

Periodic Formation of Magma Fractures and Generation of Layered Gabbros in the Lower Crust Beneath Oceanic Spreading Ridges

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There must be a transition from continuous porous flow to transient formation of melt-filled fractures in the region of magma transport below oceanic spreading centers. This transition may occur at permeability barriers that impede porous flow of melt, at the base of the oceanic crust and within the lower crust. We summarize evidence for formation of melt-filled lenses at the base of the crust in the Oman ophiolite, and evidence indicating that melt is very efficiently extracted from these lenses, probably in fractures. We also discuss the possible formation of melt lenses elsewhere in the oceanic lower crust. We then present a simple physical model for the periodic formation of melt-filled fractures originating in a melt lens beneath a permeability barrier. Finally, quantitative models show how modal layering in lower crustal gabbros can form as a result of the periodic pressure changes associated with fracture formation. An ancillary result of chemical modeling is the quantification of a thermal gradient in the lower crust of the Oman ophiolite during igneous accretion beneath a spreading center, from 1165 to 1195°C near the dike/gabbro transition to ~1240°C near the crust/mantle transition. This and other data for the Oman gabbros support models in which much of the lower crust forms by crystallization in sills at a variety of depths, from the dike/gabbro transition to the base of the crust.

1. INTRODUCTION

Many models for melt extraction from the mantle and igneous accretion of the oceanic crust at spreading ridges emphasize continuous porous flow processes (e.g., McKenzie [1984]; Spiegelman [1993a,b]; Aharonov et al. [1995, 1997]) and, although this is debated, there is evidence in support of the hypothesis that melt transport in the melting region is dominantly by focused porous flow (e.g., Kelemen et al. [1995a,b, 1997a]). However, it is clear that the formation of sheeted dikes is periodic, comprising the "quantum event of upper crustal accretion"

[Delaney et al. 1994]. Thus, somewhere in the region of melt migration there must be a transition from continuous to punctuated flow. In this paper, we propose that the two periodic phenomena in oceanic crust, sheeted dikes and layered gabbros, may have a common physical basis. The transition from continuous to punctuated melt transport may occur mainly at the base of the lower crustal, layered gabbros, where a permeability barrier blocks porous ascent of melt, giving rise to gradually increasing melt pressure, in excess of lithostatic pressure. We present a physical model in which excess melt pressure is periodically relieved by formation of melt-filled fractures. This model can quantitatively explain the genesis of melt-filled fractures and layering in gabbro. We conclude with a discussion of various kinematic models for the formation of the lower oceanic crust, with an emphasis on constraints from igneous petrology and ophiolite gabbro compositions.

1.1. Transition from Continuous to Periodic Melt Flow

Nicolas (e.g., [1986, 1990]) and Maaloe (e.g., [1981]) have proposed that the transition from porous flow to flow in periodically formed fractures occurs in the mantle melting region. However, some recent work reaffirms the idea that, though melt-filled fractures do form beneath ridges, melt extraction from the adiabatically upwelling mantle may occur mainly by porous flow throughout the melting region [Kelemen *et al.* 1995a,b 1997a]. In this view, melt-filled fractures in the asthenosphere are not required, given available field, geochemical and physical data, to explain the genesis of mid-ocean ridge basalt (MORB) or residual mantle peridotites. However, none of the data can be used to rule melt-filled fractures in the melting region [Kelemen *et al.* 1997a], and the presence or absence of such fractures is a matter of debate.

In any case, it is certain that porous flow is an important melt transport process at very shallow depths in the mantle beneath ridges; Boudier and co-workers (e.g., Boudier and Nicolas [1995]) and Ceuleneer and co-workers (e.g., Ceuleneer [1990]; Ceuleneer and Rabinowicz [1992]) have emphasized the prevalence of "impregnated peridotite", formed by crystallization of melt moving by porous flow, in the crust/mantle transition zone of the Oman ophiolite (Moho transition zone, or MTZ). Similar features are also found in the shallow mantle (MTZ?) at the East Pacific Rise, as drilled at Hess Deep [Gillis, Mével, Allan *et al.* 1993; Dick and Natland 1996; Boudier *et al.* 1996a].

Gabbroic sills are abundant in parts of the MTZ in the Oman ophiolite [Benn *et al.* 1988; Ceuleneer 1990; Boudier and Nicolas 1995; Boudier *et al.* 1996b]. Geochemical study of these sills, as well as "impregnations", in the Oman MTZ shows that they are very pure "cumulates"; i.e., they are magmatic rocks but they do not have plausible liquid compositions. Instead, they formed by partial crystallization of silicate liquids, from which most of the melt was extracted [Benoit *et al.* 1996; Kelemen *et al.* 1997b; Korenaga and Kelemen 1997]. Their compositions are determined by the stoichiometry of minerals that crystallized from a melt. These characteristics are shared by lower crustal, layered gabbros in the Oman ophiolite and in gabbros dredged from the mid-ocean ridges [Smewing 1981; Pallister and Hopson 1981; Pallister and Knight 1981; Browning 1982, 1984; Juteau *et al.* 1988; Meyer *et al.* 1989]. Thus, the MTZ and the lower crust commonly include regions of high melt fraction, which undergo partial crystallization, from which melt has later been extracted to form other parts of the crust.

Theoretically, melt-filled fractures originating in the mantle might propagate all the way through the crust to the sea floor (e.g., Nicolas [1986, 1990]). However, the rarity of primitive, mantle derived magmas forming dikes and lavas in ophiolites and along the mid-ocean ridges (summaries in Sinton and Detrick [1992]; Langmuir *et al.* [1992]; Kelemen *et al.* [1997b]), and the abundance of

gabbro sills and "impregnated" peridotites in the MTZ, suggests that propagation of melt-filled fractures from the mantle source to the sea floor has not occurred.

Instead, the evolved compositions of dikes and lavas suggest that they are derived from crustal "magma chambers" where primitive, mantle-derived melts have been modified by crystal fractionation. This hypothesis is supported by geochemical studies showing that gabbro sills in the MTZ, and lower crustal gabbros, are in major and trace element exchange equilibrium with the liquids that formed sheeted dikes and lavas in the upper crust [Pallister and Knight 1981; Kelemen *et al.* 1997b]. Even if magma transport in fractures below the MTZ is an important melt migration mechanism, the presence of anelastic, melt-rich regions at the MTZ would generally stop propagation of these fractures. Instead any melt ascending in melt-filled fractures in the mantle would be temporarily trapped in sills. In this paper, we concentrate on melt extraction at and above the MTZ.

Kelemen *et al.* [1997b] and Korenaga and Kelemen [1997, 1998] proposed that diffuse porous flow over vertical distances larger than meters to tens of meters was not the main process of melt extraction from gabbros in the MTZ and lower crust. Instead, most melt extraction must have been in fractures or channels of focused porous flow. There is little evidence for either melt-filled fractures or porous flow channels in the lower crustal section of the Oman ophiolite, though there are some centimeter-scale dikes. The general paucity of vertical melt migration features may be a consequence of transposition of initially steep features, such as dikes, due to ductile extension of the crust beneath a spreading ridge (e.g., Quick and Denlinger [1993]).

It may also be that melt-filled fractures in the lower oceanic crust rarely form dikes at all. Melt extraction in cracks involves an initial pressure decrease followed by near adiabatic ascent. In the lower crust, rocks surrounding melt-filled fractures will be hot, close to the magmatic temperature. Under these circumstances, conductive heat loss to the walls will be slow, and ascending silicate liquids will be undersaturated in solid phases. Only in the latest stage of melt transport in fractures, as fracture width and melt velocity decrease, will crystallization occur in lower crustal conditions. In some cases, this may form cm-scale dikes. If fracture width decreases to the crystal size in surrounding rocks, crystallization of melt may "heal" the fracture without a trace. In addition, if rocks surrounding fractures are permeable, porous flow of melt into the wall rock may also heal fractures with no trace.

1.2. Physical Models for Periodic Formation of Melt-Filled Fractures From Magma Chambers

In this paper, we adopt the hypothesis that periodic or intermittent melt extraction arises by formation of melt-filled fractures. Physical models for periodic melt extraction from magma chambers at the MTZ and within the crust

have been previously proposed by *Gudmundsson* (e.g., [1986, 1990]) and *Ida* [1996]. *Gudmundsson* emphasized the role of tensile stresses due to lithospheric spreading, and did not incorporate the effects of magmatic influx on melt pressure, whereas in this paper we emphasize the role of continuous influx, with plate-scale tension due to spreading entering the problem only in the sense that it decreases the tensile stress necessary to fracture rocks overlying a magma chamber. The model of *Ida* [1996] is similar to that adopted in this paper, except that *Ida* used a viscous rheology for the rocks hosting melt-filled fractures, whereas we use an elastic rheology for rocks surrounding fractures. In *Ida's* formulation, melt-filled fractures close by viscous flow of the wall rocks. By contrast, in our model fractures become very narrow due to elasticity of the wall rocks. When crack width and melt velocity become very small, melt in fractures solidifies to form dikes or, if the fractures are very narrow, the two walls simply anneal together.

An analogous physical problem is the development and episodic fracture of overpressured fluid "compartments" in sedimentary basins (e.g., *Hunt* [1990]; *Dewers and Ortoleva* [1988; 1994]). *Nur and Walder* [1992] proposed a simple analytical model to account for this process which resembles the one in this paper in some respects. However, the rheology of rocks in sedimentary basins is considerably different from that adopted for oceanic gabbros here.

A mechanism involving periodic "plugging" and "unplugging" of magmatic conduits has also been proposed by *Whitehead and Helfrich* [1991], who emphasize the role of temperature dependent magmatic viscosity. Our model can be viewed as a simplified, limiting version of the *Whitehead and Helfrich* approach. They included continuous changes in melt viscosity, whereas we assume a step function in viscosity; melt viscosity is constant up to the solidification temperature, at which point the viscosity becomes "infinite". Also, we assume the fracture opens and closes elastically, in response to changes in melt pressure, and may heal completely by crystallization, whereas *Whitehead and Helfrich* assumed a constant conduit width.

Our specific hypothesis is that permeability barriers at the MTZ are created by crystallization of multiply saturated basaltic melts within a porous medium. Rapid crystallization reduces porosity and permeability [*Korenaga and Kelemen* 1997; *Kelemen et al.* 1997b]. Other permeability barriers may be produced within the oceanic lower crust via a variety of crystallization and melt extraction mechanisms. Magma ascending by porous flow beneath these permeability barriers will pond and form melt-filled sills. In these sills, continuous influx of magma leads to increasing magmatic overpressure that periodically exceeds the tensile strength of the overlying rock, giving rise to melt-filled fracture. We present some simple physical models of this proposed mechanism, and show how it could give rise to the kinds of igneous modal layering that are observed in the Oman ophiolite in sills in the MTZ (e.g., *Boudier et al.* [1996b]) and in lower crustal gabbros [*Pallister and*

Hopson 1981; *Smewing* 1981; *Browning* 1982; *Juteau et al.* 1988]. If magma supply to the MTZ is mainly by continuous porous flow, then sills at this level record the transition from porous flow to periodic fracture. In any case, once sills are present, they are unlikely to be traversed by melt-filled fractures, and will instead be fed by porous flow and/or melt-filled fractures from below.

1.3. Origin of Modal Layering in Gabbros

In the light of our proposed model, the presence of layered gabbros in the MTZ and in the lower crust of the Oman ophiolite can be viewed as complementary to the presence of sheeted dikes in the upper crust, in that both arise from periodic formation of melt-filled fractures. Periodic pressure change has been previously proposed as a mechanism for the formation of igneous modal layering (e.g., *Cameron* [1977]; *Lipin* [1993]; also please see Section 5.5, "Pressure fluctuations" in *Naslund and MacBirney* [1996]). However, this is by no means a unique explanation for the origin of layering in gabbroic plutons. Other proposed mechanisms include (a) infusions of new magma into a pre-existing magma chamber, (b) periodic gravitational crystal settling due to "avalanches" or to intermittent magmatic convection, (c) sidewall crystallization from vertically stratified, chemically independent, "double diffusive" convection cells in a large magma chamber, (d) kinetic mechanisms for spatially and/or temporally periodic crystal nucleation and growth, and (e) intrusion and crystallization of sills within previously crystallized gabbroic rocks (e.g., review in *Naslund and MacBirney* [1996]). We believe that there is ample evidence that each of these mechanisms has operated to produce igneous modal layering at some times and places, and we will not attempt an exhaustive evaluation of each of them in this paper.

Many of these explanations are most applicable to magma chambers of considerable vertical extent, emplaced within relatively cold wall rocks. *Korenaga and Kelemen* [1997] conclude that periodic pressure change, together with subsequent infusions of new magma, is the most likely mechanism for formation of layering in the Oman MTZ sills. In this paper, we present liquid line of descent models that quantify this process.

