High-Magnesian Andesite Produced by Two-Stage Magma Mixing: a Case Study from Hachimantai, Northern Honshu, Japan

T. OHBA*, Y. KIMURA AND H. FUJIMAKI

DEPARTMENT OF MINERALOGY, PETROLOGY, AND ECONOMIC GEOLOGY, GRADUATE SCHOOL OF SCIENCE, TOHOKU UNIVERSITY, ARAMAKI AZA AOBA, AOBO-KU, SENDAI, 9808578, JAPAN

RECEIVED APRIL 6, 2006; ACCEPTED NOVEMBER 24, 2006; ADVANCE ACCESS PUBLICATION JANUARY 5, 2007

High-magnesian andesite occurs at Hachimantai, northern Honshu, Japan. Disequilibrium zoning features indicate that the phenocryst minerals were derived from three different magmas. Chemical compositions and zoning profiles are accounted for by two-stage magma mixing: the first mixing occurred between a crystal-free basalt magma and a more differentiated olivine basalt magma; the second stage occurred by mixing between the resultant of the first-stage mixing and a hypersthene-augite andesite magma. Mass balance of phenocryst crystals shows that end-member compositions were 52.5 wt % SiO2 and 10 wt % MgO for the mafic end-member and 57 wt % SiO2 for the felsic end-member of the second-stage mixing. Phenocryst minerals of the first-stage mixing end-member indicate the similarity of the end-member composition to that of basalts from nearby volcanoes. The counterpart aphyric magma in the first-stage mixing was more magnesian than the estimated mafic end-member. Calculations of the phase equilibria of similar basalts from nearby volcanoes and comparison of results with previous phase equilibrium experiments showed that the olivine basalt end-member of the first stage was hydrous and situated at a depth where the pressure was less than 2 kbar. Two-pyroxene thermometry estimates are about 1050°C for the pyroxenes derived from the felsic end-member of the second-stage mixing, and about 1180°C for groundmass pyroxenes. Crystallization temperatures of 1170–1230°C are estimated for minerals from the mafic end-member of the second-stage mixing based on phase equilibrium calculations. These similar temperature estimates between the groundmass and the mafic end-member imply achievement of thermal equilibrium between end-members preceding crystallization. The magma plumbing system of the eastern Hachimantai is illustrated by a recent volcanic event, involving lateral dike intrusion toward a pressure source. The encounter of a laterally migrating basalt dike and an andesite magma chamber triggered the magma mixing that produced the high-magnesian andesite. The model can account for the relation between the petrological model and surface distribution of volcanic rocks. The infrequency of such mixing-derived high-magnesian andesite stems from the rarity of high-magnesian basalt as a potential mixing end-member in northern Honshu.

KEY WORDS: high-magnesian andesite; Hachimantai; Northern Honshu; high-magnesian basalt; two-stage magma mixing

INTRODUCTION

The genesis of high-magnesian andesites is often studied separately from that of other andesites. A high-magnesian andesitic rock with SiO2 >52% and MgO >8% is classified as a boninite if TiO2 <0.5%. If TiO2 >0.5%, the rock name is determined by the total alkalies-silica (TAS) classification diagram (Le Bas, 2000). Here we simply use the term high-magnesian andesite for andesitic volcanic rocks with SiO2 >52% and MgO >8%, as 8 wt % MgO is high for an andesite rock regardless of its TiO2 content. The genesis of high-magnesian andesites is interesting because they have peculiar compositions that can be equilibrated with mantle peridotite. High-magnesian andesite magma is considered to be produced by several geological processes, including partial melting of hydrous mantle peridotite (Tatsumi, 1981, 1982), interaction between picrite-komatiite magma and the crust (Sparks, 1986), and interaction between basalt and subducting sediments (Shimoda et al., 1998). High-magnesian

*Corresponding author. Telephone: +81-22-795-6636. Fax: +81-22-795-6673. E-mail: ohbatu@mail.tains.tohoku.ac.jp

© The Author 2007. Published by Oxford University Press. All rights reserved. For Permissions, please e-mail: journals.permissions@oxfordjournals.org
andesite production might not necessarily be accounted for by a single process, because many types of high-magnesian andesites exist with widely differing compositions (Crawford et al., 1989).

The Japanese islands are well known for the presence of high-magnesian andesites, but their occurrence is limited. Boninites occur only in the Izu-Bonin (Izu-Ogasawara) islands, whereas other high-magnesian andesites (sanukites) occur in the Setouchi district and the Seinan Islands. Nevertheless, they reportedly occur, although much less commonly, outside these high-magnesian andesite suites. For example, rare high-magnesian andesites were found at two locations in Hokkaido (Okamura et al., 1993; Wada & Goto, 1993). High-magnesian andesite also seems to be limited temporally as well as spatially. In the Japanese islands, high-magnesian andesites have been widely sampled from Tertiary volcanic rock units but not from Quaternary volcanoes. Andesite magmas are common in Japanese Quaternary volcanoes, but their MgO contents are typically much less than 6 wt %, except for those of one andesite sample from Hachimantai, northern Honshu (Ohba & Umeda, 1999); this andesite has an exceptionally high MgO content (7.1 wt %) for its SiO₂ content (54 wt %). In developing petrogenetic models it is important to study such rare samples that occur outside the main high-magnesian andesite suites.

Petrographically disequilibrium characteristics are common in andesites from northern Honshu, typically explained as the consequence of magma mixing. Although mixing between basic and felsic magmas can theoretically produce a magnesium andesite, high-magnesian andesite is very rare in northern Honshu. The magnesian andesite from Hachimantai also exhibits disequilibrium characteristics (Ohba & Umeda, 1999), suggesting that it is the product of magma mixing.

GEOLOGICAL SETTING
Hachimantai is located at the volcanic front of the northern Honshu volcanic arc (Fig. 1a). Around Hachimantai, Quaternary composite cones and calderas are clustered in a c. 500 km² region called Sengan. The term Hachimantai originally indicated a small peak (1614 m) in the northern Sengan area, but is now commonly used as the name for the entire area. The volcanic field has been traditionally regarded as a single feature, the Hachimantai volcano. Notwithstanding, the field actually comprises several small cones of different ages (Suto, 1992; Ohba & Umeda, 1999). Here, we use the term ‘Hachimantai volcanic field’ for that area traditionally called the Hachimantai volcano.

The Hachimantai volcanic field comprises two major andesite cones (Hachimantai and Chausudake), two major basalt cones (Nishimoriyama and Maemoriyama), and numerous minor dacitic, andesitic, and basaltic cones.

Fig. 1. (a) Location of Hachimantai volcanic field, northern Honshu, Japan. (b) Simplified geological map of the eastern Hachimantai volcanic field (Ohba & Umeda, 1999). EBM, Ebisumori; HCM, Hachimantai (1614 m peak); CD, Chausudake; DM, Daikokumori; YD, Yanomunedake; NY, Nishimoriyama; MY, Maemoriyama; APD, Appidake; 1, andesite (0.6–0.6 Ma); 2, dacite (younger than 0.7 Ma); 3, high-magnesian andesite; 4, basalt–basaltic andesite (0.3–0.4 Ma); 5, basalt (0.2–0.4 Ma).
in an east–west direction between Maemoriyama and northern flank of Chausudake.

Previous geochronological studies (paleomagnetic: Suto & Mukoyama, 1987; Suto, 1992; K–Ar studies: Ohba & Umeda, 1999; thermoluminescence (TL): Takahama et al., 2001; tephra: Doi, 2000) show that magmatism occurred in at least three periods: 0.7–1.0 Ma, 0.3–0.4 Ma, and 0.1–0.2 Ma. The lower Chausudake cone consists of andesites dated at 0.7–1.0 Ma (Ohba & Umeda, 1999). Dating studies using K–Ar and TL methods indicate that the upper Chausudake cone and the minor cones were mostly olivine-rich porphyritic andesite. Exposures of fresh volcanic rock are limited because of vegetation and hydrothermal alteration. Samples were collected from lava flows and lava blocks around the Ebisumori summit and along gullies on the cone surface. The Ebisumori cone is composed mainly of olivine-rich andesite and the collected samples were mostly olivine-rich porphyritic andesite.

