Modeling chemical geodynamics of subduction zones using the Arc Basalt Simulator version 5

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ABSTRACT

Arc Basalt Simulator version 5 (ABSS) is a forward geochemical and petrological model that can be used to examine the element mass balance in primary arc magmas including the source and nature of slab materials and flux melting of the mantle-wedge peridotite. The inverse problem approach using ABSS allows the estimation of intensive and extensive geophysical variables in arc magma genesis. The intensive variables are slab dehydration depth ($P_{\text{slab-dehydration}}$) and temperature ($T_{\text{slab}}$) and mantle melting pressure ($P_{\text{mantle}}$) and temperature ($T_{\text{mantle}}$). The extensive variables are the amount of slab liquid added to the mantle ($F_{\text{slab-liq}}$) and the degree of melting of the mantle ($F_{\text{mantle}}$) along with the amounts of water in the slab liquid ($X_{\text{H}_2\text{O}_{\text{slab-liq}}}$), mantle ($X_{\text{H}_2\text{O}_{\text{mantle}}}$), and magma ($X_{\text{H}_2\text{O}_{\text{melt}}}$). Subordinate geochemical variables that also can be estimated using ABSS include the degree of chemical reaction between slab liquids and the solid slab ($\%R$); slab liquid fractions derived from igneous oceanic crust ($F_{\text{IOC}}$), sediment ($F_{\text{SED}}$), and metasomatized mantle peridotite layers ($F_{\text{MMP}}$); and the degree of depletion of the mantle wedge ($\%\text{SMORB}_{\text{mantle}}$). The mass balances for 26 incompatible elements, six major elements including H$_2$O and Sr, Nd, Hf, and Pb isotopes are calculated based on the same scheme. Monte Carlo calculations are used to estimate the aforementioned variables by fitting the calculated magma composition to observed values. This paper describes the ABSS calculation scheme and presents examples of its successful use. The geophysical variables determined for these example cases are compared with those estimated by other methods. The spatial variations of the magma productivity and implications for the location of the volcanic front are also discussed.

INTRODUCTION

Constraints on intensive and extensive geophysical variables in subduction zones have been derived from geophysical observations and modeling (Kirby et al., 1996; Nakajima and Hasegawa, 2003; Nakajima et al., 2005), numerical geodynamic modeling (Gerya and Yuen, 2003; Horiiuchi and Iwamori, 2016; Peacock, 1996; Syracuse et al., 2010; van Keken et al., 2002; van Keken et al., 2011; Wada and Behn, 2015; Wilson et al., 2014), and petrological and geochemical modeling (Kelley et al., 2010; Kimura et al., 2010; Kimura and Nakajima, 2014; Kimura et al., 2014; Plank et al., 2009). In these models, water plays the key role in magma genesis (Grove et al., 2006; Pearce et al., 2005; Plank et al., 2013; Stolper and Newman, 1994) and geochemistry (Gerya and Yuen, 2003; Horiiuchi and Iwamori, 2016; van Keken et al., 2002; Wada and Behn, 2015; Wilson et al., 2014).

Water in subduction zones fundamentally originates from oceans. The carriers are sediment (SED), altered igneous oceanic crust (AOC), partially altered igneous oceanic crust (IOC) and gabbros (GAB), and hydrated oceanic mantle (SlbP) (Bebout, 1996; 2007; Hacker, 1996, 2008; Ranero et al., 2003; Staudigel et al., 1998). Water liberated from the subducted oceanic lithosphere alters the geophysical variables of the slab (Kirby et al., 1996; Kita et al., 2006; Shinoda et al., 2013; Tsuchi et al., 2008) and lowers the solidus temperature (Hermann and Spandler, 2008; Schmidt et al., 2004; Skora and Blundy, 2010). Water penetrating the mantle wedge lowers the mantle viscosity (van Keken et al., 2002) and solidus temperature (Green et al., 2014; Grove et al., 2006). The geochemical responses are changes in the mantle-wedge flow due to the water-bearing mantle rheology (van Keken et al., 2002; Wilson et al., 2014) and the emergence of Rayleigh-Taylor instability by the formation of buoyant serpentinized mantle at the bottom of the mantle wedge (Gerya, 2011; Gerya and Yuen, 2003). Residual water in the slab and mantle returns to the deep Earth beyond the subduction zones (Hacker, 2008; Iwamori, 2000; Kimura and Nakajima, 2014; van Keken et al., 2011), which can also affect the entire mantle convection process (Karato, 2010) and therefore the Earth’s thermochemical evolution (Hirschmann, 2006).

The characteristic chemistry of arc magmas is the result of subduction zone geochemistry; the controlling factors include geophysical variables, such as pressure ($P$) and temperature ($T$), and the extent of melt and fluid mobility in the slab and mantle. Arc magmas are assumed to be the products of chemical reactions of their source materials, which are governed by the slab and mantle conditions and conservation of mass in the system. Thus, geochemistry can be used to estimate the geophysical variables by chemically mass balancing the system.

To accomplish this, a reasonable forward geochemical and petrological model is required; such a model must thoroughly simulate the geochemical mass balance during dehydration and melting of the slab and flux melting of the mantle. Geochemical variables can be used to parameterize geochemical mass balance calculations (Kimura et al., 2008; Kimura et al., 2010; Kimura et al., 2014; Shaw, 2000). Inverse modeling is applicable using the forward model by altering the parameters in fitting elemental and/or isotopic abundances of a calculated magma to those observed. Once a reasonable fit is
accomplished, the geophysical model parameters are compared with geo-
physical observations (Kimura et al., 2009; Kimura et al., 2010; Kimura and
Nakajima, 2014; Kimura et al., 2014).

Some geochemical inversion models estimate the values of geo-
chemical variables for the slab and mantle. For example, the H2O/Ce ratio in arc magmas is used to estimate the slab surface temperature (TSS) during slab defluxing at a depth of 4 GPa (Cooper et al., 2012; Plank et al., 2009). The phase relations determined by experimental databases and thermodynamic models are used to estimate mantle melting temperatures (Tmelt), pressures (Pmelt), and water contents (XH2O_perid) in addition to the degree of melting (Fmelt) in the sub-arc mantle (Kimura and Ariskin, 2014; Kuritani et al., 2014a; Kuritani et al., 2014b; Lee et al., 2009). Forward calculations using trace-element compositions of magmas have been used to estimate the same variables in the sub-arc mantle (Kelley et al., 2010). The Arc Basalt Simulator (ABS) models estimate a wider range of geophysical parameters of the slabs and mantles for a variety of subduction zones. Parameters provided by the program include: slab TSS and PSS, mantle Tmelt and Pmelt XH2O_perid and XH2O_melt; water content of the slab liquid XH2O_slab liq; peridotite XH2O_perid; and arc magma XH2O_melt. Contributions of flux fractions from different slab layers (Fsed(AOC) SED and Fsed(MwP)); and depletion of the source mantle by means of melt extraction (%MOB), (Kimura et al., 2009; Kimura et al., 2010; Kimura and Nakajima, 2014; Kimura et al., 2014). A summary of the acronyms used in this study and the parameters used in the models are presented in Table 1.

The ABS series models thoroughly formulate the petrological and geo-
chemical mass balance of both slab and mantle wedge for major and trace element, and Sr-, Nd-, Hf-, Pb-isotope compositions. The evolution of the ABS models includes a prototype (Kimura and Yoshida, 2006), followed by thoroughly realized models ABS1 (Sendjaja et al., 2009), ABS2 (Kimura et al., 2009), ABS3 (Kimura et al., 2010), and ABS4 (Kimura et al., 2014). The first model in the public domain was ABS2, which considered only aqueous slab fluids and mantle melting with fluid fluxing. The ABS3 model introduced slab melting; mantle melting with slab-melt fluxing was also parameterized. The ABS4 model introduced a multi-layer slab model that simulated the steep ther-
mal gradient in the slab and chromatographic reactions of slab liquids that occur with slab solids across the slab.

The ABS models have been successfully used to model various magma types from present-day to ancient subduction zones. The ABS3 and ABS4 models have been applied to NE and SW Japan in addition to the infant stage and present-day Izu-Bonin Arc (Kimura et al., 2009; Kimura et al., 2010; Kimura and Nakajima, 2014; Kimura et al., 2014; Li et al., 2013). The applications have been recently extended to the Kamchatka Arc (Portnyagin et al., 2015), Southern Chile Arc (Jacques et al., 2013; Jacques et al., 2014), New Zealand Arc (Rooney and Deering, 2014), ancient Mediterranean subduction systems (Becaluva et al., 2013; Mattioli et al., 2012), and Earth’s ancient subduction systems back to the Archean (3.5 Ga) (Kimura et al., 2016).

This paper presents a description of the newly refined ABS5 model and its applications to cold NE Japan, hot SW Japan, and infant Izu-Bonin arcs. Geo-

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**SUBDUCTION ZONE MAGMATISM**

This section describes the fundamental subduction zone magma genesis model used in ABS5. The chemistry of subduction zone magmas differs from that of other tectonic settings such as mid-ocean ridges (MORs) and ocean islands. This phenomenon is attributed to the contribution of water-rich chemical flux from the subducting slab in addition to decompression melting of relatively dry mantle peridotite dominant in other tectonic settings such as MORs and hotspot ocean islands.

**Subducting Slab**

In descending order, the subducting slab consists of ocean-floor sediment (SED); variously altered oceanic crust (AOC) of mid-ocean ridge basalt (MORB) lavas, dikes (DIKE), and holocrystalline gabros (GAB); and lithospheric mantle (SlbP) (Fig. 1). These components are severely to partially hydrated, which is partially due to sub-seafloor hydrothermal alteration developed near the spreading ridges and low-temperature seawater alteration in the cooled oce-
anic plate (Kelley et al., 2003; Staudigel et al., 1996). Bending of the oceanic plate at the outer rise also rehydrates the slab at depth, possibly due to the SlbP layer (Fujie et al., 2016; Ranero et al., 2003) (“1” in Fig. 1A).

Prograde metamorphism and dehydration occur when the slab of the oceanic plate begins to subduct. This process begins with pore collapse in the smectite-illite and prehnite-pumpellyite facies (Saffer and Tobin, 2011), followed by the green schist and amphibolite facies (Bebout, 2007) and ultra-
high-pressure dry eclogite metamorphic facies (Hacker, 2008; Hacker et al., 2003; van Keken et al., 2011) (Fig. 1A). Slab liquids include aqueous fluid, supercritical liquid, and melt, depending on the P-T conditions and the solid and liquid compositions (Herrmann et al., 2006; Kawamoto et al., 2012). Here-
after, the slab-derived fluid phase is referred to as “slab liquid” unless other-
wise noted.