In extending this idea to formation of "normal" oceanic crust, it is necessary to add the caveat that obvious subhorizontal, modal layering is rare or absent in available samples of gabbros dredged and drilled from the mid-ocean ridges. However, layering is rarely visible in small samples of Oman gabbros, and this is also true of our observations of diamond drill core from layered gabbros in East Greenland such as the Skaergaard and Kap Edvard Holm intrusions. Generally, outcrops at least 10's of centimeters in diameter are required to discern modal layering, which is evident because of its lateral continuity along strike, and complementary lack of continuity across strike. Also, drilling recovered samples from upper gabbros and from

shallow mantle, formed along the East Pacific Rise (e.g., Gillis, Mével, Allan *et al.* [1993]), and some mafic gabbros have been dredged from Hess Deep and from fracture zones in the Pacific (e.g., Hekinian *et al.* [1992, 1993]; Constantin *et al.* [1996]), very few samples of lower crustal gabbros from the middle of fast- to medium-spreading ridge segments have been obtained to date. Due to the compositional similarity of gabbro samples from the oceans (e.g., Meyer *et al.* [1989]; Hekinian *et al.* [1993]; Dick and Natland [1996]) and gabbros in Oman, we infer that layered gabbros will, in fact, prove to be common beneath mid-ocean ridges.

1.4. Sills in the MTZ in the Oman Ophiolite

In this paper, we rely on evidence from layered gabbros in the lower crust [Smewing 1981; Pallister and Hopson 1981; Pallister and Knight 1981; Browning 1982, 1984; Juteau *et al.* 1988] and in gabbroic sills in the MTZ in the Oman ophiolite [Benn *et al.* 1988; Ceuleneer 1990; Boudier and Nicolas 1995; Boudier *et al.* 1996b; Kelemen *et al.* 1997b; Korenaga and Kelemen 1997] to constrain processes in the lower crust and at the MTZ on-axis beneath fast- to medium-spreading oceanic spreading ridges.

The MTZ sills play a particularly large role in our reasoning. An important question is whether some of the gabbros in the MTZ might be huge inclusions of lower crustal gabbro enclosed within intrusive dunites. We find this implausible on the following grounds. (1) Interdigitated contacts between gabbro lenses and surrounding dunite preclude the possibility that the gabbro lenses are imbricate fault slices of lower crustal gabbro (e.g., Figure 2 in Korenaga and Kelemen [1997]). (2) Modal layering within the gabbroic lenses, and the external contacts of the lenses, are always parallel to the regional trend of the crust/mantle transition, and never show rotation with respect to other lenses or layering in lower crustal gabbros. If the lenses were xenoliths that had been intruded by a magma, some of them would be rotated. (3) Benoit *et al.* [1996] report compositionally similar gabbro lenses within residual, shallow mantle harzburgites (within about 1 vertical km below the crust/mantle transition). These cannot be xenoliths. (4) While small dunite intrusive bodies are associated with some wehrlite intrusions into gabbros in Oman, in general the formation of true dunites by intrusion of a mixture of melt plus olivine, followed by extraction of all but a few percent of the melt, is a rare event that is unlikely to form the great thicknesses of dunite observed in the Oman MTZ. Thus, we remain convinced that most or all of the gabbro lenses in the Oman MTZ were emplaced as sills intrusive into surrounding peridotite.

Sills in the Oman MTZ are generally 0.1 to 10 meters thick and 10's to 1000's of meters long, with aspect ratios (height/length) less than 0.01. They show evidence for flattening by magmatic compaction (pure shear) and/or magmatic deformation (simple shear), so that their current shapes are not the same as their shapes during magmatic

crystallization. However, it seems likely that they have not undergone more than 10:1 flattening. Also, it is important to realize that sills may have been built up by successive crystallization from several melt lenses. Each lens would probably have had a similar horizontal extent, compared to the sills, but individual lenses may have had a smaller vertical dimension. Thus, during magmatic crystallization the melt lenses were sub-horizontal, tabular bodies with heights less than 10 meters and aspect ratios (height/length) less than 0.1 [Korenaga and Kelemen 1997].

A somewhat independent estimate of "magma chamber height" can be derived from geochemical data. The magnitude of compositional variation of layered gabbro sections within MTZ sills and lower crustal gabbros in Oman permits estimation of the relative mass of initial liquid to crystals. Given the current thickness of such gabbro sections, and estimates of the amount of flattening, one can estimate "magma chamber heights". Using this technique, Browning [1982] and Korenaga and Kelemen [1997] estimated that "magma chambers" within both MTZ sills and lower crustal gabbros were on the order of 0.1 meters to 10's of meters. Since igneous layers in these sections commonly extend for more than 100 meters, this again indicates that the "magma chambers" were sub-horizontal sills, both in the MTZ and in the sites of crystallization of lower crustal gabbros.

1.5. Sills in the MTZ Beneath Fast- to Medium-Spreading Ridges

We infer that the sills in the Oman MTZ are similar to sills in the MTZ beneath fast- to medium-spreading ridges on the following grounds. The Oman ophiolite contains pillow lavas and sheeted dikes with major and trace element characteristics similar to MORB. The presence of a continuous layer of sheeted dikes underlying pillow lavas indicates that the igneous crust of the ophiolite formed at a submarine spreading ridge. In detail, the rare earth elements (REE) and other incompatible trace elements have lower concentrations at a given Cr concentration than most MORB (e.g., Alabaster *et al.* [1982]; Pearce *et al.* [1981]). Also, in some of the northern massifs, andesitic lavas comprise part of the extrusive section. These characteristics suggest that the ophiolite may have formed in a super-subduction zone setting. However, strong similarities between the compositions of Oman lavas in general and MORB are compelling evidence that petrogenetic processes in formation of the ophiolite were similar to processes operating at a normal mid-ocean ridge. On the basis of radiometric age data, subdued crustal thickness variations, a general lack of paleo-fracture zones, and other geological observations, it is probable that the ophiolite formed at a fast- to medium-spreading ridge (e.g., Tilton *et al.* [1981]; Nicolas [1989]).

An essential caveat is that ophiolites intrinsically preserve features formed "off-axis", in the oceanic lithosphere, and features formed during emplacement of the

ophiolite along a compressional plate margin, as well as features formed "on-axis" beneath a submarine spreading ridge. In the case of the Oman ophiolite, igneous ages and emplacement ages are virtually the same, within error [Boudier *et al.* 1985; Michard *et al.* 1991]; Hacker *et al.* 1996], so that on-axis structures may have been affected by the emplacement process. For these reasons, it is possible that the sills in the MTZ in Oman are not characteristic of the MTZ beneath a normal mid-ocean ridge.

However, we hypothesize that sills along the MTZ are, in fact, an on-axis characteristic of normal fast- to medium-spreading ridges on the following grounds:

(1) Major and trace element geochemistry of the gabbroic sills indicates that their constituent minerals are in exchange equilibrium with the liquids that formed the sheeted dikes and lavas of the ophiolite [Kelemen *et al.* 1997b]. This is also true of minerals in lower crustal gabbros [Pallister and Knight 1981; Kelemen *et al.* 1997b]. Mass balance requires that the liquid extracted from gabbroic sills in the MTZ and/or from lower crustal gabbros must comprise a substantial proportion of the crust (e.g., Browning [1982]; Pallister [1984]; Kelemen *et al.* [1997b]). Thus, the liquids extracted from the sills and/or lower gabbros must be represented by the sheeted dikes and overlying lavas; most or all of the gabbros must have formed beneath a spreading ridge.

(2) Nearly solid gabbros are present immediately above the Moho within a few km of the ridge axis at fast-spreading ridges such as the East Pacific Rise (e.g., Vera *et al.* [1990]). There is likely to be substantial intergranular porous flow of melt near the MTZ. Intergranular flow of melt leads to solid-liquid equilibrium over distances larger than the solid grain size [Spiegelman and Kenyon 1992]. Thus, the MTZ must be the level of plagioclase (\pm clinopyroxene) saturation in ascending magmas at fast- to medium-spreading ridges (see next section). Thus, gabbroic rocks should become abundant within the MTZ on-axis at fast- to medium-spreading ridges.

(3) Seismic data indicate high melt proportions near the MTZ beneath active spreading ridges ([Dunn and Toomey 1997]; and submitted manuscript, Crawford *et al.* 1998). Other workers suggest that a highly reflective Moho, both on-axis and up to 30 km off-axis, may represent the emplacement of plutons at this level [Garmann 1989, 1994]. Thus, there is evidence for accumulation of melt near the MTZ at fast-spreading, mid-ocean ridges.

On this basis, we proceed with our hypothesis that sills form on-axis in the MTZ at fast- to medium-spreading ridges, and play an important role in crustal accretion.

2. CRYSTALLIZATION OF ASCENDING BASALT NEAR THE MTZ

For the purposes of this paper, we define the oceanic crust as that portion of the oceanic plates composed of >95% plagioclase-rich rocks (troctolite, olivine gabbro, gabbro, gabbronorite, etc.) plus fine-grained basalt over vertical

distances greater than 10 meters. The MTZ represents the region in which plagioclase-bearing rocks are mixed with mantle peridotite, and neither rock type comprises more than 95% of the rock over vertical distances greater than 10 meters. By this definition, beneath slow spreading ridges the MTZ may be present over a considerable depth interval (e.g., Cannat [1996]). However, at fast spreading ridges, the geothermal gradient is likely to be nearly adiabatic within the ascending mantle all the way up to the MTZ (e.g., Sleep [1975]). In such a thermal structure, the MTZ will be a narrow depth interval immediately beneath the crust, perhaps a few hundred meters thick, in which any basalt ascending by porous flow becomes saturated in plagioclase (\pm clinopyroxene). If this were not the case, and melts saturated only in olivine came into contact with gabbroic rocks above the MTZ, the liquids would dissolve plagioclase (plag) \pm clinopyroxene (cpx), moving the Moho upward, until the liquids themselves became saturated in plag \pm cpx. Conversely, if melts became saturated with plag \pm cpx below the MTZ, they would form gabbroic rocks and the MTZ would extend downward.

When mantle-derived basaltic liquids similar to MORB become saturated in plag \pm cpx, their mass drops rapidly with continued cooling. By comparison, olivine \pm spinel fractionation at higher temperatures consumes relatively little liquid mass per $^{\circ}\text{C}$ of cooling. In Figure 1, we illustrate results of calculations using several different petrological models for the crystallization of MORB. In each, we have taken a melt composition proposed to be a mantle-derived liquid parental to MORB, and modeled fractional crystallization during cooling from above its liquidus through the temperature of plag and cpx saturation. As can be seen, crystallization rates on the order of 1 to 3%/ $^{\circ}\text{C}$ are predicted over a 20 to 40 $^{\circ}\text{C}$ interval below the temperature of olivine + plag (\pm cpx) saturation. The results of our calculations are consistent with experimental results summarized by Sinton and Detrick [1992] in their Figure 7.

The high rate of crystallization below olivine + plag (\pm cpx) saturation, in % liquid/ $^{\circ}\text{C}$, is due to the pseudo-eutectic nature of multiply-saturated basalt at crustal pressures, analogous to the isobarically invariant eutectic composition in the simple system forsterite (olivine) - anorthite (plag) - diopside (cpx). Cooling of olivine-saturated liquids similar to parental MORB in this system leads to minor olivine crystallization, followed by co-saturation in olivine + plag, and then olivine + plag + cpx. At the eutectic in the simple system, all of the magma is consumed by crystallization at constant temperature before cooling can continue. In the natural system of course, olivine + plag + cpx saturation is not truly eutectic, but the analogy with the simple system explains the large decrease in magma mass over a small cooling interval below the temperature of multiple saturation.

