Compositional Continuity and Discontinuity in the Horoman Peridotite, Japan, and its Implication for Melt Extraction Processes in Partially Molten Upper Mantle

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RECEIVED JANUARY 6, 2003; ACCEPTED SEPTEMBER 4, 2003

A hypothetical model is proposed to explain the origin of compositional discontinuities in the layering observed in orogenic lherzolites. The observed collinearity of the whole-rock peridotite compositions is best explained in terms of partial melting and melt segregation. The presence of chemical discontinuities implies that melt segregation includes an abrupt and discontinuous process. A key concept in the model is the topological transformation of melt geometry in partially molten rocks responding to the equality and inequality of the fluid pressure and solid pressure, which may be realized in a gravitational field. It is emphasized that the percolation threshold is a critical boundary, beyond which a rapid microstructural change occurs in response to the change of local fluid pressure, thus causing a rapid increase of permeability. The model implies that the mode of melting is closer to batch melting than to fractional melting in the upper mantle.

KEY WORDS: critical phenomenon; partial melting; percolation threshold; Horoman peridotite; melt segregation

INTRODUCTION

The Horoman peridotite is a well-characterized orogenic lherzolite massif exposed in the Hidaka metamorphic belt, Hokkaido, Japan. It has a well-developed layered structure and the structure of compositional layering has been studied in detail (Fig. 1) (Nuida, 1974, 1984; Obata & Nagahara, 1987; Frey et al., 1991; Takahashi, 1991, 1992; Takazawa et al., 1999, 2000). Obata & Nagahara (1987) found that the compositional variation of the Horoman peridotite is basically gradational and continuous and forms a wavy and rhythmic pattern of layering. Based on these observations, they proposed a simple conceptual model that explains that such a layered structure was caused by the segregation of partial melt, in which melt percolation and the compaction of the solid matrix occurred concomitantly in ascending masses of upper-mantle material. This model of partial melting was subsequently substantiated and elaborated with additional analytical data by Takazawa et al. (2000).

There are several places in the Horoman peridotite body, however, where the whole-rock composition changes rather abruptly and discontinuously, which cannot be explained by current models of percolation. This paper focuses on these chemical discontinuities and reconsiders the origin of the pattern of compositional variation observed in the Horoman peridotite, in particular the origin of the discontinuities. We consider that the observed compositional variation is best explained in terms of partial melting and melt segregation processes, as was originally proposed by Obata & Nagahara (1987). The presence of compositional discontinuities implies that melt segregation is not necessarily a continuous process, but may instead be an abrupt or episodic process. We explore a possible mechanism by which such discontinuous behavior during melt segregation may result from discontinuous melt percolation in partially molten zones in the upper mantle. The key concept of our new model is the role of gravity, which can induce
variations of pressure difference between the melt and the surrounding solid, which in turn influence the microstructure and thus the permeability of the partially molten rocks through recrystallization.

The model proposed is novel and hypothetical; however, it is potentially important because the discontinuous, non-linear behavior of melt percolation may have significant implications for the mode of melt segregation.
in the upper mantle and, therefore, for the interpretation of the chemistry of primary melts. We use the words ‘melt’, ‘fluid’ and ‘liquid’ interchangeably unless specific distinction is necessary.

**GEOLOGICAL OBSERVATIONS**

Figure 2 is a stratigraphic section through the Horoman peridotite body showing the compositional variation of the whole rocks and the composition of olivine. Although the overall pattern is wavy and gradational as emphasized by Obata & Nagahara (1987), there are several places where the whole-rock composition varies rather abruptly over a short distance as confirmed by Obata & Nishimoto (1992). To investigate the more detailed structure Takazawa et al. (2000) selected an ~140 m section (the Bozu section, in Fig. 1) and examined the compositional variation with much more dense and detailed sampling. They confirmed that the compositional discontinuities do exist and are not just a sampling problem (Fig. 3). Important features observed in the Bozu section are that whole-rock composition varies abruptly across the harzburgite–lherzolite boundary and the lherzolite–plagioclase lherzolite boundary, and the N-type and E-type plagioclase lherzolite boundaries. Despite such discontinuities, the whole-rock compositions define a good collinear trend in major elements such as Al₂O₃ and CaO plotted vs MgO (Fig. 4) as emphasized previously by Obata & Nagahara (1987). The compositional collinearity is best explained in

![Stratigraphic section through the Horoman peridotite body](https://academic.oup.com/petrology/article-abstract/45/2/223/1522108/0x0)
terms of partial melting and melt segregation processes but the compositional jumps are difficult to explain by any conventional model, and may imply a discontinuous behavior of melt percolation in the upper mantle.