The slab P-T path is a fundamental factor determining the chemistry of slab liquids (Fig. 1B). The governing factors are the mineralogic mode (X) of the residual slab and the T-dependent partition coefficient (D) of elements between minerals and liquids (Kimura et al., 2009; Kimura et al., 2014). A strong tempera-
ture gradient forms in the slab. The isotherms are subparallel to the slab-mantle interface due to the conductive heat of the high-T mantle wedge (Syracuse et al., 2010; van Keken et al., 2002). Therefore, slab layers have different P-T paths depending on the depth of the mantle-slab interface (van Keken et al., 2011).
<table>
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<tr>
<th>Acronyms</th>
<th>Definition</th>
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<tr>
<td><strong>Element</strong></td>
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<tr>
<td>REE</td>
<td>Rare-earth element</td>
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<td>HFSE</td>
<td>High field strength element</td>
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<tr>
<td>LILE</td>
<td>Large ion lithophile element</td>
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<td><strong>Rock and mineral type</strong></td>
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<tr>
<td>MORB</td>
<td>Mid-ocean ridge basalt</td>
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<tr>
<td>OIB</td>
<td>Ocean-island basalt</td>
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<tr>
<td>HMA</td>
<td>High-magnesium andesite</td>
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<tr>
<td>ADK</td>
<td>Adakite</td>
</tr>
<tr>
<td>BON</td>
<td>Boninite, a high-Mg andesite occurring in Bonin (Ogasawara) Island region</td>
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<tr>
<td>LK-TH</td>
<td>Low-K tholeiite</td>
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<tr>
<td>HK-CA</td>
<td>High-K calc-alkaline</td>
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<tr>
<td>TTG</td>
<td>Tonalite-trondhjemite-granodiorite</td>
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<tr>
<td>Ol</td>
<td>Olivine</td>
</tr>
<tr>
<td>Cpx</td>
<td>Clinopyroxene</td>
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<tr>
<td>Opx</td>
<td>Orthopyroxene</td>
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<tr>
<td>Gt</td>
<td>Garnet</td>
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<tr>
<td>Sp</td>
<td>Spinel</td>
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<td><strong>Earth's chemical reservoir</strong></td>
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<tr>
<td>PM</td>
<td>Primitive mantle</td>
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<tr>
<td>DMM</td>
<td>Depleted MORB source mantle</td>
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<tr>
<td>GLOSS</td>
<td>Global subduction sediment</td>
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<tr>
<td><strong>Slab material</strong></td>
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<tr>
<td>MwP</td>
<td>Mantle wedge-base peridotite layer subducted with the slab</td>
</tr>
<tr>
<td>SED</td>
<td>Sediment layer in the slab</td>
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<tr>
<td>AOC (UBAS)</td>
<td>Altered oceanic crust (upper basalt) layer in the slab</td>
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<tr>
<td>LBAS</td>
<td>Lower basalt layer in the slab</td>
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<tr>
<td>DIKE</td>
<td>Dike layer in the slab</td>
</tr>
<tr>
<td>UGAB</td>
<td>Upper gabbro layer in the slab</td>
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<tr>
<td>LGAB</td>
<td>Lower gabbro layer in the slab</td>
</tr>
<tr>
<td>IOC</td>
<td>Igneous oceanic crust (including AOC, LBAS, DIKE, UGAB, and LGAB) layer in the slab</td>
</tr>
<tr>
<td>SlbP</td>
<td>Slab peridotite layer in the slab</td>
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<tr>
<td><strong>Geochemical model code</strong></td>
<td></td>
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<tr>
<td>ABS5</td>
<td>Arc Basalt Simulator version 5 coded by Kimura et al. (this paper)</td>
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<tr>
<td>PRIMACalc2</td>
<td>Primary Magma Calculator version 1 coded by Kimura and Ariskin (2014)</td>
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<tr>
<td>pMELT</td>
<td>Thermodynamic mantle melting model coded by Ghiorso et al. (2002)</td>
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<tr>
<td><strong>Unit</strong></td>
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<tr>
<td>GPa</td>
<td>Giga Pascal</td>
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<tr>
<td>T</td>
<td>Temperature in °C</td>
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<tr>
<td>T_55</td>
<td>Slab surface temperature in °C</td>
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<tr>
<td>P_55</td>
<td>Slab surface depth in Gpa</td>
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<tr>
<td>T_slab</td>
<td>Slab temperature in °C</td>
</tr>
<tr>
<td>P_slab</td>
<td>Slab depth in Gpa</td>
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<tr>
<td>F_slab</td>
<td>Degree of partial melting of slab layer either by fraction or by % as noted</td>
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<tr>
<td>T_perid</td>
<td>Mantle peridotite temperature in °C</td>
</tr>
<tr>
<td>P_perid</td>
<td>Mantle peridotite depth in Gpa</td>
</tr>
<tr>
<td>F_perid</td>
<td>Degree of partial melting of peridotite either by fraction or by % as noted</td>
</tr>
<tr>
<td>XH_2O</td>
<td>H_2O content (fraction)</td>
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<tr>
<td>Fliq(AOC)</td>
<td>Fraction of AOC liquid</td>
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<tr>
<td>Fliq(SED)</td>
<td>Fraction of SED liquid</td>
</tr>
<tr>
<td>Fliq(MwP)</td>
<td>Fraction of MwP liquid</td>
</tr>
<tr>
<td>Fliq(slblq)</td>
<td>Fraction of slab liquid in the mantle wedge</td>
</tr>
<tr>
<td>β</td>
<td>Open-system melting parameter given by $\beta = b/(a + b)$, where $a$ and $b$ are slab liquid and mantle melt</td>
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</tbody>
</table>
This thermal structure results in different deflux profiles in each slab layer (Kimura et al., 2014; Kimura et al., 2016). In addition, buoyant slab liquids travel upward in the slab and react with the overlying layers. This leads to the formation of vertical chromatographic reaction columns in the slab, where the slab itself undergoes subduction at a rate of a few to tens of centimeters per year. Therefore, the discrete solid and liquid flows result in complex reactions in the slab (Gorman et al., 2006; Kimura et al., 2014; Kimura et al., 2016) (Fig. 1A).

Mantle Wedge

The mantle wedge consists of peridotite with inverted geotherms in the lower half of the wedge resulting from cooling by the cold subducting slab (Fig. 1A). The mantle dragged downward by the subducting slab causes upwelling of the deep rear-arc (RA) asthenosphere, which develops a hot and partially molten mantle-wedge core from which arc magmas are segregated (Grove et al., 2006;
Slab liquids migrate upward through the subsolidus mantle beneath the hot mantle-wedge core (Mibe et al., 1999; Poli and Schmidt, 1995; Schmidt and Poli, 1998; Wilson et al., 2014). The subsolidus mantle is thought to be partly dragged down with the slab (Iwamori, 2000; Tatsumi and Eggins, 1995). This occurs beyond the region of slab-mantle mechanical coupling indicated by cessation of the seismicity at the slab interface in the depth range >55–60 km in the case of NE Japan Arc (Hasegawa et al., 2009; Kimura and Nakajima, 2014; Kita et al., 2010) (Fig. 1A). The slab liquids are transported either by crack flow or porous flow depending on the viscosity of the slab fluid and/or melt, and the dihedral angles of the mantle minerals and hydrous minerals formed in the low-temperature mantle (Kimura and Nakajima, 2014; Mibe et al., 1999; Nakajima et al., 2005; Tatsumi and Eggins, 1995) eventually migrate upwards to the hot core of the mantle wedge.

Flux melting of the mantle occurs in the hot mantle-wedge core depending on $T_{\text{perid}}$, $P_{\text{perid}}$ and $X\text{H}_{2}\text{O}_{\text{perid}}$ (Green et al., 2014; Grove et al., 2006). The distribution of slab-derived water leads to complexity in the melting and melt transport (Kimura and Nakajima, 2014; Wada and Behn, 2015; Wilson et al., 2014). Moreover, the addition of slab flux essentially increases the abundances of some fluid- and/or melt-mobile incompatible elements in arc magmas. In contrast, the slab-derived water enhances mantle melting, which can result in the dilution of incompatible elements in the resulting melt (Kimura and Nakajima, 2014; Kimura et al., 2014). The extraction and coalescence of such magma occur in the two-phase flow field (Wilson et al., 2014) of slowly convecting mantle on the order of a few centimeters per year, where liquids and melts migrate more rapidly on the order of a few to tens of centimeters per year (Turner et al., 1997) to meters per year (Turner et al., 2001). Modeling the complexity in detail is beyond the scope of the ABS model, although some reactions between solids and liquids are considered.

This paper focuses on the compositions of primary arc magmas. Thus, the effect of the arc crust (Davidson, 1996; Davidson et al., 2005; Davidson et al., 2007; Kimura and Yoshida, 2006; Turner and Langmuir, 2015a) is not considered and will be discussed in a separate paper. Moreover, the serpentinite mantle diapir model (Gerya, 2011; Gerya and Yuen, 2003; Marschall and Schumacher, 2012) is not considered because it does not effectively explain systematic across-arc geochemical variations found in many subduction systems (Kimura and Nakajima, 2014; Turner et al., 1997). However, the roles of melt diapirs formed by slab melting (Kimura et al., 2014; Tsuchiya and Kanisawa, 1994) are discussed because some felsic to intermediate arc magmas (e.g., adakite and high-Mg andesite) are thought to be the products of slab melts that subsequently interacted with the overriding wedge mantle (Defant and Drummond, 1990; Kimura et al., 2014; Li et al., 2013; Martin, 1999; Moyen, 2009; Tatsumi and Hanyu, 2003).

**ARC BASALT SIMULATOR VERSION 5**

The ABS4 and ABS5 models are based on the subduction zone model described above. Updates of the ABS5 model include a refined parameterization for mantle melting and a new user interface to better fit the calculations. This section discusses the mass balance, intensive and extensive geophysical variables, calculation backgrounds, and strategy for estimating geophysical variables using ABS5. Details about the model description, operation guide, and worksheet data structure in the Excel spreadsheet of ABS5 are given in the Supplemental Tutorial [see footnote 2].

**Mass Balance Model Structure: Slab and Mantle Two-Box Model**

ABS5 is a two-box model consisting of geochemical calculations for the slab and mantle. Because of the steep thermal gradient across the slab, ABS5 and ABS4 models use eight slab layers that include slab peridotite (SlbP), lower gabbronor (LGAB), upper gabbronor (UGAB), dikes (DIKE), lower basalt (LBAS), upper basalt (UBAS) or AOC and SED, and mantle-wedge peridotite (MwP) (Fig. 1A). The uppermost MwP layer simulates the mantle wedge dragged down with the slab after mechanical coupling between the slab and mantle (Kimura and Nakajima, 2014; Kimura et al., 2014).

The mantle wedge melts through the addition of the slab liquid to the adiabatically upwelled mantle asthenosphere. The molten mantle is considered to be a single open box maintained by influxes and outfluxes of the arc magma and residual peridotite (Ozawa and Shimizu, 1995) (Fig. 1A). A complete mass balance is maintained in the mantle box but not with the slab influx into and outflux out of this box.

The influx of the mantle and slab liquid into the molten mantle wedge is only identified by the mass balance when the production rate of the primary magma (mass of outflux magma at a given time interval) is given (Kimura et al., 2009). Additional determinations of the slab subduction rate and amount of water and elements in the slab provide the whole arc mass balance including slab input, magma output, and deep mantle return (Kimura and Nakajima, 2014). Nevertheless, the two-box model in the static mode can provide the element abundances of the primary arc magma and controlling geophysical parameters without obtaining additional geodynamic variables.

**Intensive and Extensive Variables of the Two-Box Model**

This section describes important intensive and extensive variables of the two-box model.

**Variables of Slab Defluxing: $T_{SS}$ and $F_{slab liq}$**

Intensive geophysical variables that control dehydration and melting of the slab are the slab-surface temperature ($T_{SS}$) and slab-liquid fraction ($F_{slab liq}$). First, $T_{SS}$, which represents the thermal structure of the slab, is an important intensive geophysical variable that controls the slab liquid geochemistry. The slab $P$-$T$ path differs between arcs due to differences in the tectonic setting (Syracuse et al., 2001). The extraction and coalescence of such magma occur in the two-phase flow field (Wilson et al., 2014) of slowly convecting mantle on the order of a few centimeters per year, where liquids and melts migrate more rapidly on the order of a few to tens of centimeters per year (Turner et al., 1997) to meters per year (Turner et al., 2001). Modeling the complexity in detail is beyond the scope of the ABS model, although some reactions between solids and liquids are considered.

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The dehydration and melting of the slab directly depend on \( T_{\text{es}} \); they are calculated continuously over a depth range of 0.5–6.0 GPa at steps of 0.1 GPa (Fig. 1B). The stability of hydrous silicate minerals is largely \( T \) dependent based on the steep Claysius-Clapeyron slopes in this pressure range (Fig. 1B) (Hacker et al., 1996, 2008). This dehydration and vapor-saturated solidi are also strongly \( T \) dependent (Fig. 1B) (Hermann and Spandler, 2008). Thus, the composition of the slab liquid released from \( F_{\text{es}} \) is primarily determined by the slab \( T_{\text{es}} \) at that depth.

Detailed modeling using ABS4 further supports the above-mentioned view. The slab liquid composition is primarily \( T_{\text{es}} \) dependent and almost independent of the prior \( P-T \) history of the slab before the slab reached the depth \( (P_{\text{es}}) \) of dehydration or melting. This is not surprising because the net element loss from the slab layers is small for most elements, and their bulk composition therefore does not significantly change during prograde metamorphism (Kimura et al., 2016). This is true for the natural observations of high-\( P \) (HP) and ultra-high-\( P \) (UHP) metamorphic rocks of SED and IOC origin (Bebout, 2007; Behn et al., 2011). This feature allows ABS models to estimate \( T_{\text{es}} \) from the geochemistry. In contrast, the \( F_{\text{es}} \) estimation is geodynamic model dependent and only weakly correlated with \( T_{\text{es}} \) and the composition of the slab liquid. The rationale for the \( F_{\text{es}} \) estimation in ABS5 should be discussed together with geophysical observations, which will be made subsequently.

Second, the slab-liquid fraction \( (F_{\text{slab liq}}) \) can be aqueous fluid, supercritical liquid, or melt. It is an important extensive variable, as is its water content. The slab liquid is a mixture of liquids from the topmost slab layers, AOC and SED, and MwP as \( F_{\text{slab liq}} = F_{\text{slab liq}}(\text{AOC}) + F_{\text{slab liq}}(\text{SED}) + F_{\text{slab liq}}(\text{MwP}) \) (Kimura et al., 2009). This assumption holds true even if the slab-mantle interface forms a mélangé complex because the mélangé consists of the above-mentioned three components (Bebout, 2007). The composition of \( F_{\text{slab liq}} \) is a mixture of these slab liquids, each of which integrates the effects of chromatographic reactions in the slab layers. The water content in each liquid is given by the calculations of dehydration and/or melting in the slab. The total water flux \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \) in the \( F_{\text{slab liq}} \) is also calculated based on the equation above.

**Variables of Mantle Melting:** \( T_{\text{perid}}, P_{\text{perid}}, F_{\text{slab liq}}, X_{\text{H}_2\text{O}_{\text{slab liq}}}, \text{and } F_{\text{perid}} \)

Intensive geophysical variables controlling the partial melting of the mantle wedge are \( T_{\text{perid}} \) and \( P_{\text{perid}} \); the extensive variables include slab-flux fraction \( F_{\text{slab liq}} \) and water in the slab flux \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \). These four variables control \( F_{\text{perid}} \), which is another important extensive variable for the mantle wedge. All of these variables determine the geochemical mass balance in the mantle wedge.

The entire ABS5 mass balance is defined by \( F_{\text{es}}, \) which is common in both the slab and mantle boxes. In the static ABS5 model, \( F_{\text{es}} \) is used as a free parameter. Additional geophysical variables of the ABS5 model are explained for slab and mantle boxes in the following sections including a brief background calculation. Readers can also refer to the Supplemental Tutorial [see footnote 2] for more details.