Cooling, crystallizing liquids at the MTZ will become increasingly evolved, with higher Fe/Mg for example. In this way, they will diverge from the composition of magmas in equilibrium with mantle peridotite. Where they

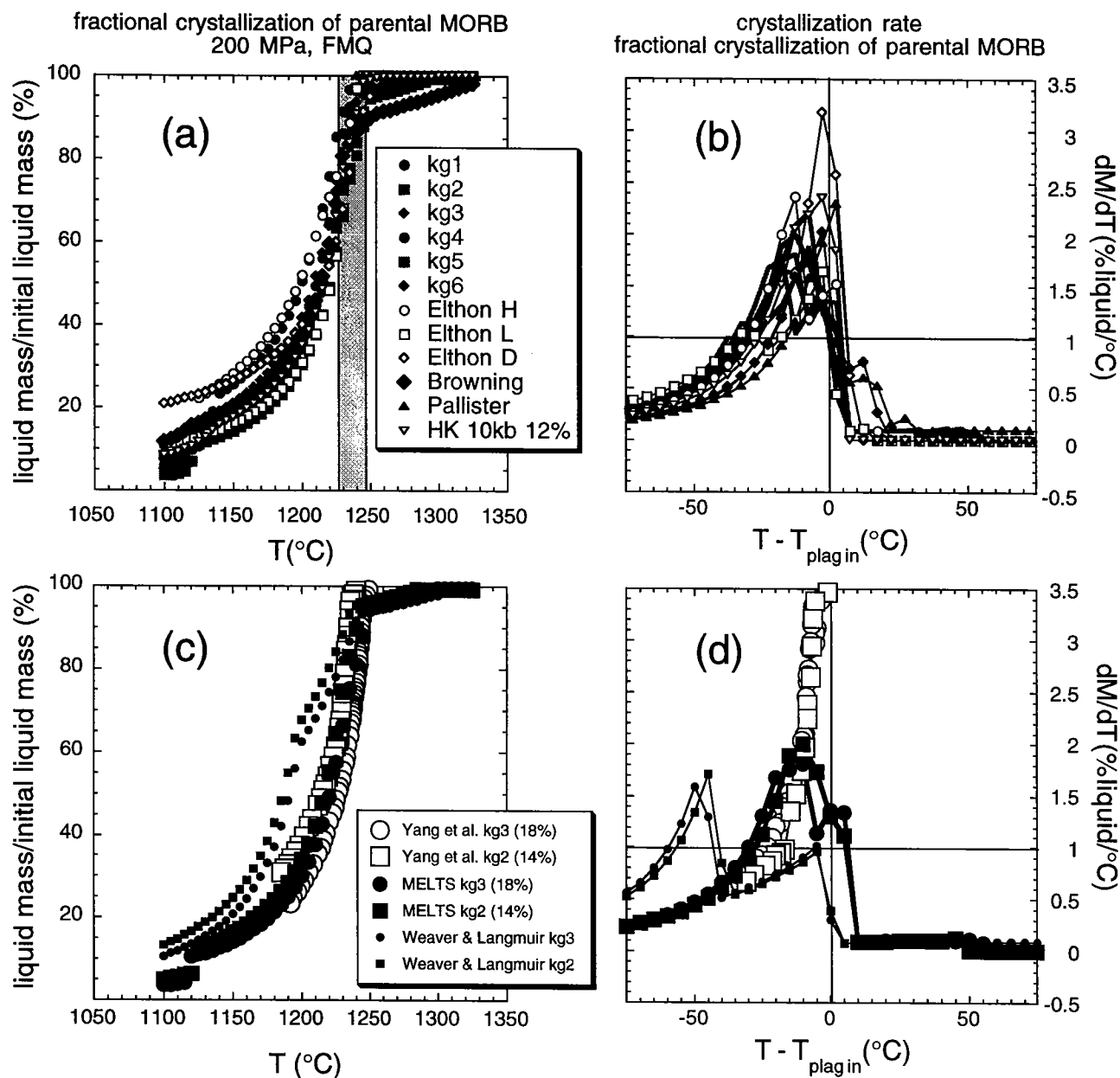


Figure 1. Modeling results for fractional crystallization of proposed, parental mid-ocean ridge basalt (MORB) liquids, H₂O-free at 200 MPa along the Fayalite-Magnetite-Quartz oxygen fugacity buffer. The main result is that the crystallization rate, in % crystallized/°C (labelled dM/dT, %/°C), increases dramatically to more than 1%/°C at the temperature of plagioclase saturation in all of these liquid compositions. The range of plagioclase saturation temperatures is indicated by a vertical grey bar in Figure 1a. Figures 1a and 1b illustrate results for a variety of proposed parental MORB compositions, calculated using the MELTS silicate liquid solution model of *Ghiorso and Sack [1995]*. Compositions labeled "kg" are calculated liquid compositions for polybaric fractional melting of the mantle, based on parameterization of experimental results [*Kinzler and Grove 1991*]. kg2 is calculated for an average of 14% partial melting, and kg3 is for an average of 18% partial melting. The compositions labeled "Elthon" are proposed parental MORB liquids based on study of MORB glasses, oceanic gabbros and peridotites, and experimental and theoretical studies [*Elthon et al. 1992*]. The compositions labeled Browning and Pallister are based on mass balance calculations and average rock compositions from the Oman ophiolite [*Browning 1982; Pallister 1984*]. The composition labeled HK 10kb 12% is an experimental liquid composition from *Hirose and Kushiro [1993]*, produced by 12% batch melting of mantle peridotite. Figures 1c and 1d illustrate results for liquid compositions kg2 (14% melting) and kg3 (18% melting), using several different models [*Weaver and Langmuir 1991; Ghiorso and Sack 1995; Yang et al. 1996*]. Although the results of the various models differ, particularly in the temperature of calcic pyroxene (cpx) saturation, for our purposes they are similar in predicting a large increase in the rate of crystallization after cooling below the temperature of plagioclase saturation.

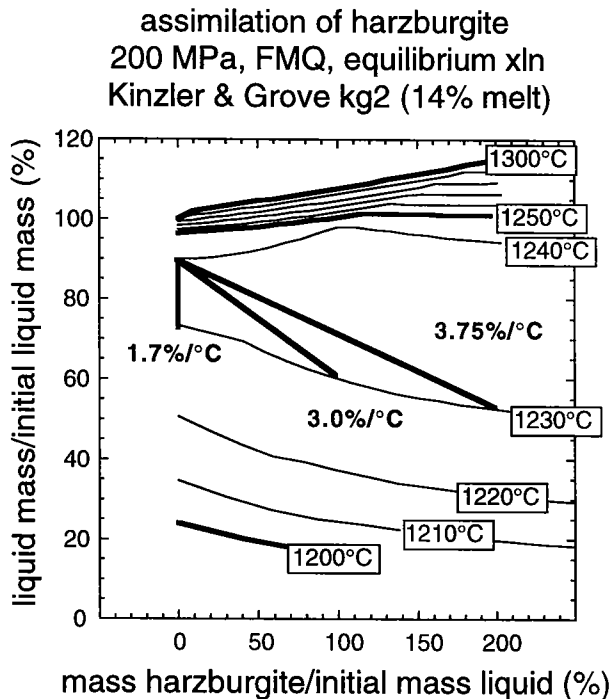


Figure 2. Modeling results for equilibrium crystallization of parental MORB liquid composition kg2 plus varying proportions depleted mantle harzburgite, calculated using the MELTS model, H₂O-free at 200 MPa along the Fayalite-Magnetite-Quartz oxygen fugacity buffer (FMQ). The kg2 liquid composition is from Kinzler and Grove [1991]. The harzburgite reactant is an average composition for shallow, residual mantle peridotite samples dredged from the mid-ocean ridges, from Dick [1989]. Curves show the results of isothermal melt/rock reaction at a variety of temperatures. As previously predicted by Kelemen [1990], and experimentally verified by Daines and Kohlstedt [1994], reaction between residual mantle harzburgite and olivine-saturated basaltic liquid at constant temperature dissolves pyroxene and precipitates a smaller mass of olivine, producing an increase in the liquid mass. Just below the temperature of plagioclase saturation, at 1240°C, reaction produces a liquid saturated only in olivine [Kelemen 1990], and liquid mass increases during isothermal reaction. By contrast, where liquids are well into the olivine + plagioclase (\pm clinopyroxene) saturated region, isothermal reaction produces a decrease in melt mass [Kelemen 1986, 1990]. Thus, below the temperature of plagioclase saturation, combined cooling and reaction generally produces a larger crystallization rate than cooling alone.

come into contact with mantle peridotite, the combined effect of cooling and melt/rock reaction will decrease liquid mass still further compared to the effect of cooling alone. This process has been extensively discussed in previous papers (e.g., Kelemen [1986, 1990]), and is illustrated quantitatively in this paper in Figure 2. Reaction between crystallizing liquid and mantle harzburgite nearly doubles the rate of crystallization with cooling.

3. PERMEABILITY BARRIERS AT THE MTZ

As discussed in the previous section, multiply saturated liquids crystallize tens of percent of their mass over a 20 to 40°C temperature interval below plagioclase saturation. We infer that this temperature is attained by melts migrating by porous flow at the depth of the MTZ. Sparks and Parmentier [1992] and Spiegelman [1993c] proposed that crystallization of cooling magma entering the conductively cooled lithosphere forms a permeability barrier in the mantle. In a variant on these ideas, Kelemen *et al.* [1997b] proposed that the depth of multiple saturation immediately beneath a ridge axis may become a permeability barrier, restricting the ascent of magmas rising by porous flow and leading to the formation of sills in the MTZ.

Korenaga and Kelemen [1997] quantified the conditions under which this can occur. If there is a forced flux of porous melt flow across an interval in which the melt is crystallizing within its pore space, there are two possible end-member consequences: (1) the solid matrix will deform to maintain constant porosity at constant melt pressure, or (2) the porosity will decrease and the melt pressure will rise. The outcome depends upon the relative rates of viscous deformation required to keep the pore space open and of crystallization in the pore space, which in turn depends upon the flux of melt and the rate of cooling. Given that oceanic crust at fast- to medium-spreading ridges attains nearly 100% of its thickness within about 2 km of the ridge axis (e.g., Vera *et al.* [1990]), it is relatively straightforward to estimate the flux of melt that must pass through the MTZ beneath a spreading ridge. The cooling rate is far more difficult to determine.

One way to constrain the cooling rate is via seismic data indicating that lower crustal gabbro immediately above the Moho is essentially solid (porosity < 5%) within about 5 km of the East Pacific Rise at 9°N (e.g., Vera *et al.* [1990]; Toomey *et al.* [1990]). Given a half spreading rate of about 0.05 m/yr, an initial magmatic temperature of 1300 to 1250°C, and a solidus temperature (porosity < 5%) for lower crustal gabbros of ~1200°C (Figure 3), this yields cooling rates $\sim 10^{-3}$ °C/yr, and lateral temperature gradients ~ 0.01 to 0.02 °C/m. On the basis of this estimate, it is clear that thermal energy is efficiently removed from the oceanic lower crust near spreading ridges, and that this can drive igneous crystallization.

Another way to constrain the cooling of the crust beneath a spreading ridge is via comparison of near-solidus temperatures for MTZ sills, lower gabbros, upper gabbros, and lavas from the Oman ophiolite (Figure 3). For an on-axis porosity of ~10%, calculated temperatures for MTZ sills are ~1240°C and temperatures for lower crustal gabbros range from ~1205 to 1225°C. For an on-axis porosity of 30 to 10%, calculated temperatures for the shallowest gabbros are ~1190 to 1170°C. These estimates are discussed in more detail in section 8.2 of this paper.

Use of these values to constrain the on-axis geotherm requires the assumption that all of the rocks in the Oman