Microscopic observation was carried out with a polarizing microscope and back-scattered electron (BSE) images were obtained by scanning electron microscopy (SEM; JSM 5410, JEOL) in the Graduate School of Science, Tohoku University. Microprobe analyses were performed by energy-dispersive spectroscopy (EDS; LINK ISIS, Oxford Co. Ltd) attached to the scanning electron microscope and wavelength-dispersive electron microscope analysis (WDS-EPMA; JXM 8800 Superprobe, JEOL) at Tohoku University. Polished thin sections were prepared for microscopic observations and microprobe analyses. Minerals were identified using a polarizing microscope for large silicate minerals and EDS-SEM for opaque and fine-grained silicate minerals. Point counting using a polarizing microscope was carried out for modal analyses. The oxide-ZAF correction method (Armstrong, 1988) was used for quantitative microprobe analysis. Line-scanning by EDS-SEM was also carried out on zoned crystals. Representative mineral compositions are reported in Table 1. The complete dataset is available as Supplementary Data, which can be downloaded from http://www.petrology.oxfordjournals.org.

Whole-rock compositions were determined using X-ray fluorescence spectroscopy (XRF; RIX2000, Rigaku Corp.) in the Graduate School of Science, Tohoku University, using the glass bead method. After grinding and ignition, 1 g of sample powder was mixed with 5 g of lithium tetraborate flux and fused at 1150°C. The resultant glass bead was irradiated using X-rays from a Rh–W dual-target tube. Accelerating voltage and current were 50 kV and 50 mA, respectively. Quantitative analyses of major (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, and P) and trace (Ni, Cr, Zr, Y, Nb, Rb, Sr, Ba, and V) elements were performed using calibration curves determined by measurement of standards from the GSJ Igneous Rock Series. The calibration curves were corrected using empirical matrix factors. Ferrous iron contents of whole-rocks were determined by KMnO4 titration. Weight loss after 110°C heating is recorded as H2O(−). The H2O(+) values were calculated from the loss on ignition (LOI) after correction of FeO for oxidation during ignition. Whole-rock compositions are reported in Table 2.

RESULTS

Petrography and mineral chemistry

Textures

The studied samples are porphyritic andesites containing phenocrysts of plagioclase, olivine, augite, hypersthene, titanomagnetite, ilmenite, and chromian spinel. Ilmenite is rare in the samples. Most phenocrysts are euhedral (polygonal or skeletal), indicating crystallization from a melt. Dissolution textures are common in plagioclase, shown by the presence of sieve textures or mottled cores and in pyroxenes by corroded cores surrounded by a euhedral mantle overgrowth. Phenocrysts are commonly 0.3–3 mm in size; chromian spinel is smaller (<0.1 mm) than other minerals. The groundmass comprises abundant small crystals (<0.05 mm) and minor glass. The groundmass mineral assemblage resembles that of the phenocrysts, except for the presence of pigeonite.

Phenocryst minerals occur both as aggregates and as discrete crystals. The aggregates comprise plagioclase, augite, hypersthene, and Fe–Ti oxides, or some combination of these minerals. Neither olivine nor chromian spinel occurs in the aggregates. In the aggregates, crystals are commonly included within each other, indicating their near-simultaneous crystallization from the melt. Olivine phenocrysts commonly contain spinel inclusions. Spinel inclusions are rarely included in discrete plagioclase phenocrysts. These spinel-bearing plagioclase phenocrysts never aggregate with pyroxene or Fe–Ti oxide minerals, but spinel-free plagioclase phenocrysts are commonly found in aggregates with these minerals.

Modal compositions vary among samples. Phenocryst abundance varies from 15 to 35 vol. % (Fig. 2).
Abundances of the different minerals vary with total phenocrystal abundance. Olivine abundance is weakly negatively correlated with total phenocryst abundance: phenocryst-poor rocks typically contain more olivine than phenocryst-rich rocks. Olivine reaches 5 vol. % in phenocryst-poor rocks, whereas only trace amounts occur in the highly porphyritic rocks. Conversely, the modal abundance of plagioclase increases from 9 to 21 vol. % with increasing total phenocryst proportion. The proportion of the other minerals is rather scattered, but increases roughly with the total phenocryst content.

**Plagioclase**

Plagioclase phenocrysts are categorized into one of two types: uniformly cored anorthitic plagioclase or complexly zoned plagioclase. The former are spinel-bearing plagioclases in which the core composition is more than An$_{90}$ (Fig. 3a). The anorthite cores are compositionally uniform and are surrounded by a thin rim of An$_{25}$ composition. The studied samples contain only traces of this type of plagioclase and the complexly zoned plagioclase is the most abundant phenocryst type. Each crystal is divisible into three zones: a chemically complex core, a homogeneous mantle of An$_{60}$ and a very thin rim of An$_{25}$ (Figs 3b and 4). The cores exhibit widely varying compositions of An$_{90}$–An$_{65}$ each crystal involves both calcic and sodic parts in its core. These cores exhibit various textures including mottled extinction, dusty zones, and clear crystals. The complexly zoned type plagioclase is either found in aggregates with pyroxenes and Fe–Ti oxide minerals or has these minerals as inclusions.

**Pyroxenes**

Pyroxene phenocrysts exhibit reverse zoning with iron-rich cores and more magheseian rims (Fig. 5). In a typical hypersthene phenocryst, a core with Mg-number ($[\text{Mg}]/(\text{Mg} + \text{Fe}) 	imes 100$) 55–58 is surrounded by a magnesian rim with Mg-number $= 73$. Similarly, augite phenocrysts show a core of Mg-number $= 59–65$, surrounded by a rim of Mg-number $= 79–80$.

### Table I: Representative mineral analyses

| Mineral   | Occurrence | Sample | SiO$_2$ | TiO$_2$ | Al$_2$O$_3$ | V$_2$O$_3$ | Cr$_2$O$_3$ | FeO$^*$ | MnO | MgO | CaO | NiO | Na$_2$O | K$_2$O | Total | Mg-no. | An | Cr/(Cr + Al) |
|-----------|------------|--------|---------|---------|-------------|-------------|-------------|---------|-----|-----|-----|-----|--------|--------|-------|-------|----------------|
| **Group 1** |
| plagioclase core EB-ch3 | 47.49 | 33.41 | - | - | 0.50 | - | - | 0.03 | 17.14 | - | 1.77 | 0.04 | 100.36 | 84.1 |
| core EB-ch3 | 54.10 | - | 27.95 | - | - | 0.51 | - | - | 0.02 | 11.40 | - | 4.80 | 0.19 | 99.03 | 56.1 |
| mantle EB-ch3 | 54.11 | - | 29.22 | - | - | 0.59 | - | - | 0.05 | 12.05 | - | 4.46 | 0.18 | 100.72 | 59.3 |
| rim EB-ch3 | 50.22 | - | 30.00 | - | - | 0.93 | - | - | 0.18 | 14.80 | - | 2.80 | 0.07 | 98.99 | 74.5 |
| augite core EB-ch1 | 51.07 | 0.41 | 1.31 | n.d. | - | 12.42 | 0.44 | 12.18 | 22.13 | - | n.d. | - | 99.96 | 63.6 |
| rim EB-ch1 | 50.79 | 0.51 | 3.49 | n.d. | - | 6.93 | 0.31 | 14.91 | 22.66 | - | n.d. | - | 99.60 | 79.3 |
| hypersthene core EB-ch1 | 52.31 | 0.22 | 0.62 | - | - | 26.53 | 0.73 | 18.87 | 1.69 | - | - | 100.86 | 56.9 |
| rim EB-ch1 | 54.48 | 0.17 | 1.41 | - | - | 16.00 | 0.42 | 25.24 | 2.27 | - | - | 99.99 | 73.8 |
| magnetite EB-ch2 | n.d. | 14.54 | 2.53 | 0.95 | 0.22 | 75.16 | 0.30 | 1.27 | - | - | - | - | 95.87 | 5.0 | 0.06 |
| olivine core EB-to0 | 39.97 | - | n.d. | - | - | 12.82 | 0.16 | 46.72 | 0.15 | 0.24 | - | - | 100.65 | 86.7 |
| rim EB-to0 | 38.01 | - | n.d. | - | - | 21.47 | 0.31 | 38.95 | 0.19 | 0.08 | - | - | 99.01 | 76.4 |
| spinel core EB-sk3 | 0.56 | 23.63 | - | 34.47 | 34.92 | 0.35 | 5.77 | n.d. | 0.12 | - | - | 99.81 | 22.8 | 0.49 |
| rim EB-sk3 | 0.05 | 6.19 | 4.34 | - | 20.06 | 63.22 | 0.36 | 0.97 | 0.02 | n.d. | - | - | 95.21 | 2.7 | 0.76 |
| inclusion EB-sk3 | 0.05 | 0.40 | 21.82 | - | 37.65 | 26.01 | 0.27 | 12.13 | n.d. | 0.14 | - | - | 98.48 | 46.4 | 0.54 |
| **Group 2** |
| plagioclase core EB-ch3 | 44.97 | - | 35.32 | - | - | 0.55 | - | n.d. | 18.54 | - | 0.96 | 0.01 | 100.35 | 91.5 |
| rim EB-ch3 | 50.05 | - | 30.33 | - | - | 0.85 | - | - | 0.16 | 15.06 | - | 2.80 | 0.09 | 99.35 | 74.8 |
| olivine core EB-sk3 | 38.81 | - | n.d. | - | - | 20.76 | 0.31 | 40.93 | 0.12 | 0.05 | - | - | 100.98 | 77.9 |
| mantle EB-sk3 | 39.99 | 0.04 | - | - | 14.58 | 0.20 | 45.70 | 0.16 | 0.15 | - | - | 100.80 | 84.8 |
| rim EB-sk3 | 38.46 | 0.04 | - | - | 20.77 | 0.30 | 40.48 | 0.17 | 0.09 | - | - | 100.30 | 77.7 |
| spinel inclusion EB-ak5 | 0.49 | 36.96 | - | 16.48 | 31.96 | 0.17 | 12.82 | 0.06 | 0.01 | - | - | - | 98.95 | 41.7 | 0.23 |