**Calculations of Slab Dehydration and/or Melting**

Prograde metamorphism and dehydration and/or melting of the slab are the most important geochemical processes in subduction zone magmatism. Element partitioning in these processes essentially determines the arc magma signatures. Therefore, ABS5 uses the (1) slab \( P-T \) path derived from the geodynamic model, (2) thermodynamic model of PerpleX for the subsolidus slab mode, (3) experimentally parameterized mode for solids-liquidus intervals, and (4) partition coefficients of slab liquids and minerals determined primarily by experiments. The following calculations apply to 26 incompatible elements and Sr, Nd, Hf, and Pb isotope from a fully MORB source mantle (DMM), which are assumed to have been fully hydrated in the beginning of subduction (Hacker, 2008). The calculations are only valid for subsolidus conditions. The lookup tables for \( F_{\text{slab liq}}, \) and residual mineral modes \( (X), X_{\text{H}_2\text{O}_{\text{slab liq}}}, \) and \( F_{\text{slab liq}} \) (Fig. 2) and are used for calculations along the prograde metamorphism of the designated slab \( P-T \) paths (Fig. 2). The entire lookup table database and graphical representation of the results are shown in the [Slab Mode] worksheet in the ABS5 model of the Supplemental Software [see footnote 1].
Figure 2 (on this and following page). (A) and (B) Modal compositions in the pressure-temperature (P-T) space for a subducted slab. The slab materials are igneous oceanic crust (IOC), gabbro (GAB), sediment (SED), and depleted peridotite (PERID). The color codes indicate the modal composition with the gray field indicating the absence of phases. The thin lines indicate the slab P-T paths of eight slab layers calculated by the geodynamic model for the Nankai (SW Japan) subduction zone. The modal compositions are given by weight. The slab modes in the lower panels are calculated for NE Japan and SW Japan slab geotherms. The slab layers are MwP—mantle-wedge peridotite; SED—sediment; AOC—altered oceanic crust (UBAS), DIKE—dike (LBAS); UGAB—upper gabbro; LGAB—lower gabbro; and SlbP—slab peridotite (oceanic lithosphere mantle). The modal compositions are given in the bottom lines. Abbreviations: Gt—garnet; Cpx—clinopyroxene; OI—olivine; Opx—orthopyroxene; SiO$_2$—silica minerals; Plag—plagioclase; Chl—chlorite; Bio—biotite; Phen—phengite; and Amp—amphibole; Law—lawsonite; Zo—zoisite; Fe-Ti—iron–titanium oxide; H$_2$O—residual water in the slab; and Melt—melt fraction in the slab.
Element Partitioning

The element partitioning between the slab solids and liquids during slab dehydration and/or melting is calculated based on the modal composition ($X_i$) and $F_{\text{slab}}$. The batch melting equation was used,

$$C_i = Cs[(D_{\text{slab}} \times F_{\text{slab}})(1 - D_{\text{slab}})],$$

where $C_i$ is the fluid and/or melt composition, $Cs$ is the solid composition, $D_{\text{slab}}$ is the bulk partition coefficient, and $F_{\text{slab}}$ is the degree of dehydration and/or melting. $D_{\text{slab}}$ is calculated by,

$$D_{\text{slab}} = \sum_i [D_i \times X_i],$$

where $D_i$ is the partition coefficient of an element in the mineral phase $a$, and $X_i$ is the modal composition of the mineral.

Figure 3 shows the changes of $D_{\text{slab}}$ in the slab layers along the NE Japan slab $P-T$ path. The $Ds$ value is largely $T$ dependent, although $X_i$ also affects $D_{\text{slab}}$. The partition coefficients used for most metamorphic minerals are temperature dependent in the ABS3–ABS5 models; details of the $Ds$ used in ABS3–ABS5 are given elsewhere (Kimura, 2012; Kimura et al., 2009; Kimura et al., 2010). They are also provided in the Supplemental Tutorial [see footnote 2] and Supplemental Software [see footnote 1].

During the subsolidus dehydration mode, $F_{\text{slab}}$ is determined by the fraction of dehydrated water calculated by the mass balance in Perple_X based on the residual mineralogical mode of hydrous minerals (Fig. 2). Once a slab layer undergoes melting, $F_{\text{slab}}$ is defined by the degree of partial melting (Fig. 2). In this region, the $H_2O$ fraction in the slab melt is calculated simply by using $D_{H_2O} = D_{\text{w}} = 0.01$ (Dixon et al., 2002). This model reproduces the water content of felsic slab melts that were generated at $T = 680$ °C–1000 °C and $P = 2.5$–5.0 GPa based on experiments for metapelitic and metagraywacke (Hermmann and Spandler, 2008).

Chromatographic Reaction

Slab liquid departing from the slab layer reacts with the overlying layers. The ABS4 and ABS5 models calculate this effect using a chromatographic reaction in a vertical 1-D column at each 0.1 GPa step in the 0.5–6 GPa range. This simulates the reactions between descending slab solids and vertically upwelling slab liquids. The mass balance of this chain reaction has been described for ABS4 (Kimura et al., 2014). A schematic figure is shown in the Supplemental Tutorial [see footnote 2] and the [SDMS4.0] worksheet of the ABS5 in the Supplemental Software see [footnote 1] in which all of the calculations are made.

Chromatographic reactions may occur either completely or incompletely (Wada et al., 2012); this factor is simulated by the “reactivity” between the liquids and solids ($\%R_{\text{slab}}$) (Kimura et al., 2014). The $\%R$ in the ABS4 and ABS5 models alters $F_{\text{slab}}$ proportionally in each calculation step (e.g., if $\%R_{\text{slab}} = 110$, then $F_{\text{slab}} = F_{\text{slab}} \times 1.1$). This indicates additional reactions in the slab with $\%R_{\text{slab}} > 100$ and fewer reactions with $\%R_{\text{slab}} < 100$. This factor is also used as an extensive variable. Although the role of $\%R_{\text{slab}}$ is subordinate in mass balance calculations, it is effective for the isotope mass balance that determines $F_{\text{slab}}$ (AOC), $F_{\text{slab}}$ (SED), and $F_{\text{slab}}$ (MwP). See the Supplemental Tutorial [see footnote 2] and Supplemental Software [see footnote 1].

Slab-Liquid Uptake

The slab-liquid uptake is determined by the depth of the slab surface using $P_{\text{slab}}$ (GPa). The ABS4 and ABS5 models assume the uptake of the slab liquids from the top three (AOC, SED, and MwP) layers in weight fractions: $F_{\text{slab}}$ (AOC), $F_{\text{slab}}$ (SED), and $F_{\text{slab}}$ (MwP). These intensive ($F_{\text{slab}}$, $P_{\text{slab}}$) and extensive ($F_{\text{slab}}$ (AOC), $F_{\text{slab}}$ (SED), and $F_{\text{slab}}$ (MwP)) variables fundamentally determine the slab-liquid composition, which is important for the ABS5 trace-element mass balance. The slab-liquid composition is calculated in the [SDMS4.0] worksheet in ABS5. See the Supplemental Tutorial [see footnote 2] and Supplemental Software [see footnote 1].

Major-Element Composition of the Slab Liquid

The major-element composition of the slab liquid is addressed in a different manner assuming two discrete modes: an aqueous fluid mode and a silicic melt mode. Aqueous fluid contains various amounts of trace elements within 100 wt% of the $H_2O$ matrix. Silicic melt also contains various amounts of trace elements in a rhyolitic major-element matrix of $SiO_2 = 72.05$, $TiO_2 = 0.41$, $Al_2O_3 = 16.76$, $FeO = 2.40$, $MgO = 0.29$, $CaO = 1.24$, $Na_2O = 3.92$, and $K_2O = 2.87$ wt%. This felsic melt composition is calculated based on the average of the experimental melts of hydrous SED and IOC (MORB) (Moyen and Stevens, 2006) for ABS3 (Kimura et al., 2010).

The rationale of this assumption is a petrological invariant point when the degree of slab melting is small ($F_{\text{slab}} < 20\%$) (Kimura et al., 2010). However, there is not a single invariant point at a higher degree of melting ($F_{\text{slab}} > 20\%$). The melt composition will vary for different slab materials such as volcaniclastic, pelagic, chert, and carbonate sediments. This simple assumption introduces potential errors in estimating the major-element composition of the primary arc magmas. The ABS5 model uses the trace-element composition for the mass balance calculations in the mantle wedge. The errors with respect to the major-element composition are thus limited. Moreover, ABS4 model calculations reasonably reproduce adakitic dacite and high-Mg andesite, not only their trace elements but also their major elements (Kimura et al., 2014).

Calculations of Flux Melting of the Mantle

ABS3–ABS5 use (1) an open-system melting equation (Ozawa and Shimizu, 1995) for the flux melting of the mantle; (2) melting parameterization of peridotite derived from experiments (Katz et al., 2003); and (3) residual mineral
modes and major-element compositions of the melts parameterized based on pMELTS thermodynamic model calculations (Ghiorso et al., 2002) with adjustments of the experimental database (Kimura et al., 2010). The ABS5 model uses newly updated lookup tables including the melting mineralogy and major-element compositions of the melt to provide a better melting profile. The following calculations apply to 26 incompatible elements and Sr, Nd, Hf, and Pb isotopes. The treatment of major elements is also described.

**Open-System Melting**

The open-system melting equation of Ozawa and Shimizu (1995) is used for trace-element calculations. This equation requires the degree of melting of a peridotite \( (F_{\text{perid}}) \), residual peridotite mineral mode \( (X_\text{a}) \), partition coefficients between minerals and melt \( (D_i) \), and an open-system melting parameter \( \beta \) defined by:
\[ \beta = \frac{a}{a + b}, \]  

where \( a \) is the influx fraction from the slab and \( b \) is the fraction of partial melt from the mantle, as shown by the full open-system melting equation in the Supplemental Tutorial [see footnote 2].

In the ABS5 model, \( F_{\text{perid}} \) is determined by experimental parameterization (Katz et al., 2003), which requires the mantle temperature \( T_{\text{perid}} \) (°C) and pressure \( P_{\text{perid}} \) (GPa) with \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \). The \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \) value is derived from the slab flux fraction \( F_{\text{perid}} \) and water in the slab flux \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \). The \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \) value is assumed to be 100 wt% in an aequous slab liquid and equal to \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \) in a slab melt, as noted above. The uppermost row in Figure 4 graphically represents the map of \( F_{\text{perid}} \) in the \( P-T \) space of five different water contents (dry, 0.1, 0.5, 1.0, and 3.0 wt% \( X_{\text{H}_2\text{O}_{\text{slab liq}}} \)) of the depleted MORB source mantle DMM (Workman and Hart, 2005). The same set is prepared for the primitive mantle (PM) (Sun and McDonough, 1989) in ABS5.

The mantle fertility depends on the location of clinopyroxene out in the solidus-liquidus intervals (Katz et al., 2003). The ABS5 model newly incorporates the choice between DMM and PM for the initial melt wedge composition, as mentioned above. The subsolidus clinopyroxene mode of \( X_{\text{cpx}} = 18 \) wt% and \( X_{\text{ov}} = 15 \) wt% is given for PM and DMM, respectively, in the Katz et al. (2003) model. \( F_{\text{perid}} \) was calculated according to the [Katz+] worksheet. See the Supplemental Tutorial [see footnote 2] and Supplemental Software [see footnote 1].

**Residual Mantle Mode**

The residual mantle mode \( X_{r} \) was calculated for olivine, clinopyroxene, orthopyroxene, garnet, and spinel at 0.5–3.0 GPa (Fig. 4). The pMELTS thermodynamic model (Ghiorso et al., 2002) with Adiabat_1ph frontend version 1.8 (Smith and Asimow, 2005) was used to calculate the residual mode for the PM and DMM compositions (Sun and McDonough, 1989; Workman and Hart, 2005). Systematic discrepancies occurred in olivine and orthopyroxene modes compared with the experimental results (Lambart et al., 2012). The discrepancies were adjusted by applying a correction factor:

\[
X_{\text{corrected}} = X_{r} \times 0.99 \times \exp \left\{ \left[ -0.002 \times P\text{(GPa)}^2 + 0.0135 \times P\text{(GPa)} - 0.0115 \right] \times F(%)) \right\}
\]

and renormalization of the total sum of the modal composition. This simple parameterization compensates for the discrepancy when \( X_{r} \) is calculated as a function of \( F_{\text{perid}} \) (Fig. 5A) from the parameterization of Katz et al. (2003).

The same adjustment was made for slab-melt–fluxed mantle melting. Therefore, the major-element compositions of DMM and PM were modified by additions of the felsic slab melt noted above at 0, 5, 10, 15, 20, and 25 wt% of the mantle mass. Examples of the mantle modes with various slab-melt additions are shown in Figure 4 for DMM at \( F_{\text{perid}} = 0\%–30\% \). The ABS5 model uses new lookup tables for DMM and PM and calculates \( X_{r} \) from a given \( F_{\text{perid}} \) determined by the melting parameterization. The melting mode of the mantle inherently contains errors from the fixed slab-melt composition as noted above.

The liquidus-solidus intervals of the DMM and PM fluxed by slab melt differ from that of genuine peridotite. However, conditions of clinopyroxene and garnet out did not significantly differ up to 20% slab-melt addition (Kimura and Kawabata, 2015; Mallik and Dasgupta, 2012). The addition of felsic slab melt lowered the solidus temperature, but the residual mantle phases were always olivine-clinopyroxene-orthopyroxene plus an alumina-bearing phase. The melting profile did not significantly differ with ~25% slab-melt addition at 1–3 GPa (Mallik et al., 2016).