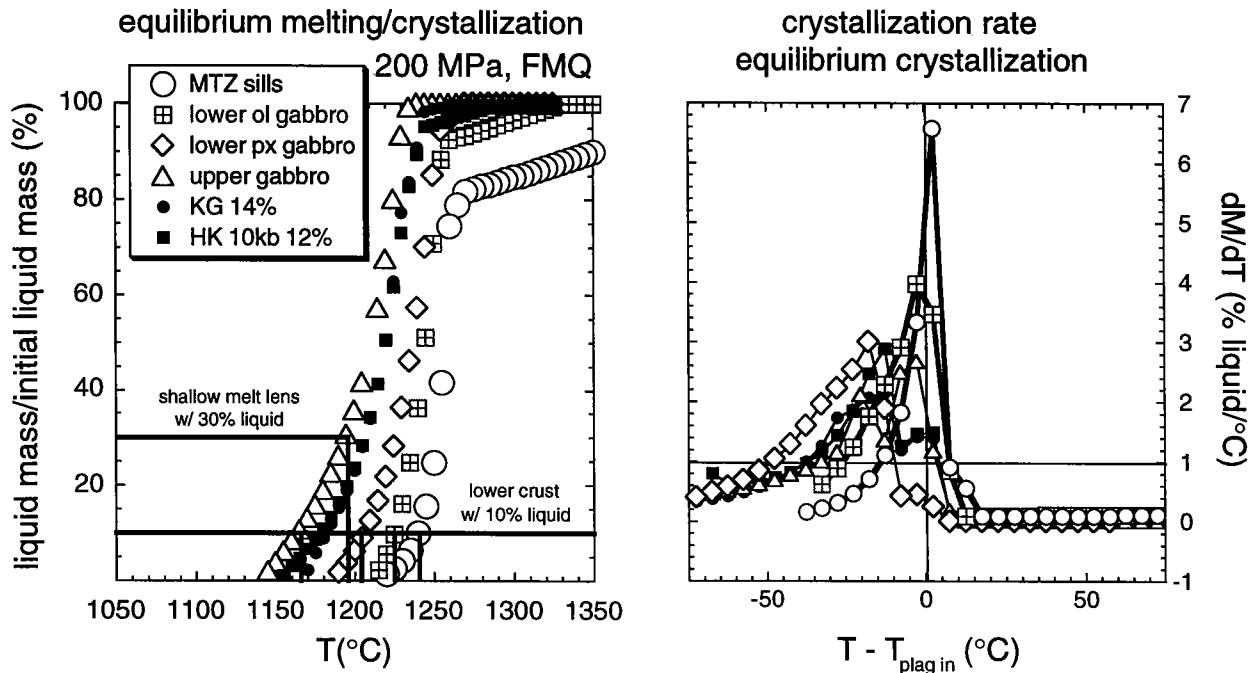


Figure 3. Modeling results for equilibrium crystallization of gabbro compositions from the Oman ophiolite, calculated using the MELTS model, H_2O -free at 200 MPa along FMQ. The composition for sills from the Moho Transition Zone (MTZ) is an average of 26 samples from *Korenaga and Kelemen [1997]*. The other gabbro compositions are averages from *Browning [1982]*. Equilibrium crystallization results for two parental MORB liquid compositions are shown for comparison. Because equilibrium crystallization consumes liquid more rapidly than fractional crystallization, all of the calculated crystallization curves are steeper than for fractional crystallization (compare Figure 1). The upper gabbro composition shows crystallization behavior similar to the liquid compositions. Averages from the layered, lower gabbros, and gabbro sills in the MTZ, have nearly vertical melting curves over the range from 10 to >80% liquid. The melting curves are at progressively higher temperatures for samples from the upper gabbros, the lower gabbros, and the MTZ, which probably indicates a temperature gradient of 45 to 75°C beneath the spreading ridge during formation of the lower crust in Oman.

crustal section formed on-axis. If they did, then the temperatures imply an average vertical temperature gradient of $\sim 0.01^\circ\text{C}/\text{m}$. While it is unclear how this average vertical temperature gradient is distributed within the crust, it is apparent that the base of the crust is hotter than the average lower crust, and this in turn is hotter than the depth at which the uppermost gabbros crystallize, suggesting a relatively smooth temperature gradient. Thus, heat removal beneath the ridge axis may be sufficient to drive crystallization of melt near the MTZ.

It is difficult to be more precise about the magnitude of heat flux away from the lower crust beneath spreading ridges. Y. J. Chen and J. Phipps Morgan (personal communication 1998) have suggested that the rate of heat removal is small, and therefore can drive only a limited amount of basalt crystallization near the base of the crust, $\sim 10\%$ of the total crustal mass. However, this is based on assumptions concerning the amount and spatial distribution of hydrothermal circulation that are difficult to justify in detail, as Chen and Phipps Morgan would agree (e.g.,

Phipps Morgan and Chen [1993], p. 6287). If hydrothermal circulation in the lower crust 2 km off-axis beneath fast-spreading ridges is more vigorous than they have assumed, then more crystallization could occur near the base of the crust.

Using the oceanic crustal flux, cooling rates of 0.01 to $0.02^\circ\text{C}/\text{m}$, and rates of crystallization of multiply saturated basalt of 1% to $3\%/^\circ\text{C}$, *Korenaga and Kelemen [1997]* found that crystallization can fill porosity more rapidly than viscous flow of the matrix maintains constant porosity, so that the result is a low porosity, permeability barrier.

Implicit in this discussion is the inference that the permeability barrier at the MTZ must be subhorizontal or concave downward beneath a spreading ridge, essentially parallel to sub-horizontal isotherms. While this inference is in accord with generalized thermal models for the oceanic crust (e.g., *Sleep [1975]*), the vigor and geometry of hydrothermal convection could lead to locally steep isotherms, and this is a subject which needs to be evaluated in detail. If isotherms near the MTZ are indeed subhorizontal, magma ascending

due to buoyancy by porous flow will accumulate beneath permeability barriers. This may be responsible for formation of both gabbroic sills and more diffuse, "impregnated peridotites", in the MTZ.

Korenaga and Kelemen found that pressure in melt ascending due to buoyancy beneath a sub-horizontal permeability barrier may increase for two reasons: (1) due to forced flux of melt combined with elastic deformation of the solid matrix, and/or (2) by localizing the pre-existing vertical pressure gradient in the fluid within the region immediately surrounding the barrier. The second type of pressure effect has been numerically modeled in a simplified, rigid system by *Aharonov et al.* [1997].

Although the initiation of the permeability barrier and an underlying layer rich in accumulated melt requires porous flow, porous flow need not be the only mechanism of melt transport beneath the MTZ. If melt ascends in fractures, the presence of a weak layer or melt-filled lens will tend to create an anelastic barrier to fracture propagation. Thus, sills in the MTZ, once formed, will effectively trap all ascending melt, regardless of the mechanism of melt transport in the underlying mantle.

4. PERMEABILITY BARRIERS WITHIN CRYSTALLIZING GABBROS

Permeability barriers may form within crystallizing, lower crustal gabbros as well as along the MTZ. For example, it has been proposed that anorthosite bands, composed almost entirely of plag, form permeability barriers in layered intrusions (e.g., *Boudreau and McCallum* [1986]; *Boudreau* [1988]). Anorthosite bands are observed in lower crustal gabbros in Oman (e.g., *Pallister and Hopson* [1981]) and in the Bay of Islands ophiolite (e.g., *Bédard* [1991]). Such bands may be permeability barriers within the crystallizing oceanic lower crust.

Another point where permeability barriers might arise within crystallizing gabbros is at the point of low calcium pyroxene saturation. (In this paper, low calcium pyroxene is symbolized as "opx", since it is often orthopyroxene). At this point liquid compositions are analogous to an isobarically invariant reaction point in a simple system, in this case the point forsterite-cpx-opx in the system forsterite - diopside - silica, and forsterite-plag-opx in the system forsterite-anorthite-silica. Thus, as for ol-plag-cpx co-saturation, crystallization rates (%/°C) may increase dramatically at the level of plag-cpx-opx saturation, reducing porosity where liquids ascend by porous flow.

Finally, permeability barriers may arise within "cumulate" gabbros simply as a result of melt extraction. Most lower crustal gabbros in the Oman ophiolite are composed of unzoned, refractory minerals crystallized from basaltic magma at high temperature; the remaining liquid was later extracted [*Pallister and Hopson* 1981; *Browning* 1982, 1984; *Juteau et al.* 1988; *Kelemen et al.* 1997b; *Korenaga and Kelemen* 1997]. After melt extraction, these

cumulate gabbros had a very small intergranular porosity, even at temperatures where basaltic liquid compositions are more than 60% liquid (Figure 3). Thus, after melt extraction, cumulate gabbros could constitute low porosity barriers to the ascent of magma by porous flow.

In all of these cases, accumulation of large liquid fractions beneath permeability barriers will create low viscosity, anelastic layers - in the limit of 100% porosity, these will be sub-horizontal, melt-filled lenses - that cannot be traversed by propagating fractures from below. Thus, once one or more lenses are present in the crust, they will act as barriers or traps for liquid ascending by melt-filled fracture. Instead, liquid from any fractures propagating from below will be trapped in these lenses, as previously suggested by *Gudmundsson* [1986].

5. MODEL FOR BEHAVIOR OF A CONTINUOUSLY FILLED MELT LENS

In this section we investigate a simplified model of the behavior of a viscoelastic melt lens below a sub-horizontal layer of low permeability rock. Stresses in the lens may rise due to continuous melt influx from below and accumulation of liquid. The rising stress may finally result in hydrofracture. If the stress state is isotropic, fracturing may preferentially occur in the horizontal direction (forming or extending melt lenses) due to the presence of a strong, sub-horizontal layer above. However, beneath mid-ocean ridges the tectonic environment is expected to produce a horizontal extensional component, such that $\sigma_H < \sigma_V$, where σ_H and σ_V are the horizontal and vertical stress components, respectively. In addition to tectonic stresses, the inflation of a melt-filled lens may generally produce a horizontal extensional component on the walls of the lens, as demonstrated by calculations for lacoliths (see *Turcotte & Schubert* [1982], p. 120).

As the lens inflates, the horizontal stress in the overlying layer will have an increasing extensional component, until a stress state is reached where vertical fractures will form. Magma flowing through vertical fractures exits the lens and relieves the elevated stress. As melt flow rates in the fracture drop, the fracture closes elastically. As the fracture becomes very narrow and melt flow velocities decrease, melt may crystallize to form a dike, or may "leak" out by porous flow. In either case, under many circumstances the fracture may be considered to close, or heal.

We loosely describe the melt-filled body beneath a permeability barrier as a lens because permeability barriers in igneous systems may often be parallel to isotherms, and in many geological systems isotherms are affected by conductive cooling to the Earth's surface are sub-horizontal. Also, evidence from sills in the Oman ophiolite crust/mantle transition zone indicates that they crystallized from melt-filled lenses with vertical dimensions more than 100 times smaller than their horizontal dimensions. We describe the melt lens as having some initial (deflated) volume V_0 , and

an additional volume contribution derived from elastic inflation of the lens by incoming melt, V_e . A 2D melt lens is drawn schematically in Figure 4.

To simulate the dynamic behavior of the system, we need to know the relationship of melt flux to the melt pressure in the lens. We use a relationship between the elastic expansion and the magma pressure, appropriate for elongate bodies such as cracks and lacoliths [Rubin 1995a],

$$V_e \propto \frac{l^2}{G} p, \quad (1)$$

where l is the horizontal length of the lens, G is the elastic stiffness of the host rock, and $p = P - \rho gh$ is the melt pressure in excess of lithostatic pressure (ρ is the density of the overlying crustal column, g is the gravitational constant, h is the depth beneath the Earth's surface). The proportionality coefficient depends on geometry, and has a value of one for cracks [Rubin 1995b].

We use equation (1) and the fact that the volume of the melt lens changes due to input and output melt fluxes to describe how the pressure in the chamber evolves with time. We make a simplifying assumption that as the melt lens inflates or deflates, its lateral extent remains constant. Given these constraints

$$\frac{l^2}{G} \frac{\partial p}{\partial t} = Q_0 - Q_1 - \frac{1}{v} p, \quad (2)$$

where Q_0 is a continuous volumetric influx of melt and Q_1 is the volumetric flux out of the lens through fractures that form at the top of the lens. The last term in equation (2) is a pressure relaxation term, with v being a relaxation parameter, allowing the chamber to exhibit Maxwell viscoelasticity (e.g., Wilkinson [1960]). The relaxation parameter allows for irreversible volume changes in the melt lens that arise from porous flow of melt outward into the surrounding rock, viscous deformation of the walls of the melt lens, and crystallization of melt.

In what follows, we introduce three postulates for our coupled melt lens-fracture model. The first is an assumption regarding when the surrounding rock will fracture and what width of fracture will form. The second postulates laminar melt flow in fractures, and the final assumption approximates the conditions under which a fracture will heal.

(1) We assume that the low permeability layer overlying the magma chamber will fracture when $p > p_{\max}$, where p_{\max} is a critical melt pressure in excess of lithostatic stress that is needed to fracture the rock. In addition, we assume that a fracture instantaneously reaches its final length, h , and neglect effects due to fracture propagation. Once the fracture is open, we assume that low permeability rock surrounding the fracture behaves elastically (simple Hookean spring

model) with respect to the open fracture of width a , with elastic stiffness G (e.g., Rubin [1995a]), giving

$$a = \frac{h}{G} (p + \Delta\sigma_H). \quad (3)$$

The deviation of the horizontal stress from lithostatic stress is $\Delta\sigma_H = \rho gh - \sigma_H$, and may be calculated given a specific geometry of the melt lens and the tectonic stresses.

(2) We assume that once a fracture is open, melt flow through the fracture is laminar and constant with height, so it can be modeled with an average melt velocity

$$\omega = -\frac{a^2}{12\mu} \left(\frac{\partial p}{\partial z} + \Delta\rho g \right) \approx \frac{a^2}{12\mu h} (p + \Delta\rho gh) \quad (4)$$

where ω is melt velocity, μ is the melt viscosity, $\Delta\rho g$ is the buoyancy term arising from different densities of melt and solid, and z is depth taken to be positive upward. The pressure difference between melt and surrounding rock at the top of the crack is taken to be zero.

