Dusty and honeycomb plagioclase: indicators of processes in the Uchino stratified magma chamber, Izu Peninsula, Japan

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ABSTRACT

Two types of melt-inclusion-bearing plagioclase coexist in the Uchino basalt, Izu peninsula, Japan. (1) dusty plagioclase with an Ab-rich core mantled by a dusty zone consisting of more An-rich plagioclase and fine melt inclusions; and (2) honeycomb plagioclase with a central core containing larger melt inclusions. The dusty plagioclase probably forms through partial dissolution of Ab-rich plagioclase, as dusty zones truncate the concentric zonal structure of the Ab-rich inner core. The honeycomb plagioclase is formed by skeletal growth under supercooling conditions. This is indicated by the euhedral geometry of the boundary between the inner inclusion-free and the outer inclusion-bearing zones of the honeycomb plagioclase which has a concentric inner core. Bimodal distribution of the chemical compositions of phenocryst cores and assemblages in crystal clots suggest that hybridization of basaltic and andesitic magmas took place before crystallization of phenocryst rims. The core-rim boundaries of phenocrysts derived from the andesitic magma and the basaltic magma have a rounded and an euhedral geometry, respectively. This suggests the andesitic magma was heated and the basaltic magma was cooled prior to hybridization. These features are consistent with the development of a thermal diffusive interface between overlying andesitic magma and underlying basaltic magma in a stratified magma chamber.

Introduction
A common feature of plagioclase morphology in hybrid* rocks is the existence of two types of melt inclusions: (1) dusty plagioclase with relatively fine melt inclusions; and (2) honeycomb plagioclase with large melt inclusions (Larsen et al., 1938; Kuno, 1950; Sakuyama, 1979, 1981; Gerlach and Grove, 1982; Bloomfield and Arculus, 1989). A consensus of opinion for the origin of dusty plagioclase is by partial dissolution of Ab-rich plagioclase through a reaction with melt in equilibrium with more An-rich plagioclase (Kuno, 1936, 1950; Larsen et al., 1938; Macdonald and Kat- sura, 1965; Eichelberger, 1978; Sakuyama, 1979; Bloomfield and Arculus, 1989). The origin of honeycomb plagioclase is, however, a matter of debate, contended by models of partial dissolution (Kuno, 1950; Dungan and Rhodes, 1978; Gerlach and Grove, 1982) and skeletal growth under supercooling conditions (Kuo and Kirkpatrick, 1982; Anderson, 1984). The honeycomb plagioclase in MORBs from DSDP Legs 45 and 46 has been proposed to be formed both by dissolution (Dungan and Rhodes, 1978) and by skeletal growth (Kuo and Kirkpatrick, 1982) during magma mixing. Lofgren and Norris (1981) demonstrated experimentally that honeycomb plagioclase can form through partial dissolution of Ab-rich plagioclase in the MORB melt which is in equi-
librium with more An-rich plagioclase. However, plagioclase dissolution experiments in the Di-Ab-An system do not show any evidence for the formation of honeycomb plagioclase (Tsuchiyama, 1985). In order to understand the evolution of hybrid magmas, it is, therefore, essential to clarify whether honeycomb plagioclase is formed by dissolution or by skeletal growth processes. Furthermore, an explanation for the coexistence of dusty and honeycomb plagioclase, which may provide constraints on processes in a magma chamber, has remained uncertain.

The Uchino volcano is a scoria cone of the Higashi-Izu monogenetic volcano group which has erupted a basaltic lava flow (0.0067 km$^3$, Hamuro, 1978; Fig. 1). The Higashi-Izu volcano group is composed of more than 70 monogenetic volcanoes scattered in a circle of 30 km diameter within which most volcanic activity took place between 30,000 and 3000 y B.P. (Aramaki and Hamuro, 1977). The ejecta
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of this volcano group belong to high-alumina basalt series (Kuno, 1959) and are composed of olivine basalt, olivine-clinopyroxene basalt and andesite, olivine-clinopyroxene-orthopyroxene andesite, olivine-clinopyroxene-orthopyroxene-hornblende andesite and dacite, and orthopyroxene-hornblende dacite. Coexistence of dusty and honeycomb plagioclase is commonly observed in basalts and andesites in this group.

This paper describes compositional and morphological relationships among phenocrystic phases in the Uchino basalt containing both dusty and honeycomb plagioclase and discusses a mechanism for the formation of dusty and honeycomb plagioclase. This is then set within a framework of magmatic processes occurring in a chemically stratified magma chamber.