The calculation scheme is shown in the Supplemental Tutorial [see footnote 2] and the lookup tables and calculations of \( X_{r} \) are given in the [Per_Mode_Melt] worksheet of the ABS5 in the Supplemental Software [see footnote 1]. This function reflects a major update in the ABS5 model. The effects on the calculation results will be discussed subsequently.

**Element Partitioning and Calculations**

The partition coefficients \( D_{\text{a}} \) for mantle minerals use \( D_{\text{a}} \) values in water-rich mantle (Green et al., 2000); they are the same for ABS3 (reasons have been discussed elsewhere; Kimura et al., 2009). The \( D_{\text{a}} \) values for 26 trace elements based on open-system melting calculations are given in the Supplemental Tutorial [see footnote 2] and in the [OSM] worksheet in Supplemental Software [see footnote 1]. The same calculation strategy is used for Sr, Nd, Hf, and Pb isotopes in the [IOSM] worksheet.

**Major-Element Composition of the Mantle Melt**

The calculation scheme for the major-element composition is essentially consistent with that for trace elements but uses a different approach. The ABS5 model uses major-element compositions calculated by pMELTS for the slab-melt added DMM and PM mantle primarily used for \( X_{r} \) determination. This is an important function of the ABS4 and ABS5 models accessing major elements of the mantle melts because major elements are sensitive to shallow or slab-melt–fluxed silica saturation in addition to deep undersaturation or that occurring with a lower amount of slab melt. The ABS4 model first introduced this parameter for slab-melt–fluxed DMM; ABS5 extends this function to both PM and DMM. Experimental adjustment is also needed for the SiO2 and MgO contents of the melts due to the overestimation of the olivine stability in pMELTS at > 1.0 GPa. Corrections are made using:

\[
\text{SiO}_2\text{correct} = \text{SiO}_2 \times [0.05229 \times P\text{(GPa)} + 0.967133],
\]

and

\[
\text{MgOcorrect} = \text{MgO} \times [0.11458 \times P\text{(GPa)} + 1.113095],
\]
Figure 4. Melting relations and residual modal compositions for the depleted mantle. Top row: P-T slab-melt degree (F) relations of dry and water-bearing (0.1–3 wt%) depleted mantle. The calculations follow the parameterization in Katz et al. (2003). Lower five rows: residual mineralogical mode of depleted mantle and mantle fluxed by slab melt at 0, 5, 10, 15, and 20 wt% of the mantle at given P-T conditions. Once the mantle P-T and F are given, the modal compositions (X) are obtained. The degree of melting (0–56 wt%) and modal composition (0–1 fraction) are shown by the color codes; the gray field indicates the absence of phases. The thin white lines indicate isopleths. Abbreviations: Ol—olivine; Cpx—clinopyroxene; Opx—orthopyroxene; Gar—garnet; and Sp—spinel.
Due to the use of a fixed major-element composition for the slab melt and experiments. Please visit http://doi.org/10.1130/GEOS01468.S3 or the full-text article on www.gsapubs.org to view Supplemental Figure S1.}

Similar melting experiments have been examined in recent experiments (Mallik et al., 2016) that used a mixture of 75 wt% fertile KLB1 peridotite with 25 wt% slab melt and 2–6 wt% water contents. In that study, $F_{\text{perid}}$ and the major-element composition of their partial melts were determined. The experimental results showed that all major-element compositions were similar to those calculated by ABS5 at $P_{\text{perid}} = 2$ GPa, $T_{\text{perid}} = 1200 \, ^\circ\text{C}$–1300 $^\circ\text{C}$, and $F_{\text{perid}} = 18\%$–30% under olivine + orthopyroxene residue conditions (Supplemental Fig. S1†). In contrast, 3-GPa conditions in ABS5 stabilized a larger amount of garnet than those in the experiments. The Al$_2$O$_3$ and CaO concentrations were lower in ABS5 melt due to the high partitioning of these elements to garnet, which resulted in higher SiO$_2$ contents in the melts (Supplemental Fig. S1). The pMELTS model always stabilized garnet more than the experiments at $P_{\text {perid}} > 2.3$ GPa. Because of this discrepancy, ABS5 underestimated $P_{\text{perid}}$ in the higher-pressure range (2.3–3.0 GPa) in its mass balance. However, $P_{\text{perid}}$ estimated from ABS3–ABS5 was almost always <2.3 GPa (Kimura et al., 2010; Kimura et al., 2014). Therefore, this problem did not affect the ABS5 results in most cases. Due to the use of a fixed major-element composition for the slab melt and the discrepancy from the experimental results, the ABS4–ABS5 major-element calculations might contain errors. However, the results reasonably reproduce major-element compositions of primary arc magmas including adakitic dacite and high-Mg andesite (Kimura et al., 2014).

The ABS4 studies were conducted on the NE and SW Japan subduction zones (Kimura and Nakajima, 2014; Kimura et al., 2014). These two arcs represent extremes of old-cold and young-hot subduction in terms of tectonics and geodynamics (Kimura et al., 2014; Peacock and Syracuse et al., 2010; Wang, 1999). Moreover, their primary magma types differ, with low-K tholeiitic (LKT) and high-K calc-alkaline (HKCA) basalts found in the former and high-Mg andesite (HMA) and adakitic dacite (ADK) found in the latter (Kimura et al., 2014; Kimura et al., 2016). Southwest Japan is a potential analogue for an Archean subduction system with a high potential mantle temperature ($T_p = 1600 \, ^\circ\text{C}$), which produced adakitic tonalite-trondhjemite-granodiorite (TTG) continental crust (Kimura et al., 2016). The ABS3 model was applied to boninite (BON), a REE and Ti-depleted high-Mg andesite that erupted in the inception stage of the Bonin Arc at ca. 48 Ma (Li et al., 2013). These magma types cover much of the range of primary arc magmas worldwide.

**Figure 5.** Comparisons between model and experimental results of mantle melting. (A) Residual mode compositions of modified pMELTS and experimental results. (B) Melt compositions of modified pMELTS and experimental results. Comparisons are made for enriched mantle compositions. Adjustments are made for the olivine mode and SiO$_2$ and MgO in the melts. The experimental data are taken from the literature (Baker and Stolper, 1994; Falloon et al., 1999; Walter, 1998).

**ABS5 ANALYSIS**

The ABS4 studies were conducted on the NE and SW Japan subduction zones (Kimura and Nakajima, 2014; Kimura et al., 2014). These two arcs represent extremes of old-cold and young-hot subduction in terms of tectonics and geodynamics (Kimura et al., 2014; Peacock and Syracuse et al., 2010; Wang, 1999). Moreover, their primary magma types differ, with low-K tholeiitic (LKT) and high-K calc-alkaline (HKCA) basalts found in the former and high-Mg andesite (HMA) and adakitic dacite (ADK) found in the latter (Kimura et al., 2014; Kimura et al., 2016). Southwest Japan is a potential analogue for an Archean subduction system with a high potential mantle temperature ($T_p = 1600 \, ^\circ\text{C}$), which produced adakitic tonalite-trondhjemite-granodiorite (TTG) continental crust (Kimura et al., 2016). The ABS3 model was applied to boninite (BON), a REE and Ti-depleted high-Mg andesite that erupted in the inception stage of the Bonin Arc at ca. 48 Ma (Li et al., 2013). These magma types cover much of the range of primary arc magmas worldwide.
Figure 6. Melting relations and generated melt composition of the depleted mantle. Top row: $P-T-F$ relations of dry and water-bearing (0.1–3 wt%) depleted mantle. The calculations follow the parameterization of Katz et al. (2003). Lower five rows: melt compositions for SiO$_2$, Al$_2$O$_3$, FeO, MgO, and CaO. The degree of melting (0–56 wt%) and magma composition (0–60 wt%) is indicated by color codes; gray fields indicate the absence of phases. The addition of felsic slab melts to the mantle peridotite increases SiO$_2$ and Al$_2$O$_3$ in the peridotite melts, whereas MgO and CaO decrease with increasing slab-melt flux (lower five rows). The initial water content is assumed to be zero under all conditions shown. The effect of water is simply treated by increasing the degree of partial melting $F$ at a given $P-T-XH_2O$ (top row) to the lower five rows in Arc Basalt Simulator version 5 (ABS5) calculations. The thin white lines indicate isopleths.
The following subsections compare the magma genesis models using ABS5. The same primary magma and source material compositions were used for the comparison with ABS3 and ABS4 results (Table 2). The calculation method and primary magma compositions are reported elsewhere (Kimura and Nakajima, 2014; Kimura et al., 2014a; Li et al., 2013). The sensitivity to errors inherited from primary magma estimations is also examined below.

Cold Subduction Zone: NE Japan

The NE Japan arc has LKTH basalt in the volcanic front (VF) and HKCA basalt in rear-arc (RA) volcanoes. These basalts from the Iwate (VF) and Sannomegata (RA) volcanoes are well studied in terms of phase equilibria (Kuritani et al., 2014a; Kuritani et al., 2014b). Seismic tomography (Hasegawa et al., 1991; Nakajima and Hasegawa, 2003; Nakajima et al., 2005; Tsuji et al., 2008; Zhao et al., 1992) and receiver function (Kawakatsu and Watada, 2007) studies for this arc have led to its relatively well-understood substructure. The deep slab and mantle seismicity is related to slab dehydration; the migration of water through the slab and mantle has also been investigated (Kita et al., 2006; Nakajima et al., 2013; Shiona et al., 2013). The correlations of these geophysical observations and petrological and geochemical processes were examined in terms of the ABS5 model perspective (Kimura and Nakajima, 2014).

The new calculation results using the ABS5 model are given in Table 3; the results for the trace-element abundances and isotopic compositions are shown in Figure 7. The ABS5 values are indicated first in the results provided here and in the following subsections; the ABS4 values are provided second (in parentheses). The ABS5 model produces fairly good fits for the Iwate LKTH basalt compositions. The Iwate VF basalt (Figs. 7A and 7F–7I) was generated from a and the Sannomegata HKCA basalts in terms of major, trace, and isotope compositions. The ABS5 model produces fairly good fits for the Iwate LKTH basalt deduced from phase petrology (Umino et al., 2015). Because of this ancient and extreme subduction setting, the hot SW Japan slab geotherm was applied (Li et al., 2013; Umino et al., 2015); a cold subduction model cannot account for the BON chemistry.

Infant Subduction Zone: Bonin

Boninite (BON) reported from the island Izu-Bonin-Mariana Arc (ca. 48 Ma) is another type of HMA with an extremely depleted incompatible element composition (Ishizuka et al., 2006; Shiraki and Kuroda, 1977; Umino et al., 2015). Hot Subduction Zone: SW Japan

A high-Mg andesite (HMA) from the Setouch HMA and an adakitic dacite (ADK) from the Daisen Volcano were chosen as representative magmas from SW Japan. Both magma types were from the VF of the arc. The HMA (Figs. 7C and 7F–7I) was generated from a deep-slab flux $P_{\text{slab}} = 4.0$ (3.6) GPa at $T_{\text{SS}} = 922$ (907) °C. Mantle melting occurred deep at $P_{\text{perid}} = 2.2$ (1.7) GPa and $T_{\text{perid}} = 1234$ (1186) °C, which is somewhat similar to that in the NE Japan RA basalts. The slab liquid was melt rich with $X_{\text{H}_2\text{O}} = 10.3$ (17.3) wt% and results in a smaller degree of mantle melting $F = 5.8$ (78%)%. These conditions left a water-rich garnet-bearing mantle as the residue with $X_{\text{G}_\text{ar}} = 0.44$ (0.58), $X_{\text{Cpx}} = 0.16$ (0.12), and $X_{\text{Gar}} = 0.01$ (0.01). The estimated major elements were SiO$_2$ = 56.6 (61.0) wt% and MgO = 9.0 (6.0), which is consistent with the definition of HMA (Tatsumi, 1982). The ADK generation (Figs. 7D and 7F–7I) also required a deep-slab flux with $P_{\text{slab}} = 4.4$ (4.5) GPa and $T_{\text{SS}} = 943$ (945) °C. Mantle melting occurred at a depth similar to that of the HMA at $P_{\text{perid}} = 2.1$ (2.1) GPa and $T_{\text{perid}} = 1243$ (1207) °C. The slab liquid was water-bearing slab melt with $X_{\text{H}_2\text{O}} = 1.1$ (1.9) wt%, which resulted in the lowest degree of mantle melting with $F = 1.4$ (2.9)%. The slab-melt–rich flux left wehrlitic mantle as the residue with $X_{\text{O}_\text{px}} = 0.34$ (0.40), $X_{\text{Cpx}} = 0.18$ (0.13), and $X_{\text{Gar}} = 0.02$ (0.05). The results are almost identical to those from ABS4 (Table 3). The estimated major elements were SiO$_2$ = 61.3 (64.2) wt% and MgO = 5.0 (3.1) wt%. An MgO-rich dacitic composition with an extremely heavy rare-earth element (HREE)–depleted trace-element composition is mentioned elsewhere with respect to the mantle origin of the ADK (Defant and Drummond, 1990; Drummond and Defant, 1990).
TABLE 2. TARGET PRIMARY MAGMA COMPOSITIONS OF NE JAPAN, SW JAPAN, AND BONIN ARCS AND SOURCE MATERIAL COMPOSITIONS OF THE SUBDUCTED SLAB AND MANTLE WEDGE

