 field survey. We thank J. Wakabayashi and K. Putirka for their thoughtful and constructive 918 reviews and S. Straub for her editorial efforts. This work was supported by Grant-in-Aid for JSPS 919 Research Fellow (Grant Number 17J06812) and JSPS KAKENHI (Grant Number 17H02982). 920 921 **REFERENCES CITED** 922 Abbate, E., Bortolotti, V., Passerini, P., and Principi, G. (1985) The rhythm of Phanerozoic 923 ophiolites. Ofioliti, 10, 109-138. 924 925 Arevalo Jr., R. and McDonough, W.F. (2008) Tungsten geochemistry and implications for understanding the Earth's interior. Earth and Planetary Science Letters, 272, 656-665. 926 927 Arevalo Jr., R., McDonough, W.F., and Luong, M. (2009) The K/U ratio of the silicate Earth: insights into mantle composition, structure and thermal evolution. Earth and Planetary 928 Science Letters, 278, 361-369. 929 930 Asimow, P.D., Hirschmann, M.M., and Stolper, E.M. (1997) An analysis of variations in isentropic melt productivity. Philosophical Transactions of the Royal Society of London. 931 Series A: Mathematical, Physical and Engineering Sciences, 355, 255-281. 932 Bagnold, R.A. (1954) Experiments on a gravity-free dispersion of large solid spheres in a 933 Newtonian fluid under shear. Proceedings of the Royal Society of London. Series A, 225, 934 49-63. 935

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1256

## 1257 Figure Captions

Fig. 1: (a): Simplified geological map of the northwestern part of southern Kitakami Mountains 1258 showing distributions of the Miyamori and Hayachine ultramafic-mafic complexes in the 1259 Hayachine-Miyamori (HM) Ophiolite, Ordovician-Devonian strata, and Silurian-Devonian 1260 granitic rocks (after Ehiro and Suzuki, 2003; Kawamura at al., 2013; Ozawa, 1984; Maekawa, 1261 1262 1981). Gray rectangle in (a) represents the plot region of (b). (b): Geological map of the southern 1263 part of the Miyamori ultramafic-mafic complex located to the east of the Hizume-Kesennuma 1264 fault (modified after Ozawa et al., 2013). Localities of examined ultramafic dikes are shown by filled stars with locality numbers in (b). The dikes occur in the Cumulate Member near the 1265 1266 contact with the Tectonite Member.

1267

1268 Fig. 2: (a) Field occurrence of the ultramafic dike (DK) in the host dunite (HD) at locality #1 in

1269 Fig. 1b. The inset in the upper left corner shows an outline of the dike. This dike shows

- 1270 porphyritic texture with olivine (dark spots) and minor clinopyroxene phenocrysts set in the
- 1271 matrix. It shows zoning in phenocryst modes; olivine phenocrysts are abundant in the central

1272	darker zone and fewer in the brighter marginal zones. The dike has short branches, indicated with
1273	arrows. (b): A view of a polished slab taken from the central part of the dike (a rectangle marked
1274	with b in (a)) showing abundant olivine phenocrysts (~ 50 vol%; Table 1) with euhedral-
1275	subhedral morphology. (c): A view of a polished slab in the marginal part of the dike (a rectangle
1276	marked with c in (a)) showing a smaller amount of phenocrysts (~20 vol%; Table 1) and their
1277	shape-preferred orientation (outlined with white dashed lines) elongate parallel to the contact. A
1278	few millimeter-thick branch, shown with a white arrow, is present.
1279	
1280	Fig. 3: Photomicrograph of thin sections made from ultramafic dikes (DKSH2 and RYO1): (a)
1281	Wide view of the center of DKSH2 in an open nicol, (b) magnified view of the area between an
1282	olivine phenocryst and a plagioclase pseudomorph in RYO1 under open nicol, and (c) that under
1283	crossed nicols. In (a), olivine (ol) phenocrysts are euhedral to subhedral and set in the matrix
1284	consisting mostly of polycrystalline amphibole (amph). Orthopyroxene (opx), partially (or

totally) altered into darker fine-grained aggregate, occur rimming olivine, and plagioclase 1285 1286 pseudomorph (pl) occur in the middle of the amphibole matrix. Euhedral-subhedral chromian spinel (sp) grains are included in olivine or amphibole. In (b) and (c), orthopyroxene is mostly 1287

unaltered and occur rimming the olivine phenocryst. 1288

1289

Fig. 4: Whole-rock Al<sub>2</sub>O<sub>3</sub> and Fe<sub>2</sub>O<sub>3</sub> contents in wt% of examined samples of dikes plotted 1290 1291 against MgO wt%. Triangle, circle, square, and diamond indicate examined dikes: DKSH1, DKSH2, RYO1, and RYO2, respectively (Fig. 1b; Table 1). Symbols with a dot, vertical line, 1292 plus, and cross denote samples from the edge, margin, and center of dikes, and rootless block 1293

1294	without information on sample position in the dike, respectively. Large symbols indicate weakly
1295	altered, and small symbols indicate samples showing intermediate or strong alteration. The solid
1296	lines connect two pairs of samples with different degrees of alteration, but the other features,
1297	such as sampling location in the dike and igneous modal abundance, are the same.
1298	
1299	Fig. 5: CI-chondrite normalized whole-rock trace element and rare earth element (REE) patterns
1300	of 10 samples from 4 dikes (localities, #1, #2, #3, and #4; Fig. 1b, Table 1). Six samples are from
1301	dikes with homogeneous phenocryst distribution, shown in (a) and (c), and four samples from
1302	dikes zoned in terms of phenocryst abundance, shown in (b) and (d). The CI-chondrite
1303	composition is after McDonough and Sun (1995). Symbols are the same as in Fig. 4. Shaded
1304	symbols with a dashed line indicate samples with a weak alteration.