*Total iron as FeO.
Mg-number $= \text{Mg}/(\text{Mg} + \text{Fe}^{2+}) \times 100$; –, not determined; n.d., not detected.
### Table 2: Whole-rock analyses of Ebisumori andesites

<table>
<thead>
<tr>
<th>Sample</th>
<th>EB-ch1</th>
<th>EB-ch3</th>
<th>EB-ch5</th>
<th>EB-ch6</th>
<th>EB-eb9</th>
<th>EB-eb11</th>
<th>EB-eb13</th>
<th>EB-sk3</th>
<th>EB-sk5</th>
<th>EB-sk7</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Major elements (wt %)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>54.2</td>
<td>54.28</td>
<td>54.78</td>
<td>54.46</td>
<td>54.97</td>
<td>54.74</td>
<td>55.79</td>
<td>55.47</td>
<td>55.27</td>
<td>54.94</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.66</td>
<td>0.63</td>
<td>0.65</td>
<td>0.64</td>
<td>0.71</td>
<td>0.69</td>
<td>0.68</td>
<td>0.7</td>
<td>0.69</td>
<td>0.69</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>15.51</td>
<td>15.56</td>
<td>15.71</td>
<td>15.68</td>
<td>16.04</td>
<td>16</td>
<td>16.42</td>
<td>16.68</td>
<td>16.32</td>
<td>16.7</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>3.74</td>
<td>3.66</td>
<td>3.79</td>
<td>4.42</td>
<td>3.79</td>
<td>5.17</td>
<td>3.13</td>
<td>3.79</td>
<td>5.5</td>
<td>4.13</td>
</tr>
<tr>
<td>FeO</td>
<td>4.89</td>
<td>5.06</td>
<td>5.27</td>
<td>4.49</td>
<td>5.04</td>
<td>4</td>
<td>5.32</td>
<td>4.82</td>
<td>3.34</td>
<td>4.5</td>
</tr>
<tr>
<td>MnO</td>
<td>0.16</td>
<td>0.16</td>
<td>0.15</td>
<td>0.15</td>
<td>0.15</td>
<td>0.16</td>
<td>0.15</td>
<td>0.15</td>
<td>0.15</td>
<td>0.15</td>
</tr>
<tr>
<td>MgO</td>
<td>7.3</td>
<td>8</td>
<td>7.47</td>
<td>7.58</td>
<td>6.15</td>
<td>6.41</td>
<td>6.2</td>
<td>6.19</td>
<td>6.42</td>
<td>6.49</td>
</tr>
<tr>
<td>CaO</td>
<td>8.84</td>
<td>9.11</td>
<td>8.73</td>
<td>8.99</td>
<td>8.94</td>
<td>9.06</td>
<td>8.69</td>
<td>7.97</td>
<td>8.3</td>
<td>8.65</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2</td>
<td>1.99</td>
<td>2.07</td>
<td>2.06</td>
<td>2.16</td>
<td>2.28</td>
<td>2.1</td>
<td>2.2</td>
<td>2.14</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>0.61</td>
<td>0.57</td>
<td>0.62</td>
<td>0.62</td>
<td>0.57</td>
<td>0.8</td>
<td>0.75</td>
<td>0.8</td>
<td>0.71</td>
<td></td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.06</td>
<td>0.07</td>
<td>0.06</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>H₂O(+)</td>
<td>0.95</td>
<td>0.96</td>
<td>0.92</td>
<td>0.71</td>
<td>0.68</td>
<td>0.42</td>
<td>0.19</td>
<td>1.03</td>
<td>0.02</td>
<td>0.28</td>
</tr>
<tr>
<td>H₂O(−)</td>
<td>0.63</td>
<td>0.73</td>
<td>0.51</td>
<td>0.52</td>
<td>0.76</td>
<td>0.45</td>
<td>0.5</td>
<td>0.84</td>
<td>1.43</td>
<td>1.12</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>99.47</td>
<td>100.76</td>
<td>100.47</td>
<td>100.35</td>
<td>100.08</td>
<td>100.18</td>
<td>100.22</td>
<td>100.54</td>
<td>100.51</td>
<td>100.65</td>
</tr>
<tr>
<td><strong>Trace elements (ppm)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ni</td>
<td>118</td>
<td>143</td>
<td>174</td>
<td>130</td>
<td>79</td>
<td>80</td>
<td>64</td>
<td>68</td>
<td>66</td>
<td>68</td>
</tr>
<tr>
<td>Cr</td>
<td>372</td>
<td>403</td>
<td>374</td>
<td>349</td>
<td>241</td>
<td>254</td>
<td>253</td>
<td>259</td>
<td>267</td>
<td>273</td>
</tr>
<tr>
<td>Zr</td>
<td>71</td>
<td>67</td>
<td>68</td>
<td>75</td>
<td>83</td>
<td>82</td>
<td>87</td>
<td>92</td>
<td>90</td>
<td>87</td>
</tr>
<tr>
<td>Y</td>
<td>18</td>
<td>25</td>
<td>31</td>
<td>20</td>
<td>27</td>
<td>23</td>
<td>25</td>
<td>26</td>
<td>22</td>
<td>23</td>
</tr>
<tr>
<td>Nb</td>
<td>3</td>
<td>3</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Rb</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
<td></td>
</tr>
<tr>
<td>Sr</td>
<td>183</td>
<td>186</td>
<td>186</td>
<td>186</td>
<td>196</td>
<td>207</td>
<td>216</td>
<td>221</td>
<td>219</td>
<td>226</td>
</tr>
<tr>
<td>Ba</td>
<td>203</td>
<td>211</td>
<td>206</td>
<td>225</td>
<td>224</td>
<td>208</td>
<td>208</td>
<td>208</td>
<td>215</td>
<td></td>
</tr>
<tr>
<td>V</td>
<td>203</td>
<td>204</td>
<td>204</td>
<td>196</td>
<td>207</td>
<td>223</td>
<td>216</td>
<td>221</td>
<td>228</td>
<td>227</td>
</tr>
</tbody>
</table>

(continued)
Round iron-rich cores overgrown by euhedral magnesian mantles are also apparent in back-scattered electron (BSE) images (Fig. 6). The cores are compositionally uniform and the magnesian rims are normally slightly zoned. The boundary between the core and rim is sharp because of the abrupt compositional gap. No compositional differences exist between pyroxenes in aggregates and discrete phenocrysts. Groundmass pyroxenes exhibit wide compositional ranges in which the most magnesian compositions are similar to those of the phenocryst rims. Groundmass pigeonite is more iron-rich (Mg-number = 65–51) than groundmass hypersthene (Mg-number = 75–65).