<table>
<thead>
<tr>
<th>Region</th>
<th>NE Japan</th>
<th>SW Japan</th>
<th>Bonin (Ogasawara)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>IWT</td>
<td>SAN</td>
<td>ADK</td>
</tr>
<tr>
<td>SiO₂</td>
<td>49.72</td>
<td>47.5</td>
<td>63.43</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.71</td>
<td>0.82</td>
<td>0.67</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.86</td>
<td>15.47</td>
<td>17.67</td>
</tr>
<tr>
<td>FeO</td>
<td>10.42</td>
<td>9.38</td>
<td>1.10</td>
</tr>
<tr>
<td>MgO</td>
<td>13.04</td>
<td>13.94</td>
<td>2.56</td>
</tr>
<tr>
<td>CaO</td>
<td>9.26</td>
<td>9.8</td>
<td>5.02</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.67</td>
<td>1.93</td>
<td>4.25</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.06</td>
<td>0.79</td>
<td>2.00</td>
</tr>
<tr>
<td>H₂O</td>
<td>99.62</td>
<td>99.62</td>
<td>99.67</td>
</tr>
</tbody>
</table>

| Rb       | 0.979    | 15.000   | 0.004           | 0.050              |
| Ba       | 49.20    | 402.00   | 0.080           | 0.056              |
| Th       | 0.169    | 3.750    | 0.001           | 0.001              |
| U        | 0.044    | 0.925    | 0.001           | 0.001              |
| Nb       | 0.632    | 2.002    | 1.046           | 0.001              |
| Ta       | 0.035    | 0.001    | 0.002           | 0.001              |
| K        | 0.146    | 0.001    | 0.001           | 0.001              |

| K       | 521.0    | 611.0    | 16.0            | 80.0               |
| La       | 1.500    | 12.400   | 0.060           | 0.060              |
| Ce       | 3.800    | 24.600   | 0.193           | 0.193              |
| Pr       | 1.990    | 4.340    | 0.012           | 0.012              |
| Sr       | 238.00   | 401.00   | 0.026           | 0.026              |
| Nd       | 3.090    | 0.000    | 0.000           | 0.000              |
| Sm       | 1.040    | 3.300    | 0.130           | 0.130              |
| Eu       | 0.210    | 0.000    | 0.000           | 0.000              |
| Gd       | 1.370    | 3.280    | 0.030           | 0.030              |
| Tb       | 0.245    | 0.010    | 0.000           | 0.000              |
| Dy       | 1.630    | 3.100    | 0.048           | 0.048              |
| Y        | 8.80     | 16.10    | 0.290           | 0.290              |
| Ho       | 0.350    | 0.640    | 0.100           | 0.100              |
| Er       | 1.030    | 1.880    | 0.300           | 0.300              |
| Tm       | 0.150    | 0.050    | 0.000           | 0.000              |
| Yb       | 0.988    | 1.790    | 0.031           | 0.031              |
| Lu       | 0.152    | 0.270    | 0.050           | 0.050              |

| Rb/Sr    | 0.07419  | 0.07319  | 0.07290          | 0.07061             |
| Ba/Sr    | 0.51283  | 0.51296  | 0.51302          | 0.51302             |
| Th/U     | 30.360   | 33.330   | 33.330           | 33.330              |
| Nd/Sm    | 0.70426  | 0.70426  | 0.70426          | 0.70426             |
| Eu/Sm    | 0.02000  | 0.02000  | 0.02000          | 0.02000             |
| Gd/Yb    | 1.569    | 0.569    | 0.569            | 0.569               |
| Tb/Dy    | 4.577    | 4.577    | 4.577            | 4.577               |
| Ho/Lu    | 0.117    | 0.117    | 0.117            | 0.117               |

| Abbreviations: IWT—Iwate; SAN—Sannomegata; ADK—adakite from Daisen; HMA—high-Mg andesite from Setouchi; BON—low-Ca boninite from the Hahajima Seamount; X(SibP)—ocean plate lithosphere mantle; X(DIKE)—oceanic crust; X(AOC)—altered oceanic crust; X(SED)—slab sediment; X(Mwp)—mantle wedge peridotite. Data were taken from the literature (Kimura and Nakajima, 2014; Kimura et al., 2014; Li et al., 2013). Iron is given as FeO.
### Table 3: ABSS Calculation Results and Comparison with Results of the ABSS3 and ABSS4 Models

<table>
<thead>
<tr>
<th>Sample</th>
<th>IWT</th>
<th>SAN</th>
<th>HMA</th>
<th>ADK</th>
<th>BON</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVG/SD</td>
<td>AVG</td>
<td>1SD</td>
<td>ABS4</td>
<td>AVG</td>
<td>1SD</td>
</tr>
</tbody>
</table>

**ABS-fitting parameters**

- Slab P (Gpa)
  - 3.1 ± 0.2
- Slab Tsl (°C)
  - 742 ± 13
- %Slab
  - 148 ± 11
- Fm (OAC)
  - 0.66 ± 0.077
- Fm (SED)
  - 0.076 ± 0.037
- Fm (DMM)
  - 0.263 ± 0.096
- n (PERID)
  - 0.000 ± 0.000
- Pm (Gpa)
  - 1.5 ± 0.3
- Tsl (°C)
  - 1258 ± 38
- %MORB
  - 0.00 ± 0.00
- Fm (ABSS4)
  - 3.2 ± 0.4
- H2O% in SLAB liq
  - 28.6 ± 9.3
- H2O% in PERID
  - 0.91 ± 0.35
- F (%) PERID
  - 21.9 ± 0.9
- β (%)
  - 11 ± 2

**Major and trace elements**

- SiO2
  - 47.82 ± 1.65
- TiO2
  - 0.50 ± 0.01
- Al2O3
  - 12.40 ± 0.97
- FeO
  - 11.38 ± 1.41
- MgO
  - 16.24 ± 1.55
- CaO
  - 10.26 ± 0.36
- Na2O
  - 1.24 ± 0.00
- K2O
  - 0.13 ± 0.01
- H2O
  - 4.22 ± 1.74
- Rb
  - 2.329 ± 0.321
- Sr
  - 134 ± 17
- Nd
  - 3.073 ± 0.064
- Sm
  - 1.058 ± 0.027
- Eu
  - 0.413 ± 0.012
- Tb
  - 0.293 ± 0.009
- Dy
  - 2.099 ± 0.068
- Y
  - 13.240 ± 0.424

(continued...)

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**Note:** The table continues with similar entries for various chemical elements and their concentrations, along with their corresponding uncertainties. The data represent the chemical geodynamics of subduction zones, as analyzed by Kimura.
### ABS3 and ABS4 versus ABS5

Table 3 and Figure 8 summarize comparisons made between the ABS5 and ABS3 and ABS4 results for intensive and extensive geophysical variables for arc magma genesis. The ABS5 results were overall identical to those from ABS3 and ABS4. An important conclusion is that Pslab for arc magma genesis. The ABS5 results were overall identical to those from ABS3 and ABS4 versus ABS5.

The estimation of the primary magma composition is not an easy task and therefore is expected to contain errors (Kimura and Anskin, 2014). Thus, the ABS5 solutions should be affected by these errors. To test this uncertainty, the primary magma compositions of LKTH and HKCA basalts from Iwate and Sannomegata were artificially altered by multiplication by 0.8 and 1.2.

### Sensitivity of the ABS5 Model

The range of each window is manually provided by the operator. The calculation residues are confined by applying fitting windows due to the extremely narrow conditions for successful fits. In the Monte Carlo calculations, the calculation residues are confined by applying fitting windows to each element and/or isotope abundance in the [CONTROL PANEL] worksheet. The range of each window is manually provided by the operator. The calculation results that match all fitting windows are recorded in the [SUMMARY] worksheet. Further confinement is manually possible by sorting the summary results. See details of the calculations in the Supplemental Tutorial [see footnote 2].

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**Table 3. ABS5 Calculation Results and Comparison with Results of the ABS3 and ABS4 Models (continued)**

<table>
<thead>
<tr>
<th>Sample</th>
<th>IWT</th>
<th>SAN</th>
<th>HMA</th>
<th>ADK</th>
<th>BON</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVG/SD</td>
<td>AVG</td>
<td>1SD</td>
<td>ABS4</td>
<td>AVG</td>
<td>1SD</td>
</tr>
<tr>
<td>Major and trace elements (continued)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ho</td>
<td>0.475</td>
<td>0.015</td>
<td>0.317</td>
<td>0.702</td>
<td>0.046</td>
</tr>
<tr>
<td>Er</td>
<td>1.431</td>
<td>0.044</td>
<td>0.961</td>
<td>2.030</td>
<td>0.155</td>
</tr>
<tr>
<td>Sm</td>
<td>0.222</td>
<td>0.006</td>
<td>0.153</td>
<td>0.294</td>
<td>0.026</td>
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<tr>
<td>Eu</td>
<td>1.437</td>
<td>0.040</td>
<td>1.005</td>
<td>1.792</td>
<td>0.182</td>
</tr>
<tr>
<td>Yb</td>
<td>0.228</td>
<td>0.006</td>
<td>0.159</td>
<td>0.290</td>
<td>0.031</td>
</tr>
<tr>
<td>Lu</td>
<td>0.228</td>
<td>0.006</td>
<td>0.159</td>
<td>0.290</td>
<td>0.031</td>
</tr>
<tr>
<td>Isotopes</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$^{187}$Os/$^{187}$Nd</td>
<td>0.51297</td>
<td>0.00004</td>
<td>0.51290</td>
<td>0.51290</td>
<td>0.00005</td>
</tr>
<tr>
<td>$^{207}$Pb/$^{206}$Pb</td>
<td>18.507</td>
<td>0.066</td>
<td>18.611</td>
<td>18.539</td>
<td>0.058</td>
</tr>
<tr>
<td>$^{30}$Ar/$^{38}$Ar</td>
<td>15.624</td>
<td>0.013</td>
<td>15.668</td>
<td>15.596</td>
<td>0.016</td>
</tr>
<tr>
<td>Residual mantle mode</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ol</td>
<td>0.653</td>
<td>0.045</td>
<td>0.645</td>
<td>0.599</td>
<td>0.009</td>
</tr>
<tr>
<td>Opx</td>
<td>0.321</td>
<td>0.042</td>
<td>0.340</td>
<td>0.287</td>
<td>0.007</td>
</tr>
<tr>
<td>Cpx</td>
<td>0.002</td>
<td>0.004</td>
<td>0.007</td>
<td>0.164</td>
<td>0.002</td>
</tr>
<tr>
<td>Gar</td>
<td>0.000</td>
<td>0.000</td>
<td>0.004</td>
<td>0.005</td>
<td>0.004</td>
</tr>
<tr>
<td>Sp</td>
<td>0.004</td>
<td>0.001</td>
<td>0.004</td>
<td>0.035</td>
<td>0.001</td>
</tr>
<tr>
<td>H$_2$O/Ce</td>
<td>7852</td>
<td>2690</td>
<td>1463</td>
<td>441</td>
<td>8916</td>
</tr>
</tbody>
</table>

Abbreviations: ABS5—Arc Basalt Simulator version 5; IWT—Iwate; PERID—peridotite; SAN—Sannomegata; ADK—adakite from Daisen; HMA—high-Mg andesite from Setouchi; BON—low-Ca boninite from the Hahajima Seamount; AVG—average; 1 SD—one standard deviation; ABS4—results from ABS4 (Kimura et al., 2014; Kimura and Nakajima, 2014); ABS3—results from ABS3 (Li et al., 2013); Ol—olivine; Opx—orthopyroxene; Cpx—clinopyroxene; Gar—garnet; Sp—spinel. See acronyms and parameters in Table 1 and in the text. Iron is given as total FeO.

Note: Boldface values are used for calculated results.
Figure 7. Arc Basalt Simulator version 5 (ABSS) model calculation results and comparison with target magmas. (A)–(E) show trace elements. The black solid lines indicate the target magmas; red lines with open circles indicate ABSS calculation results; and thin red lines indicate minimum and maximum values. The values are normalized using the primitive mantle composition results of Sun and McDonough (1989). (F)–(I) show isotopes. The solid symbols represent observed values, and open symbols represent ABSS calculations. The error bars present two standard deviations. The error bars of the isotopic compositions from the ABSS model represent ranges (minimum and maximum) of the calculated values. Abbreviations: Min—minimum; Max—maximum; Avg—average; OBS—observed; CALC—calculated; LKTH—low-K tholeiite; MKCA—medium-K calc-alkaline; HMA—high-Mg andesite; ADK—adakitic dacite; BON—boninite.
This simulates the errors of the olivine maximum fractionation at ±20%, which covers the entire range of olivine fractionation and/or accumulation during the estimation of primitive magmas. These artificial and original basalts were calculated by ABS5 using the same fitting strategy (e.g., width of fitting windows). The results are shown in Table 4; representative results are given in Figure 9. Most of the resultant intensive and extensive variables were not affected. However, the mantle melting temperature \( T_{\text{perid}} \) systematically differed in LKTH, showing lower values for high-abundance basalt (IWT+20) and higher values for low-abundance basalt (IWT–20). In contrast, \%MORB_{\text{ext}} \) differed in HKCA, indicating a less depleted source for high-abundance basalt (SAN+20) and a more depleted source for low-abundance basalt (SAN–20). These variations fully compensate the errors of the primary basalts. These differences were fairly small with \( T_{\text{perid}} = 1260 \pm 20 \) °C for LKTH and \%MORB_{\text{ext}} = 1 ± 1 wt% from DMM for HKCA (Fig. 9).