1305

Fig. 6: Present-day values of <sup>87</sup>Sr/<sup>86</sup>Sr and <sup>143</sup>Nd/<sup>144</sup>Nd of whole-rocks and separated 1306 clinopyroxene phenocrysts for examined 4 ultramafic dikes (a) and other mafic and ultramafic 1307 rocks in the Hayachine-Miyamori ophiolite from the literature (b). Symbols and paring lines in 1308 (a) are the same as those used in Fig. 4 except for star in the square, which denotes clinopyroxene 1309 separated from a sample indicated by arrows. Plotted in (b) are literature data for pyroxenes 1310 separated from the chromite-bearing Ultramafic Suite (CRUS) (solid inverted triangle), 1311 aluminous spinel-bearing Ultramafic Suite (ASUS) (open inverted triangle) (Yoshikawa and 1312 Ozawa, 2007), the Cumulate Member of the Miyamori complex (thick cross) (Ozawa et al., 1313 2015) and the hornblende-rich mafic-ultramafic rocks (cross) (Shibata and Ozawa, 1992). The 1314

60

rectangle in (b) represents the plot region of (a). The area for mid-ocean ridge basalt (MORB) in 1315 (b) is according to the compilation by Yoshikawa and Nakamura (2000). 1316

1317

Fig. 7: Spatial variations of Mg# of minerals (a) and NiO wt% in olivine (b) crossing the contact 1318 between ultramafic dikes and host dunite for DKSH2 (locality #2; Fig. 1b). The distance is 1319 measured from the boundary taken positive towards the center of the dike and negative towards 1320 1321 the dunite host. Vertical lines indicate the positions of the dike-host boundary, where the igneous modal abundance abruptly changes. Horizontal lines indicate the lowest grain-averaged values of 1322 the central homogeneous part of DKSH2, which are 0.833 in Mg# and 0.20 wt% in NiO. 1323 1324 Fig. 8: Chemical compositions of spinel plotted in the triangular diagram of Cr, Al, and  $Fe^{3+}$  in 1325 examined four ultramafic dikes. Circle, square, triangle, and diamond are the same as those used 1326 in Fig. 4. Open and solid symbols indicate euhedral-subhedral grain included in 1327 olivine/amphibole and anhedral grain occurring near plagioclase, respectively. Cross indicates 1328 1329 spinel in the host dunite.

1330

Fig. 9: K<sub>2</sub>O/TiO<sub>2</sub> ratio of amphibole in wt% forming the matrix of examined ultramafic dikes. 1331 1332 The data fields for amphibole from the aluminous Spinel Ultramafic Suite (ASUS), the Chromite-

1333 bearing Ultramafic Suite (CRUS), and the Cumulate Member (Cumulate) of the Hayachine-

1334 Miyamori ophiolite (Ozawa, 1988) are also shown. The ranges of magma compositions from

various tectonic settings are after Ozawa (1988). Circle and diamond are the same as those used 1335

1336 in Fig. 4. The data of amphibole in the ultramafic dikes are plotted in the field of normal mid-

61

1337	ocean ridge basalt. Abbreviations: NMORB, normal mid-ocean ridge basalt; OIPB, oceanic intra-
1338	plate basalt; BABB, back-arc basin basalt; EMORB, enriched mid-ocean ridge basalt; IAB,
1339	island arc basalt; CFB, continental flood basalt; CRB, continental rift basalt; CAB, continental
1340	alkali basalt; KIMB, kimberlite; HMA, high magnesian andesite; LAMP, lamproite.
1341	
1342	Fig. 10: CI-chondrite normalized trace element patterns of clinopyroxene and amphibole (a) and
1343	rare earth element (REE) pattern of plagioclase pseudomorph (fine-grained aggregate mostly
1344	consisting of grossular) (b) in weakly altered dike DKSH2. These minerals consist of the matrix
1345	of olivine and clinopyroxene phenocrysts. The ranges of 7 analyses of amphibole from 2 samples
1346	(DKSH2MW and DKSH2CW), 5 analyses of clinopyroxene from 2 samples (DKSH2MW,
1347	DKSH2CW), and 4 analyses of plagioclase pseudomorph from 1 sample (DKSH2CW) are shown
1348	by filled or shaded areas. The REE pattern of plagioclase pseudomorph in (b) has a remarkable
1349	positive Eu anomaly and depletion in heavy REEs, which confirms the plagioclase origin of the
1350	aggregate. The CI-chondrite composition is after McDonough and Sun (1995).
1351	
1352	Fig. 11: Whole-rock Al <sub>2</sub> O <sub>3</sub> (a), MgO (b), and Fe <sub>2</sub> O <sub>3</sub> (c) contents of two pairs of samples from
1353	dike DKSH2 plotted against their loss on ignition (LOI). Each pair, connected by a solid line,
1354	was sampled within ~4 cm distant in the same dike and has the same igneous modal composition
1355	and texture but a difference in alteration degree (Table 1). Symbols with a vertical line and a
1356	cross indicate margin and center of the dike, respectively. Data with large symbols are from less
1357	altered, and those with small symbols are from more altered samples. The gray vertical bands

indicate the range of initial water content (1.31 - 1.36 wt%) estimated from the water contents of

1359	amphibole and the igneous modal abundances of each sample. Correction of alteration was made
1360	by linearly extrapolating the tie lines to the expected initial water content as shown by the dashed
1361	lines. The restored original major elements contents (open circles) and the mean slopes of the tie
1362	lines are listed in Tables 4 and S1, respectively.