The peridotites we deal with in this paper exclusively belong to the Main-Harzburgite–Lherzolite (MHL) suite of Takahashi (1991), except for a few samples in the section illustrated in Fig. 2 (see figure caption). Petrographically the peridotites exhibit porphyroclastic textures and show many features of sub-solidus deformation and recrystallization (Ozawa & Takahashi, 1995; Takazawa et al., 2000). The harzburgites consist of olivine, orthopyroxene and minor amount of clinopyroxene (<3%) and chromian spinel (<0.5%). The lherzolites (referred to as ‘spinellherzolites’ in some of the literature; e.g. Takahashi, 1991; Ozawa & Takahashi 1995) consist of olivine, orthopyroxene, clinopyroxene and spinel. They contain fine-grained (~0-1 mm) mineral aggregates or seams with the same mineral assemblage as the host. The seams are typically ~1 mm thick and several centimeters long, occupying up to 9 modal %, and define a macroscopically well-developed foliation, which is typically parallel to the megascopic compositional layering. They usually contain several clots of pyroxene–spinel symplectite that is thought to be after garnet (Takahashi & Arai, 1989). The plagioclase lherzolites consist of olivine, orthopyroxene, clinopyroxene, plagioclase, and minor amounts of chromian spinel (2–3%) and titaniferous pargasite (~1%). The plagioclase typically occurs in the fine-grained seams (2–9 modal %) and is associated with olivine, chromian spinel, and minor amount of orthopyroxene, clinopyroxene and titaniferous pargasite. The E-type plagioclase lherzolites contain fewer seams and less plagioclase than the N-type plagioclase lherzolites (Fig. 3a).

**MICROSTRUCTURE AND PERMEABILITY OF PARTIALLY MOLTEN ROCKS**

An important macroscopic parameter that governs the velocity of the porous flow in partially molten mantle is the permeability. The permeability is a function of porosity (melt fraction), grain size and the microstructure of the solid and melt network. Generally speaking, the permeability increases as the porosity increases and as grain size increases. Even for the same melt fraction, the permeability may greatly vary according to the microstructure of the pores and the connectivity of the melt channels. The connectivity is related to both the microgeometry of the individual melt-filled pores and the melt fraction; this has been discussed in a number of studies (e.g. Holness, 1997, and references therein).

In partially molten, super-solidus conditions in the upper mantle, recrystallization is generally considered to
be rapid enough, compared with melting reactions and with melt migration along grain boundaries, that the mantle rocks are closely maintained at 'texturablequilibrium'. For this reason the nature of texturable equilibrium has been well studied experimentally and theoretically for mantle systems as well as industrial materials such as metals or ceramics, and the principles that govern the processes are well understood (Bulau et al., 1979, and references therein).

Texturable equilibrium in an idealized system
An important geometrical parameter that characterizes the shape of melt droplets and/or melt tubules is the dihedral angle, $\theta$, which is defined by the force balance of surface tensions (numerically equivalent to the surface energy) between the solid--solid and solid--melt interfaces. For its significance for the microstructure of the partial melt and transport kinetics, a number of experimental studies have been performed to directly measure the dihedral angle for relevant mantle materials (Holness, 1997, and references therein).