The melting temperature of the mantle is extremely sensitive to low degrees of partial melting (\( F_{\text{perid}} = 3\%–4\% \)) to keep garnet in the garnet residue at \( P_{\text{perid}} = 2.3 \) GPa. Therefore, \( T_{\text{perid}} \) is almost invariant to maintain \( T_{\text{perid}} \) for HKCA. In contrast, \( T_{\text{perid}} \) and thus \( F_{\text{perid}} \) represent the control of the incompatible element abundances for LKTH. A ±20% error of the trace-element composition is expected in natural rocks; it does not occur in back calculations in primary magma estimations (Kimura and Ariskin, 2014). The ABS5 model is robust against inherited errors of primary basalts.

Moreover, mantle-derived andesite to dacite magmas, such as BON, HMA, and ADK, are also considered in the ABS3–ABS5 models. The primary magma composition of these magmas is much harder to estimate than that of basalts (Kimura et al., 2014; Tatsumi and Hanyu, 2003). The mantle equilibrium with these silica-rich magmas was calculated with an internally consistent forward model in ABS5 for both DMM and PM sources. The ABS5 model can suggest coherence (or incoherence) of these magmas with the mantle sources.

**ORIGIN OF ARC MAGMAS**

The calculation results of the ABS5 model provide insight into both the source conditions of the slab and mantle and the elemental mass balance controlled by these conditions. The implications of these calculations for arc magma genesis are given below.
TABLE 4. ABS5 SENSITIVITY TEST FOR THE PRIMARY MAGMA COMPOSITION

<table>
<thead>
<tr>
<th>Sample</th>
<th>IWT</th>
<th>IWT-20</th>
<th>IWT+20</th>
<th>SAN</th>
<th>SAN-20</th>
<th>SAN+20</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVG/SD</td>
<td>AVG</td>
<td>1SD</td>
<td>AVG</td>
<td>1SD</td>
<td>AVG</td>
<td>1SD</td>
</tr>
</tbody>
</table>

### ABS-fitting parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slab P (GPa)</td>
<td>23.237</td>
</tr>
<tr>
<td>Slab T (°C)</td>
<td>110</td>
</tr>
<tr>
<td>F$_{\text{MORB}}$%</td>
<td>0.1</td>
</tr>
<tr>
<td>H$_2$O% in SLAB liq</td>
<td>17.3</td>
</tr>
<tr>
<td>H$_2$O% in PERID</td>
<td>0.46</td>
</tr>
<tr>
<td>F (%) PERID</td>
<td>21.7</td>
</tr>
<tr>
<td>Slab T (°C)</td>
<td>1262</td>
</tr>
</tbody>
</table>

### Major and trace elements

<table>
<thead>
<tr>
<th>Element</th>
<th>IWT</th>
<th>IWT-20</th>
<th>IWT+20</th>
<th>SAN</th>
<th>SAN-20</th>
<th>SAN+20</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVG/SD</td>
<td>AVG</td>
<td>1SD</td>
<td>AVG</td>
<td>1SD</td>
<td>AVG</td>
<td>1SD</td>
</tr>
</tbody>
</table>

(continued)

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parameters are given in Table 1.

0.235

1.483

0.024

1.458

0.000

0.228

1.165

0.037

1.634

0.126

1.861

0.155

1.678

0.099

2.261

0.123

Yb

1.483

0.080

1.218

0.038

1.620

0.093

1.589

0.182

1.492

0.151

1.910

0.139

Lu

0.235

0.013

0.193

0.006

0.257

0.015

0.256

0.031

0.239

0.026

0.310

0.024

HFSEs, including Nb, Ta, Zr, and Hf, are generated from mantle melts. In

that most medium rare-earth elements (MREEs) and HREEs (Nd-Lu) and

Fperidot of the mantle (\(F_{\text{peridot}}\)) without slab flux (Davidson, 1996; Pearce et al., 2005). Rationale

remaining light rare-earth elements (LREEs) and large ion lithophile elements

and high field strength elements (HFSE) represent mantle melt, and (2) the

mantle melt are: (1) element abundances of heavy rare-earth elements (HREEs)

Breakdown of the Mass Balance

The basic concepts of trace-element mass balance between slab liquid and

melt components of arc magmas. Figure 10 shows the magma, slab-fluid, source-mantle, and mantle-melt compositions calculated by ABS5. The

mantle-melt compositions are estimated from the given degree of melting of the

mantle (\(F_{\text{mantle}}\)) without slab flux (\(F_{\text{mantle-slab}} = 0\); Table 3). The results show that most medium rare-earth elements (MREEs) and HREEs (Nd-Lu) and

HFSEs, including Nb, Ta, Zr, and Hf, are generated from mantle melts. In contrast, LREEs (La-Pr) and LILeS are mostly produced in the slab. This is

true for both the shallow melting of a depleted mantle in the spinel stability field (LKT) and deep melting of relatively undepleted mantle in the garnet stability field (ADK). The exceptions, HMA and BON, showed Nb-Ta and Zr-Hf components dominated by slab liquids, resulting in elevated positive Zr-Hf anomalies and elevated Nb-Ta regions that characterize these high-Mg andesites.

The good fit of the Nb-Ta (and Zr-Hf) concentrations in the models is related
to the phase stability of TiO2 mineral phases (e.g., rutile and titanomagnetite) in subducted materials (Kimura et al., 2016; Xiong et al., 2011). The ABS5 model uses fixed partition coefficients for these minerals and modal compositions calculated from Perple_X phase equilibria using fixed source compositions. The stability of these minerals largely depends on the source compositions and impacts the behavior of these elements in magmas. Although the ABS5 model does not require adjustment in most cases, the code allows the adjustment of the modal TiO2 minerals. These can be made after fitting REEs and LILeS. The same assumption is made for zircon and allanite, which affect Zr-Hf and U-Th in the slab liquid, respectively. See details in the Supplemental Tutorial [see footnote 2].
Origin of the Arc Magma Signature

The abundances of LILEs, REEs, and some HFSEs in the slab liquid are strongly affected by $T_{\text{ss}}$, $F_{\text{perc}}$, $F_{\text{int}}$ (AOC), $F_{\text{int}}$ (SED), and $F_{\text{int}}$ (MwP), whereas HFSEs and HREEs are more strongly controlled by the composition of DMM/PM, %MORB<sub>ext</sub>, and the degree of melting of the mantle determined by $P_{\text{perc}}$, $F_{\text{perc}}$, and $F_{\text{int}}$.

Slab-Temperature Control

The residual slab minerals in the slab liquid consist of chlorite, epidote, zoisite, lawsonite, amphibole, clinopyroxene, garnet, TiO<sub>2</sub> minerals, and silica minerals in GAB, AOC, and SED. Serpentine, chrysotile, brucite, antigorite, chlorite, amphibole, olivine, clinopyroxene, and garnet were found in SlbP and MwP (Figs. 1 and 2). The stability of water-bearing minerals is strongly $T$ dependent at 0.5–6 GPa and 150 °C–1100 °C (Figs. 1 and 2) (Hacker, 2008; Hacker et al., 2003). Therefore, both dehydration and element redistribution in the slab are fundamentally controlled by $T_{\text{ss}}$ (Fig. 2). The $T$ dependence of the partition coefficient is also important for major slab minerals, such as garnet, clinopyroxene, and amphibole, because the $D$s value varies at least two orders of magnitude (Hermann and Spandler, 2008; Kessel et al., 2005), which accounts for the slab $T$ range of 150 °C–900 °C (Fig. 3), as noted above. Therefore, $T_{\text{ss}}$ controls the slab-liquid composition. Figure 10F clearly shows that the slab liquid from a deep and high-$T_{\text{ss}}$ slab beneath NE Japan is almost identical in these two cases and is the basic key in determining the geochemical arc signature.

The correlation of the arc magma chemistry with $T_{\text{ss}}$ has been discussed by examining a global arc magma database and geophysical variables. Some geochemical factors, such as $H_2O/Ce$, correlate with $T_{\text{ss}}$ (Cooper et al., 2012; Plank et al., 2009), whereas other factors, such as Na, Sr, HFSEs, and MREEs–HREEs,
Figure 10. Mass balance of the slab liquid and mantle melt calculated by the Arc Basalt Simulator version 5 (ABSS5) model. (A)-(E) Slab-derived liquid and mantle-derived melt are mixed in the open-system melting region in the mantle wedge to generate various arc magmas. The slab-liquid compositions also differ by arcs depending on the slab geotherm, which determines the residual mineral mode, defluxing profile, and partition coefficient of the minerals. The red (blue) lines indicate slab geotherms in SW (NE) Japan. See Table 1 for acronyms.
and their ratios do not correlate (Turner and Langmuir, 2015b; Turner et al., 2016). This is consistent with the ABS5 mass balance, which illustrates that most fluid mobile elements (arc signatures), including water and LILEs, are $T_{ss}$ controlled, whereas HFSEs and MREEs-HREEs are mantle controlled (Fig. 10).

### Slab-Material Control

The different contributions of slab materials are additional (subordinate) controlling factors for the signatures of arc magmas. The SED has arc geochemical signatures with elevated LILEs and depleted HFSEs and HREEs, which are inherent from its upper crustal sources (Plank and Langmuir, 1988; Plank et al., 2007). The MORB and fresh IOC have fairly uniform compositions (Jenner and O’Neill, 2012), with element abundances of approximately one order of magnitude higher than that of DMM (Workman and Hart, 2005). The AOC compositions differ from MORB but are only enriched in LILEs (e.g., Rb, U, and Sr) added from the seawater (Hirahara et al., 2015; Kelley et al., 2003). Both SED and IOC are sources of arc magma, as indicated by $F_{oc}(\text{AOC})$, $F_{oc}(\text{SED})$, and $F_{oc}(\text{MwP})$ in Table 3. The generation of the arc (upper crustal) signature is therefore a result of subduction zone element fractionation mainly due to the composition of SED and $D_{\text{bulk}}$ of the slab (Kimura et al., 2016; Porter and White, 2009). This is evident from the calculated compositions of the slab liquids and their large contributions to arc magmas (Figs. 10A–10F). Different contributions of slab materials lead to subtle differences in incompatible element abundances, as indicated by the similar trace-element patterns of LKTH, HKCA, HMA, ADK, and BON in Figure 10.

In contrast, the isotopic compositions of the slab materials are highly variable (Class and Lehnert, 2012; Plank and Langmuir, 1998). They cover almost the entire range of isotope variations of the Earth’s oceanic basin materials including DMM (MORB), terrigenous SED (EM II), OIB sourced lavas, and volcaniclastic sediments (EM I and HIMU). The ABS5 model uses Sr, Nd, Hf, and Pb isotopes to identify the contributions of these materials for mass balance calculations (Figs. 7F–7H). These compositions provide key information to identify the different contributions of slab materials when examining the mass fractions of $F_{oc}(\text{AOC})$, $F_{oc}(\text{SED})$, and $F_{oc}(\text{MwP})$ (Table 3). For example, recycled terrigenous SED dominates the SW Japan Arc (Kimura et al., 2014), whereas BON is generated from HIMU-type subducted volcaniclastics from adjacent Cretaceous OIBs (Li et al., 2013). The contribution of subducted ocean-island basalt from the Emperor Seamount Chain is also discussed for the Kamchatka Arc (Portnyagin et al., 2015).

### Mantle-Wedge Conditions and Material Control

Figure 10 shows that the element abundances of HREE and HFSE and the element ratios between HREEs and HFSEs reflect the composition of the source mantle (PM or DMM and \%MOEB$_{ext}$) and its melting conditions ($T_{\text{mantle}}$, $P_{\text{mantle}}$, and $F_{\text{mantle}}$). A fertile PM source results in a flatter trace-element pattern of the mantle melt (Kimura et al., 2014), as indicated by the mantle-melt compositions for HMA and ADK in SW Japan (Figs. 10C and 10D). A greater degree of melting in HMA ($F_{\text{mantle}} = 5.8\%$) results in a more depleted LREE-LILE mantle melt compared with ADK ($F_{\text{mantle}} = 1.4\%$). The greater HREE depletion in ADK is the result of dominant residual garnet ($X_{\text{gar}} = 0.01$ for HMA and $X_{\text{gar}} = 0.02$ for ADK) at smaller degrees of melting at depth ($P_{\text{mantle}} = 2.1–2.2$ GPa) (Kimura et al., 2014).

In contrast, the REE and HFSE concentrations in the LKTH magmas from NE Japan require a DMM source and a relatively high degree of melting ($F_{\text{mantle}} = \sim 22\%$) in the shallow spinel stability field (Fig. 10A). More depleted DMM is needed to reproduce low HREE and HFSE abundances for HKCA mantle melt formed under lower degrees of mantle melting ($F_{\text{mantle}} \sim 4\%$) in a garnet stabilization field (Fig. 10B). More depleted mantle (\%MOEB$_{ext}$ = 4% from DMM) is also needed to account for the extremely low HREE abundances in LILE-HFSE-LREE ultra-depleted BON mantle melt (Fig. 10E). This is an advantage of ABS models using the degree of mantle depletion as a fitting parameter.