1363

Fig. 12: Alteration-corrected whole-rock Al<sub>2</sub>O<sub>3</sub> (a) and Fe<sub>2</sub>O<sub>3</sub> (b) contents plotted against MgO for nine samples of the examined four dikes. Symbols are the same as those used in Fig. 3. The restored compositions form linear trends.

1367

Fig. 13: Chondrite-normalized trace element patterns in primary melts being in equilibrium with mantle peridotite estimated for the four dikes (a), and the range of chondrite-normalized rare earth element (REE) pattern (b). The REE patterns of mid-ocean ridge basalts and ocean island basalts after Sun and McDonough (1989) are shown in (b) for comparison. Symbols in (a) are the same as those used in Fig. 4. All trace element and REE data are after correction of the influence of secondary modifications.

1374

Fig. 14: Comparison of the chondrite-normalized rare earth element (REE) patterns of modeled and observed (solid lines with symbols; see Fig. 4) of the primary melts in equilibrium with mantle peridotite for four dikes (DKSH2, (a); DKSH1, (b); RYO1, (c); and RYO2, (d)). The adopted models are based on a one-dimensional steady-state decompressional melting model of Ozawa (2001) with adjustment of model parameters to best reproduce the observation. Effects of two of the most critical model parameters, mantle source composition and strength of melt

63

1381	separation, are shown for accumulated melt and primitive mantle, thick solid line; accumulated
1382	melt and MORB mantle source, solid line; accumulated and depleted MORB mantle source,
1383	dotted line; and instantaneous melt and MORB mantle source, dashed line (Table 8). The
1384	complete lists of optimized model parameters are listed in Tables S3, S4, S5, and S6 for the
1385	examined four dikes. The REE abundances in the primitive mantle are after McDonough and Sun
1386	(1995), and those of the mantle source of mid-ocean ridge basalt (MORB) and depleted MORB
1387	are after Workman and Hart (2005). The CI-chondrite composition is after McDonough and Sun
1388	(1995).

1389

Fig. 15: Normative composition of the primary melts estimated from 4 dikes (Table 7; DKSH1, 1390 open triangles; DKSH2, open circles; RYO1, open squares; and RYO2, open diamonds) plotted 1391 1392 in the ternary diagram of Ne'- Qtz'- Ol'. Pressure isopleths defining the compositions of partial melts of the anhydrous system are after Sakuyama et al. (2009) for the low-pressure region ( $\leq 3.0$ 1393 GPa). For high-pressure above 3.0 GPa, the isopleths are based on data of the high-pressure 1394 melting experiment of Walter (1998) (see Fig. S6). The projection scheme is after Irvine and 1395 Barager (1971): Ne'= Nepheline + 0.6Albite; Qtz'= Quartz + 0.4 Albite + 0.25 Orthopyroxene; 1396 Ol'=Olivine + 0.75 Orthopyroxene. 1397

1398

1399 Fig. 16: Relationship between normative values of Ne' of primary melt  $(C_m)$  and degree of

- 1400 melting (f) for a given Na<sub>2</sub>O content of mantle source ( $C_s$ ) at two pressure ranges (1 3 GPa,
- 1401 dashed lines, and 4 6 GPa, dotted lines). The dashed lines ( $C_m = aC_s / (0.04+0.96f, a=11)$ ) are
- 1402 based on the experiments at 1 3 GPa (Laporte et al, (2004); Hirose and Kushiro (1993)

1403	recalculated according to Hirschmann (2000)) and the dotted lines ( $C_m = aC_s / (0.1 + 0.9f, a = 11)$
1404	are based on experiments at 4 - 6 GPa (Walter, 1998), The gray region represents the range for
1405	the MORB mantle source (0.28 - 0.3 Na <sub>2</sub> O wt%: Workman and Hart, 2005; Takahashi, 1986) at 3
1406	- 6 GPa. Open and solid symbols connected with each other represent Na <sub>2</sub> O content of the
1407	primary melts and estimated degrees of melting for four dikes (DKSH1, triangle; DKSH2, circle;
1408	RYO1, square; and RYO2, diamond).
1409	
1410	Fig. 17: Slab breakoff model in the latest stage of the Hayachine-Miyamori ophiolite evolution
1411	during the Cambrian-Ordovician time constrained by melt generation depths, mantle source
1412	compositions, and melting mechanism for the primary melt, the composition of which was

1412 compositions, and melting mechanism for the primary melt, the composition of which was
1413 estimated from the four ultramafic dikes in this study. Passive upwelling (thick arrow) of the sub1414 slab mantle from >170 km depth is assumed to have been triggered by slab breakoff (Ozawa et
1415 al., 2015). Thin gray arrow shows melt transport through a fast path. Abbreviations: E, east; W,
1416 west.