For a simple, idealized material in which the solid crystals are isotropic (with respect to surface energy) and uniform in grain size, it is known that, at texturable equilibrium, the melt resides preferentially at corners where four grains meet, and along the edges where three grains meet unless the dihedral angle is equal to zero (Fig. 5a). It is also known that for dihedral angles less than 60° the melt is always interconnected, no matter how small is the melt fraction, thereby forming a three-dimensional, continuous melt network. It is well established that the dihedral angle is below 60° for the olivine--basalt melt system (e.g. Waff & Bulau, 1979; Toramaru & Fujii, 1986). It can, therefore, be concluded that basaltic melt is always interconnected at any infinitesimally small melt fraction in a partially molten mantle system. As soon as partial
melting takes place in the upper mantle, and if textural equilibrium is attained instantaneously, the melt is interconnected from the onset of the partial melting. This led some workers to conclude that the separation of partial melt from partially molten mantle is very efficient and that melt of, say, more than 3%, cannot be maintained in partially molten mantle, for example beneath mid-oceanic ridges (e.g. McKenzie, 1984, 1989).

In natural, more complex systems, however, certain non-zero values of melt fraction may have to be achieved for melt percolation to occur. Toramaru & Fujii (1986) demonstrated that the dihedral angle is greater than 60° for pyroxene--basaltic melt pairs and, by applying percolation theory, they demonstrated that percolation thresholds exist for pyroxene-bearing peridotite systems. This ‘pyroxene effect’, however, may be diminished in hydrous systems because the dihedral angle decreases as the melt becomes hydrous (Fujii et al., 1986; von Bargen & Waff, 1988).

On the contrary, Faul (1997) and Cmi'ral et al. (1998) emphasized the significance of the effect of anisotropy of olivine surface energies on the distribution of partial melt. Faul (1997) demonstrated in his experiments in the olivine--basalt system that a substantial portion of the melt resides as low aspect ratio, disk-shaped inclusions along grain boundaries, probably as a result of the anisotropy of olivine surface energies. Based on this observation and applying percolation theory, Faul (1997) argued that although the melt may be interconnected for all melt fractions along the grain edges where three grains meet, these tubules are very thin and, therefore, the rock is practically impermeable until the melt disks on grain boundaries grow large enough and are sufficiently connected to each other at some higher melt fraction. A conclusion is, therefore, that there is a generally non-zero value of the critical melt fraction, $\phi_c$, for percolation to occur, ‘the percolation threshold’, and therefore, that a small amount of partial melt below $\phi_c$ may be retained in the mantle. The actual value of $\phi_c$ will depend on the mineral mode, grain size and other microstructural factors, and is difficult to estimate precisely but is generally considered to be in the range 2–5% (Faul, 1997).

**Fluid pressure vs solid pressure**

Another important factor to be mentioned regarding the development of textural equilibrium is the force balance between the fluid pressure and the solid pressure across the interface. At textural equilibrium under hydrostatic conditions, a small pressure difference must be present between the fluid and the solid phases because of the presence of curvature and surface tension forces at the interface. This pressure difference is balanced with the surface tension as follows:

$$\Delta P = P_f - P_s = \sigma \left( \frac{1}{r_1} + \frac{1}{r_2} \right)$$

where $P_f$ is the fluid pressure and $P_s$ the solid pressure; $\sigma$ is the surface tension between the melt and crystalline solid; $r_1$ and $r_2$ are the principal radii of curvature measured in two orthogonal directions at the point of interest. The value $[(1/r_1) + (1/r_2)]$ represents the mean curvature, which has to be constant everywhere in the system at textural equilibrium. For $\theta < 60°$, because the interface is convex inward toward the melt (i.e. the mean curvature is negative), $\Delta P$ is negative; that is, the fluid pressure is less than the solid pressure (Fig. 5a). This pressure difference is very small in magnitude compared with the lithostatic pressure, but it is critical in governing the shape and thus connectivity of the melt.