**Table 2**: Continued

<table>
<thead>
<tr>
<th>Sample: EB-894</th>
<th>F.E</th>
<th>M.E.</th>
<th>M.E.melt</th>
<th>IW40</th>
<th>IWL16</th>
<th>CHA1</th>
<th>CHA2</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>56.49</td>
<td>57.08</td>
<td>52.05</td>
<td>53.34</td>
<td>51.83</td>
<td>53.09</td>
<td>58.39</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.68</td>
<td>0.73</td>
<td>0.64</td>
<td>0.69</td>
<td>0.79</td>
<td>0.77</td>
<td>0.75</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>16.5</td>
<td>16.99</td>
<td>14.82</td>
<td>16.12</td>
<td>17.57</td>
<td>16.88</td>
<td>15.8</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>3.75</td>
<td>-</td>
<td>-</td>
<td>2.39</td>
<td>-</td>
<td>2.63</td>
<td>2.6</td>
</tr>
<tr>
<td>FeO</td>
<td>4.33</td>
<td>7.39</td>
<td>9.49</td>
<td>10.09</td>
<td>6.97</td>
<td>9.39</td>
<td>4.91</td>
</tr>
<tr>
<td>MnO</td>
<td>0.15</td>
<td>0.15</td>
<td>0.16</td>
<td>0.16</td>
<td>0.16</td>
<td>0.18</td>
<td>0.15</td>
</tr>
<tr>
<td>MgO</td>
<td>4.84</td>
<td>4.36</td>
<td>10.11</td>
<td>7.16</td>
<td>7.19</td>
<td>6.53</td>
<td>4.63</td>
</tr>
<tr>
<td>CaO</td>
<td>8.4</td>
<td>7.68</td>
<td>10</td>
<td>10.88</td>
<td>10.26</td>
<td>9.61</td>
<td>7.52</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.19</td>
<td>2.4</td>
<td>1.78</td>
<td>1.93</td>
<td>1.96</td>
<td>2.22</td>
<td>2.52</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.79</td>
<td>0.94</td>
<td>0.35</td>
<td>0.38</td>
<td>0.2</td>
<td>0.22</td>
<td>0.96</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.09</td>
<td>0.08</td>
<td>0.06</td>
<td>0.07</td>
<td>0.06</td>
<td>0.08</td>
<td>0.1</td>
</tr>
<tr>
<td>H₂O(+l)</td>
<td>1.2</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.93</td>
</tr>
<tr>
<td>H₂O(−l)</td>
<td>0.29</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.33</td>
</tr>
<tr>
<td>Total</td>
<td>99.69</td>
<td>97.79</td>
<td>99.46</td>
<td>100.00</td>
<td>99.4</td>
<td>98.97</td>
<td>100.00</td>
</tr>
<tr>
<td>Ni</td>
<td>31</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Cr</td>
<td>98</td>
<td>76</td>
<td>507</td>
<td>243</td>
<td>93</td>
<td>104</td>
<td>-</td>
</tr>
<tr>
<td>Zr</td>
<td>97</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Y</td>
<td>31</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Nb</td>
<td>5</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Rb</td>
<td>14</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Sr</td>
<td>251</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Ba</td>
<td>248</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>V</td>
<td>198</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

*Total Fe as FeO.*

F.E., estimated felsic end-member; M.E., estimated mafic end-member; M.E.melt, olivine-free melt of mafic end-member; IW40, olivine basalt from Iwate volcano (Ishikawa et al., 1984); IWL16, AD 1732 lava of Iwate volcano (Takagi et al., 2005); CHA1, CHA2, augite–hypersthene andesite from Chausudake.

**Fig. 2.** Modal abundance of each phenocryst mineral vs total phenocryst abundance.

**Fig. 3.** Plagioclase compositions: (a) complexly zoned plagioclase; (b) uniformly cored anorthitic plagioclase; (c) groundmass plagioclase.

The boundary between the core and rim is sharp because of the abrupt compositional gap. No compositional differences exist between pyroxenes in aggregates and discrete phenocrysts. Groundmass pyroxenes exhibit wide compositional ranges in which the most magnesian compositions are similar to those of the phenocryst rims. Groundmass pigeonite is more iron-rich (Mg-number = 65–51) than groundmass hypersthene (Mg-number = 75–65).
Olivine

Olivine phenocrysts commonly have compositionally uniform or slightly zoned Mg-rich cores (Fo\textsubscript{85–90}) surrounded by very thin more iron-rich rims (<20\,\mu m) that are normally zoned (Fo\textsubscript{70–80}); however, rarely more Fe-rich cores also occur (Fo\textsubscript{78–80}). Within the thin rim, the chemical composition continuously varies without any sharp boundary, in contrast to the discontinuous zoning exhibited by the pyroxenes. The olivines with more Fe-rich cores exhibit reverse zoning in which the cores are surrounded by thick magnesian mantles that have similar compositions to those of the normally zoned olivine cores (Fo\textsubscript{83–88}). The reverse zoning profile is gradual and broad; the magnesian mantles are as thick as typical normally zoned olivine phenocrysts (0.5–1\,\mu m). The olivine phenocrysts with Fe-rich cores have thin normally zoned rims (<20\,\mu m) outside their magnesian mantles that are similar in composition to the common olivine phenocrysts.

Oxide minerals

Typically, Fe–Ti oxide minerals occur as discrete phenocrysts, components of aggregates, inclusions in pyroxenes and plagioclase, and groundmass minerals. Titanomagnetite is the most common Fe–Ti oxide mineral, and ilmenite is rare. The core compositions of the titanomagnetite phenocrysts fall in the range Usp\textsubscript{43–50}. Groundmass titanomagnetite is similar to the phenocrysts in titanium content (Usp\textsubscript{35–39}). Ilmenite compositions are Ilm\textsubscript{81–88} Hm\textsubscript{12–19} in phenocryst cores, and Ilm\textsubscript{88–96} Hm\textsubscript{4–12} in the groundmass.

Chromian spinel occurs as discrete microphenocrysts and inclusions in magnesian olivine (Fo\textsubscript{85–90}) (Fig. 7). Chromian spinel inclusions exhibit no chemical zoning, whereas the discrete spinel microphenocrysts are strongly zoned (Fig. 6c). These chromian spinel inclusions are chromium-rich in terms of the chromium–aluminum ratio [Cr/(Cr+Al) = 0.56–0.77]. The cores of the zoned spinel microphenocrysts are similar in composition to the inclusions in olivine. The rims are rich in iron and chromium and can be classified as chromium-titanomagnetite. The rims differ compositionally from other titanomagnetite phenocrysts in terms of chromium abundance.

Chromium-poor spinel occurs as rare inclusions in iron-rich olivine and plagioclase. These inclusions are more aluminous [Cr/(Cr+Al) = 0.07–0.26] than the chromian spinel described above. The reversely zoned olivine phenocrysts include spinel in their iron-rich cores, whereas their magnesian mantles include chromian spinel. Plagioclase phenocrysts that contain spinel inclusions are anorthitic. The spinel inclusions in iron-rich olivine cores and anorthitic plagioclase are compositionally similar.

Whole-rock chemistry

The studied samples are rich in MgO (6–8 wt %) for their silica content (53.5–57 wt % SiO\textsubscript{2}, Table 1). These MgO-rich andesites are anomalous for northern Honshu,
where andesites typically contain 5 wt % MgO or less, and even the MgO contents of the basalts do not exceed 7 wt % (Fig. 8). Nickel contents at 64–177 ppm are also higher than those in typical basalts in the region (15–80 ppm). Similarly, the chromium contents of the high-Mg andesites (240–403 ppm) are much higher than those of typical basalts (15–150 ppm).

These high-MgO andesites do not belong to a boninite suite, although two samples have MgO contents as high as those of boninites (>8 wt % according to Le Bas [2000]). High-Mg samples contain more than 0.5 wt % TiO₂ in Japan this type of rock is known as sanukite, and typically occurs in the Setouchi district of southeastern Japan. Some high-Mg rocks from Ebisumori resemble sanukite rather than boninite in terms of their MgO and TiO₂ contents (Fig. 8). Other than the two high-magnesian samples, the remaining samples are of less magnesian andesite. Within the wide compositional spectrum of the studied rocks (Fig. 8), the low-magnesian samples are typical of the andesites of northern Honshu, whereas the high-MgO samples are compositionally similar to sanukite.

The Ebisumori samples define linear trends on Harker variation diagrams (Figs 8 and 9). Their SiO₂ contents are negatively correlated with MgO. The high-MgO samples are still andesites rather than basalts. Similarly, FeO, CaO, Ni and Cr contents show negative linear correlations with silica contents. Alkali elements (Na₂O and K₂O) are positively correlated with silica content (Fig. 9).

Silica contents correlate with increasing modal phenocryst abundances (Fig. 10). Figure 2 shows that there is a weak negative correlation between olivine abundance and total phenocryst content. Consequently, whole-rocks that are poorer in SiO₂ have more abundant olivine, and fewer phenocrysts overall. Accordingly, the high-magnesian rocks are poorer in phenocrysts than the low-MgO rocks. The lavas from Ebisumori exhibit a wide spectrum from olivine-rich, phenocryst-poor high-magnesian andesites to olivine-poor, highly porphyritic, low-MgO, more siliceous andesites.