(3) We next approximate the conditions under which a fracture of length h , with magma flowing with velocity ω , will close, "heal", or solidify. Two competing processes occur during melt flow through a fracture: heating of the wall due to advection of hot melt, and cooling of the melt

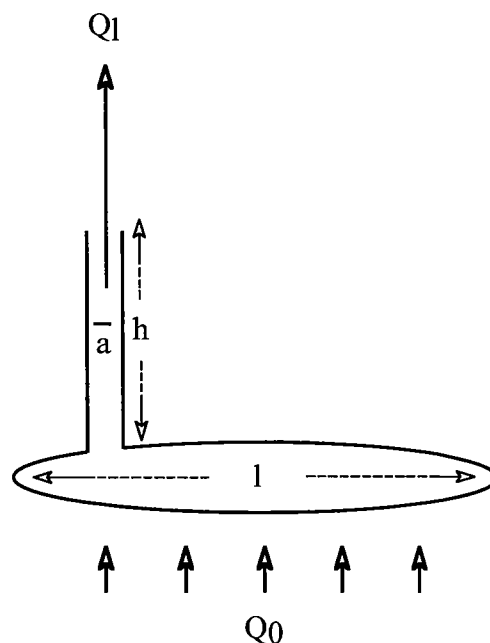


Figure 4. Schematic illustration of a melt lens below a low permeability layer, with an open crack allowing for melt flux out of the top of the lens.

due to contact with colder, surrounding rock. The latter process leads to crystallization along the walls and decreasing fracture width. In our dynamic problem, pressure and flow rates in the fracture are both varying with time, due to decreasing pressure and volume in the melt lens as melt is removed. An analytical solution for solidification of a dike when both the pressure and the flux are varying is not yet available. However, solutions for solidification of melt driven by a constant pressure (e.g., [Bruce and Huppert, 1989; Lister and Dellar, 1996]) can be used here with some simplifying assumptions. Lister and Dellar [1996] find that if melt is driven by a constant pressure in the source region, there exists a critical Peclet number, Pe_c , below which solidification will occur, and above which melt-back will occur. The Peclet number is the ratio of the time scale for melt transport through the fracture, $t_t \sim h/\omega$, and the time scale for solidification to form a dike, $t_s \sim a^2/\kappa$, where κ is the diffusivity of heat, and is thus defined by equation (2.6) of Lister and Dellar [1996] as

$$Pe = \frac{(p + \Delta\rho gh)a^4}{h^2\mu\kappa} \quad (5)$$

When flow rate is slower than solidification rate, the Peclet number of the system is smaller than the critical value and the fracture will solidify. Although in reality solidification occurs over some time interval, we will approximate it here as occurring instantaneously if the Peclet number is below Pe_c . The value of Pe_c depends on the melt temperature in the lens, T_0 , the temperature of the rock surrounding the fracture, T_∞ , the solidification temperature, T_L , the latent heat of crystallization, L , the specific heat capacity, c , and diffusivity of heat, κ . Reasonable parameters for our problem may be $L = 8 \times 10^5$ J/kg, $\kappa = 10^{-6}$ m²/s, $c = 730$ J/kg°C, $T_0 = 1230^\circ\text{C}$, $1230^\circ\text{C} > T_\infty > 1220^\circ\text{C}$, and $T_L \sim 1225$ to 1230°C , of which the most uncertain are the temperatures of the melt and the surrounding rock. Using these values and Figure 5 of Lister and Dellar [1996], we determine that Pe_c is of order 1 for our system.

In order to use Pe_c as a criterion for solidification, we note that our system has a continuously changing Peclet number, since both driving pressure and fracture width change with time due to depletion of the source reservoir. When a fracture is formed, the pressure is high and the fracture is wide, so that the value of the instantaneous Peclet number as described in equation (5) is large. As the pressure drops, the value of the instantaneous Peclet number is also reduced. Thus, our approximate fracture healing criterion is that the fracture will heal when and if

$$Pe = Pe_c \sim 1 \quad (6)$$

In mathematical terms, the governing equations for the

fracture width, the pressure in the chamber, and the flow in the fracture (equations (3)-(6)) exhibit hysteresis:

for $p \geq p_{\max}$

$$\rightarrow a = \frac{h}{G}(p + \Delta\sigma_H), \quad \omega = \frac{a^2}{12\mu h}(p + \Delta\rho gh); \quad (7a)$$

for $p < p_{\max}$ & $\omega a^2 > \omega_c a_c^2$

$$\rightarrow a = \frac{h}{G}(p + \Delta\sigma_H), \quad \omega = \frac{a^2}{12\mu h}(p + \Delta\rho gh); \quad (7b)$$

and for $p < p_{\max}$ & $\omega a^2 \leq \omega_c a_c^2$

$$\rightarrow a = 0, \quad \omega = 0, \quad (7c)$$

where a_c and ω_c are fracture width and average melt velocity when $Pe = Pe_c$, i.e., when melt in the fracture solidifies to form a dike. From equations (4), (5) and (6),

$$\omega_c a_c^2 \sim \frac{kh}{12}. \quad (8)$$

Note that when $a=0$, the width of the melt-filled fracture is 0, but a residual dike of finite width, formed by earlier crystallization along the fracture walls, may remain in the rock record.

5.1. Non-dimensionalization

We non-dimensionalize the problem for ease of solution. The primed non-dimensional variables are: $p' = p/p^*$, $p'_{\max} = p_{\max}/p^*$, $e' = \Delta\sigma_H/p^*$, $a' = a/a^*$, $t' = t/t^*$, $Q'_0 = Q_0 t^*/l^2$, $\omega' = \omega/\omega^*$, where the starred variables are characteristic dimensional quantities: $p^* = \Delta\rho gh$, $\omega^* = a^2 p^*/(12\mu h)$, $t^* = h/\omega^*$, $a^* = hp^*/G$. Finally we define a characteristic Peclet number, $Pe^* = p^* a^{*4}/(\kappa h^2 \mu)$. The non-dimensional equations are then

$$\gamma p' = Q'_0 - \delta \omega' a' - R p'; \quad (9)$$

for $p' \geq p'_{\max}$

$$\rightarrow a' = p' + e', \quad \omega' = a'^2 (p' + 1); \quad (10a)$$

for $p' < p'_{\max}$ & $Pe^* \omega' a'^2 > 1$

$$\rightarrow a' = p' + e', \quad \omega' = a'^2 (p' + 1); \quad (10b)$$

and for $p' < p'_{\max}$ & $Pe^* \omega' a'^2 \leq 1$

$$\rightarrow a' = 0, \quad \omega' = 0, \quad (10c)$$

where

$$\gamma = \frac{p^*}{G} \quad (11)$$

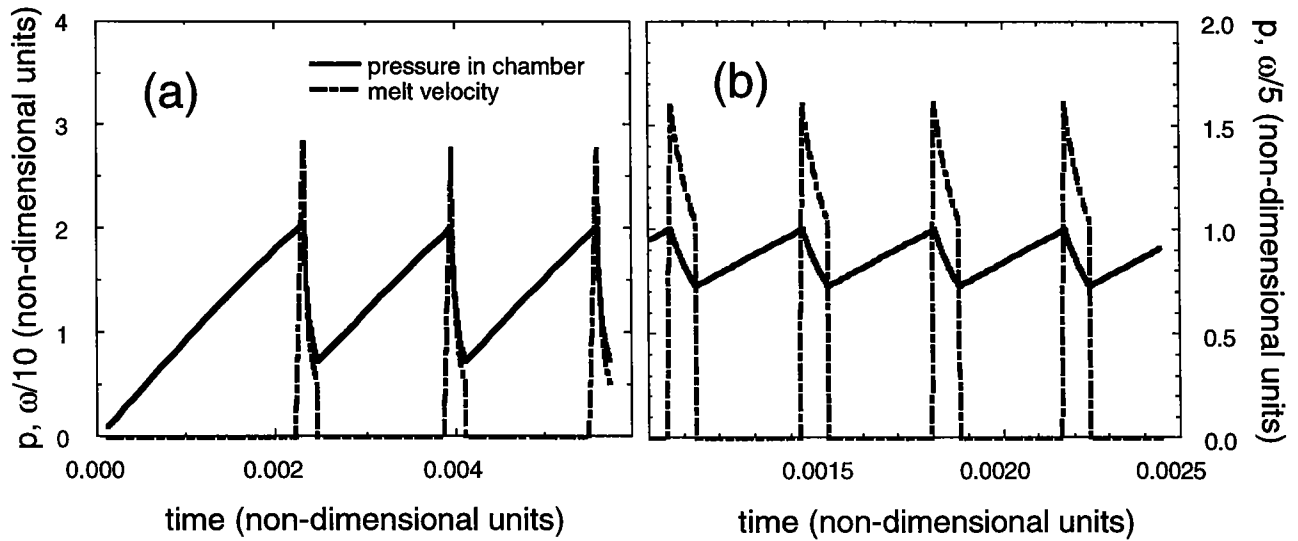


Figure 5. Melt pressure and flow velocity numerically calculated for $\gamma = 5 \times 10^{-5}$, $R = 0.005$, $Q'_0 = 0.05$, and $\delta = 0.02$. (a) has $p'_{max} = 2$ and (b) has $p'_{max} = 1$. For both cases, oscillatory behavior is established, in which pressure (solid line) rises in the melt lens until it reaches $p' = p'_{max}$. At that point a fracture is opened, and melt flow (dashed line) is maximal. As time progresses and pressure is released, flux through the fracture is reduced, until finally the healing criteria is reached and flow in the dike ceases due to crystallization or annealing of the fracture walls. At this time, pressure in the melt lens begins to rise again. In (a), the cycles are more asymmetric than in (b), because the pressure rise time in (a) is longer due to higher p'_{max} , as explained in the text.

is a coefficient expressing the inflation and deflation ability of the melt lens,

$$R = \frac{p^* t^*}{l^2 v} \quad (12)$$

measures the relative amount of relaxation of pressure on the time scale of melt flow, and

$$\delta = \frac{a^* h}{l^2} \quad (13)$$

5.2. Solution:

We use the initial conditions of $p'(0)=0$ and $a'(0), \omega'(0)=0$, and numerically calculate solutions for equations (9) and (10). We find that there are three regimes for solutions:

(1) If $Q'_0/R > p'_{max}$ (see equation (9)), the rock surrounding the melt lens will fracture. There are then two possibilities: 1a. oscillating fracture and healing, or 1b. a continuously open fracture and steady state melt flow.

(1a) In the case of oscillatory fracture, initial melt velocity in the fracture is high, but as the pressure is released the velocity decreases, and the crack heals before the melt velocity reaches a steady value. Then the pressure increases

again, and the process repeats itself in a cyclic manner (Figure 5).

The time $\Delta t'_r$ for the rise of pressure between cycles is obtained by solving equation (9) with a closed fracture, ($a'=0$), and an initial condition of $p'=0$,

$$\Delta t'_r = \frac{\gamma}{R} \ln \left(\frac{Q'_0}{Q'_0 - R p'_{max}} \right) \quad (14)$$

Where “viscous” pressure dissipation is negligible, $R \rightarrow 0$, and the rise in pressure between cycles is linear:

$$\Delta t'_r = \gamma \frac{p'_{max}}{Q'_0} \quad (15)$$

Similarly, the decay time $\Delta t'_d$ of pressure due to melt expulsion through the fracture can be calculated from equation (9). Although the rise time, $\Delta t'_r$, increases without bound with increasing p'_{max} , the decay time, $\Delta t'_d$, approaches a constant value as p'_{max} increases, with increasingly longer rise times as can be seen in Figure 5.

(1b) The second possible scenario following fracture of the rock surrounding the melt lens is an open fracture, with steady state width and melt velocity, in which the output flux is equal to the input flux (Figure 6). In our model we have not included melt-back, but this is possible for case

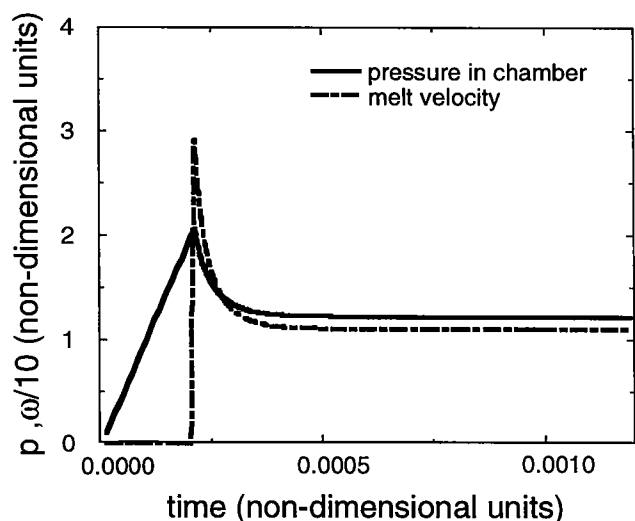


Figure 6. Pressure and flow velocity calculated numerically with flux into the melt lens, Q'_0 , ten times greater than in Figure 5. Otherwise, all values are the same as in Figure 5a. Initially, melt pressure (solid line) rises in the melt lens until it reaches p'_{max} . Then a fracture is opened and melt flow (dashed line) is maximal. As time progresses, pressure drops. Outflux through the fracture is reduced and reaches a steady value equal to Q'_0 . As a consequence, the crack width and melt velocity are never small enough to allow for the fracture to heal.