1417

Fig.18: Relationship between chondrite normalized ratio of Nd and Sm ((Nd/Sm)n) and that of
Dy/Yb ((Dy/Yb)n) for the ultramafic dikes (Table 5) of the Hayachine-Miyamori ophiolite as
stars with errors and the other Cambrian-Ordovician subduction-related ophiolites, which are (a)
Bymarka (Slagstad, 2003); Solund-Stavfijord (Furnes et al., 2012); Kudi (Fang, 1998; Wang et
al., 2002); Cabo Oltegal Complex (Sánchez-Martínez et al., 2007); Star Lake (Lissenberg et al.,
2005), and (b) Bay of Islands (Kurth et al., 2005); Jiugequan (Xia and Song, 2010);

1424 Annieopsquotch (Lissenberg et al., 2005). Also shown are trends of accumulated partial melts

1425	produced by fractional melting of the MORB mantle source (Workman and Hart, 2005). The
1426	melting model assumed steady-state decompressional melting with continuous melt separation
1427	starting from various depths of the garnet stability field (Ozawa, 2001). The adopted melt
1428	separation rates ( $\gamma$ ) are 0.99, 0.96, 1.00, and 1.00 for the reaction extents ( $t$ ) of 0.001, 0.02, 0.15
1429	and 0.2, respectively. Solid curves represent the weight percentage of the degree of melting in the
1430	garnet stability field in the total melting (PMGF), and the dashed curves represent the degree of
1431	melting in weight fraction (F). Compiled literature data of the normal mid-ocean ridge basalts
1432	(NMORB) (Arevalo Jr. and McDonough, 2008; Arevalo Jr. et al., 2009; Davis et al., 2008; Hall
1433	et al., 2006; Mahoney et al., 2002; Nauret et al., 2006; Sims et al., 2002; Sun et al., 2003; Sun et
1434	al., 2008) and those of normal back-arc basin basalts (NBABB) (Arevalo Jr. and McDonough,
1435	2008; Pearce et al., 2005; Pearce et al., 1994) are also shown for comparison. The CI-chondrite
1436	composition is after McDonough and Sun (1995).

Dike		DKSH1		DKSH2					RYO1	RYO2
Phenocryst		. 1								π <del>-1</del>
distribution		zoned '				homogeneous	zoned '			
Position Sample	edge DKSH1ES	margin DKSH1MI	center DKSH1CI	edge DKSH2EI	DKSH2MI	Irgin DKSH2MW	Ce DKSH2CI	nter DKSH2CW	rootless rock RYO1I	rootless rock RYO2I
(wt%) SiO <sub>2</sub>	40 14	41 63	42 59	42 86	42 76	42 62	42 81	42 53	43 41	41 22
TiO₂	1 07	0.44	0.45	0.68	0.69	0.66	0.76	0.62	0.56	0.56
Al <sub>2</sub> O <sub>2</sub>	6.55	5.90	4.32	4 24	5 27	7 50	4 64	7 39	7 43	8 42
Fe <sub>2</sub> O <sub>2</sub>	13.82	12 24	11.83	11 23	11 17	11.50	11.38	11 43	9.98	9.32
MnO	0.15	0.19	0.23	0.19	0.18	0.17	0.19	0.18	0.18	0.21
MgO	25.24	29.75	31.67	32.38	30.31	27.99	30.97	28.13	22.25	19.38
CaO	13.01	9.64	8.86	8.39	9.43	8.61	9.15	8.79	15.94	20.86
Na <sub>2</sub> O	0.01	0.16	0.03	0.00	0.14	0.83	0.06	0.81	0.23	0.00
K <sub>2</sub> O	0.01	0.03	0.02	0.02	0.03	0.11	0.03	0.11	0.01	0.00
$P_2O_5$	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.03
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
Loss on ignition	0.20	5.22	5.40	0.35	5.43	2.72	6.10	3.02	4.39	5.87
Ni	599	927	983	943	919	836	931	780	498	379
Cr	1406	1588	1729	1570	1457	1376	1534	1277	1529	1371
Ва	96	110	101	98	119	112	104	133	94	93
Nb	0.04 1 22	0.01	0.08	0.02	0.04 1 NQ	0.13	0.04	0.08 1 11	0.02	0.01
Та	0.07	0.40	0.03	0.05	0.07	0.07	0.09	0.07	0.06	0.05
La	1.68	1.44	1.48	0.93	1.07	1.07	1.40	1.06	1.05	0.74
Ce	4.08	3.17	3.11	2.90	3.25	3.15	4.05	3.08	2.78	2.59
Pr	0.81	0.52	0.48	0.58	0.61	0.62	0.76	0.59	0.53	0.48
Sr	309	146	146	175	195	198	164	223	87	70
Nd	4.90	2.61	2.53	3.66	3.97	4.00	4.40	3.58	3.37	3.18
Hf	0.91	0.47	0.49	0.63	0.68	0.68	0.76	0.62	0.68	0.58
Zr	20.45	13.20	13.73	14.99	16.45	17.20	20.13	16.44	16.89	14.47
Fu	0.71	0.82	0.87	0.40	0.50	0.52	0.63	0.48	0.44	0.50
Gd	2.58	1.27	1.26	1.71	1.84	1.77	2.07	1.69	1.72	1.69
Tb	0.45	0.21	0.20	0.28	0.31	0.30	0.34	0.27	0.28	0.30
Dy	3.03	1.48	1.54	1.97	2.07	2.01	2.42	1.87	1.92	2.09
r Ho	0.62	7.80	8.1Z 0.31	9.99	0.42	0.40	0.48	9.63	0.40	9.94 0.44
Er	1.73	0.88	0.90	1.13	1.20	1.15	1.38	1.06	1.17	1.17
Tm	0.24	0.12	0.13	0.15	0.17	0.16	0.19	0.15	0.15	0.15
Yb	1.55	0.83	0.83	1.01	1.12	1.08	1.27	0.97	0.98	1.00
LU	0.22	0.12	0.12	0.15	0.16	0.16	0.18	0.15	0.14	0.15
Modal composit Phenocrysts	ion (vol%)									
olivine	14.00	25.67	31.19	34.83	25.31		27.18		14.20	7.65
orthpyroxene	3.50	14.01	11.76	2.52	13.07		10.95		9.03	12.24
clinopyroxene	2.46	1.38	2.75	-	00.00		00.40		7.53	4.12
l otal Matrix	19.96	41.06	45.70	37.35	38.38		38.13		30.76	24.01
amphibole	36.77	41.49	41.14	60.66	56.50		54.52		29.79	18.82
orthovroxene	36.01	14.20	7.61	0.13 1.32	1.87		2.34		15.71	24.41 7 17
plagioclase	-	1.72	2.69	-	2.02		2.40		18.41	25.29
spinel	7.26	1.52	2.87	0.53	1.23		2.61		1.42	0.29
Total	80.04	58.94	54.30	62.65	61.62		61.87		69.24	75.99
Extent of replac	ement by see	condary low-t	emperature m	inerals <sup>3</sup>	intermediate	week	intormadict	weak	intermediate	intormediat-
Phenocrysts	suong	mennediate	mennediate	IIIIeIIIIediate	mennediate	weak	mennediate	weak	mermediate	mennediaté
olivine	×	0	0	0	0	0	0	0	Δ	Δ
orthpyroxene	×	×	×	×	×	Δ	×	$\Delta$	0	0
clinopyroxene	Δ	0	0	-	-	-	-	-	0	0
amphibole	×	×	×	×	×	0	×	0	^	٨
clinopyroxene	Δ	0	0	0	0	0	0	0	0	0
plagioclase	-	×	×	×	×	×	×	×	×	×
spinel	$\triangle$	0	0	0	0	0	0	0	0	0