Petrography of the Uchino basalt

Back-scattered electron images were made using a Hitachi S550 electron microprobe analyzer with a GW 30A back-scattered electron detector. Chemical compositions of minerals were analyzed using a Hitachi S550 electron microprobe analyzer with a Kevex X-ray detector. Complete analyses were made for all phases following the methods of Mori and Kanehira (1984), except plagioclase, for which most analyses were based on only Si and Ca, anorthite content of plagioclase being calculated from ratios of Ca/(Ca+Si) raw intensity using a calibration line from working standards. Whole-rock major elements contents were determined with a Rigaku 3550 X-ray fluorescence analyzer following the analytical procedures of Goto and Tatsumi (1991).

Phenocryst phases are plagioclase, olivine, and clinopyroxene, which make up 14.9 volume % of the rock (Table 1). Some phenocrysts form crystal clots (glomeroporphyritic aggregates; Table 2). The groundmass is homogeneous and composed of plagioclase, clinopyroxene, olivine, titanomagnetite and small amounts of glass. The whole-rock major elements composition is listed in Table 1.

Plagioclase

Plagioclase phenocrysts can be divided into three textural types; clear, dusty and honeycomb (Fig. 2).
Fig. 2. Diagrammatic illustration of phenocryst plagioclase types.

Clear plagioclase

Clear plagioclase lacks melt inclusions, is characterized by a concentric zonal structure defined by euhedral growth surfaces (Fig. 3A), and is 0.4–1.2 mm in length. An abrupt change in chemistry divides the clear plagioclase into two distinct regions, a core and a rim. The outermost fringe of the core is An$_{84-88}$, whereas the innermost part of the rim is An$_{74-79}$ (Figs. 4 and 5A; Table 3). The rims are normally zoned towards the edge of the phenocryst. Some clear plagioclase form crystal clots with olivine, euhedral Mg-rich clinopyroxene and honeycomb plagioclase (Table 2).

ter and almost completely devitrified into aggregates of Ab-rich plagioclase, clinopyroxene and titanomagnetite. Dusty zones within the plagioclase occur either as discrete zones forming the outer part of the core (Fig. 2C) or as patches within the core. In the former case, the inner core has a concentric zonal structure which is truncated by the dusty zone (Fig. 3C). The highest anorthite content measured in the central parts of the cores is An$_{79}$.

Rims surrounding the dusty zones are normally zoned towards the edge of the phenocryst (Figs. 3B, C and 5C). The anorthite content of the innermost part of rim of dusty plagioclase is the same as that of the other plagioclase phenocryst types (Fig. 4). The outer perimeters of the dusty zones have rounded forms (Figs. 2C, 3B and C), a feature commonly recognized in the dusty plagioclase elsewhere (e.g., Macdonald and Katsura, 1965; Eichelberger, 1978; Gerlach and Grove, 1982). Some dusty plagioclase form crystal clots with rounded Fe-rich clinopyroxene (Table 2; Fig. 3D).

Dusty plagioclase

Dusty plagioclase is 0.2–1.3 mm in length and is characterized by an Ab-rich core (An$_{88-72}$) mantled by a dusty zone composed of more An-rich plagioclase (An$_{78-84}$) and fine melt inclusions (Figs. 2C, 3B and C). The melt inclusions are commonly 10–30 μm in diame-
Fig. 3. (A) Backscattered electron images (BEI) of clear plagioclase. The brightness of the images is proportional to anorthite content. Analyses were performed at marked points (numerals represent anorthite content). Note the concentric zonal structure defined by several euhedral growth surfaces. (B, C) BEI of dusty plagioclase. (B) Dusty zone consists of An-rich plagioclase (gray), clinopyroxene and titanomagnetite (white). The boundary between the dusty zone and the rim has a rounded geometry (arrow). (C) Concentric zonal structure of Ab-rich plagioclase is truncated by the dusty zone. (D) BEI of rounded Fe-rich clinopyroxene forming a crystal clot with dusty plagioclase (D-PI) (A6 in Table 3). The brightness of the images is inversely proportional to Mg#. Fe-rich core is enclosed by Mg-rich clinopyroxene mantle. Internal concentric zonal structure of the Fe-rich core is truncated (arrow). Euhedral shaped hypersthene (Hy) is observed in the center.

**Honeycomb plagioclase**

Honeycomb plagioclase is 0.4–2.5 mm in length and contains many relatively large melt inclusions (commonly 10–100 μm; Fig. 6A). These melt inclusions have irregular shapes and are almost completely devitrified into aggregates of Ab-rich plagioclase (An₄₀–₆₀), which always precipitates on inclusion walls, clinopyroxene and titanomagnetite with or without olivine. The melt inclusions are commonly restricted to the inner core, which is enclosed by an inclusion-free outer core with concentric zonal structure (Figs. 2B, 6A and B). The long axis of elongated melt inclusions is commonly parallel to that of the host plagioclase crystal (Fig. 6B and E).