**Permeability**

Beyond the permeability threshold, permeability gradually increases as the melt fraction increases but its detailed behavior is very difficult to ascertain in any general form. A simple formula, often used in the literature for the permeability ($K$), is

$$K = \frac{d^2 \phi^a}{C}$$

Fig. 5. Two possible geometries of melt for a same value of dihedral angle ($\theta$), less than 60°. (a) Fluid pressure is less than solid pressure; (b) fluid pressure is greater than or equal to solid pressure. From Waff (1980).
where \( d \) is the grain size, \( \phi \) is the melt fraction, and \( n \) and \( C \) are numerical constants that depend on detailed geometry of the pores. Von Bargen & Waff (1986) theoretically calculated the permeability for texturally equilibrated pore geometries assuming laminar flow of the melt through melt channels and demonstrated that formula (2) is a good approximation as long as the melt is interconnected. When the permeability threshold \( \phi_c \) is non-zero, equation (2) should be modified by replacing \( \phi \) with \( \phi - \phi_c \) (Maaløe & Scheie, 1982). Actual measurements of the permeability of various synthetic porous materials may be fitted to equation (2) with a value \( n \) between two and three (Faul, 1997; and references therein). Therefore, it is reasonable to assume that the permeability function takes the general shape illustrated in Fig. 6.

**The effect of gravity in a geochemically open system**

The argument above is concerned with equilibrium in a closed system without the effect of the gravitational field. In a permeable zone in which the interstitial melt is interconnected, and therefore, the melt is mobile with respect to the solid matrix as in the partially molten upper mantle, some modification must be made to the above consideration.

At a height \( h \) above the base of the permeable zone, \( \Delta P \) may be expressed, at hydrostatic equilibrium, as

\[
\Delta P = (\rho_s - \rho_f)gh + \Delta P_0 \tag{3}
\]

where \( \rho_s \) and \( \rho_f \) are the mean density of the partially molten rocks and the density of the fluid (melt), respectively; \( g \) is the gravitational acceleration; \( \Delta P_0 \) is the difference in liquid and crystalline solid pressures existing at the base of the permeable zone, where the effect of the gravitational field is negligible. It should be noted that \( \Delta P_0 \) is negative according to relationship (1).

For small melt fractions, as considered here, the mean density of the partially molten rocks may be approximated to the density of the crystalline solid, and because the solid density is generally greater than the melt density, the negative \( \Delta P \) should decrease in magnitude with increasing elevation above the permeability boundary, and will eventually become zero at a certain height.

Waff (1980) predicted that, in a situation where the melt pressure is equal to or greater than the solid pressure, the melt geometry illustrated in Fig. 5a becomes unstable and that in Fig. 5b becomes stable. In the latter geometry, the dihedral angle \( \theta \) remains less than 60°, whereas the mean curvature is zero or even positive. Because \( \theta \) is less than 60°, the principal curvature measured perpendicular to the edge has to be negative, but the other principal curvature, measured semi-parallel to the edge, is positive so that the mean curvature may be zero or even positive. It is important to note that because this shape is possible only along the edges where three grains meet and not possible at the corners, the melt is disconnected for this shape for small melt fractions. Such a topological transformation, therefore, will result in the formation of an impermeable roof of the permeable zone.

It should be noted that the interstitial melt is stationary and the fluid pressure is hydrostatic within the permeable zone. Waff (1980), however, did not explain how mechanical equilibrium is maintained within this permeable column in which \( \Delta P \) is variable vertically. Stevenson (1986) argued that this gravity-induced vertical variation of \( \Delta P \) may be balanced with the surface tension through local melt segregation, and thereby induces variation of the melt curvature. From a simple geometric argument it may be shown that the greater the melt fraction, the smaller the curvature for a constant dihedral angle, and hence the smaller the pressure difference, provided that the grain size is kept constant.

Obviously such textural adjustment is limited to a short length scale because of the generally small melt fractions and will not exceed the order of 1 m (Stevenson, 1986).

Above this permeable zone, either of the following phenomena is considered to take place: (1) macroscopic melt segregation and compaction of the solid matrix; or (2) topological transformation of the melt geometry, forming an impermeable roof as Waff suggested. It is not clear at this point which is more likely to occur in the upper mantle in geologically realistic situations. In the case of (1), the melt fraction at the top of the permeable column is a true percolation threshold; melt segregation is a continuous process and no chemical discontinuity will result in the residual mantle. [This is the process originally considered by Obata & Nagahara (1987).] In the case