Table 1: Whole-rock chemical compositions, igneous modal compositions, alteration degrees of examined ultramafic dikes.

<sup>1</sup>Phenocrysts are concentrated in the dike center.

<sup>2</sup> Orthopyroxene always occurs rimming olivine phenocrysts.

 $^3$  O, weak (< 20 %);  $\triangle$ , intermediate (20 - 60 %); ×, extensive (60 - 100 %).

Dike	Sample	<sup>87</sup> Rb/ <sup>86</sup> Sr	<sup>147</sup> Sm/ <sup>144</sup> No	d <sup>143</sup> Nd/ <sup>144</sup> No	ł	<sup>87</sup> Sr/ <sup>86</sup> Sr		Sr <sub>l</sub> 1	εNd <sub>l</sub> 1
					2σ (10 <sup>-6</sup> )		2σ (10 <sup>-6</sup> )		
Whole rock									
두	DKSH1ES	0.0023	0.224	0.513031	4.7	0.703511	9.8	0.703495	6.0
4SF	DKSH1MI	0.0063	0.199	0.512975	7.9	0.703432	12.8	0.703387	6.4
ā	DKSH1CI	0.0026	0.200	0.512986	9.1	0.703458	12.4	0.703440	6.6
	DKSH2EI	0.0013	0.202	0.513020	5.8	0.703498	12.0	0.703489	7.1
<b>9</b>	DKSH2MI	0.0019	0.204	0.513021	5.9	0.703432	14.1	0.703419	7.0
4SF	DKSH2MW	0.0070	0.212	0.513021	5.3	0.703219	9.0	0.703169	6.5
ā	DKSH2CI	0.0027	0.204	0.513009	5.7	0.703456	8.8	0.703437	6.8
	DKSH2CW	0.0091	0.211	0.513013	5.8	0.703234	10.8	0.703169	6.4
RY01	RY01I	0.0121	0.215	0.513047	6.6	0.703693	9.9	0.703607	6.8
RYO2	RYO2I	0.0177	0.225	0.513079	7.6	0.703955	15.6	0.703829	6.8
Separated clinopyroxene									
RY01	CPXRY01I	0.0075	0.228	0.513055	6.8	0.703701	12.4	0.703648	6.1

Table 2: Sm–Nd and Rb–Sr isotopic compositions of selected samples from four ultramafic dikes (see Table 1).

<sup>1</sup> Sr<sub>1</sub> and Nd<sub>1</sub> are initial <sup>87</sup>Sr/<sup>86</sup>Sr and initial <sup>143</sup>Nd/<sup>144</sup>Nd, respectively at 499Ma (Ozawa et al., 2015).

Table 3: Results of examination of mass balance of MgO between the dike (DKSH2) and its host dunite from the Mg# profile crossing their contact shown in Fig. 7. The mass of MgO lost from the host with the initial Mg# = 0.9, which is the homogeneous Mg# of olivine in the host away from the contact, and the mass gained by the dike with assumed initial Mg# = 0.83, which is the mean Mg# at the dike center, are calculated. If the former is given as unity, the latter is calculated to be 1.15, which implies that the diffusional Mg-Fe exchange between the dike and the host took place almost in a closed system under subsolidus condition.

	host dunite	dike (DKSH2)
Observed zoning profile		
variation width (mm)	50	54
area ratio of the profile	0.31	0.69
Mg#		
mean of all analyses	0.90	0.84
mean at a flat segment far away from the contact	0.90	0.83
Modal abundance (wt%)		
olivine	99.9	37.0
amphibole	-	58.3
clinopyroxene	-	0.1
orthopyroxene	-	3.9
MgO mass balance		
olivine	1.00	0.75
amphibole	-	0.36
clinopyroxene	-	0.00
orthopyroxene	-	0.04
Total	1.00	1.15

Table 4: Major element contents of the examined four dike samples after the correction of alteration (ALT) and after the correction of high-temperature diffusional modification (High-T). The latter correction is necessary for DKSH1ES and DKSH2EI sampled taken from the dike edges, but the DKSH1ES is too altered to make the reliable correction as done for DKSH2EI.