The inner cores containing the melt inclusions have heterogeneous chemical composi-
### Table 3

Chemical compositions of representative plagioclase phenocrysts

<table>
<thead>
<tr>
<th>Name</th>
<th>PL5 — honeycomb plagioclase</th>
<th>PL14 — clear plagioclase</th>
<th>PL12 — dusty plagioclase</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SiO₂</strong></td>
<td>49.52 48.09 45.42 45.65 45.71 45.28 47.25 47.09 45.42 48.11</td>
<td>49.75 49.25 45.75 46.00 47.70 46.76</td>
<td>50.18 48.61 47.09 46.94 47.89 51.22 50.01</td>
</tr>
<tr>
<td><strong>Al₂O₃</strong></td>
<td>31.16 32.49 34.29 34.16 34.00 34.28 33.04 33.11 34.54 32.43</td>
<td>30.84 31.36 33.97 33.74 33.30 33.64 30.21 31.95 32.82 32.79</td>
<td>30.21 31.95 32.82 32.79 32.51 30.42 31.24</td>
</tr>
<tr>
<td><strong>Fe₂O₃</strong></td>
<td>0.90 0.52 0.44 0.46 0.40 0.42 0.33 0.35 0.54 0.48</td>
<td>0.83 0.71 0.44 0.49 0.47 0.49 0.94 0.54 0.62 0.94</td>
<td>0.94 0.54 0.62 0.56 0.66 0.55 0.58</td>
</tr>
<tr>
<td><strong>Na₂O</strong></td>
<td>3.08 2.42 1.34 1.48 1.48 1.24 2.14 1.99 1.26 2.28</td>
<td>3.20 2.87 1.55 1.70 2.02 1.89 3.64 2.60 1.91 2.09</td>
<td>3.64 2.60 1.91 2.09 2.36 3.90 3.35</td>
</tr>
<tr>
<td><strong>K₂O</strong></td>
<td>0.09 0.09</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td><strong>An (mol %)</strong></td>
<td>72.6</td>
<td>78.5 88 80</td>
<td>86.8</td>
</tr>
</tbody>
</table>

Locations of analyzed points are shown in Figs. 6B (PL5), 3A (PL14) and 3C (PL12)
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Fig. 4 Histogram showing anorthite content of plagioclase types in the Uchino basalt.

Melt inclusions are also found in the outer zones of inclusion-free plagioclase with concentric zonal structure (Fig. 6D). This feature is rare, and is recognized in just one grain. The interior clear plagioclase is An$_{87-89}$, which is more An-rich than the core of dusty plagioclase (An$_{58-79}$). The boundary between the inner inclusion-free zone with concentric zoning and the outer inclusion-bearing zone has a euhedral form; a feature consistent with that described in honeycomb plagioclase by Anderson (1984).

An abrupt compositional change marks the boundary between the core and rim in honeycomb plagioclase. The outermost fringe of the core is An$_{84-88}$, which is similar to the clear plagioclase. The innermost part of the rim is An$_{75-80}$, and corresponds to that of other phenocryst plagioclase (Fig. 4). The outer margin of the core is euhedral (Figs. 2B, 6A, D and E) as in the clear plagioclase. This euhedral feature is also recognized in honeycomb plagioclase elsewhere (fig. 2E in Kuo and Kirkpatrick, 1982; fig. 3B in Gerlach and Grove, 1982).

Some honeycomb plagioclase forms crystal clots with olivine, euhedral Mg-rich clinopyroxene and clear plagioclase (Table 2; Figs. 6E and F). A single chromian spinel was found in the core of honeycomb plagioclase (Fig. 6B) and its chemical composition is similar to spinel contained in Fo-rich olivine cores.

**Groundmass plagioclase**

Groundmass plagioclase is free of melt inclusions and is lath shaped, less than 0.4 mm in length, and shows gradual normal zoning. Its cores are compositionally distinct from the phenocrysts, falling within An$_{69-77}$, which is slightly more Ab-rich than the innermost parts of phenocryst rims (Fig. 4).
Fig. 5. Compositional profiles of representative plagioclase phenocrysts. Locations of analyzed points and their complete chemical compositions are shown in Figs. 3A, 6B, 3C and Table 3, respectively. Error bar of the contents of Fe$_2$O$_3$ indicate 2σ of analytical precision.