![Fig. 6. Evolution of the permeability as a function of the porosity (melt fraction) and the fluid pressure. \( \Phi_1 \) is the first percolation threshold, \( \Phi_2 \) is the second percolation threshold and \( \Phi_3 \) is the critical melt fraction for the topological transformation of the melt. It should be noted that \( \Phi_1 < \Phi_2 < \Phi_3 \). (See text for further explanation.)](https://academic.oup.com/petrology/article-abstract/45/2/223/1522108/1522108)
of (2), on the other hand, a new threshold for macroscopic percolation for the new melt topology has to be located at a significantly higher melt fraction than the conventional threshold. The critical melt fraction of this kind should be dependent upon the geometry of the melt pocket. It is definitely less than 27%, which is a theoretical value for the case of a spherical shape, and would be estimated to be about 22% (Waff, 1980), which implies that up to 20% melt may be retained in partially molten upper mantle.

Most published laboratory experiments, however, indicate that the topology depicted in Fig. 5a is the stable one and the other geometry depicted in Fig. 5b has not been observed. However, it should be noted that in these laboratory experiments, only the confining pressure is controlled and the fluid pressure is not controlled independently. To see the effect of the gravitational field, sample sizes greater than a critical size (at least 1 m in the vertical dimension) must be used (Stevenson, 1986). Such an experiment has never been performed to the authors’ knowledge.

It is rare to see equilibrium textures in naturally partially molten rocks because most mantle-derived ultramafic rocks have undergone crystallization or recrystallization in sub-solidus conditions and the original melting textures have rarely been preserved. Maaløe & Printzlau (1979) observed small, rounded isolated blebs (1–1.5 mm in size) of glass and microcrystalline material consisting mainly of diopside, olivine and spinel in many spinel lherzolite xenoliths from the Dreiser Weiher, Germany. The volume percentage of the blebs varies from a trace to 10–4%. They interpreted these blebs to represent partial melt in the upper mantle. We note that the shape of the blebs is closer to the geometry depicted in Fig. 5b than to the one in Fig. 5a. Nicolas (1989) also argued on the basis of textural observations on European lherzolite massifs (Nicolas, 1986) and lherzolite xenoliths from the Massif Central, France (Nicolas et al., 1987) that the presence of trapped melt up to 10% is possible in an ascending mantle diapir. Although some ambiguities remain in textural interpretations, these lines of evidence are consistent with the geometry depicted in Fig. 5b; we develop our line of reasoning assuming that this geometry of melt pocket may become possible in some circumstances in partially molten upper mantle.

**MELT PERCOLATION IN THE GRAVITATIONAL FIELD AND CRITICAL PHENOMENA**

Using the concept of topological transformation of the melt and of the percolation threshold, as explored in the discussion above, we now seek a mechanism for abrupt segregation of partial melt to occur in a dynamic situation of an ascending and partially molten mantle in a similar manner as considered by Obata & Nagahara (1987).

Let us consider a column of mantle ascending at a constant rate, in which pressure-release melting is taking place everywhere above a partial melting boundary (i.e. solidus) as illustrated in Fig. 7. The melt is not in motion with respect to the solid matrix below the first permeability boundary ($\phi_1$), where melt pressure is less than the solid pressure because of the melt geometry and the surface tension of the interface. Above the permeability boundary a thin permeable zone will be generated in which mechanical equilibrium is maintained and the melt is still stationary (shaded area in Fig. 7) (Stevenson, 1986). This permeable zone is terminated by the topological transformation at some critical level of melt fraction ($\phi_2$), forming an impermeable zone of a certain thickness. The melt is disconnected and the fluid pressure is equal to the solid pressure throughout the impermeable zone. This impermeable zone will be eventually terminated at a certain level because the melt fraction increases upward by the advance of the melting reaction and reaches the second threshold $\phi_3$ for the new melt geometry. Above this level, mechanical equilibrium is no longer possible and macroscopic melt segregation should occur via compaction of the solid matrix. At the base of the new permeable and compacting zone, the fluid pressure must be less than the solid pressure and, therefore, there occurs a discontinuity in fluid pressure across the new permeability boundary. This decrease of fluid pressure has an important influence on the microstructure and thus the permeability structure of the impermeable zone just below.