Dike	DKSH1				DKSH2						RYO2
Sample	DKSH1ES	DKSH1MI	DKSH1CI	DKS	H2EI	DKSH2M	DKSH2MW	DKSH2CI	DKSH2CW	RY01I	RYO2I
Correction	ALT	ALT	ALT	ALT	ALT & High-T	ALT	ALT	ALT	ALT	ALT	ALT
(wt%)											
SiO <sub>2</sub>	39.73	41.28	42.22	42.46	42.13	42.42	42.47	42.42	42.36	43.11	40.81
TiO <sub>2</sub>	0.92	0.31	0.32	0.54	0.53	0.58	0.62	0.62	0.57	0.45	0.41
Al <sub>2</sub> O <sub>3</sub>	11.06	9.53	8.11	8.42	8.40	8.75	8.66	8.75	8.84	10.62	13.05
Fe <sub>2</sub> O <sub>3</sub>	14.17	12.51	12.12	11.55	13.04	11.44	11.58	11.70	11.53	10.23	9.69
MnO	0.13	0.18	0.22	0.18	0.18	0.17	0.17	0.18	0.18	0.17	0.20
MgO	20.54	25.96	27.69	28.00	26.89	26.66	26.75	26.68	26.58	18.92	14.57
CaO	11.89	8.73	7.91	7.35	7.33	8.56	8.31	8.13	8.41	15.14	19.71
Na <sub>2</sub> O	1.32	1.21	1.13	1.22	1.22	1.16	1.17	1.25	1.24	1.16	1.35
K <sub>2</sub> O	0.16	0.15	0.15	0.16	0.16	0.14	0.14	0.16	0.16	0.12	0.15
P <sub>2</sub> O <sub>5</sub>	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.03
NiO	0.05	0.10	0.10	0.10	0.09	0.10	0.10	0.09	0.09	0.04	0.02
Cr <sub>2</sub> O <sub>3</sub>	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Total Amphibole	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
abundance (wt%)	39.32	39.39	38.95	58.29	-	54.41	54.41	52.32	52.32	31.55	19.34
H <sub>2</sub> O (wt%) <sup>1</sup>	0.98	0.98	0.97	1.46	-	1.36	1.36	1.31	1.31	0.79	0.48

<sup>1</sup>Whole-lock water content calculated from the water content of amphibole (2.5wt%) and its modal abundance.

Major element oxides are recalculated to 100 wt % volatile free.

Table 5: Major and trace element contents and normative compositions of intruded melts and chemical composition of olivine phenocrysts in equilibrium with the melts estimated for each examined dike. For two dikes (DKSH2 and DKSH1) mean values and their standard errors (±1SE) are listed.

Dike	DKS	H1	DKS	H2	RYO1	RYO2
-		1SE <sup>1</sup>		1SE <sup>1</sup>		
Intruded melt						
Major oxides (wt%)						
SiO <sub>2</sub>	43.52	0.64	44.19	0.05	44.58	41.46
TiO <sub>2</sub>	0.56	0.02	0.98	0.02	0.62	0.53
Al <sub>2</sub> O <sub>3</sub>	15.78	0.51	14.64	0.08	14.64	16.89
FeO	9.57	0.09	9.57	0.26	7.38	6.40
MnO	0.20	0.01	0.18	0.00	0.15	0.17
MgO	11.43	0.10	12.76	0.12	8.72	5.98
CaO	14.89	0.15	13.73	0.31	20.86	25.50
Na <sub>2</sub> O	2.10	0.00	2.03	0.03	1.59	1.74
K <sub>2</sub> O	0.27	0.00	0.26	0.01	0.16	0.20
$P_2O_5$	0.02	0.00	0.02	0.00	0.04	0.04
Fe <sub>2</sub> O <sub>3</sub>	1.60	0.02	1.59	0.04	1.23	1.07
NiO	0.02	0.00	0.02	0.00	0.01	0.00
Cr <sub>2</sub> O <sub>3</sub>	0.03	0.00	0.03	0.00	0.02	0.02
Total	100.00		100.00		100.00	100.00
H <sub>2</sub> O	1.74	0.04	2.27	0.05	1.08	0.62
Trace elements (ppm)						
Ва	187	1	189	9	128	120
U	0.08	0.04	0.10	0.03	0.03	0.02
Nb	0.99	0.12	1.91	0.12	0.82	0.94
Та	0.06	0.01	0.12	0.01	0.08	0.06
La	2.59	0.08	1.85	0.11	1.43	0.95
Се	5.58	0.10	5.49	0.29	3.79	3.33
Pr	0.89	0.00	1.05	0.05	0.72	0.61
Pb	0.55	0.03	1.39	0.10	2.53	1.12
Sr	259	6	319	15	119	89
Nd	4.56	0.07	6.54	0.29	4.60	4.08
Hf	0.85	0.03	1.13	0.04	0.92	0.75
Zr	23.9	0.9	28.4	1.2	23.1	18.6
Sm	1.50	0.06	2.21	0.11	1.64	1.64
Eu	0.69	0.05	0.85	0.05	0.61	0.64
Gd	2.25	0.05	3.03	0.10	2.35	2.17
Tb	0.37	0.01	0.50	0.02	0.38	0.39
Dy	2.69	0.10	3.46	0.14	2.63	2.68
Y	14.2	0.5	17.7	0.7	14.4	12.8
Но	0.55	0.02	0.70	0.03	0.55	0.56
Er	1.59	0.05	1.97	0.08	1.59	1.50
Tm	0.22	0.01	0.27	0.01	0.21	0.20
Yb	1.47	0.04	1.82	0.08	1.34	1.28
Lu	0.22	0.01	0.27	0.01	0.19	0.19
Olivine in equilibrium						
with intruded melt						
Mg#	0.88	0.00	0.89	0.00	0.88	0.85
NiO (wt%)	0.21	0.00	0.21	0.01	0.14	0.08
MnO (wt%)	0.20	0.02	0.17	0.00	0.21	0.32
Abundance (wt%)	43.6	2.0	40.1	0.3	26.8	22.2