Fig. 6. (A–E) BEI of honeycomb plagioclase. (A) Melt inclusions are observed in the inner core. The inner core is enclosed by the melt-inclusion-free outer core. (B) BEI of the region enclosed by solid line in (A). The boundary between the core and the rim has a euhedral form. The long axis of elongated melt inclusions is parallel to that of the host crystal. Sp denotes a chromian spinel inclusion (P15-S1 in Table 6). (C) BEI of the region enclosed by solid line in (B). Inner core is composed of two regions with different anorthite content (An$_{87-89}$ and An$_{77-83}$). Pl and Cpx denote plagioclase and clinopyroxene growing in melt inclusions, respectively. Melt inclusions are commonly found in the An-rich regions. (D) Melt inclusions are found in the outer zones of inclusion-free plagioclase with concentric zonal structure. The boundary between the clear inner zone and the outer inclusion-bearing zone has a euhedral geometry. (E, F) BEI of a crystal clot composed of honeycomb plagioclase (E) and euhedral Mg-rich clinopyroxene (F). (E) The long axis of elongated melt inclusions is parallel to that of the host crystal. The boundary between the core and the rim has a euhedral geometry. (F) Melt inclusions, which are devitrified into Fe-rich clinopyroxene (white), plagioclase (black) and titanomagnetite, exist over the core.
Clinopyroxene in melt inclusions of phenocryst plagioclase

Clinopyroxene growing in dusty inclusions tends to be unzoned but may show sector zoning, whereas most clinopyroxene growing in the melt inclusions of honeycomb plagioclase is chemically zoned. Chemical compositions of the most Mg-rich part of clinopyroxenes in honeycomb plagioclase and of unzoned clinopyroxenes in dusty plagioclase are plotted in Figure 7. The clinopyroxene in the inclusions of honeycomb plagioclase has higher Mg/(Mg+Fe²⁺) (Mg#) than those of dusty plagioclase (Fig. 7). This suggests the Mg# of melt inclusions in honeycomb plagioclase was higher than those in dusty plagioclase. This is consistent with the different phase assemblages, i.e., the presence of olivine in melt inclusions of honeycomb plagioclase but not in dusty plagioclase.

Pyroxene

Phenocryst and groundmass pyroxene is commonly Ca-rich clinopyroxene, though Ca-poor clinopyroxene occurs surrounding olivine and hypersthene is locally observed in the core of rounded Fe-rich clinopyroxene. Clinopyroxene phenocrysts occur as isolated grains and in crystal clots with other minerals (Table 2). Clinopyroxene phenocrysts can be divided into two types based on composition and the type of plagioclase with which it is associated in crystal clots; clinopyroxene associated with dusty plagioclase is rounded and more Fe-rich (Mg# 70–76 in core, Table 4) whereas clinopyroxene associated with clear and/or honeycomb plagioclase is euhedral and Mg-rich (Mg# 84–86 in core, Table 5).

Rounded Fe-rich clinopyroxene

Clinopyroxene in contact with the inner Ab-rich core of dusty plagioclase is 0.2–0.8 mm in diameter and can be divided into three distinct regions: a Fe-rich core (Mg# 70–76), a Mg-rich mantle (Mg# 83–85, 20–50 μm thick), and a rim (Mg# 70–83). The internal concentric zoning of the Fe-rich core is truncated across a sharp boundary by the Mg-rich mantle which, in turn, is enclosed by a narrow Fe-rich rim also across a sharp boundary (Fig. 3D). The most magnesian part of the inner core is Mg# 85. The Fe-rich cores locally have euhedral shaped hypersthene in their centers (Fig. 3D).

Chemical compositions of the innermost part of rims, the outermost part of the Mg-rich mantles and the Fe-rich rounded cores immediately inside mantle of clinopyroxene which forms crystal clots with dusty plagioclase are plotted in Figure 8A. These can be compared with all phenocryst clinopyroxene, including isolated grains, which are plotted in Figure 8C. The compositions of the mantles are always higher in Mg# and Cr₂O₃ content than the cores immediately inside mantle (Table 4; Figs. 8A and C).

Euhedral Mg-rich clinopyroxene

Clinopyroxene forming crystal clots with clear plagioclase and/or honeycomb plagioclase is characterized by relatively high Mg# (84–86) in cores compared with the cores of the rounded Fe-rich clinopyroxene (70–76). They are 0.2–1.3 mm in diameter. Some have a homogeneous core without zonal structure.
### Table 4

Representative analyses of rounded Fe-rich clinopyroxene forming crystal clots with dusty plagioclase

<table>
<thead>
<tr>
<th>Name</th>
<th>A6</th>
<th>A20</th>
<th>A27</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>R-Cpx + D-P1</td>
<td>R-Cpx + D-P1</td>
<td>R-Cpx + D-P1</td>
</tr>
<tr>
<td></td>
<td>Core</td>
<td>Mantle</td>
<td>Rm</td>
</tr>
<tr>
<td>SiO₂</td>
<td>48.20</td>
<td>50.77</td>
<td>51.18</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.93</td>
<td>0.50</td>
<td>0.60</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>5.27</td>
<td>4.57</td>
<td>1.82</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>n.d.</td>
<td>0.50</td>
<td>n.d.</td>
</tr>
<tr>
<td>FeO</td>
<td>9.69</td>
<td>5.67</td>
<td>11.21</td>
</tr>
<tr>
<td>MgO</td>
<td>13.49</td>
<td>15.96</td>
<td>15.93</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.51</td>
<td>0.52</td>
<td>0.97</td>
</tr>
<tr>
<td>Total</td>
<td>97.90</td>
<td>99.68</td>
<td>98.32</td>
</tr>
</tbody>
</table>

Core represents the core immediately inside mantle. Mantle represents the outermost part of mantle. Rm represents the innermost part of rim.