(1b) when input flux is large enough and/or hot enough, as shown by *Bruce and Huppert [1989]*. Neglecting “viscous” pressure dissipation during the time that the fracture is open, and assuming $e' \sim 1$, this steady state will occur if the input flux is high, or the solidification time is long. The criterion for continuous, steady outflux is

$$Q'_0 > \frac{\delta}{Pe^{*4/5}} \quad (16)$$

(2) If $Q'_0/R < p'_{max}$ then the system reaches a steady state (see equation (9)) where the pressure in the melt lens is elevated relative to the initial pressure, but has not reached the point where it can fracture the surrounding rock. The larger the relaxation parameter, or the smaller the melt input, the more likely it is that this steady state will be reached before fracture occurs (Figure 7).

6. SCALED CALCULATIONS, AND COMPARISON TO OBSERVED PHENOMENA

Field observations indicate that the case of oscillatory release of melt does occur beneath oceanic spreading ridges. It is possible to obtain some quantitative estimates of the physical process from the simple model we have developed

in the previous section. First, an estimate for p_{max} can be made using laboratory measurements of the strength of rocks. Using this estimate, and ignoring viscous dissipation, we can obtain the time between consecutive fracture events, Δt_r , using equation (15), and translating back into dimensional variables

$$\Delta t_r = \frac{l^2}{GQ_0} p_{max}. \quad (17)$$

The melt flux through the Moho beneath a fast-spreading mid-ocean ridge must be $\sim 600 \text{ m}^2/\text{yr}$ per unit length of the ridge. This is distributed over a cross-sectional region about 4 km wide, perpendicular to the ridge axis. If we take a typical MTZ melt lens to be 0.5 m high by 100 m long (volume of 10 m^3), there would be about 40 sills across 4 km, and the flux into each would be $15 \text{ m}^2/\text{yr}$.

The elastic stiffness of rocks, G , may be $\sim 10^{10}$ Pa (10 GPa) for gabbroic rocks at temperatures near their solidi (e.g., *Turcotte and Schubert [1982]*), though the actual value is uncertain. Like the elastic stiffness, the melt pressure required to form tensile fractures in gabbroic rocks, particularly partially molten gabbroic rocks, is poorly known. We will use $5 \cdot 10^7$ Pa (50 MPa), which is the estimated tensile strength of partially molten peridotite peridotites (e.g., *Kelemen et al. [1997a]*; *Nicolas [1986]*,

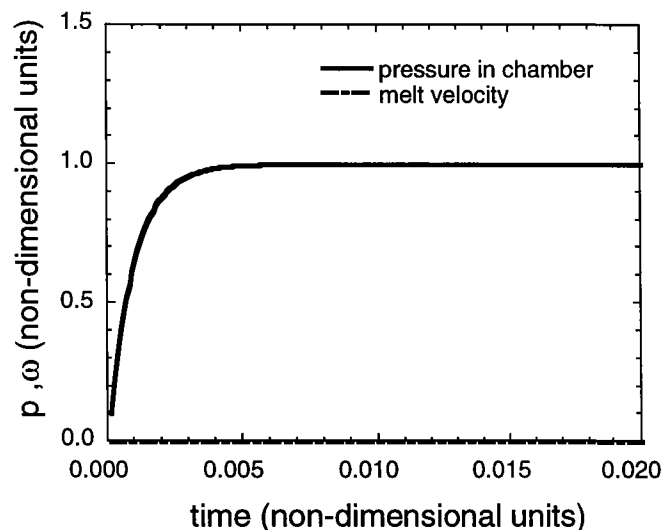


Figure 7. In this Figure, R , the “viscous dissipation coefficient”, is ten times greater than in Figure 5. Otherwise, all parameters are as in Figure 5b. In this case, a steady pressure (solid line) is established when influx is compensated by viscous pressure dissipation (due to porous flow out of the melt lens and/or viscous expansion of the lens). This pressure is less than required to fracture the overlying rock, so there is no melt flow in fractures.

1990]), assuming that externally imposed tensile stresses are negligible.

With these approximations, melt-filled fractures originating at the MTZ may be estimated to occur once every three years. If viscous dissipation of melt pressure were included in the estimation, the frequency of fracture formation would be smaller. Also note that this calculation only includes the time to reduce melt flux to the critical value for "healing". Since the healing process itself may take considerable time, the recurrence intervals estimated using equation (17) should be considered minimal.

A similar estimate can be made for the "shallow melt lens" seismically imaged near the dike/gabbro transition along the East Pacific Rise (e.g., *Harding et al.* [1989]; *Sinton and Detrick* [1992]). If igneous accretion is assumed to take place continuously over the full height of the crustal section, at this level in the crust the melt flux would be about one third of the flux through the Moho, cross-sectional magma chamber width may be $\sim 10^3$ m, G may be closer to 10^{11} Pa [*Turcotte and Schubert* 1982], and melt lens volume may be $\sim 10^4$. These values give a frequency of about one fracture per 2.5 years, similar to the rate inferred from the width of sheeted dikes (~ 1 m wide dikes at a spreading rate of 0.1 m/yr yields 1 dike/10 years).

Finally, we can estimate the volume of melt expelled during each cycle, Q_{tot} , by noting that in each cycle the flux into the lens is equal to the flux out of the lens (neglecting crystallization), since at the end of each cycle the lens returns to the same deflated state. Thus, the total melt flux expelled during each diking event will be approximately

$$V_{tot} = Q_0 \Delta t_r = \frac{l^2}{G} p_{max}. \quad (18)$$

Individual melt removal events from MTZ sills might thus have cross-sectional volumes ~ 50 m². This is of the same order as the cross-sectional volume estimated for melt lenses near the MTZ. Thus, individual fracture events may completely empty melt lenses near the base of the crust. Eruptions generated from the shallow melt lens would have cross-sectional volumes of ~ 500 m², much smaller than the shallow melt lens volume. This is consistent with the hypothesis that a shallow melt lens is present at steady-state beneath fast-spreading ridges.

7. MODAL LAYERING FORMED BY PERIODIC PRESSURE CHANGE

In this section of the paper, we present quantitative models showing how modal banding may be formed by small changes in magmatic pressure. Several different techniques can be used to calculate "liquid lines of descent" as a result of changing pressure and/or temperature in natural silicate liquid systems (e.g., *Weaver and Langmuir* [1991]; *Ghiorso and Sack* [1995]; *Yang et al.* [1996]). Each

of these techniques its own special strengths and weaknesses, but for our purposes the results are very similar; the differences between these models not important in this paper. After testing several of them (Figure 1), we have chosen to present results only from the MELTS model [*Ghiorso and Sack* 1995] which was preferable for mainly because of flexibility in specifying input parameters.

7.1. Modally Graded Layering in Gabbro Lenses

In section 7.1, proposed primitive, mantle-derived liquids parental to MORB are used in modeling. The initial temperature is the temperature of olivine saturation. It is assumed that pressure changes due to inflation of magmatic lenses and melt-filled fracture are rapid compared to the rate of cooling, so the temperature is held constant. The pressure change used in modeling is 50 Mpa. This is an estimate for tensile strength of partially molten peridotites (e.g., *Kelemen et al.* [1997a]; *Nicolas* [1986, 1990]) and - by analogy, in the absence of other estimates - we use it here for partially molten gabbros. The magnitude of the pressure change used in the model is quite arbitrary. If the initial, olivine-saturated liquid were slightly cooler, and thereby closer to plagioclase saturation at the initial pressure, then a much smaller pressure change could be used in modeling, with almost exactly the same quantitative result.

Since the liquidus for anhydrous silicate liquids has a positive P/T slope, the effect of increasing pressure at constant temperature is to cause crystallization. In this case, increasing pressure leads to olivine crystallization, followed by saturation in olivine + plagioclase (Figure 8). The predicted crystallization of plagioclase due to increasing pressure at constant temperature may seem counterintuitive to some petrologists, since increasing pressure decreases the stability of plagioclase relative to olivine. In considering only the liquidus surface for primitive basalts, it is true that increasing pressure moves the plagioclase saturation point further from the composition of mantle-derived liquids. However, it must be kept in mind that, because of the positive P/T slope of the liquidus surface, the plagioclase saturation point at a relatively high pressure is at a higher temperature than the plagioclase saturation point at a lower pressure. In the course of increasing pressure at constant temperature, mantle-derived liquids are driven below their liquidus surface and onto multiply saturated cotectics, as illustrated schematically in Figure 9. Thus, increasing pressure drives crystallization and changes the saturated phases from olivine only, to olivine + plagioclase.

One graded layer could be produced by fracture and rapid decompression, followed by fracture healing, gradual pressure increase, and crystallization. During a second isothermal decompression/compression cycle, any melt from the first cycle remaining in the melt lens would be above its liquidus and no crystallization would occur. Slow cooling, and/or influx of additional, olivine-saturated liquid, is required to return to olivine crystallization.

If crystallization occurs at the base of a magma lens, pressure increase is gradual, and pressure decrease is sudden, this process can give rise to "normal", modally graded layers, with only olivine at the base, grading into gabbro at the top, and then a sharp upper contact with the next olivine layer. "Reversely" graded layers could form simultaneously at the top of a magma lens, but might be expected to be less common since olivine will settle in basaltic liquids at MTZ pressures. Both normally and reversely graded layers are observed within gabbroic sills in the Oman MTZ, and within lower crustal layered gabbros in Oman, but normally graded layers are much more common (e.g., *Pallister and Hopson [1981]*).

7.2. Phase Layering in Lower Crustal Gabbros

In addition to layers grading from dunite to gabbro, a variety of other types of modal layering are observed in gabbroic sills and lower crustal gabbros in Oman. Particularly obvious in outcrop are pure plagioclase ("anorthosite") bands, from a few cm to > 50 cm in thickness. These layers are difficult to understand as the result of crystal fractionation during cooling of a mantle-derived basalt, since in liquids produced by such a process, plag only crystallizes along cotectics with olivine and/or cpx. However, anorthosite layers are a likely result of crystallization along an accreting surface, combined with rapid decompression of cotectic liquids followed by cooling. Because the plag primary phase volume increases in size with decreasing pressure, liquids on plag - olivine and/or plag - cpx cotectics at a high pressure will be in the plag-only stability field at a lower pressure, as illustrated in Figures 9 and 10. This process could produce anorthosite layers grading into gabbroic rocks within a magma lens, or sills composed only of plagioclase, either by subsequent removal of liquid from the crystallizing lens before it returns to olivine and/or cpx saturation, or by injection of plag-saturated liquid into previously crystallized gabbro, crystallization of an anorthosite sill, and extraction of residual liquid.

More subtle types of modal layering could also be produced by periodic pressure changes, changing the proportions of minerals crystallizing from the same liquid where cotectics are curved in PT space. In offering these examples, we do not mean to suggest that this is an exhaustive review of the possible processes in layer formation due to pressure change. On the contrary, we simply wish to provide a few examples from the rich inventory of possible processes that could form igneous modal layering in a regime of periodically changing pressure. However, where magmatic temperatures are so low that liquids are well within the domain of plag + cpx + opx + oxide saturation, formation of obvious modal layering as a result of periodic pressure change is relatively unlikely. This may help to account for the paucity of modal layering observed in "upper gabbros" near the base of the sheeted dikes in the Oman ophiolite (e.g., *Nicolas et al. [1988]*) and in oceanic

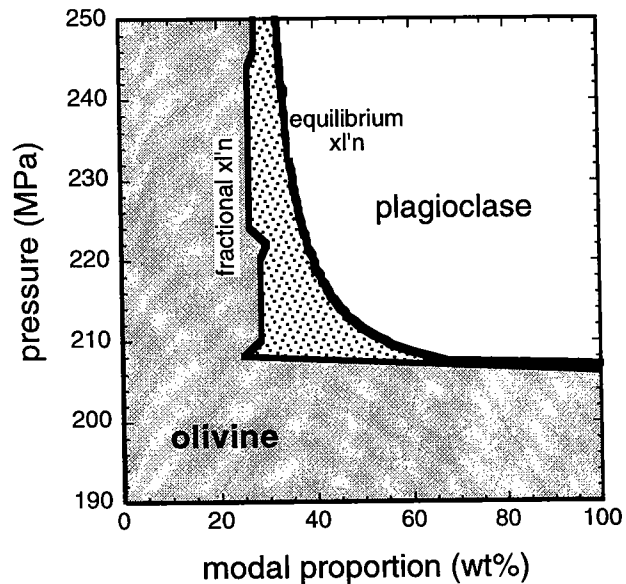


Figure 8. Results of a quantitative model for formation of an igneous layer with an olivine-rich base and a plagioclase-rich top, in terms of the modal proportion of minerals produced over a given increment of pressure increase. Crystallization is assumed to take place on the floor of a melt lens. The initial liquid composition is the calculated product of fractional crystallization of parental MORB liquid kg2 [*Kinzler and Grove 1991*] at 200 MPa from 1300 to 1244°C. At 200 MPa and 1244°C, this liquid composition is just above the temperature for plagioclase saturation. With increasing pressure, plagioclase saturation is reached. Schematic phase relations for this process are illustrated in Figure 9. Fractional crystallization produces a discontinuous change in the proportions of minerals crystallized, whereas equilibrium crystallization gradually changes the proportion of minerals in all of the products of crystallization. Production of a graded layer probably requires a process intermediate between perfect fractional and perfect equilibrium crystallization. All MELTS calculations are for H₂O-free liquids along the Fayalite-Magnetite-quartz oxygen fugacity buffer (FMQ).

crust formed at the East Pacific Rise (e.g., *Gillis, Mével, Allan et al. [1993]*).