To see the process in a time sequence, let us choose an arbitrary point fixed with respect to the solid matrix (marked ‘T’ on the left side of each column in Fig. 7) and trace the state of the rocks at this point during mantle ascent. We also follow the variation of the permeability of this point as a function of melt fraction as shown in Fig. 6. Beyond the first permeability threshold $\phi_1$, the rock becomes permeable but the melt is stationary up to the next critical melt fraction $\phi_2$, beyond which the topological transformation occurs and thereby the permeability drops to zero (b to c in Fig. 6). The rock becomes impermeable and continues to be so until the second threshold $\phi_3$ is achieved through the advance of the melting reactions. Fluid pressure is equal to the solid pressure throughout this impermeable zone. As soon as the rock reaches this critical level of melt fraction and enters the permeable zone just above, the fluid pressure suddenly drops whereas the solid pressure does not change. Because of this drop of fluid pressure relative to the solid pressure, the disconnected melt topology becomes unstable and transforms back to the connected topology of the negative $\Delta P$ (Fig. 5a) through recrystallization. Through this topological transformation, the permeability suddenly increases for this high level of melt fraction (from ‘d’ to ‘e’ in Fig. 6).
Because of the abrupt large increase of the permeability, a batch of melt, which had been retained in the partially molten rocks, starts to be segregated upward, leaving a residual area behind (phase 3 in Fig. 7). The residual zone rapidly expands downward because the base of the permeable zone, termed here the ‘permeable front’, is excavated through continuous and instantaneous structural conversion from an impermeable to a permeable structure at the front through microstructural changes responding to the drop of the fluid pressure there and the successive release of partial melt. The permeability at the base of the permeable zone is rapidly reduced as the melt fraction is reduced (from point ‘e’ to ‘b’ along the permeability function curve in Fig. 6). At the critical melt fraction ($\phi_{c2}$) the melt becomes immobilized so that the fluid pressure is raised back to the solid pressure level, which causes the transformation from the permeable structure back to the impermeable structure, and the permeability drops to zero (from ‘b’ to ‘c’ in Fig. 6). This is the termination of the first episode of melt segregation (phase 4 in Fig. 7, which is identical to phase 1).

Once the rocks become impermeable, they continue to be so until the percolation threshold ($\phi_{c3}$) is reached again with further ascent of the column and with further increase of the melt fraction by pressure-release melting. It is important to note that the process beyond the critical stage (phase 3 in Fig. 7) is catastrophic and, therefore, rapid (compared with the ascent rate); the permeability front rapidly descends by releasing the melt upward and excavating the base of the permeable zone. The released melt should migrate upward by grain-boundary percolation.

The rate of melt percolation depends on the permeability structure, which depends on the instantaneous melt fraction and the fluid pressure. It is conceivable that the melt pressure reaches or exceeds the solid pressure somewhere in the upper part of the permeable column, where the ascending melt meets a sudden decrease (or change to a smaller value) in permeability owing to a microstructural change. In this case a certain degree of melt stagnation is expected to occur and this ‘hump’ or region of higher melt fraction will continue to ascend by percolation as a solitary wave, which may be called a magmon (Scott & Stevenson, 1984). Whether this ‘magmon’ migrates all the way through the column or stagnates somewhere in the middle and forms a pool (or a layer) of melt depends on the thermal and rheological structure of the upper mantle. Alternatively, fracturing and vein formation may occur in the partially molten mantle when fluid pressure exceeds the solid pressure and when the ductility of the solid is reduced at shallower levels of the upper mantle (Maaløe, 2003).

**DISCUSSION**

The process we have envisaged above assumes a critical behavior of melt percolation. This critical behavior...
comes from a sudden increase in the permeability in response to a sudden decrease of fluid pressure upon the onset of permeability that induces a microstructural change in the partially molten rocks. Beyond the critical stage, the permeability boundary collapses and it descends rapidly to some lower horizon. During this rapid process, a batch of melt is released and migrates upward, leaving a residual zone behind. As a consequence of this catastrophic melt release, there should occur a sharp front of depletion at the upper edge of the fertile zone. Such a macroscopic geochemical discontinuity is stable and remains unchanged until the next critical stage is achieved. A snapshot of the vertical variation of melt fraction and rock fertility (i.e. the ‘melt’ component retained in the whole rock) is schematically drawn at the right side in Fig. 7. It should be noted that the rock composition is unchanged from the primary one until the depletion front where a sudden depletion occurs.