![Fig. 6. Backscattered electron images of pyroxenes and spinel. (a) Reverse zoning in hypersthene. The euhedral magnesian (dark) rim surrounds the Fe-rich (bright) rounded core. (b) Reverse zoning of augite. The texture resembles that of hypersthene. (c) Zoned spinel microphenocryst. The bright rim is chromium titanomagnetite and the dark core is chromian spinel.](https://academic.oup.com/petrology/article-abstract/48/3/627/1497307/High-Magnesian-Andesite-Produced-by-Two-Stage)
Evidence of magma mixing

Phenocryst minerals can be classified into three groups based on paragenesis, zoning patterns, and mineral compositions. Group 1 involves complexly zoned plagioclase (An50^85), hypersthene (Mg-number 55^58), augite (Mg-number 59^65), titanomagnetite (Usp32^51), and ilmenite (ilm85^96). Group 2 consists of magnesian olivine (Fo 85^90) and chromian spinel \([Cr/(Cr + Al) = 0.07^0.26]\). Group 3 consists of aluminous spinel \([Cr/(Cr + Al) = 0.49^0.54]\) and ferrous olivine \(\text{Fo78}^\text{80}\). The rims of the phenocrysts have different composition in each of these groups. Groundmass minerals show variable compositions because of localized chemical differentiation of the residual melt during final-stage solidification. Rim compositions of the pyroxene phenocrysts are similar in composition to the groundmass pyroxenes (Mg-number 65-73 for hypersthene; Mg-number 79-80 for augite), indicating that rim formation occurred at an early stage of groundmass crystallization. The Mg-number of the olivine rims is 79-81, similar to that of the augite rims, implying that the \(K_d\text{aug}\) for Mg–Fe is about one at the early groundmass crystallization stage. Similarly, the compositions of the plagioclase rims are similar to those of the calcic groundmass plagioclase (An70-80).

These mineral groups are in mutual chemical disequilibrium. Groups 2 and 3 both include olivine and spinel, but of different composition, indicating that they have crystallized under different conditions. The pyroxenes of Group 1 are not in equilibrium with either of the olivines of Groups 2 and 3. According to previous studies of Fe–Mg partitioning between olivine and pyroxenes (Matsui & Nishizawa, 1974; Perkins & Viezeuf, 1992; Berman & Aranovich, 1996; Loucks, 1996; Kawasaki, 1999), the Mg/Fe ratios of these olivine phenocrysts are too high to be in equilibrium with pyroxene at magmatic temperatures of 1050–1180°C estimated by two-pyroxene geothermometry (Brey & Köhler, 1990). None of these three groups contains accidentally incorporated minerals from crustal rocks, as their euhedral shapes imply that all these minerals crystallized from a melt. Consequently, these three groups of minerals are considered to have originated from three different magmas.

Group 1 phenocrysts were derived from a magma that was cooler and more felsic than the erupted high-magnesian andesite magma. These phenocrysts are characterized by reverse chemical zoning. The zoning features of pyroxenes (Fig. 6) imply dissolution and overgrowth caused by a change in the ambient magma composition from lower Mg/Fe to higher Mg/Fe. An increase in oxygen fugacity is also a possible explanation for the reverse zoning of the pyroxenes (Luhr & Carmichael, 1980). The compositional gap in Mg/(Fe + Mg) between the rims and cores is c. 0.2, which requires an increase of oxygen fugacity of the order of 10^6. We do not have any evidence to negate the hypothesis that an oxygen fugacity change was the cause of the reverse zoning; however, we prefer compositional change by magma mixing, as there is other evidence for magma mixing as outlined in the following discussion. If the end-member magma compositions buffer the oxygen fugacity, magma mixing could result in an oxygen fugacity change, but an increase of 10^6 is simply too large. Similarly, reverse zoning from mantle (An90) to rim (An70-80) in plagioclase implies a change in ambient magma composition or conditions. Two-pyroxene geothermometer temperatures (Brey & Köhler, 1990) are c. 1050 ± 15°C for the cores of the pyroxenes, whereas the rims and groundmass give 1180 ± 26°C.

**DISCUSSION**

Evidence of magma mixing

Phenocryst minerals can be classified into three groups based on paragenesis, zoning patterns, and mineral compositions. Group 1 involves complexly zoned plagioclase (An50-85), hypersthene (Mg-number 55-58), augite (Mg-number 59-65), titanomagnetite (Usp32-51), and ilmenite (ilm85-96). Group 2 consists of magnesian olivine (Fo85-90) and chromian spinel [Cr/(Cr + Al) = 0.56-0.77]. Group 3 consists of aluminous spinel [Cr/(Cr + Al) = 0.07-0.26], anorthite (An92-94), and ferrous olivine (Fo78-80). The rims of the phenocrysts have different composition in each of these groups. Groundmass minerals show variable compositions because of localized chemical differentiation of the residual melt during final-stage solidification. Rim compositions of the pyroxene phenocrysts are similar in composition to the groundmass pyroxenes (Mg-number 65-73 for hypersthene; Mg-number 79-80 for augite), indicating that rim formation occurred at an early stage of groundmass crystallization. The Mg-number of the olivine rims is 79-81, similar to that of the augite rims, implying that the $K_d_{aug}$ for Mg–Fe is about one at the early groundmass crystallization stage. Similarly, the compositions of the plagioclase rims are similar to those of the calcic groundmass plagioclase (An70-80).

These mineral groups are in mutual chemical disequilibrium. Groups 2 and 3 both include olivine and spinel, but of different composition, indicating that they have crystallized under different conditions. The pyroxenes of Group 1 are not in equilibrium with either of the olivines of Groups 2 and 3. According to previous studies of Fe–Mg partitioning between olivine and pyroxenes (Matsui & Nishizawa, 1974; Perkins & Viezeuf, 1992; Berman & Aranovich, 1996; Loucks, 1996; Kawasaki, 1999), the Mg/Fe ratios of these olivine phenocrysts are too high to be in equilibrium with pyroxene at magmatic temperatures of 1050–1180°C estimated by two-pyroxene geothermometry (Brey & Köhler, 1990). None of these three groups contains accidentally incorporated minerals from crustal rocks, as their euhedral shapes imply that all these minerals crystallized from a melt. Consequently, these three groups of minerals are considered to have originated from three different magmas.

Group 1 phenocrysts were derived from a magma that was cooler and more felsic than the erupted high-magnesian andesite magma. These phenocrysts are characterized by reverse chemical zoning. The zoning features of pyroxenes (Fig. 6) imply dissolution and overgrowth caused by a change in the ambient magma composition from lower Mg/Fe to higher Mg/Fe. An increase in oxygen fugacity is also a possible explanation for the reverse zoning of the pyroxenes (Luhr & Carmichael, 1980). The compositional gap in Mg/(Fe + Mg) between the rims and cores is c. 0.2, which requires an increase of oxygen fugacity of the order of $10^6$. We do not have any evidence to negate the hypothesis that an oxygen fugacity change was the cause of the reverse zoning; however, we prefer compositional change by magma mixing, as there is other evidence for magma mixing as outlined in the following discussion. If the end-member magma compositions buffer the oxygen fugacity, magma mixing could result in an oxygen fugacity change, but an increase of $10^6$ is simply too large. Similarly, reverse zoning from mantle (An90) to rim (An70-80) in plagioclase implies a change in ambient magma composition or conditions. Two-pyroxene geothermometer temperatures (Brey & Köhler, 1990) are c. 1050 ± 15°C for the cores of the pyroxenes, whereas the rims and groundmass give 1180 ± 26°C.
(groundmass temperature is estimated for magnesian samples). This suggests that the Group 1 phenocryst assemblage experienced a heating event after initial crystallization. Titanomagnetite compositions might also be an indicator of reheating; however, we are not able to use these because of the variable compositions and complex zoning patterns.

Group 2 phenocrysts are interpreted as the products of a more magnesian magma than the erupted high-magnesian andesite magma. The olivine cores are too magnesian (Fo > 85) to be in equilibrium with the erupted magma, in which the olivine composition should be Fo 64–80 at 1050–1200°C based on Fe–Mg exchange equilibrium between augite rims and olivine (Loucks, 1996). Their thin rims (Fo > 80) are less magnesian than the cores and might be equilibrated with the erupted magma. Their zoning profile, characterized by a uniform magnesian interior surrounded by very thin, more iron-rich rim, can be explained by a change in the ambient magma composition from high Mg/Fe to low Mg/Fe or to a change in fO2. The zoning is normal, but the abrupt compositional change in the thin rims cannot be explained by the crystallization differentiation of interstitial residual melt.
Chromian spinel inclusions in olivine also imply that the high Mg/Fe magma was a chromium-rich magma able to crystallize olivine and chromian spinel. The chromian titanomagnetite rims of the chromian spinel phenocrysts are caused by a change in ambient conditions including temperature, magma composition and oxygen fugacity. It is likely that diffusion of Ti and Fe resulted in the zoning rather than continuous crystal growth, as the rims are richer in chromium than the Group 1 titanomagnetite rims and groundmass titanomagnetite.