8. DISCUSSION

8.1. More Layers Than Dikes: Several Melt Lenses Within the Crust?

One might ask if each modal layer records the formation of one sheeted dike? Apparently not. If individual modal layers average about 100 meters in horizontal extent and 0.1 m in height (e.g., *Pallister and Hopson [1981]*; *Smewing [1981]*; *Browning [1982, 1984]*; *Benn et al. [1988]*; *Boudier et al. [1996b]*), over a vertical extent of layered

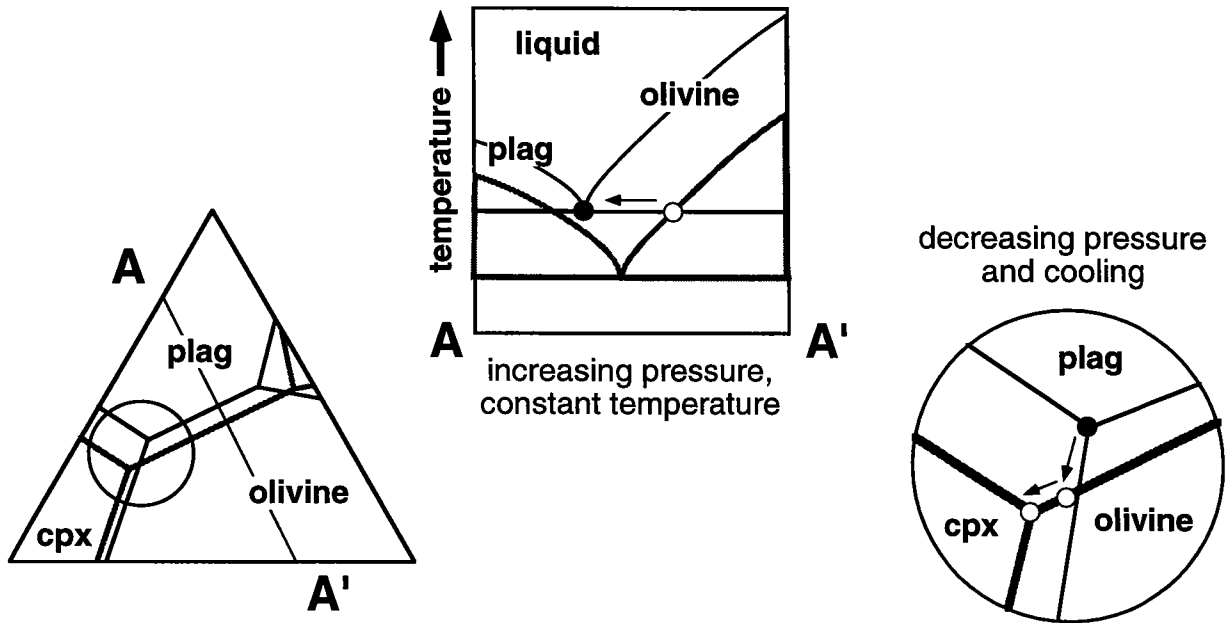


Figure 9. Schematic illustration of the system Forsterite (olivine) - Diopside (cpx) - Anorthite (plag), with cotectics at two different pressures in the range of 100 to 300 MPa, based on phase equilibrium data summarized by *Morse [1980]*. Solid lines and filled symbols indicate phase relations at relatively high pressure; grey lines and open symbols are for relatively low pressure. The temperature/composition section A-A' elucidates the process quantitatively illustrated in Figure 8. Increasing pressure at constant temperature leads to olivine crystallization, which in turn drives the liquid to the olivine-plagioclase cotectic. Crystallization on the floor of a melt lens as a result of this type of process may form graded layers with olivine-rich bases and plagioclase-rich tops. The lower inset shows phase relations for the process quantitatively illustrated in Figure 10. Cooling liquid saturated in olivine, plagioclase and cpx undergoes rapid decompression. Initially, at lower pressure the liquid is not saturated in any solid phase. Because decreasing pressure increases the size of the plagioclase primary phase volume relative to olivine and cpx, after cooling liquid at low pressure the liquid initially saturates in plagioclase. Furthermore, because decreasing pressure has also expanded the olivine primary phase volume relative to cpx, the second phase to crystallize is olivine, followed by cpx after continued cooling. This type of process may form pure plagioclase layers ("anorthosite") and olivine + plagioclase layers ("troctolite") within layered gabbro intrusions.

gabbro of the order of 3 km, there will be $\sim 10^4$ layers in a 100 m wide section of crust perpendicular to the spreading direction. In the same section, if sheeted dikes average about 1 m in width (e.g., *Pallister [1981]*), there will be only 10^2 dikes. If modal layers had an average horizontal extent of 1 km, there would be 10^4 layers and 10^3 dikes in section 1 km wide. Only if layers were more than 1 m thick and 1 km long, or if dikes were less than 0.1 m wide, could the number of modal layers be equal to the number of dikes. While the average dimensions of modal layers are uncertain, an average of 1 meter thick by 1 km long is far too large.

Therefore, if each layer within gabbros uniquely records a melt-filled fracture event, there must be ~ 10 to 100 more fracture events in the lower crust than there are sheeted dikes. This result is in accord with the estimates for fracture frequency derived using equation (17) in section 6. Such a result implies that melt-filled fractures do not all reach the level of sheeted dike formation; many may terminate at

overlying magma lenses or within other mechanical boundaries within the crystallizing crust. Thus, our preferred model for the accretion of oceanic layered gabbros at fast spreading ridges is that there are several magma lenses extant at any given time, at levels ranging from the MTZ to the base of the sheeted dikes. The lenses at the MTZ may be fed primarily by continuous porous flow, but the lenses higher in the crust are likely to be fed mainly by injection from melt-filled fractures originating from other lenses at greater depth.

The notion that there are several different depths at which gabbros are formed at mid-ocean ridges is supported by data from the Oman ophiolite that show compositional differences between "upper gabbros", "lower gabbros", and gabbroic sills in the MTZ (e.g., *Pallister and Hopson [1981]*; *Browning [1982]*; *Juteau et al. [1988]*; *Kelemen et al. [1997]*; *Korenaga and Kelemen [1997]*). In general, Mg/(Mg+Fe) and Ca/(Ca+Na) is higher in MTZ sills than

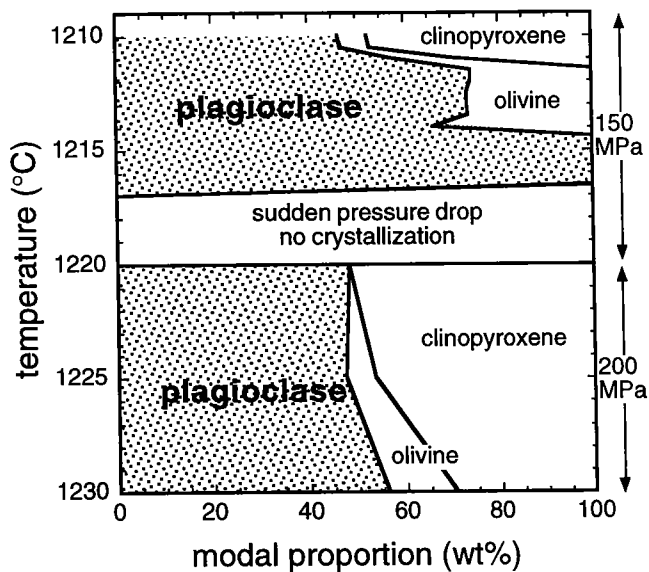


Figure 10. Results of a quantitative model for formation of an igneous anorthosite layer within gabbros. The initial liquid composition is the calculated product of fractional crystallization of parental MORB liquid kg2 [Kinzler and Grove 1991] at 200 MPa from 1300 to 1220°C. At 200 MPa, from 1230 to 1220°C, the liquid is saturated in olivine, plag and cpx. Rapid decompression to 150 MPa at 1220°C, followed by continued cooling leads to an interval of no crystallization, followed by plagioclase-in at 1216.5°C, olivine-in at 1214.5°C, and cpx-in at 1211.5°C. All MELTS calculations are for H₂O-free liquids along FMQ.

in lower crustal, layered gabbros within 2 km above the MTZ. These layered gabbros, in turn, have generally higher values than in upper gabbros, within 1 km of the base of the sheeted dikes. These data are consistent with a vertical zonation of temperature in the oceanic crust, with liquids that crystallize at higher levels attaining a larger degree of compositional fractionation as compared to primitive magmas entering the MTZ.

8.2. Crystallization Temperatures for Gabbros and Thermal Structure Beneath Ridges

Because many Oman gabbros are cumulates, i.e., products of small amounts of crystallization from a larger melt mass, after which the residual liquid was efficiently extracted, their calculated melting behavior is different from that of a basaltic liquid. Specifically, because they retain nearly cotectic proportions of olivine, cpx and plag, with nearly constant composition, they resemble eutectic melt compositions in some respects. Thus, as illustrated in Figure 3, the calculated % liquid in gabbro compositions changes dramatically from about 10% to >50% over a very narrow temperature interval. This crystallization behavior

can be used to estimate the temperature at which the cumulates formed, as already discussed in section 3.

Here, we consider the temperature/depth information in more detail. The average composition of 26 samples of sills from the MTZ [Korenaga and Kelemen 1997] gives has calculated liquid fraction of 10% at a temperature of ~1240°C. We compare this to temperatures for average crustal rock compositions compiled by Browning [1982]. Olivine gabbros (number of samples, n=6), typical of the lower, layered gabbro section, give a temperature of ~1225°C at 10% liquid. Pyroxene gabbros (n=3), with little or no olivine, are also found in the layered gabbro section. Their average composition gives ~1205°C at 10% liquid.

Browning's average for non-layered, upper gabbros (n=12) closely resembles a liquid composition (e.g., Kelemen et al. [1997b]). Such upper gabbros crystallized at lower temperatures than any of the layered gabbros. The calculated temperature at 10% liquid is ~1165°C. As noted in previous sections of this paper, seismic evidence indicates that there may be a "shallow melt lens", with more than 10% liquid, near the dike/gabbro transition directly beneath the ridge axis at some fast-spreading ridges (e.g., Harding et al. [1989]; Sinton and Detrick [1992]). Since many of these melt lenses support shear waves, Hussenoeder et al. [1996] estimate that they generally contain <30% liquid. The calculated temperature for the upper gabbro composition of Browning [1982] at 30% liquid is ~1195°C.

It should be noted that the calculated temperature values are dependent on the choice of modeling technique, pressure, oxygen fugacity, and so on. However, the temperature difference from one rock to another is more robust and better constrained by these calculations than the absolute temperatures.

Another important issue is whether all of the gabbroic rocks in Oman are coeval, or whether some formed well off-axis, so that the apparent vertical gradient in crystallization temperature is an artifact of some kind. There is increasing evidence that magmatism in the northern part of the ophiolite may have continued for more than 1 million years (e.g., Perrin et al. [1994]). This protracted magmatism is temporally related to large tectonic rotations of the northern part of the ophiolite relative to a hotspot reference frame, and relative to the southern part of the ophiolite. Unfortunately, lavas in the southern part of the ophiolite are not well exposed and there are few geochemical data for them. However, the chemical differences that distinguish the early and later volcanics ought to be discernible as two distinct compositional types in gabbros as well. In fact, this is the case in the northern part of the ophiolite (e.g., Smewing [1981]; Juteau et al. [1988]). However, geochemical data for gabbros from the southern ophiolite massifs, where we have concentrated our research, suggest that >90% of the igneous rocks form a single liquid line of descent. [Pallister and Hopson 1981; Pallister and Knight 1981; Browning 1982; Benoit et al. 1996; Kelemen et al. 1997b; Korenaga

and Kelemen 1997]. Thus, we infer that most or all of the gabbros in the southern massifs form a single suite crystallized beneath a spreading center.