Recently, Turner and co-authors (Turner and Langmuir, 2015a, 2015b; Turner et al., 2016) proposed that the Moho depth and the proxy of the thermal structure of the slab ($\Phi = \text{slab age (Ma)} \times \text{convergence rate (mm/\text{y})} \times \text{dip angle (degree)}$ (Kirby et al., 1991)) are the main controls of arc magma chemistry (Turner and Langmuir, 2015b). Their latest discussion on the South Volcanic Zone of Chile used Na$_{eq}$ (Na$_2$O content of a magma at MgO = 6.0 wt%), which correlates with Ca$_{eq}$, Sr$_{eq}$, Nd$_{eq}$, Zr$_{eq}$, La/Sm, and Dy/Yb as functions of $F_{\text{mantle}}$ (and $P_{\text{mantle}}$) (Turner et al., 2016). They considered the effect of the source-mantle composition (DMM/PM) to be small (Turner and Langmuir, 2015b; Turner et al., 2016). These proposals agree with ABS5 analyses of HFSEs and MREEs-HREEs, which are mainly controlled by the mantle condition. However, the amount of water is an essential control of the slab $T_{ss}$ that affects the mantle $F_{\text{mantle}}$. The correlation found between fluid-immobile elements and $\Phi$ partially reflects the role of water and thus partial slab control.

In contrast, the Moho depth is supposed to be the primary control of the mantle conditions (Turner and Langmuir, 2015b; Turner et al., 2016). The thickness of cold arc crust essentially controls the mantle-wedge convection and determines the thickness of the mantle lithosphere, which is impermeable to arc magmas (Wilson et al., 2014). The ABS5 model has potential to provide information about the lithosphere thickness based on $P_{\text{mantle}}$ (Table 3 and Fig. 11) by resolving the complex roles of $P_{\text{mantle}}$, $T_{\text{mantle}}$, and $X_{H-2\text{operid}}$ in $F_{\text{mantle}}$. The application of this ability will be a subject in future studies for global arc systems. Notably, a shallow magma source for VF and a deeper source for RA have been estimated for the NE Japan Arc and are consistent with the seismic observations (Kimura and Nakajima, 2014) (Fig. 1A; see below).

As discussed above, the source mantle fertility (PM/DMM, \%MOEB$_{ext}$) and melting conditions, including depth and melting degrees ($P_{\text{mantle}}$, $T_{\text{mantle}}$, $F_{\text{mantle}}$, and $X_{\text{H-2\text{operid}}}$), can be successfully solved by examining the HREE and HFSE abundances and their ratios in AB5. In addition, the Sr-, Nd-, Hf-, and Pb-isotope compositions of the mantle wedge are important factors in constraining the contribution of the mantle melt flux to the arc magmas in balance with the slab-liquid flux fractions of SED, AOC, and MwP (Jacques et al., 2013; Jacques et al., 2014).
The estimation of the mantle composition (Table 3) is usually based on the isotope composition of a backarc basin basalt (BAB) (Kimura et al., 2010; Kimura and Nakajima, 2014) or a basalt unaffected by the slab component of an arc (Kimura et al., 2014). This assumption is consistent with the results from global arc magma analyses, which indicate the close proximity of the Nd isotope in VF magmas to the backarc basin magmas (Turner and Langmuir, 2015b).

**CHEMICAL GEODYNAMICS IN SUBDUCTION ZONES**

The ABS5 inverse model can estimate important intensive variables in the geodynamic means (Table 3). These values are immediately compared with geophysical observations and arc geodynamic constraints. The ABS5 model perspectives on the chemical geodynamics in subduction zones are given below.
Implications for the Slab-Surface Temperature

There is no way to estimate $T_{SS}$ by using a geophysical observation alone. The slab $V_p$ was constrained with a petrological model and estimated $T_{SS}$ to be $\sim 750$ °C at a slab-surface depth of 80 km in NE Japan (Kimura and Nakajima, 2014; Shiina et al., 2013; Tsuji et al., 2008) (Fig. 11A). The slab $V_p$ profile used for ABS4 and ABS5 (Syracuse et al., 2010; van Keken et al., 2011) are consistent with these seismological models in NE Japan (Kimura and Nakajima, 2014; Figs. 1 and 11A). Similarly, the petrological interpretation of the slab seismicity and corresponding slab $T_{SS}$ profile and dehydration (Omori et al., 2009) are consistent with the ABS5 model results for SW Japan (Fig. 11B) (Kimura et al., 2014).

The slab-surface temperature at the slab-liquid release points is $T_{SS} = 740$ °C–940 °C (Table 3 and Fig. 11). These numbers are almost identical to the $T_{SS}$ estimate based on H$_2$O/Ce systematics (Cooper et al., 2012; Plank et al., 2009). This indicates that the experimentally determined H$_2$O/Ce thermometers for the slab dehydration (Plank et al., 2009) and whole-element mass balance of ABS4 and ABS5 (Kimura et al., 2014) are mutually consistent. The H$_2$O/Ce values in primary magmas range from 440 to 8900 for warm to cold slabs of $T_{SS} = 940$ °C–740 °C based on ABS5. These are nearly equal to previously estimated H$_2$O/Ce values for $T_{SS} = 980$ °C–730 °C (Plank et al., 2009). In the ABS5 model, the slab defluxing pressures $P_{SS}$ vary between 2.7 and 4.5 GPa (Table 3). The close match of the $T_{SS}$ irrespective of the difference in $P_{SS}$, is attributed to the $T$-dependent dehydration and element partitioning and/or mobility in the slab, as noted above (Figs 2, 3, and 10).

Implications for the Mantle-Wedge Temperature

Corner flow develops in the mantle wedge and forms a hot mantle core asthenosphere in the middle of the mantle wedge (Fig. 1). These dynamics are controlled by the mantle rheology and are affected by the thermal structure and water distribution (van Keken et al., 2002; Wilson et al., 2014). A seismic study using Q-attenuation estimated the highest mantle-wedge temperature of $T_{PSS} = 800$ °C–1200 °C (Nakajima and Hasegawa, 2003). Geodynamic models estimated a $T_{PSS}$ of 1200 °C–1450 °C (Syracuse et al., 2010).

The temperature estimated by ABS5 for the mantle asthenosphere beneath NE and SW Japan varies from $T_{PSS} = 1229$ °C–1334 °C (Table 3 and Fig. 11). The mantle-wedge temperature estimates for NE Japan determined based on the phase petrology were also almost identical with $T_{PSS} = 1230$ °C–1250 °C (Kuritani et al., 2014a; Kuritani et al., 2014b). Overall, the estimated mantle-wedge temperatures were, in descending order, geodynamic model (1200 °C–1450 °C) > geochemical model (1200 °C–1350 °C) > seismic model (800 °C–1200 °C). The phase equilibria in mantle melting, used either in major- or trace-element models, are robust in terms of the experimental background. However, such a condition is localized in the mantle-wedge core, and the volume and distribution are unknown. In contrast, seismological observations are spatially diffused and represent the overall mantle-wedge temperature.

Implications for Water in the Mantle Wedge

The existence of excess water in the mantle wedge has been suggested by seismic structures, such as $V_p/V_s$, $V_p$ attenuation and receiver functions, in NE Japan. This region is slightly above and parallel to the slab surface at the 60–120 km depth range (Kawakatsu and Watada, 2007; Nakajima et al., 2013; Tsuji et al., 2008), as shown by the “$V_p$ jump zone” in Fig. 11A. The ABS4 and ABS5 models calculated the drastic drop in $V_p$ (and $V_s$) values in the IOC and MwP layers, corresponding to the explosive slab dehydration from the SED and/or IOC layers at ~70–110 km depth in NE Japan; this can explain the wet mantle above the slab detected by seismic observations (Kimura and Nakajima, 2014) (Fig. 11A inset). In SW Japan, a slow $V_p$ mantle occurs in the forearc wedge mantle toe at 40–80 km depth (Nakajima and Hasegawa, 2007), which correlates with the intensive slab dehydration from the hot slab in ABS4 and ABS5 (Kimura et al., 2014; Fig. 11B). This is consistent with seismological observations including the diminished intra-slab seismicity and the development of low $V_p$ mantle above the seismic slab (Omori et al., 2009).

In geophysical models, it is difficult to separate the effect of water in the molten mantle, although the amounts of mantle melt (total sum of melt and/or aqueous fluid) can be estimated using seismic attenuations (Nakajima and Hasegawa, 2003; Nakajima et al., 2013). The ABS4 and ABS5 models consider the net water in the mantle asthenosphere independent from $T_{PSS}$ and $P_{PSS}$. The water content in this region is $X_{H_2O,PSS} = 0.3$–0.9 wt% for LKTH and HKCA sources in NE Japan and the HMA source in SW Japan (Table 3 and Fig. 11). These values affect mantle melting to a larger degree, at $P_{PSS}$, than $T_{PSS}$ and $P_{PSS}$. The water content in the arc magmas was $X_{H_2O,arc} = 4$–6 wt% in all magma types including LKTH, HKCA, HMA, ADK, and BON (Table 3 and Fig. 11). These results are consistent with the average maximum water content of arc magmas derived from water in olivine melt inclusions, at $X_{H_2O,arc} = 4$–6 wt% (Plank et al., 2013). This unexpected match indicates a reasonable water mass balance in the ABS4 and ABS5 models, either in mantle melting or in the water content of slab liquids.

Implications for the Melting Depth of the Mantle Wedge

Inconsistency in the depth range of mantle melting is a limitation in seismological observations and geodynamic models. Detailed seismic tomography images of NE Japan indicate that the asthenospheric mantle occurs at depths >30 km (~1 GPa) beneath the VF and >60 km (~2 GPa) beneath the RA.
As shown in Figure 11A, the mantle asthenosphere occurs in the upper half of the mantle wedge (Nakajima et al., 2005). In contrast, many geodynamic models indicate that the asthenosphere appears in the lower half of the mantle wedge (Hebert et al., 2009; van Keken et al., 2002; Wada et al., 2012), although the latest geodynamic model suggests that water transport might control the location of the asthenosphere (Wilson et al., 2014).

The ABS4 model indicates $P_{\text{perid}} = 1.0$–2.2 GPa for NE Japan and 1.6–2.5 GPa for SW Japan. The ABS5 model implies $P_{\text{perid}} = 1.5$–2.3 GPa for these arcs with an exceptionally shallow 0.6 GPa for the Izu-Mariana Infant Arc for BON (Table 3 and Fig. 11). The petrological model showed the depth of mantle melting at $P_{\text{perid}} = 1.2$–1.8 GPa in NE Japan (Kuritani et al., 2014a; Kuritani et al., 2014b), consistent with the ABS5 results. All of the petrological and geochemical models are based on phase equilibria and are independent of $T_{\text{perid}}$ and $X_{\text{H}_2\text{O}}$. The ABS3–ABS5 models have an extra advantage with respect to depth information of $P_{\text{perid}}$ and $T_{\text{perid}}$ (Fig. 11). These models are able to determine the average $P\cdot T$ of the molten mantle core from which the primary arc magma coalesces, or alternatively, above the melting region, where the melt finally equilibrates with the molten mantle. The ABS5 model generally supports seismological observations for the depth of the upper bounds of the mantle asthenosphere (Kimura and Nakajima, 2014; Kimura et al., 2014).

Unfortunately, the resolution of seismic images in SW Japan is poor because the VF is located close to the Sea of Japan coastline (Nakajima and Hasegawa, 2007). Instead, tomographic images of the Kyushu Arc show a correlation between discrete volcanic clusters and low-$V_s$ and $V_p$ mantle-wedge bodies (Zhao et al., 2012) with high attenuation (Saita et al., 2015). The mantle-derived magmas in that region are HKCA with some HMA and ADK (Shibata et al., 2013; Shinjo et al., 2000), which is intermediate between SW Japan and NE Japan and reflects the intermediate tectonic setting (Syracuse et al., 2010). The depth level of the slow mantle at 150–50 km in Kyushu correlates well with that from ABS5 results for SW Japan, where similar HMA, ADK, and HKCA basalts are generated (Fig. 11B).

**Implications for the Magma Productivity**

The degree of melting of the wedge mantle is an important factor of the melt productivity. The petrological model using $\frac{d\ln V_s}{P}$ and $\frac{d\ln V_p}{P}$ indicates the porosity and pore shape in the NE Japan mantle wedge (Nakajima et al., 2005). The estimated melt fraction $F_{\text{perid}} = 0.05$% below the RA and 5.6% below the VF. The aspect ratios of the melt are 0.001 below the RA and 0.2 below the VF, indicating elongated pores beneath the VF in contrast to the pore bodies with equal dimensions beneath the RA.

The low melting porosity determined by the seismic model reflects the average value in the mantle wedge. In contrast, the larger values from petrology and geochemistry models reflect the local conditions around the magma segregation. The larger $F_{\text{perid}}$ beneath the VF might reflect water-dominated slab flux, which is attributed to the focused slab dehydration beneath the VF (Figs. 2 and 11). Melt-rich slab liquids in the RA or in the hot SW Japan VF might lead to melt pores of equal dimensions due to the increased dihedral angle. In the former, focused fluid channel flow is a plausible mechanism for slab-liquid transfer (Wilson et al., 2014). Diapiric transfer of melt-rich slab liquid (Tsuchiya and Kanisawa, 1994) and channelized flow transfer of supercritical liquid are plausible mechanisms in the latter (Kimura and Nakajima, 2014). The seismic mantle-wedge structure of NE and SW Japan differs. The former has a planar-shaped low-$V_p$ zone parallel to the slab surface (Nakajima et al., 2005), whereas the latter has massive round-shaped, low-$V_p$ high-attenuation zones rising from the slab (Saita et al., 2015; Zhao et al., 2012; Figs. 11A and 11B). These differences might reflect the mode of slab-liquid transportation.