The Group 3 phenocryst assemblage represents the crystallization products of a basalt magma that is fairly evolved, containing Fe-rich olivine, anorthitic plagioclase and aluminous spinel. Neither the olivine nor the anorthitic plagioclase has inclusions of the other phase and they do not form aggregates together. Both, however, contain inclusions of aluminous spinel. The magnesian mantles of the more Fe-rich olivines have similar compositions to the Group 2 olivines, which implies that the olivine crystals were mixed with the magnesian magma that produced the Group 2 phenocrysts, after crystallization from a lower Mg/Fe ratio magma. The assemblage and compositions of the Group 3 minerals resemble those of typical olivine basalts from northern Honshu. Phenocryst assemblages in olivine basalts from nearby volcanoes (Iwate, Akita Komagatake, and Hakoda; Kawano & Aoki, 1960; Kawano et al., 1961; Aoki, 1989) comprise anorthite, olivine (Fo70–80), and aluminous spinel. The Cr/(Cr+Al) of the aluminous spinel in the Group 3 assemblage is less than 0.14, which is similar to the spinel compositions in olivine basalts reported by Arai (1988). Olivine basalts occur commonly in the region. It is therefore inferred from the resemblance of the Group 3 assemblage to phenocryst assemblages in these locally abundant basalts that the Group 3 minerals were derived from a similar olivine basalt magma. The Group 3 magma is considered to be relatively evolved because those olivine basalts have evolved compositions with high FeO/MgO (typically greater than 1.3).

The characteristics of the phenocryst mineral assemblage can be accounted for by mixing between a felsic magma containing Group 1 phenocrysts and mafic magmas containing Group 2 and 3 phenocrysts. Although some other processes could produce the observed disequilibrium features, magma mixing alone can account for the correlation between the groundmass composition and crystal abundance (Fig. 11). When a crystal-bearing mafic magma mixes with a crystal-bearing felsic magma, the melt composition varies with the mixing ratio. In this case, the resultant melt composition must correlate with the proportion of mafic magma-derived crystals to felsic magma-derived crystals. Figure 11 shows a correlation between olivine modal abundance and calculated groundmass composition, suggesting that an olivine-bearing mafic magma mixed with an olivine-free magma in varying proportions. It should be noted that we have excluded a number of olivine-rich, possibly accumulative, samples (olivine >5 vol. %; dotted lines in Fig. 11) from our calculations. Despite wide variations in the chemical compositions of both bulk-rocks and groundmass and in the modal proportions of the different phenocryst phases, the mineral compositions are nearly invariant.

Another magma mixing event (first-stage mixing) preceding that described above (second-stage mixing) can account for the characteristics of Group 2 and 3 phenocrysts. The compositional similarity between the Group 2 olivines and the mantles of the Group 3 more iron-rich olivines implies that the magnesian olivine (Fo85–89) crystallized after mixing between an evolved
basalt magma containing the Fe-rich olivine and a more magnesian magma. As aluminous spinel is included in the iron-rich olivine and anorthitic plagioclase, the evolved basalt must have contained phenocrysts of Fe-rich olivine, anorthitic plagioclase, and aluminous spinel (Group 3), which is a similar phenocryst assemblage to that of the olivine basalts from the nearby volcanoes. The magnesian mantles of the Fe-rich ferrous olivines are surrounded by thin ferrous olivine rims, indicating that mixing with felsic magma containing Group 1 phenocrysts occurred after the mixing event, which involved the evolved basalt magma containing Group 3 phenocrysts. As chromian spinel is included with the Group 2 magnesian olivine, we conclude that its crystallization was simultaneous with, or preceded, the crystallization of the magnesian olivine. Therefore, the magnesian end-member magma contained no phenocrysts or possibly only trace amounts of chromian spinel.

Compositions and oxygen fugacity of the end-members

The end-member compositions involved in the second-stage mixing event were estimated using the relations between bulk-rock SiO$_2$ (wt %) contents and phenocryst abundances and those between other elements. The silica contents of the end-members were estimated from modal abundance vs wt % SiO$_2$ diagrams. The systematic modal mineralogical variation with silica implies that the crystal proportions reflect the mixing ratios of felsic and mafic magmas, and that alteration of the modal proportions by crystallization, melting, or fractionation after magma mixing is negligible. Because of the negligible amount of the Group 3 phenocryst minerals, the mineral proportions reflect the mixing ratios between the felsic magma of Group 1 and the mafic magma of Group 2. The modal abundance of olivine is negatively correlated with decreasing bulk-rock SiO$_2$ (Fig. 10). Excluding some samples that were affected by olivine accumulation, the SiO$_2$ content of the felsic magma end-member can be determined from the extrapolation of the modal abundance trend to 0% olivine as the felsic end-member contains no olivine. Similarly, the SiO$_2$ content of the mafic end-member can be estimated from the intersection of the modal abundance trend with the SiO$_2$ axis where the total phenocryst proportion minus olivine equals zero. Using this method, the SiO$_2$ contents of the end-members are estimated as c. 52 wt % for the mafic end-member, and c. 57 wt % for the felsic end-member. Similarly, diagrams of phenocryst modal abundance vs elemental abundance were used to estimate the contents of other elements. The estimated felsic end-member composition is similar to that of typical andesites from the volcanic field (CHA1 and CHA2 in Table 1). The western contiguous andesite cone of Chausudake includes this type of andesite. The estimated mafic end-member is a primitive high-magnesian basalt that does not occur in the Quaternary volcanoes along the volcanic front of northern Honshu.

The phenocryst abundances in the end-members are determined by extrapolating the modal variation trends to the point where the crystal abundance in the counterpart is zero. Olivine abundance in the mafic end-member is thus c. 6 vol. %; the total phenocryst abundance in the felsic end-member is estimated as 25–30 vol. %. Plagioclase accounts for c. 20–25 vol. % in the felsic end-member, and the balance of c. 5 vol. % consists of augite, hypersthene, and Fe–Ti oxide minerals. The phenocryst abundance in the felsic end-member is also typical of andesite from the volcanic field. The silica contents of the mixed rocks are 54–57%, implying that the mixing ratio of felsic to mafic magmas ranges from 1:5 to 1:0.

The oxygen fugacity of the mafic end-member can be estimated based on the Fe$^{2+}$–Mg exchange equilibrium between olivine and melt (Ulmer, 1989) and on the ratio between FeO and Fe$_2$O$_3$ in the inferred melt. The FeO content in the melt is calculated from the FeO/MgO equilibrium between olivine and melt, then the Fe$_2$O$_3$ content is estimated as the balance of total iron. The oxygen fugacity is then calculated from the ferric–ferrous ratio and the amounts of the other elements in the melt (Kress & Carmichael, 1991). The calculations were carried out for two assumed melt compositions of bulk (E.M. in Table 1) and olivine-free melt (E.M.melt in Table 1). The olivine-free melt composition was calculated by subtracting the average olivine composition from the bulk composition. Assuming that highly magnesian (Foa90) olivine is the first phenocryst crystallized that is in equilibrium with the bulk magma composition, and that the more Fe-rich (Foa90) olivine crystals are equilibrated with the melt, oxygen fugacities are estimated as +1.0 to +2.2 log units relative to the QFM (quartz–fayalite–magnetite) oxygen buffer for both of the assumed melts. The oxygen fugacity estimates are similar to typical estimates for basaltic rocks (Haggerty, 1978; Carmichael, 1991). In the estimation, the temperature was assumed to be in the range of 1150–1350°C. The temperature of the mafic end-member will be discussed in the next section.

The oxygen fugacity of the felsic end-member was estimated using the Fe–Ti oxide oxygen geothermobarometer. Oxygen fugacity estimated from coexisting magnetite and ilmenite in the Group 1 phenocryst assemblage by the method of Anderson & Lindsley (1988) ranges from +0.5 to +1.5 log units relative to the QFM buffer. Temperature estimates calculated using this method were in the range 880–1060°C. The highest temperature estimate, calculated based on the most iron-rich ilmenite and the most titanium-rich magnetite, corresponds to the two-pyroxene temperature estimate. The assumed magma temperature is
1050°C from the pyroxene geothermometer (Brey & Kohler, 1990); oxygen fugacity at that temperature is estimated as $+1.0$ log units with respect to QFM using the Fe–Ti oxide oxygen geothermobarometer.