The model assumes that the topological transformation from a melt-connected geometry to a disconnected geometry occurs soon after the first permeability threshold is exceeded and that substantial amounts of isolated melt are generated in fertile peridotite before collapse occurs. We have reviewed in the previous section several microstructural observations from the literature that such a melt-disconnected structure might actually occur in the upper mantle. One may question then if there is any textural evidence recognized in the Horoman peridotite. As emphasized in the petrographic descriptions in the ‘Geological Observations’ section, the Horoman peridotite has undergone intensive recrystallization and deformation after the main event of partial melting and so it is difficult to make any firm conclusions. The fine-grained seams in the lherzolite and plagioclase lherzolite have been interpreted as the recrystallization products of pyroxene–spinel symplectites (Ozawa & Takahashi, 1995), but it may also be conceivable to suppose that such aggregates are the relics of crystallized melt pockets (Niida, 1984). The Horoman peridotite has a long history even after the main stage of partial melting and magmatic differentiation in the garnet lherzolite-facies stability field (~833 ± 78 Myr, Yoshikawa & Nakamura, 2000). Whether the melt droplets that presumably were formed in the fertile peridotites during the partial melting stage could preserve their morphology and survive a subsequent long sub-solidus recrystallization and deformation history is questionable.

**Geochemical aspects of the model**

In this model the mode of partial melting is considered to be close to batch melting, in the sense that the partial melt is in chemical equilibrium with the residual solid until the critical level is reached. The charged melt is released in a batch rapidly enough so that chemical modification of the melt is minimal during this release. The present model differs from the ‘critical melting’ of Maaløe (1981, 1982), who assumed a permeability threshold being static and not dynamic as in the present model. A static threshold cannot produce such abrupt motion of partial melt, nor chemical discontinuity. It just assumes a continuous overflow of infinitesimally small increment of partial melt above the fixed level of permeability. The ‘critical melting’ of Maaløe (1981, 1982) is, therefore, a continuous process and, geochemically speaking, it is a kind of fractional melting with certain amounts of interstitial melt being always present in the residual matrix. The present model perhaps more deserves to be called ‘critical melting’ in the strict sense of the word.

There has been a long-lasting controversy on the mode of partial melting in the upper mantle, i.e. batch melting vs fractional melting. Nicolas (1989) considered that significant amounts of trapped melt can be present stably in the upper mantle. On the contrary, geochemical studies of mid-oceanic ridge basalts indicate that melt can be effectively segregated from the residual crystalline phases at melt contents as small as 0-1% (Salters & Hart, 1989; Johnson et al., 1990). Geochemical studies of the Horoman peridotite and the proposed model suggest that batch melting is important for magma generation in the upper mantle.

**Multiplicity of the layering**

The process considered above produces only a single layer of discontinuity—the first one directly above the partial melting boundary (Fig. 7). Theoretically, a second layer (and perhaps a third and more layers) of discontinuity may occur at some elevated horizons in the ascending column by a series of similar processes as long as the rocks are fertile enough to produce more melt by pressure-release melting. The timing of the critical collapse of the permeability boundary in the upper level should be influenced by flow of melt from below and it is difficult to predict exact behaviors. By this series of critical melt segregations at multiple levels, we expect to generate a layered structure in the peridotite above the permeability boundary. If the separation of partial melt is complete, without stagnation midway, then we should see a stepwise succession of depletion upward. If the melt stagnation occurs, we will have a more complicated pattern characterized by a multiple layered structure.

**Geological implications and application of the model**

It is important to note that the resultant pattern of compositional layering has a distinct polarity in such a manner that a stepwise depletion always occurs at the upper edges of fertile zones. The observed apparent stratigraphic sequence in the Bozu section (Fig. 3) (i.e. from harzburgite at the base, through lherzolite to plagioclase lherzolite, in ascending order) is, however, exactly
opposite in directional pattern to the above theoretical scheme. It follows then that the whole Bozu section is structurally inverted from an original succession; that is, downward is the original upward direction. On a wider scale both polarities are mixed in the Horoman peridotite (Fig. 2). Statistically there are more reversed patterns than normal patterns.