The ABS5 model results show the degree of partial melting of $F_{\text{perid}} = 10$–25% beneath the VF and 3%–5% beneath the RA of NE Japan. The SW Japan mantle has a lower $F_{\text{perid}} = 2$%–5% beneath the VF. This difference reflects slab-fluid–dominant flux melting in the NE Japan VF in contrast to the melt-dominant flux melting in the NE Japan RA and SW Japan VF (Table 3 and Fig. 12A). A different approach using the element mass balance of REEs and Ti implies $F_{\text{perid}} = 7$%–23% for the VF in the Marianas Arc. The high $F_{\text{perid}}$ has been attributed to $X_{\text{H}_2\text{O}}$ in the mantle wedge based on the studies on olivine melt inclusions and modeling (Kelley et al., 2010) (Fig. 12). The ABS5 model results compare well with the results from Marianas Arc (Kimura and Nakajima, 2014; Kimura et al., 2014).

The magma productivity is essentially a function of $T_{\text{perid}}$, $P_{\text{perid}}$, and $X_{\text{H}_2\text{O}}$. The first two conditions are within the confined ranges of $T_{\text{perid}} = 1200$ °C–1350 °C and $P_{\text{perid}} = 1.0$–2.3 GPa shown above. In contrast, the water in the mantle varies by two orders of magnitude, with $X_{\text{H}_2\text{O}} = 0.8$–1.0 wt% (Fig. 12). The $X_{\text{H}_2\text{O}}$ therefore fundamentally controls the magma productivity in the colder arcs, such as NE Japan and the Marianas, where the old-cold Pacific plate subducts.

Slightly different features occur in hot SW Japan, where the slab-melt productivity dominates the amount of mantle melt in the primary magmas by a factor of three (HMA to four (ADK; Figs. 12A and 12B) (Kimura et al., 2014). However, the total magma productivity of most arcs is still overall $X_{\text{H}_2\text{O}}$ dependent, as indicated by the positive correlation between $X_{\text{H}_2\text{O}}$ and the total magma production, $F_{\text{perid}} + F_{\text{slab-liq}}$ calculated by ABS5 (Fig. 12B). This conclusion is important because (1) the total slab water influx to a subduction zone typically controls the magma productivity of an arc, and (2) the location of slab defluxing might control the spatial distribution of volcanoes and thus VF and the width of the magmatic arc.

**Implications for the Formation of a Volcanic Front**

The correlation between the slab-surface depth ($h = 120$ ± 20 km) and location of volcanic front (VF) (Gill, 1981) has been a matter of debate. Recent discussions about the formation of VF concern its relation or non-relation to the behavior of water in the mantle wedge. A first model considers the $T_{\text{perid}}$ control of the location of slab dehydration and thus the location of the VF (Cooper et al., 2012; Wilson et al., 2014). A second model attributes the VF location
to the location of chlorite breakdown in the lower part of the mantle wedge (Grove et al., 2009). A third model considers both the thermal structure in the mantle wedge and location of slab water supply (Syracuse and Abers, 2006). A fourth model considers the role of the lithosphere thickness that controls the mantle convection (Turner et al., 2016). A fifth model considers lithospheric lids for the location of melt coalescence in the relatively dry mantle lithosphere (England and Katz, 2010a, 2010b). The ABS5 models for NE and SW Japan have implications with respect to this fundamental geodynamic issue.

**Slab-Surface Temperature Does Not Indicate VF**

The first-order observation from the ABS5 model is that the overall slab deflux depth $P_{SS} = 89–149$ km is common in NE and SW Japan (Table 3 and Fig. 11). The depth range almost correlates with the slab-surface depth beneath the VF in subduction zones worldwide of $h = 80–196$ km, with a major peak at $90–150$ km (Syracuse and Abers, 2006; Syracuse et al., 2010; Fig. 11 left insets). This relation suggests that the slab defluxing depth, either fluid or melt, fundamentally controls the location of the VF. The $T_{SS}$ is less closely associated with the VF as demonstrated by, for example, the $T_{SS}$ value of 742 °C–766 °C at $P_{SS} = 2.9–4.0$ GPa beneath NE Japan, whereas $T_{SS} = 922 °C–950 °C$ at $P_{SS} = 3.7–4.5$ GPa beneath SW Japan, despite having a common VF location ($h = 90–150$ km; Fig. 11). The slab “defluxing” depth does not correlate with the slab dehydration depth because intensive dehydration takes place in the forearc region in SW Japan. As a reflection of the thermal structure, slab dehydration occurs in shallower regions of the SW Japan forearc, much closer to the trench than the VF, which is indicated by the slow-$V_p$ mantle toe (Nakajima and Hasegawa, 2007) and forearc hot spring water with a high $^3He/^He$ ratio, indicating the interaction of slab fluid with the overriding mantle (Kimura et al., 2014; Sano et al., 2009; Fig. 11B). In contrast, the NE Japan forearc has a relatively high-$V_p$ dry mantle toe (Kimura and Nakajima, 2014; Fig. 11A). The same applies for arcs worldwide based on seismological observations (McCory et al., 2014).

**Slab and Mantle Coupling Used to Locate VF**

The mechanism that locates VF depends on the location of slab “defluxing.” The ABS5 model suggests that all arc magmas are sourced in the mantle wedge, although the slab-fluid fraction varies widely from $\beta = 0.05$ in the mantle melt (BON) to $\beta = 0.11$ (LKTH), $\beta = 0.96$ (MKCA), $\beta = 1.54$ (HMA), and $\beta = 4.18$ (ADK; Table 3). These results indicate that mantle asthenosphere should commonly exist above the “defluxing” slab at $P_{SS} = 89–149$ km (Fig. 11).

The penetration of the mantle asthenosphere beneath arcs is accomplished by slab-mantle coupling, which creates corner flow in a steady-state geodynamic model (van Keken et al., 2002; Wada and Behn, 2015; Wada et al., 2012; Wilson et al., 2014). However, it is difficult to assume a common geo-dynamic configuration for NE and SW Japan, as noted above. In NE Japan, slab-mantle coupling is assumed to occur at a depth of $50–55$ km, as indicated by the depth limit of down-dip plate interface seismicity (Hasegawa et al., 2009; Kita et al., 2006). In contrast, the intra-slab upper plane seismicity of the double seismic zone continues down to a slab depth of $140$ km (Kita et al., 2010). The NE Japan magmatic arc is located between the two seismic structures at a slab depth of $50–140$ km (Fig. 11A). The depth range...
correlates with the region of slab dehydration for NE Japan from the ABS5 model (Fig. 11A, inset). Steady-state normal and planar corner flow induced by slab-mantle coupling might start at a slab depth of ~50–55 km. Penetration of the mantle asthenosphere is observed up to the VF location and is regarded as the upwelling flow portion of the corner flow (Hasegawa et al., 2005). This implies that the combination between the slab dehydration and asthenosphere corner flow determines both the location of the VF and width of the magmatic arc in NE Japan (Fig. 11A).

In contrast, the depth of slab melting appears to determine the VF in SW Japan. Interplate seismicity does not occur in the mantle beneath SW Japan (Seno and Yamasaki, 2003) beyond the depth of low-frequency tremors (Obara, 2002). Intensive slab dehydration occurs between the tremors and VF in SW Japan, where high 3He/4He forearc hot springs occur (Fig. 11B). The VF almost corresponds with the depth of diminishing intra-slab upper plane seismicity at a slab depth of ~100 km (Kimura et al., 2014; Nakajima and Hasegawa, 2007), the same as in Kyushu (Saita et al., 2015). At this slab depth, slab melting is supposed to occur based on the ABS5 model (Fig. 11B). Mantle corner flow can also be induced by the down-going of the aseismic slab (Zhao et al., 2012), which explains the high mantle \( T_{\text{perid}} \) of 1240 °C–1250 °C for ADK and HMA comparable with \( T_{\text{perid}} = 1230 \text{ °C–1250 °C} \) for LKTH and HKCA in NE Japan. The massive supply of buoyant slab melt might account for local diapir upwelling, which might lead to the even spacing of the ADK dacite volcanic centers in SW Japan controlled by the Rayleigh-Taylor instability (Kimura et al., 2014). Massive low-V mantle bodies beneath the volcanic clusters in Kyushu (Saita et al., 2015) would correlate with the diapirs (Fig. 11B). The high \( T \) in the mantle wedge supports mantle corner flow, even with lower-T slab-melt diapirs (\( T = 920 \text{ °C–950 °C} \), Fig. 11B). The mantle convection in the region forms complex downwelling corner flow and upwelling diapir flow. Such steep thermal (complex convection) and compositional (slab melt versus mantle) gradients would account for the diverse magma types, ADK, HMA, HKCA, high-Nb basalt, and even shoshonite, along the VF in the hot subduction zone of SW Japan (Kimura et al., 2014).

Interactions between the slab flux and mantle corner flow are common in NE and SW Japan arcs, although the slab flux varies from hydrous liquid (NE Japan) to melt (SW Japan), and the induced mechanism of the corner flow varies from hydrous slab coupling (NE Japan) to aseismic slab coupling (SW Japan; Fig. 11). It is difficult to constrain the exact h-VF relation by using a simple geodynamic model; however, a hydrous high-\( T \) mantle wedge is commonly needed. Case three (Syracuse and Abers, 2006) is the most preferred model; nevertheless, the mechanism is hard to constrain. The ABS5 results suggest that the first-order observation of the \( h-P_{\text{SS}} \) correlation in the depth range of \( h = 80–150 \text{ km} \) is critical. The VF can be located at any site above the \( h \) range, although the interaction with the mantle corner flow controls its location in a specific arc. The critical isotherm of the asthenosphere is \( T_{\text{perid}} = 1250 \text{ °C} \) ± 50 °C in the depth range of \( P_{\text{perid}} = 2.3–0.6 \text{ GPa} \) based on the ABS5 model.

The chemical geodynamic factors based on the ABS5 model can constrain the parameters of geodynamic models (Fig. 11). The combined roles of the slab, mantle, and overriding lithosphere are all key components of the model. The magma production rate is another interesting geodynamic factor that appears to be controlled by water from the slab (Fig. 12). Implications drawn from the ABS5 model with respect to the location of the volcanic front and other arc features need to be reviewed with some skepticism because they are highly dependent on the 1-dimensional (D) melting model. Future studies should include 3-D mantle geodynamics using either steady-state or instability models.

**CONCLUSIONS**

The ABS5 model is a spreadsheet-based numerical mass balance model that is useful for modeling the chemical geodynamics of subduction zones worldwide. It is a petrological and geochemical model that forwardly calculates the prograde metamorphism of the slab and provides slab-liquid compositions. Primary arc magma compositions are calculated based on an open-system, mantle-melting model that uses the slab liquid for flux melting. Thirty major- and trace-element concentrations, including H2O and Sr, Nd, Hf, and Pb isotopic compositions, in the subducted slabs SED, AOC, GAB, SlbF, wedge-mantle peridotite, and primary arc magmas are examined. Inversion calculations using this forward model revealed geophysical intensive and extensive variables in the slab and mantle. This allowed the determination of independent constraints of these important subduction conditions based on petrology and geochemistry information. The calculation results for various primary arc magmas, such as LKTH, HKCA, HMA, ADK, and BON, show that the slab liquids were released at \( T_{\text{SS}} = 740 \text{ °C–950 °C} \) and \( P_{\text{SS}} = 2.8–4.5 \text{ GPa} \). Mantle melting occurred at \( T_{\text{perid}} = 1229 \text{ °C–1370 °C} \) and \( P_{\text{perid}} = 2.3–0.6 \text{ GPa} \). The melting conditions were \( F_{\text{perid}} = 1.4%–28.2% \) and \( X_{\text{H2O\perid}} = 0.07–0.91 \text{ wt%} \) with slab-liquid fractions of \( F_{\text{slab-lq}} = 2.0–10.3 \text{ wt%} \). The resultant arc magma compositions reasonably matched those of natural primary magmas. The water contents of the magmas were \( X_{\text{H2O\mag}} = 3.07–6.10 \text{ wt%} \), which are consistent with those observed in olivine melt inclusions. This successful modeling allowed the further discussion of comparisons with geophysical observations of the slab and mantle beneath the cold NE Japan and hot SW Japan subduction zones. The results show that the location of the VF at \( h = 80–150 \text{ km} \) is essentially controlled by the depth range of slab defluxing and the penetration of the mantle asthenosphere, which is induced by corner flow induced by the slab. The mantle convection mode might differ between hot and cold subduction zones due to buoyancy-induced (diapirc) upwelling of slab melts in the former in contrast to downwelling corner flow dominant in the latter. The control of the magma productivity was inferred to be related to the amount of water derived from the slab and added to the mantle wedge to induce melting. This occurs because the \( RT \) conditions in the mantle wedge cover a narrow range, whereas the supplied slab water flux varies by two orders of magnitude and determines the melting degree of the mantle and slab. Overall, the ABS5 model demonstrates that the water in subduction zones controls many geo-
dynamic aspects of the arc magma genesis and its geodynamic background.
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