**Temperature, pressure, water contents, and heat transfer**

The physical and chemical properties of end-member magmas involved in the second-stage mixing (chemical composition, crystal content, oxygen fugacity, and temperature) are relatively well constrained except for the temperature of the mafic end-member, whereas the characteristics of the end-members of the first-stage mixing remain poorly determined. The only clue to the composition of the differentiated olivine basalt magma (Group 3 phenocrysts) is its petrographic similarity to that of olivine basalts from nearby volcanoes. As stated above, differentiated olivine basalts occur at the Hakkoda, Iwate, and Akita Komagatake volcanoes (Kawano & Aoki, 1960; Kawano et al., 1961; Aoki, 1989). The olivine basalts broadly resemble each other in terms of whole-rock composition except for the variation in Fe/Mg ratio, which is reflected in the Fo content of the olivine phenocrysts. The Group 3 olivine compositions (Fo78–80) are less magnesian than those of Group 2 olivine phenocrysts (Fo85–90), but are still magnesian relative to olivine in the regional olivine basalts (Fo63–85, Aoki, 1989). We presume here that the end-member containing the Group 3 phenocrysts was similar in composition to the least differentiated basalts from the nearby volcanoes. For the purposes of comparison, we examine basalts (FeO/MgO from 1.3 to 1.6) from Iwate volcano, which have been well studied petrographically (Ishikawa et al., 1984; Nakagawa, 1987) and experimentally (Takagi et al., 2005). Equilibrium phase relations were calculated using MELTS (Ghiorso & Sack, 1995) for two basalts (IW40 and IWL16 in Table 1) to estimate the undetermined properties that control the phase relations of the Group 3 magma. IW40 is an olivine basalt that has the least differentiated composition among the data reported by Ishikawa et al. (1984). The other basalt (IWL16) was used by Takagi et al. (2005) for phase equilibrium experiments; this is a pyroxene-phric basalt, but its composition resembles those of the pyroxene-free olivine basalts from the volcano. For example, the MgO content of IWL16 is 5.5 wt%, which is within the range of pyroxene-free olivine basalts (4.2–6.9 wt%). The composition of IWL16 is slightly more magnesian than that of IW40 in terms of FeO/MgO and CaO/Na2O ratios; FeO/MgO is 1.6 in IWL16 and 1.3 in IW40, and CaO/Na2O is 4.3 in IWL16 and 5.2 in IW40. Trial-and-error calculations for these compositions with varying water contents suggest that water contents of at least 3 wt% are required to match the composition of observed anorthite plagioclase phenocrysts. The anorthite content in plagioclase is known to increase with increasing water content in the melt (Housh & Luhr, 1991). Takagi et al. (2005) demonstrated that crystallization of calcic plagioclase from basalt magma requires high water contents at pressures of 2–3 kbar. Anorthite-rich plagioclase in arc basalts has sometimes been regarded as a xenocryst because of the obvious disequilibrium with the melt (Kimata et al., 1995; Anma-Miyasaka & Nakagawa, 2002); however, Takagi et al. (2005) have argued that arc tholeiite magmas with normal compositions can be in equilibrium with An-rich plagioclase (An90–94) at high P H2O, except for extremely high-An plagioclase (An96–98). The anorthitic plagioclase of Group 3 phenocrysts (An92–94) from this study can be equilibrated with a normal tholeiitic basalt at high P H2O.

Phase relations in $P$–$T$ space are shown for IWL16 at nickel–nickel oxide (NNO) and 5% initial H2O (Fig. 12a), and for IW40 at QFM and 5% initial H2O (Fig. 12b). The illustrated phase relations are similar for IWL16 and IW40, but the spinel crystallization temperature of IWL16 is considerably lower than that of IW40. The disparity is attributable to the difference in oxygen fugacity. At a fixed oxygen fugacity, the spinel crystallization temperatures become closer.

Our intention here is to reproduce the observed Group 3 phenocryst assemblage (olivine + plagioclase + aluminous spinel); however, for simplicity, we ignore the presence of spinel because assuming an appropriate value of oxygen fugacity allows spinel crystallization. The stability field of olivine + plagioclase occurs at pressures as low as 1 kbar under water-saturated conditions (Fig. 12). H2O contents greater than 3 wt% result in little change of the olivine + plagioclase stability field. At pressures higher than ~1 kbar, the crystallization temperature of pyroxene becomes higher than that of plagioclase (Fig. 12). However, the experimental phase equilibria determined

---

Fig. 12. Phase relations of the assumed low-MgO basalt end-member magma (the first stage) in $P$–$T$ space calculated by MELTS for (a) IWL16 (pyroxene-bearing basalt), used in the phase equilibrium experiments of Takagi et al. (2005) and (b) IW40, olivine basalt reported by Ishikawa et al. (1984).
at 2 kbar by Takagi et al. (2005) include the stability field of olivine + plagioclase under hydrous conditions (≥3 wt % H₂O). In the experiments, in contrast to the wide stability field of olivine + plagioclase at pressures below 1 kbar, the field shrinks at 2 kbar and finally disappears at higher pressures. Despite the discrepancy between the MELTS calculation and the experiments of Takagi et al. (2005), it can be concluded that the assemblage olivine + plagioclase is stable at pressures lower than 2 kbar under hydrous conditions (≥3 wt % H₂O). Corresponding temperatures are 1040–1180°C.

Phase relations were calculated using MELTS for the end-member magmas of the second mixing stage (Fig. 13). These calculations were carried out at varying temperatures and water contents, and for an oxygen fugacity of +2/0 log units relative to the QFM buffer. Pressures lower than 2 kbar (0.5–1 kbar) were assumed for our interpretation as this reflects the crystallization conditions inferred for the end product of the preceding mixing episode. The calculation revealed that the observed assemblage of olivine + plagioclase is stable across a wide range of temperatures under hydrous conditions; however, olivine is unstable under anhydrous conditions or at low water contents, and is replaced by orthopyroxene near the liquidus (Fig. 13). The temperature of olivine + spinel stability depends on the water content of the magma rather than pressure. At 1 kbar, the stable crystallization temperature range is between 1170 and 1230°C at water contents as low as 1 wt %, and between 1120 and 1170°C under H₂O-saturated conditions. Consequently, temperature estimates for the mafic end-member fall into the range 1120–1230°C. Phase equilibrium calculations for the felsic end-member were made with varying water contents at T = 1040–1080°C, P = 1 kbar, log fO₂ = QFM + 1. The modeling at 12–15 wt % H₂O and 1040°C reproduced the observed Group 1 phenocryst abundances (20–25 vol. % plagioclase, several per cent of orthopyroxene and clinopyroxene, and a few per cent Fe–Ti oxides).

We next discuss heat transfer during second-stage magma mixing based on the estimated temperatures of the end-members and of the mixed magma. The crystallization temperature of groundmass pyroxene is estimated as c. 1180°C for magnesian samples for which the mixing ratio of mafic magma to felsic magma is c. 1:5. An estimated temperature of 1050°C from Group 1 phenocryst pyroxenes represents the temperature of the felsic end-member magma. When a simple heat balance is assumed without any crystallization or fusion, the temperature of the mafic end-member is estimated as 1270°C; however, this temperature is too high compared with the olivine–spinel crystallization conditions estimated using MELTS (1120–1230°C). Notwithstanding, temperature estimates differ little between the two-pyroxene thermometry conditions of groundmass pyroxenes (1180°C) and the olivine–spinel crystallization temperature determined from MELTS; hence, the assumption of simple heat balance seems inadequate. The similarity in temperature between the mafic end-member and the groundmass is accounted for by thermal equilibration between the two mixing magmas preceding the crystallization of olivine and spinel. Thermal equilibration occurs before any chemical interaction because thermal diffusion is typically three to five orders of magnitude greater than chemical diffusion when two magmas mingle (Sparks & Marshall, 1986). Cooling by thermal equilibration causes crystallization in the hot magma. In the mafic end-member, olivine and spinel are considered to have crystallized after thermal equilibration was achieved, but before chemical hybridization occurred. During thermal equilibration, the felsic end-member must have been heated and the minerals might thereby have partially fused or re-equilibrated. However, most of the Group 1 phenocrysts remain and re-equilibration is apparent only in thin rims, indicating that the duration of heating was too short for fusion or thermal re-equilibration. The only evidence of fusion is the round iron-rich cores of the pyroxene phenocrysts;
however, a compensated magnesian rim surrounding the iron-rich cores might have preserved the crystal proportion.

**Comparison with the 1998 magma intrusion event at Iwate**

In addition to the above petrological model (Fig. 14a), the magma plumbing system beneath Ebisumori can be further constrained by comparison with the latest magma intrusion event in the Sengan area. Tanaka et al. (2002) reported that dike intrusions were observed geodetically and seismologically at Iwate volcano from February to September 1998. The magma commenced rising beneath the eastern basaltic cone. Then, after initially rising vertically, it migrated laterally westward. In addition to the dike, other point pressure sources were observed at the western end of the volcano. The point pressure might be a magma chamber, but results of that study merely suggested that the origin of the point pressure source was related to hydrothermal activity at the southwestern flank. The point pressure was a stationary source of isotropic expansion. If the point pressure was a magma chamber, the form of intrusive body is likely to be a laccolith, as this is a common intrusive form for intermediate magmas and can hemispherically deform the overlying rocks (Pollard & Johnson, 1973). Assuming that the point pressure and the dike are, respectively, an andesite magma chamber and a basaltic intrusions, a similar magma plumbing system can account for the evolution of the Hachimantai magmatic system as discussed above.