### Table 5

Representative analyses of euhedral Mg-rich clinopyroxene forming crystal clots

<table>
<thead>
<tr>
<th>Name</th>
<th>A13 Cpx + C-Pl</th>
<th>A1 Cpx + Ol + C-Pl</th>
<th>A17 Cpx + Ol + H-Pl</th>
<th>A19 Cpx + H-Pl</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Core</td>
<td>Rm</td>
<td>Core</td>
<td>Rm</td>
</tr>
<tr>
<td>SiO₂</td>
<td>51.62</td>
<td>50.99</td>
<td>52.03</td>
<td>51.99</td>
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<tr>
<td>TiO₂</td>
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<td>0.63</td>
<td>0.30</td>
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<tr>
<td>Al₂O₃</td>
<td>3.55</td>
<td>2.65</td>
<td>3.71</td>
<td>2.74</td>
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<tr>
<td>Cr₂O₃</td>
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<td>n.d.</td>
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<td>n.d.</td>
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<tr>
<td>FeO</td>
<td>5.06</td>
<td>9.48</td>
<td>5.34</td>
<td>9.52</td>
</tr>
<tr>
<td>MgO</td>
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<td>16.21</td>
<td>16.92</td>
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<tr>
<td>Na₂O</td>
<td>0.41</td>
<td>0.78</td>
<td>0.51</td>
<td>0.72</td>
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<tr>
<td>Total</td>
<td>99.33</td>
<td>98.87</td>
<td>100.11</td>
<td>99.81</td>
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</tbody>
</table>

Core represents the outermost part of core. Rm represents the innermost part of rim.

and others exhibit sector zoning. Melt inclusions are locally found within the core and are almost completely devitrified to Fe-rich clinopyroxene, plagioclase and titanomagnetite (Fig. 6F). The core is euhedral and enclosed by the narrow rim.
The chemical compositions of the innermost part of the rim and the outermost part of the core of clinopyroxene forming crystal clots with clear plagioclase and/or honeycomb plagioclase are plotted in Figure 8B. The outermost part of the core has almost the same range of chemical compositions (Mg# 84–86, 0.3–0.8 wt.% Cr2O3) as those of the mantle region of the rounded Fe-rich clinopyroxene (Mg# 83–85, 0.3–0.5 wt.% Cr2O3) (Tables 4 and 5; Fig. 8C), and the rims of all phenocryst clinopyroxene are scattered in the same range (Mg# 70–83, Fig. 8C). The Fe-rich clinopyroxene growing in the melt inclusions of the euhedral Mg-rich clinopyroxene (Fig. 6F) also has the same range of chemical composition as the rims of clinopyroxene phenocrysts.

**Groundmass clinopyroxene**

Groundmass clinopyroxene is less than 0.2 mm in diameter. The compositions of the center of groundmass clinopyroxene are scattered within the same range as those of the innermost rims of the clinopyroxene phenocrysts.

**Olivine**

Olivine forms both isolated grains and crystal clots with other minerals (Table 2) but no olivine forms crystal clots with dusty plagioclase or rounded Fe-rich clinopyroxene. Clinopyroxene olivine is 0.2–1.5 mm in length and shows gradual normal zoning at its margin. Forsterite content of the central part of olivine which is more than 0.3 mm in length is commonly between Fo83–86 whereas smaller olivines are more Fe-rich. The forsterite content of the outer margins is about Fo69, which is the

<table>
<thead>
<tr>
<th>Name</th>
<th>O4-S1</th>
<th>A2660-S1</th>
<th>O11-S1</th>
<th>A25O952-1</th>
<th>P15-S1</th>
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<tr>
<td>Fo#</td>
<td>85 1</td>
<td>85 6</td>
<td>83 2</td>
<td>83 0</td>
<td>88</td>
<td>85</td>
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<tr>
<td>SiO2</td>
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<td>0.99</td>
<td>0.94</td>
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<tr>
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<tr>
<td>Al2O3</td>
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<tr>
<td>Cr2O3</td>
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<td>Fe2O3</td>
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<td>MgO</td>
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<tr>
<td>CaO</td>
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<td>n.d.</td>
<td>n.d.</td>
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<tr>
<td>Total</td>
<td>100.78</td>
<td>99.18</td>
<td>100.01</td>
<td>98.93</td>
<td>98.72</td>
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</table>

Fo# is forsterite content of host olivine around an inclusion
FeO, Fe2O3 are estimated from stoichiometry.