When we try to interpret the observed pattern, the factor we must bear in mind is the effect of deformation. There is substantial textural and structural evidence that indicates the peridotite had suffered intensive plastic deformation in the upper mantle, even after solidification, and thus that the original structure may have been severely distorted. The layered structure may have been substantially thinned by stretching during plastic flow in the mantle (Ozawa, 1997). Toramaru et al. (2001) argued for such a deformation process and presented a kinematic model for the formation of the layered structure by means of the repetition of stretching and isoclinal folding in an ascending diapir in the upper mantle. Therefore, we have to be very cautious in interpreting the observed structural pattern. Nevertheless, we emphasize that the geochemical pattern and discontinuous local pattern can be preserved surviving the structural disturbances and thus is useful for constraining the modeling.

Considering the measured section of the Horoman peridotite (Fig. 3), the degree of partial melting has been estimated from geochemical data, assuming that the most fertile plagioclase lherzolite is the starting primary composition, to be 4–12% for N-type plagioclase lherzolite, 16–18% for E-type plagioclase lherzolite and 23–25% for harzburgite (Takazawa et al., 2000). Therefore, by applying the present model, the step ‘height’ of the chemical discontinuity is inferred to be of the order of 10% in the scale of melt fraction.

The geochemical modeling is not yet fully successful for certain trace elements, however, and problems remain unsolved (Takazawa et al., 2000). For example, some enrichment of La observed at the boundary between the harzburgite and lherzolite cannot be explained (Fig. 3c). Such a geochemical anomaly may be ascribed to the chromatographic effect of percolating melt or melt wall-rock reactions that may have produced the layered structure (Takazawa et al., 2002), which is a totally different process from that considered in this paper.

It is conceivable that layers with such compositional discontinuities are successively generated in zones of ascending mantle. The formation of each layer may be affected by the formation of the layer directly below through the upward influx of percolating melt from the lower layers. This is a one-way interaction between neighboring layers of criticality. Because of such interactions, the whole system may work as a highly non-linear dynamic system and the formation of the layered structure as a whole can be regarded as a self-organized critical phenomenon in a non-equilibrium dynamic system (e.g. Nicolis & Prigogine, 1977; Bak & Chen, 1991). It should be noted that, in this dynamic system, the gravitational field plays a central role. It is suggested that the observed compositional discontinuities in the Horoman peridotite represent such boundaries, created by series of catastrophic critical processes. The formation of multiple layering may be a very complex process and we have not yet achieved a full understanding of it. The assessment of the effect of large-scale deformation of the peridotite is a necessary task.

The geochemical implication of the model is that the mode of melting in the upper mantle is closer to batch melting than to fractional melting. Because of the complex behavior of the non-linear dynamic system, it may be very difficult to explain the entire structure observed in the Horoman peridotite. We must also bear in mind the effect of large-scale deformation, such as flow folding and stretching, which may have occurred on global scales in the upper mantle. Nevertheless, we believe that the proposed model provides a useful means to interpret the geochemical stratigraphy observed not only in the Horoman peridotite, but also in other orogenic lherzolite massifs worldwide. The amount of partial melt that may be discharged in a single catastrophic event is inferred be about 10 vol. %, based upon geochemical studies of the Horoman peridotite.

ACKNOWLEDGEMENTS
This work is an outcome of a Japan–US collaborative research project supported by the Japan Society for the Promotion of Science and the US National Science Foundation. We thank Fred Frey (MIT) and Nobumichi Shimizu (WHOI) for their many years of collaboration and continuous help during the research, and David Kohlstedt (University of Minnesota), David Stevenson (Caltech) and Atsushi Yamaji (Kyoto University) for their stimulating discussions and comments during the preparation of the manuscript. The paper has been significantly improved by critical reading and constructive criticisms of two anonymous reviewers, which were useful in clarifying the model and the writing. However, should any errors or ambiguities remain, they are the sole responsibility of the authors.