Major element oxides are recalculated to 100 wt % volatile free.

<sup>1</sup> standard error of the mean

Table 6: Concentrations of highly incompatible trace elements in amphibole and clinopyroxene consisting the matrix of DKSH2. The trace element concentrations of the matrix estimated from the weight fraction and chemical compositions of the amphibole and clinopyroxene listed in the first and second columns, respectively, are shown in the last column.

	Amph	ibole	Clinopy	roxene	estimated matrix		
	Average	1SE <sup>1</sup>	Average	1SE <sup>1</sup>		1SE <sup>1</sup>	
(ppm)							
Ва	38.4	3.0	5.4	2.7	37.1	2.9	
U	0.11	0.01	0.04	0.00	0.10	0.01	
Nb	2.57	0.18	0.07	0.01	2.47	0.17	
Та	0.17	0.02	0.01	0.00	0.17	0.02	
Pb	0.44	0.03	0.06	0.01	0.42	0.03	
Sr	243	11	36	5	235	10	
Hf	0.94	0.04	1.01	0.04	0.94	0.04	
Zr	31.1	1.3	33.2	0.5	31.2	1.3	

<sup>1</sup> standard error of the mean
Table 7: Major and trace element concentrations and normative composition of the primary melts in equilibrium with the residual peridotites estimated for each examined dike. The chemical compositions of olivine in the residual peridotites are also listed. The mean values and their  $\pm 1SE$  are shown for two dikes (DKSH2 and DKSH1) with multiple samples (Table 1).

Dike	DKSH1		DKSH2		RYO1	RYO2
-		1SE <sup>1</sup>		1SE <sup>1</sup>		
Primary melt						
Major oxides (wt%)						
SiO <sub>2</sub>	43.28	0.58	43.85	0.06	44.15	41.36
TiO <sub>2</sub>	0.51	0.01	0.88	0.02	0.55	0.46
Al <sub>2</sub> O <sub>3</sub>	14.37	0.50	13.13	0.10	12.94	14.66
FeO	9.64	0.10	9.54	0.25	7.64	6.98
MnO	0.20	0.01	0.18	0.00	0.16	0.18
MgO	14.74	0.18	16.52	0.19	13.37	11.54
CaO	13.56	0.17	12.31	0.30	18.45	22.13
Na <sub>2</sub> O	1.92	0.00	1.82	0.02	1.41	1.51
K <sub>2</sub> O	0.24	0.00	0.23	0.00	0.14	0.17
P <sub>2</sub> O <sub>5</sub>	0.02	0.00	0.02	0.00	0.04	0.04
Fe <sub>2</sub> O <sub>3</sub>	1.45	0.02	1.43	0.04	1.09	0.93
NiO	0.04	0.00	0.05	0.00	0.04	0.03
Cr <sub>2</sub> O <sub>3</sub>	0.03	0.00	0.02	0.00	0.02	0.02
Total	100.00		100.00		100.00	100.00
H <sub>2</sub> O	1.59	0.03	2.05	0.04	0.96	0.54
Trace elements (ppm)						
Ва	171	2	33	3	114	105
U	0.07	0.04	0.09	0.01	0.03	0.02
Nb	0.91	0.11	2.23	0.15	0.73	0.82
Та	0.06	0.01	0.15	0.01	0.07	0.05
La	2.37	0.07	1.67	0.10	1.28	0.83
Ce	5.10	0.08	4.95	0.26	3.38	2.91
Pr	0.81	0.01	0.95	0.04	0.64	0.54
Pb	0.51	0.03	0.38	0.03	2.25	0.98
Sr	236	5	212	9	106	78
Nd	4.17	0.05	5.90	0.21	4.10	3.57
Hf	0.78	0.03	0.85	0.04	0.82	0.65
Zr	21.88	0.80	28.14	1.14	20.56	16.28
Sm	1.37	0.05	1.99	0.10	1.46	1.43
Eu	0.63	0.05	0.76	0.05	0.54	0.56
Gd	2.05	0.04	2.73	0.09	2.09	1.90
Tb	0.33	0.01	0.45	0.02	0.34	0.34
Dy	2.46	0.09	3.12	0.12	2.34	2.35
Y	12.98	0.44	15.93	0.62	12.86	11.18
Но	0.50	0.01	0.63	0.02	0.49	0.49
Er	1.45	0.05	1.78	0.07	1.42	1.31
Tm	0.20	0.01	0.25	0.01	0.18	0.17
Yb	1.35	0.03	1.64	0.07	1.20	1.12
Lu	0.20	0.00	0.24	0.01	0.17	0.17
Norm components <sup>2</sup>						
Ne'	23.3	0.5	20.9	0.4	23.3	22.7
Qtz'	0.3	0.2	2.1	0.4	0.0	0.0
Ol'	76.4	0.7	77.0	0.3	76.7	77.3
	-					