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**Table 6**

Representative analyses of spinel inclusions in olivine phenocrysts and honeycomb plagioclase (P15-S1, Fig. 6B)
same as the homogeneous groundmass olivine grains. All olivine crystals have reaction rims of Ca-poor clinopyroxene with Mg# between 67 and 69. Olivine phenocrysts contain cubic chromian spinels in the cores and titanomagnetite towards their rims. The chromian spinel inclusions in olivine are similar in composition to spinel which is contained in the core of honeycomb plagioclase (Table 6; Fig. 6B).

**Magmatic processes in a chemically stratified magma chamber**

The innermost rims of phenocryst plagioclase have a limited range of anorthite content, but plagioclase immediately inside these rims can be divided into two groups with different anorthite content: clear and honeycomb plagioclase, and dusty plagioclase (Fig. 4). There are also two types of clinopyroxene phenocrysts: normal-zoned Mg-rich and reverse-zoned Fe-rich clinopyroxene. The clear and honeycomb plagioclase locally form crystal clots with olivine and Mg-rich clinopyroxene, whereas the dusty plagioclase locally forms crystal clots with the Fe-rich clinopyroxene (Table 2). Incorporation of so-called xenocryst might be one possible interpretation of these bimodal distributions of phenocryst compositions and the mineral assemblages. However, it is unlikely that Ab-rich plagioclase and Fe-rich pyroxene are xenocrysts, as the Ab-rich core of the dusty plagioclase of the Uchino basalt ($\text{An}_{58-70}$) is more An-rich than plagioclase ($\text{An}_{40-60}$) of the tonalitic xenoliths observed in the ejecta of the Higashi-Izu volcano group (Hamuro, 1985; Miyajima, 1990). Alternatively, the bimodal distributions of phenocryst compositions and the mineral assemblages of the Uchino basalt may be explained by mixing, prior to crystallization of the rims of phenocryst plagioclase, between a mafic magma containing olivine+Mg-rich clinopyroxene+Ab-rich plagioclase and a felsic magma containing Fe-rich clinopyroxene + hypersthene + Ab-rich plagioclase (Table 7).

**TABLE 7**

<table>
<thead>
<tr>
<th>Phenocryst assemblage in each magma prior to hybridization</th>
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<tbody>
<tr>
<td><strong>andesitic magma</strong></td>
</tr>
<tr>
<td>rounded Fe-rich clinopyroxene (Mg# = 85–70), hypersthene (Mg# = 78–74)</td>
</tr>
<tr>
<td><strong>basaltic magma</strong></td>
</tr>
<tr>
<td>olivine (Fo = 86–83, chromian spinel inclusion)</td>
</tr>
<tr>
<td>euhedral Mg-rich clinopyroxene (Mg# = 86–84)</td>
</tr>
<tr>
<td>honeycomb plagioclase (An = 84–88)</td>
</tr>
<tr>
<td>clear plagioclase (An = 84–88)</td>
</tr>
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</table>

The Sukumoyama basalt ($\text{SiO}_2 \sim 47.4$ wt.%, $\text{MgO} \sim 8.3$ wt.%) and the Dainoyama andesite ($\text{SiO}_2 \sim 58.4$ wt.%, $\text{MgO} \sim 3.8$ wt.%), which are members of this volcano group and 8 and 7 km distant from the Uchino basalt, are potential representatives of the mafic and felsic end component, respectively. The Sukumoyama basalt has olivine phenocrysts containing chromian spinels, an identical to those occurring in the Uchino basalt and the Dainoyama andesite contains clinopyroxene, hypersthene and plagioclase which are within the same compositional range of Fe-rich clinopyroxene cores, hypersthene contained in the Fe-rich clinopyroxene, and the Ab-rich core of the dusty plagioclase in the Uchino basalt. The bulk chemical compositions of the Uchino basalt fall on a mixing line between the Sukumoyama basalt and the Dainoyama andesite in the proportion 7:3, apart from the minor components TiO$_2$ and P$_2$O$_5$, therefore it is reasonable that these represent the end components in the mixing which produced the Uchino basalt. The liquidus temperature of the Sukumoyama basalt have been experimentally determined as 1230°C by Hamuro (1985). Temperature of the andesitic magma can be estimated to be 1050°C on the basis of the two-pyroxene geothermometer (Lindsley, 1983) using hypersthene and Fe-rich clinopyroxene pairs in the Uchino basalt.