One additional point that should be considered is that, even if all of the Oman gabbros are coeval and formed beneath a spreading ridge, their igneous temperatures do not necessarily reflect temperatures along a steady-state geotherm beneath that ridge. This is most true for upper parts of the crustal section, where wall rocks would have been relatively cold, and magma transport was dominated by brittle intrusion of dikes and sills. Thus, in the upper crust directly beneath a ridge, there probably is no steady state geotherm, and instead every depth may be characterized by a variety of temperatures over time. In contrast, for the MTZ and lower gabbros, magmatic temperatures and wall rock temperatures were probably quite similar. For example, *Korenaga and Kelemen [1997]* calculated thermal Peclet numbers for diffuse, porous, intergranular melt flow near the Moho at fast-spreading ridges, and concluded that liquid and solid would maintain the same temperature during such a process.

In summary, we believe that vertical reconstructions of composition and temperature based on average data for many samples from Oman are likely to be representative of a coeval suite of rocks formed beneath a spreading ridge. Thus, the data from the Oman ophiolite are consistent with a lower crust below the ridge axis that had systematic vertical variation in both composition and temperature, with a temperature gradient of ~ 45 to 75°C from the base of the crust to the "shallow melt lens" below the base of the sheeted dikes.

8.3. Igneous Accretion of the Lower Oceanic Crust

On the basis of field data from Oman and seismic data from the East Pacific Rise, it has been proposed that the lower oceanic crust is formed by partial crystallization of gabbros in a single, shallow melt lens near the dike/gabbro transition, followed by ductile flow downward and outward from this melt lens [*Nicolas et al. 1988, 1993; Quick and Denlinger 1993; Phipps Morgan and Chen 1993; Henstock et al. 1993*], as originally proposed on theoretical grounds by *Sleep [1975]* and *Dewey and Kidd [1977]*. We shall refer to these as "conveyor belt" models. Because the gabbros do not preserve structures formed by crystal plastic deformation, it was proposed deformation occurred by intracrystalline glide coupled with pressure solution, in partially molten gabbros with porosities greater than 10% (e.g., *Nicolas [1992]; Nicolas and Ildefonse [1996]*).

Boudier et al. [1996] and *Kelemen et al. [1997b]* have questioned this end-member hypothesis on the grounds that some structures in the lower, layered gabbros such as modally graded layering could not survive the large shear strains required by "conveyor belt" models. In addition, *Kelemen et al. [1997b]* showed that the lower, layered

gabbros were compositionally nearly identical to gabbroic sills in the MTZ; thus, it is possible to form layered gabbroic cumulates in sills near the Moho. Furthermore, *Kelemen et al.* showed that lower crustal gabbros are compositionally different from upper gabbros. Thus, both groups proposed emplacement of gabbroic sills near the Moho and within the lower crust, as well as just below the dike/gabbro transition. Figure 11 shows these models.

Schouten and Denham [1995], reacting to data from the East Pacific Rise and the Oman ophiolite, produced a model with two "conveyor belts" originating from two sills, one at the Moho and the other near the dike/gabbro transition. Our illustration of this hypothesis, Figure 11a, is thermally quite an extreme case, because it requires crystallization of about 50% of the crustal mass at the base of the crust, where heat removal by conduction and hydrothermal convection may not be efficient, especially beneath a steady-state melt lens at a fast-spreading ridge.

Boudier et al. [1996] and *Quick and Denlinger [1993]* proposed that lower crustal sills were emplaced into a gabbro "conveyor belt" originating near the dike/gabbro transition. *Kelemen et al. [1997b]* proposed an alternative hypothesis in which the entire lower crust is constructed of "sheeted sills", each emplaced at its current depth in the crustal section, without vertical transport by ductile flow.

Whereas *Kelemen et al. [1997b]* proposed the "sheeted sill" model as one of several, equally plausible alternatives, we now tentatively prefer it, for several reasons.

(1) It seems unlikely on thermal grounds that 50% of the crust crystallizes along the Moho, as in Figure 11a.

(2) As noted in an earlier section of this paper, *Korenaga and Kelemen [1998]* argued that correlation between vertical variation of mineral compositions, coupled with the lack of a systematic, vertical compositional trend, precludes large amounts of melt transport by porous flow through lower crustal gabbros in Oman. If the main mechanism of melt transport through the lower crust were by diffuse porous flow, then the observed correlations, for example between forsterite content in olivine and anorthite content in plagioclase, would have been obliterated by solid/liquid exchange reactions and diffusion.

Therefore, although small amounts of diffuse porous flow must have played a role in melt extraction from cumulate gabbros, the main mechanism of melt transport through the lower crust must have been in zones of focused flow, such as melt-filled fractures. This seems to rule out the presence of a large (ca. 10%), interconnected melt porosity throughout the lower oceanic crust, such as is required in the "conveyor belt" hypotheses to produce low bulk viscosities. It remains possible that formation of vertical chemical variation postdates "conveyor belt" deformation.

(3) Analysis in sections 3 and 8.2 suggests that there was a gradient of 45 to 75°C from the dike/gabbro transition to the MTZ beneath the ridge axis during formation of the Oman ophiolite, as in Figure 11. As shown by *Phipps*

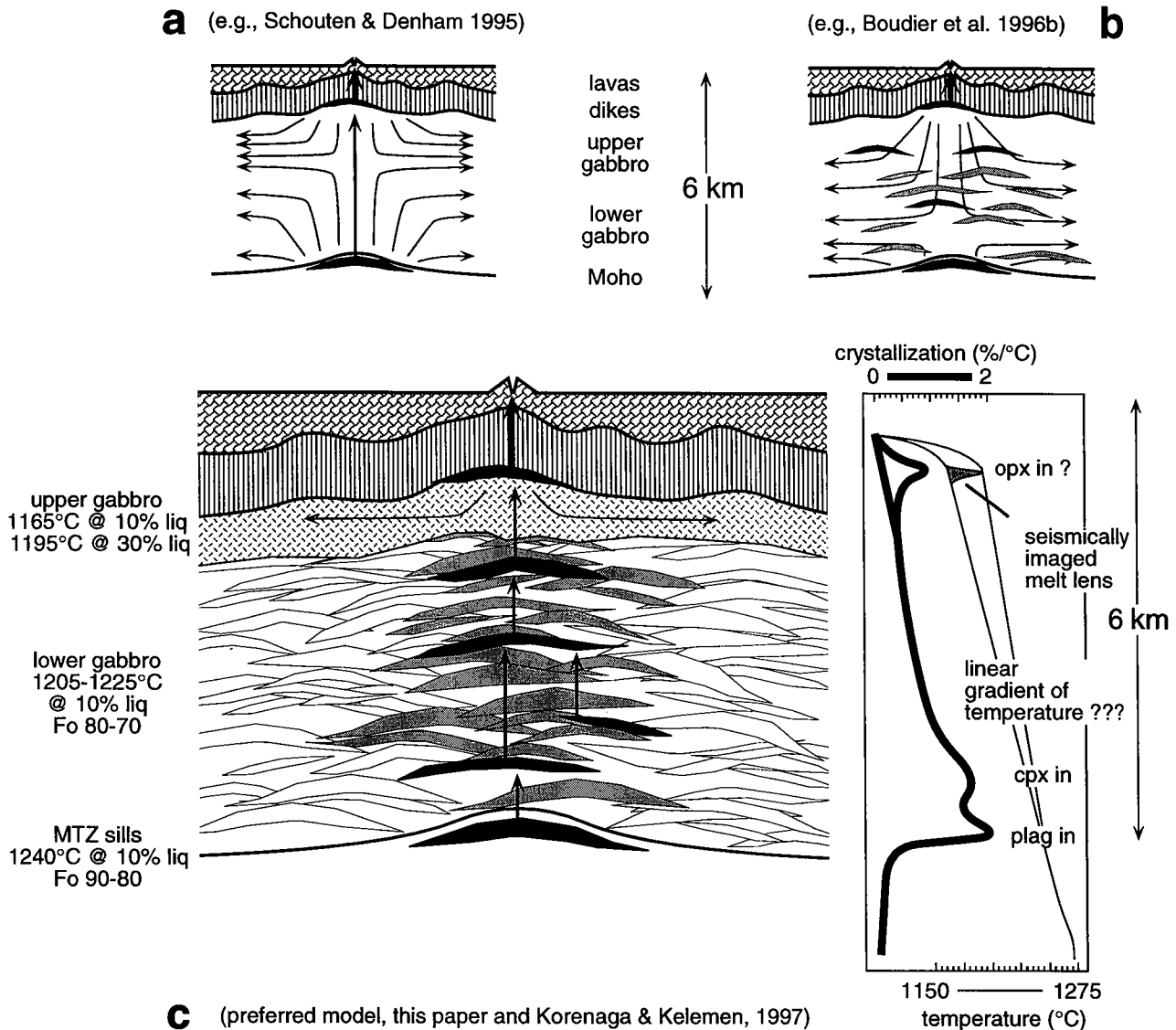


Figure 11. Recent models for the formation of the oceanic lower crust, based on seismic observations from the East Pacific Rise and field constraints from the Oman ophiolite. (a) is redrawn from *Schouten and Denham [1995]*, who illustrated a scenario in which gabbros crystallize mainly in two melt lenses, one in the MTZ and one near the dike/gabbro transition. (b) is redrawn from *Boudier et al. [1996]*. They proposed that most gabbros crystallize in a melt lens near the dike/gabbro transition, but as these move downward and away from the melt lens by melt-lubricated ductile deformation they are intruded by lower crustal, gabbroic sills. A similar scenario was suggested by *Quick and Denlinger [1993]*. (c) is redrawn from *Kelemen et al. [1997b]*, who proposed an end-member scenario in which there is no vertical ductile transport of gabbro in the oceanic lower crust, which is instead constructed of sheeted sills, emplaced at approximately their final depth below the sea-floor. Similar proposals have also been made by J. Bédard and co-workers (e.g., *Bédard et al. [1988]*; *Bédard [1991]*). We tentatively conclude that (c) is most consistent with observations from Oman, including the thermal gradient we infer from crystallization calculations for gabbro compositions (Figure 3). On the lower right, we schematically illustrate the inferred thermal gradient and vertical distribution of crystallization rates in the crust forming beneath a spreading ridge axis. A linear temperature gradient is adopted for simplicity. Along such a gradient, there could be three local maxima in the crystallization rate, corresponding to the appearance of plag, cpx and opx on the liquidus of cooling, crystallizing basaltic magma. Each of these might form a permeability barrier, with melt lenses ponding beneath it. Formation of melt lenses below these barriers, and elsewhere in the lower crust, followed by replenishment of these lenses with relatively hot magma from below, could dramatically perturb the linear thermal gradient illustrated here. This, in turn, would give rise to still more variations in crystallization rate and permeability in a vertical section below the axis of an active spreading ridge.

Morgan and Chen [1993], "conveyor belt" processes may result in small, nearly adiabatic temperature gradients in the lower crust beneath a ridge axis.

(4) In section 8.1, we suggested that - if both modal layers in gabbros and sheeted dikes record melt fracture events - the fact that the number of layers is much larger than the number of dikes suggests that there are many melt lenses at different depths within the crust at any given time.

In light of these observations and inferences, we suggest that the "sheeted sill" model accounts better for the composition of lower crustal gabbros in the Oman ophiolite than any "conveyor belt" model proposed to date.

9. CONCLUSION

Periodic formation of melt-filled fractures can result from magma accumulation beneath a permeability barrier in a visco-elastic solid medium. Periodic pressure change, caused by continuous influx and periodic extraction of liquids, can potentially explain a variety of modal layering phenomena in gabbroic rocks, not only in the oceanic crust but perhaps also in some layered intrusions. This hypothesis seems particularly likely to apply to the generation of the oceanic crust, where sheeted dikes preserve a clear record of periodic melt-filled fracture and, at least in Oman, have the compositional characteristics of liquids extracted from lower crustal, layered gabbros. Thus, the two most evident features indicative of periodic crystallization in igneous crust formed at oceanic spreading centers, modal layering in lower crustal gabbros and sheeted dikes, may be complementary features formed by the same physical process.

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