#### Olivine in equilibrium

· · · ·						
	0.57	0.00	0.50	0.00	0.50	0.50
NiO(wt%)	0.37	0.00	0.38	0.00	0.38	0.38
Mg#	0.90	0.00	0.91	0.00	0.91	0.91
<u></u>						

<sup>1</sup> standard error of the mean

<sup>2</sup> C. I. P. W norm mineral components calculated from primary melt (Ne'= nepheline +

0.6albite, Ol'= olivine + 0.75olthopyroxene and Qtz'= Quartz + 0.4albite +

0.25olthopyroxene)

Major element oxides of the estimated primary melts are recalculated to 100 wt % volatile free.

Table 8: Percentage of melting in the garnet stability field (PMGF) in the total melting degree that reproduces the rare earth element pattern of primary melt for each dike (Fig. 14) by a steady-state one-dimensional decompressional melting model (Ozawa, 2001). The PMGFs are optimized for several sets of two of the model parameters: strength of melt separation shown in the first column and mantle source shown in the second column.

Dike	Strength of melt separation	Mantle source <sup>1</sup>	PMGF (%)	
		PM	95.7	
DKSH	accumulated	MS	88.7	
		DMS	87.3	
	instantaneous	MS	99.9	
DKSH2		PM	89.2	
	accumulated	MS	76.6	
		DMS	75.5	
	instantaneous	MS	99.9	
RY01		PM	97.2	
	accumulated	MS	90.1	
		DMS	93.0	
	instantaneous	MS	99.9	
RYO2		PM	98.0	
	accumulated	MS	93.9	
		DMS	93.9	
	instantaneous	MS	99.9	

<sup>1</sup> PM, primitive mantle; MS, MORB mantle source; DMS, depleted MORB mantle source.

Table 9: Segregation conditions (pressures, temperatures, and degrees of melting) of estimated primary melts for the examined four dikes and their solidus conditions (pressures and temperatures) and mantle potential temperatures. The solidus conditions were estimated by assuming an adiabatic decompressional melting. The standard errors listed in the last column (1SE) were estimated by propagating the uncertainty in the pressure and temperature estimation, which overwhelms the other uncertainties.

Dike	DKSH1	DKSH2	RY01	RYO2	
					1SE
Segregation conditions					
before H <sub>2</sub> O correction					
T (°C)	1490	1510	1470	1440	30
P (GPa)	4.7	4.7	4.8	4.9	1.0
after H <sub>2</sub> O correction					
T (°C)	1430	1430	1430	1420	30
P (GPa)	5.0	5.0	4.9	5.0	1.0
degree of melting <sup>1</sup> (wt%)					
Ne' norm	7.1	8.7	7.1	7.5	3.5
REE ratios	18.5	16.8	19.7	16.3	3.8
Solidus condition					
<i>T</i> (°C)	1440	1460	1450	1440	30
P (GPa)	5.3	5.4	5.3	5.3	1.0
Mantle potential temperature (°C)	1350	1360	1350	1340	40

<sup>1</sup> The degree of melting was estimated by two methods. One is based on the dependence of Ne' normative composition on melting degree (Ne' norm) and the other is based on the relationships between REE ratios (REE ratios). We adopted the melting degree based on the Ne' norm to calculate solidus conditions and potential temperatures. See the main text for the reasons and more details.



Post-Devonian strata



Motai metamorphic rocks

fault

] Tertiary and Quatenary strata

Cretaceous granitic rock



Jurassic accretionary complex



## Miyamori complex



hornblende-rich mafic-ultramafic intrusion



ultramafic dike



Cumulate member



Tectonite member





Cretaceous strata

Paleozoic strata













OKSH2
intermediately altered
△ DKSH1
○ DKSH2
□ RYO1
◇ RYO2
· edge
I margin
+ center
× unknown













X spinel in the host

ultramafic dike euhedral-subhedral spinel inclueded in olivine/amphibole △ DKSH1 □ RYO1 ○ DKSH2 ◇ RYO2

anhedral spinel near plagioclase
DKSH2







Fig.10









### loss on ignition (wt%)







### La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Tm Yb Lu

13 FIG



accumulated melt — primitive mantle — MORB mantle source … Depleted MORB mantle source instantaneous melt --- MORB mantle source







mantle source for this study (Na<sub>2</sub>O wt%: 0.28 - 0.30) at 3 - 6 GPa primary melt Ne' norm degree of melting ▲ DKSH1  $\Delta - DKSH1$ • DKSH2 O- DKSH2 RYO1 D-RYO1  $\sim RYO2$  $\mathbf{A}$  RYO2







#### slab breakoff

#### Mantle wedge



# △ Star Lake (Canada) ※ Bay of Islands (Canada) ▽ Annieopsquotch (Canada) + Jiugequan (China) ■ MORB mantle source