Theoretical and experimental fluid dynamics studies (e.g., Campbell and Turner, 1985) indicate that an influx of a dense basaltic magma into a lighter and cooler andesitic
magma will not form a hybrid magma, but result in the formation of a stratified magma chamber. In such a magma chamber, the andesitic magma layer is heated while the basaltic magma layer is cooled (e.g., Huppert and Sparks, 1980). Numerous petrographic studies of mafic inclusions in felsic lavas have demonstrated several lines of evidence for heat transfer between mafic and felsic magmas, including: rapid growth morphology of crystals, vesiculation of inclusion upon cooling and crystallization, and resorption texture of phenocrysts due to superheating of host felsic lava (Eichelberger, 1980; Bacon and Metz, 1984; Koyaguchi, 1986; Green, 1988). Mixing of magmas may have taken place in the Uchino volcano through disruption of such a stratified magma chamber.

Partial dissolution in an andesitic layer prior to hybridization

The truncation of the concentric zonal structure of Ab-rich plagioclase cores (An$_{58-72}$) by dusty zones composed of An-rich plagioclase (An$_{78-84}$) and fine melt inclusions (Fig. 3C) suggests the Ab-rich plagioclase reacted with a melt in equilibrium with more An-rich plagioclase, and that the dusty zone forms through re-equilibration. This interpretation is the same as that of previous workers (Kuno, 1936; Larsen et al., 1938; Macdonald and Katsura, 1965; Eichelberger, 1978; Sakuyama, 1979; Bloomfield and Arculus, 1989). In experiments, Tsuchiyama (1985) recognized two types of dissolution: (1) plagioclase partially dissolves to form dusty zone through a reaction between an Ab-rich plagioclase and melt in equilibrium with An-rich plagioclase; and (2) plagioclase simply dissolves into a surrounding melt undersaturated with respect to plagioclase. The rounded shape of the outer margin of the dusty zone (Figs. 3B and C) may be attributed to its simple dissolution origin and suggests the melt surrounding the Ab-rich plagioclase was heated to temperature above its liquidus. The Fe-rich core of clinopyroxene which locally forms crystal clots with dusty plagioclase is also rounded (Fig. 3D). On the other hand, the outer margins of the cores of honeycomb plagioclase, clear plagioclase, and Mg-rich clinopyroxene are euhedral shaped (Figs. 3A, 6A, D and E). Prior to magma hybridization, phenocrysts contained in the andesitic magma had rounded outer margins and those in the basaltic magma had euhedral outer margins. This is consistent with heating of the andesitic magma and cooling of the basaltic magma and suggests the andesitic and basaltic magmas were juxtaposed in stratified layers.

Anorthite content of plagioclase equilibrated with the hybrid magma is between An$_{74-80}$, as represented by the uniform compositions of the innermost rims of all phenocryst plagioclase (Fig. 4). The anorthite content of plagioclase in dusty zones is An$_{78-84}$, which suggests the dusty zones were not produced after hybridization but had been formed prior to it. Partial dissolution of Ab-rich plagioclase can be explained in a schematic phase diagram in the Di-Ab-An system (Fig. 9). A melt, $M_1$, is equilibrated with crystal $P_1$ at temperature $T_1$, the $M_1$ and $P_1$ pair is then brought to temperature $T_2$ ($T_2 > T_1$) following formation of the stratified magma chambers. Superheating causes $P_1$ to begin to dissolve into the melt (simple dissolution according to Tsuchiyama’s terminology), and the melt around $P_1$ may change composition until reaching the plagioclase liquidus surface at $T_2$ ($M_2$ in Fig. 9) before $P_1$ dissolves out. $M_2$ would be equilibrated with more An-rich plagioclase ($P_2$) than $P_1$. At this stage, reaction relationships between a relic of $P_1$ (now rounded shape) and $M_2$ is the same as the relationships between Ab-rich plagioclase and melt equilibrated with more An-rich plagioclase. Partial dissolution (according to Tsuchiyama’s terminology) would take place along with formation of dusty zones in the outer margin of $P_1$. 
Honeycomb plagioclase is likely to have formed through growth processes under supercooling conditions rather than by partial dissolution. This is indicated by the euhedral geometry of the boundary between the inner inclusion-free and the outer inclusion-bearing zones of the honeycomb plagioclase which has a concentric inner core (Fig. 6D). This feature is distinct from that of the dusty zone which truncates the zonal structure of the Ab-rich inner core. Anderson (1984) also described the same feature and has suggested a skeletal growth origin for honeycomb plagioclase.

Dungan and Rhodes (1978) and Kuo and Kirkpatrick (1982) described honeycomb plagioclase in MORBs from DSDP Legs 45 and 46 and observed that the inner inclusion-bearing cores are more Ab-rich (An75-78) than the inclusion-free outer cores (An85). Dungan and Rhodes (1978) suggested such reverse zoning is evidence for a dissolution origin, whereas Kuo and Kirkpatrick (1982) explain this reverse zoning by skeletal growth (Lofgren, 1974). Compositional heterogeneity is recognized in the Ab-rich region of the honeycomb plagioclase of the Uchino basalt (Fig. 6C), suggesting that the Ab-rich region is unlikely to be a relict core. Instead, the following observations provide evidence for a process involving crystallization of the Ab-rich region under supercooled conditions and subsequent overgrowth of the An-rich region under smaller degrees of supercooling:

(1) Each of the Ab-rich and An-rich regions of the inner core of honeycomb plagioclase has an identical extinction angle, indicating the crystallographic continuity within an individual grain.