The Iwate and eastern Hachimantai volcanic fields have similarities in vent alignment and rock types, as well as spatial proximity. Eastern Hachimantai and Iwate comprise sub-parallel west–east-trending volcanic chains at a spacing of ~12 km. The western ends of the volcanic terminate in a north–south-trending zone of uplift that is likened to the backbone of northern Honshu. The main volcanic chains are located to the east of the uplifted zone (Ohba et al., 2003). Vents are aligned approximately parallel to the regional principal horizontal stress direction. In both volcanic chains, andesite cones are situated at the western end, and basalt–basaltic andesite cones are located in the middle and east of the chains. In the middle and east of the chains, basalt–basaltic andesite magmas usually erupt to build basaltic cones, but, as the Iwate 1998 event illustrated, lateral migration along the chains seems to occur occasionally.

Analogous to the Iwate 1998 event, a basalt dike from the eastern basalt feeding area beneath Nishimoriyama is presumed to have migrated laterally toward the stationary andesite magma chamber beneath Chausudake. In the 1998 event at Iwate, the lateral dike did not reach the point pressure source. Assuming that the lateral basalt dike encountered and penetrated the andesite magma chamber beneath Chausudake, magma mixing would occur there and the mixed magma would erupt between the andesite and basalt cones, at exactly the location of the Ebisumori cone. First-stage mixing is inferred to have occurred in the basalt-feeding area that is located beneath the middle and eastern parts of the volcanic chain because it requires mixing of two basalt magmas. That mixing is considered to have occurred when a migrating basalt dike overtook a preceding basalt dike. The model shown in Fig. 14b is based upon the 1998 Iwate event. This represents
a petrological model for the production of high-magnesian andesite in the Hachimantai volcanic field.

Comparison with other high-Mg andesite suites

In high-magnesian andesite suites such as those of the Ogasawara (Bonin) islands, high-magnesian andesite occurs as a major component. In these suites, primary high-magnesian andesite is considered to be derived directly from the mantle. Unlike those suites, there seems to be no primary high-magnesian andesite magma at the volcanic front of northern Honshu, a mature volcanic arc associated with thick crust. No high-magnesian andesite occurrence has been reported there, except for the Ebisumori andesite, which is suggested by this study to have resulted from magma mixing between a felsic and a high-MgO basalt magma.

The infrequency of occurrence of high-magnesian andesites in northern Honshu stems from the rarity of the mafic end-member basalt. As emphasized above, high-magnesian andesite that contains more than 8 wt % MgO is rare in the Quaternary volcanoes of northern Honshu. Ebisumori is the only occurrence of high-magnesian andesite, and its volume is less than 0.3 km$^3$ (Ohba & Umeda, 1999). The volume of Quaternary volcanic products along the volcanic front of northern Honshu is c. 960 km$^3$; therefore, high-magnesian andesite accounts for only 0.03% of the erupted magmas. However, magma mixing is a common phenomenon and is especially observed in andesite magmas (Wada, 1985; Nakagawa, 1991; Ban & Yamamoto, 2002).

The high-magnesian aspect of the Ebisumori samples originates from the high MgO content of the basaltic end-member (10-2 wt % MgO). Although basalt magmas are considered to be the mafic end-members of magma mixing at other volcanoes, the MgO contents of the mixed andesites are much lower than those of the Ebisumori high-magnesian andesite, indicating that the end-member magmas are less magnesian than the end-members of the Ebisumori mixed rocks. The mafic end-member of magma mixing in the rest of northern Honshu is considered to be a less magnesian basalt.

The distinctive compositional characteristics of the mafic end-member are consistent with those of the regional basalts. Although many Quaternary basaltic volcanoes exist in northern Honshu, high-magnesian basalts (MgO > 8 wt %) occur at only one volcano, Oshima-Ohshima, which is located in the back-arc region (Yamamoto et al., 1977; Yamamoto, 1988). Even amongst the Neogene basalts, which are a major eruptive product within northern Honshu, high-magnesian basalt has been described at only two locations: Ryosen (Shuto et al., 1985) and Tomari (Takimoto & Shuto, 1994; Takimoto, 1996). The magnesian basalts from these localities have compositions that are interpreted to be in equilibrium with mantle peridotite. It is therefore considered that ascent of such undifferentiated basalt has been an extremely rare event. Therefore, the ascent of the high-magnesian basalt magma to shallow crustal levels in eastern Hachimantai can also be considered as a highly unusual event, which explains the infrequency of high-magnesian andesite produced by magma mixing in northern Honshu.

Origin of the high-magnesian basalt magma

The assumed high-magnesian basalt appears to be an ordinary mantle-derived magma composition despite its infrequent occurrence. The second-stage mafic end-member magma and the first stage high-magnesian end-member (crystal-free primitive magma) are plotted on normative $\text{di-ol-silica}$ and $\text{pl-ol-silica}$ ternary diagrams in Fig. 15 (Walker et al., 1979) together with representative high-magnesian basalts (MgO > 8 wt %) from other arcs. Assuming that the composition of the differentiated basalt end-member of the first stage is within the range of similar basalts from nearby volcanoes (dashed line in Fig. 15), the magnesian, crystal-free end-member magma shifts only slightly in the opposite direction to the mafic end-member of the second-stage mixing (shaded area in Fig. 15), as phenocryst phases from the differentiated end-member are rare in the samples. High-magnesian basalts from subduction zones define elongated trends in these diagrams. These basalts are rich in magnesium and are interpreted as being in equilibrium with mantle peridotite. Therefore, the trends are not fractional crystallization related but probably represent variable degrees of peridotite melting. The trend on the $\text{di-ol-silica}$ diagram is similar to the range of chemical variation in the melt formed during peridotite melting experiments (Hirose & Kushiro, 1993). The inferred primitive magma is plotted at the silica-rich end of the trend, indicating the highest degree of mantle partial melting among subduction zone magmas. Hirose & Kushiro (1993) noted that the normative hypersthene content of partial melts increases with increasing melting degree at a given pressure. Compared with the experimental studies of Hirose & Kushiro (1993) and Kushiro (1998), the primitive basalt was assumed to be produced at 5–20 kbar by 10–25% partial melting of mantle peridotite. As the crustal thickness is about 40 km beneath the volcano, the pressure range narrows to 13–20 kbar. Based on phase equilibrium studies, Kushiro (1998) demonstrated that primitive mid-ocean ridge basalts (MORBs) are produced at pressures of 10–15 kbar, similar to the range envisaged for the primitive magmas at Hachimantai. Olivine fractionation is required to produce primitive MORB compositions, whereas the primitive basalt composition of Hachimatai can be accounted for solely by partial melting of peridotite. In fact, the estimated mafic end-member
composition is close to the experimental melt compositions at 10–15 kbar pressure and at 1350°C of Hirose & Kushiro (1993).

Figure 16 illustrates K$_2$O and Na$_2$O characteristics of the mafic end-member compared with typical high-magnesian arc basalts. The estimated end-member is poor in K$_2$O and Na$_2$O compared with arc basalts, possibly suggesting a high degree of partial melting of mantle peridotite. However, the origin of intra-arc variations in K$_2$O, Na$_2$O, trace elements, and Sr–Nd–Pb isotopes remains controversial, and many hypotheses have been proposed (Tatsumi & Eggins, 1995; Kersting et al., 1996; Churikova et al., 2001; Kimura & Yoshida, 2006). Our result merely demonstrates that primitive basalt magmas are being generated beneath the volcanic front of northern Honshu at a similar depth and melt fraction to MORB. Intra-arc variations involving increasing alkali element contents compared with MORB presumably result from other geological factors, such as variable melting degrees, modified mantle source compositions and increasing alkali-bearing fluid supply from the subducting slab.

**ACKNOWLEDGEMENTS**

We gratefully acknowledge the important technical assistance rendered by Y. Itoh, T. Ohyama and H. Kawanobe, and the valuable discussion afforded by M. Nakamura, T. Hasenaka and K. Aoki. We thank J. Foden and B. Stewart for careful reviews and thoughtful comments. We also thank J. Gamble, the editor, for his careful handling of the manuscript.

**SUPPLEMENTARY DATA**

Supplementary data for this paper are available at Journal of Petrology online.

**REFERENCES**