(2) The Ab-rich region forms a network-like structure, consistent with rapid growth.

(3) The melt inclusions are trapped only in the An-rich region, suggesting overgrowth onto the network-Ab-rich-plagioclase.

Under much smaller degrees of supercooling, the outercore of the honeycomb plagioclase and the core of the clear plagioclase crystallized (Fig. 5). Finally, it is notable that the boundary between Ab-rich and An-rich regions is sharp (Fig. 6C), yet the duplication of this reverse zoning in skeletal plagioclase in crystallization experiments shows gradual compositional change (Lofgren, 1974). The origin of this discrepancy remains uncertain.

Kuo and Kirkpatrick (1982) suggested that the honeycomb plagioclase in MORB from DSDP Legs 45 and 46 grew skeletally in a supercooled hybrid magma. However, such honeycomb plagioclase could have crystallized in the mafic layer before hybridization, for chemical compositions of the melt inclusions are more primitive (higher Mg#) than those of whole rocks (Dungan and Rhodes, 1978; Rhodes et al., 1979). If the skeletal plagioclase had crystallized after hybridization (Kuo and Kirkpatrick, 1982), the chemical compositions of the
melt inclusions should show a similar compositional range to those of the whole rocks. Moreover, the inclusion-free outer core of this honeycomb plagioclase has identical compositions (An$_{85}$) to the cores of clear plagioclase which are too An-rich (An$_{83-86}$) to have crystallized from liquids with the observed whole-rock compositions (An$_{72-74}$, Rhodes et al., 1979). This further suggests that this honeycomb plagioclase crystallized from a primitive magma prior to hybridization. The above interpretation may not mean that all melt-inclusion-bearing plagioclase has been produced by skeletal growth in the MORBs, as evidence for plagioclase of dissolved origin does also exist. Such plagioclase has a homogeneous Ab-rich core (An$_{68}$) surrounded, across an irregular boundary, by a zone composed of relatively small melt inclusions and An-rich plagioclase (An$_{68-73}$) (fig. 2F in Rhodes et al., 1979).

Mixing of magmas may have occurred shortly before the Uchino basalt eruption. Plagioclase with double dusty zones is rare in the Higashi-Izu monogenetic volcano group. This suggests that mixing of magmas takes place only once. Koyaguchi (1985) and Blake and Campbell (1986) suggested that mixing of magmas can take place through a conduit during ascent of the magmas. This model is concordant with the observation that mixing of magmas has taken place only once shortly before eruption. Koyaguchi and Blake (1989) simulated hybridization in a squeezed conduit and concluded that disruption of andesite/basalt stratification and mixing can occur. Koyaguchi (1987) speculated hybridization can be efficiently achieved in a rising magma batch when the volume of magma is smaller than the conduit volume. Withdrawals from chemically stratified magma chamber with limited hybridization have been often observed in relatively voluminous eruptions (e.g., Smith, 1979; Vogel et al., 1989). The volume of the Uchino basalt is so small (0.0067 km$^3$, Hamuro, 1978) that the conduit may have been rootless allowing efficient hybridization in the squeezed conduit to produce homogeneous magma.

Conclusion

The following model is advanced to explain the observed compositional and morphological features of the phenocrysts. A hot basaltic magma was injected into a cooler andesitic magma chamber where Ab-rich plagioclase and Fe-rich clinopyroxene had crystallized. A chemically stratified magma chamber was produced. The andesitic magma layer was heated and the basaltic magma layer was cooled. In the andesitic magma layer, the Ab-rich plagioclase and the Fe-rich clinopyroxene dissolved into the surrounding melts, and in the basaltic magma layer, honeycomb plagioclase grew skeletally under supercooled conditions. Hybridization took place in a rising magma batch and a homogeneous magma was produced, from which chemically identical rims grew on the margin of all plagioclase phenocrysts types.

Coexistence of dusty and honeycomb plagioclase are common in the other hybrid rocks (Larsen et al., 1938; Kuno, 1950; Rhodes et al., 1979; Sakuyama, 1979, 1981; Gerlach and Grove, 1982; Bloomfield and Arculus, 1989) and these examples may also be explained by the same mechanism proposed here.

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References


Sakuyama, M., 1979. Evidence of magma mixing; petrological study of Shirouma Oike calc-alkaline andesite