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.

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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 McBirney* [*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 my 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 (± 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 [*Garmany 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 (± 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) ± clinopyroxene (cpx), moving the Moho upward, until the liquids themselves became saturated in plag ± cpx. Conversely, if melts became saturated with plag ± 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 ± cpx, their mass drops rapidly with continued cooling. By comparison, olivine ± spinel fractionation at higher temperatures consumes relatively little liquid mass per °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%/°C are predicted over a 20 to 40°C interval below the temperature of olivine + plag (± 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 (± cpx) saturation, in % liquid/°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

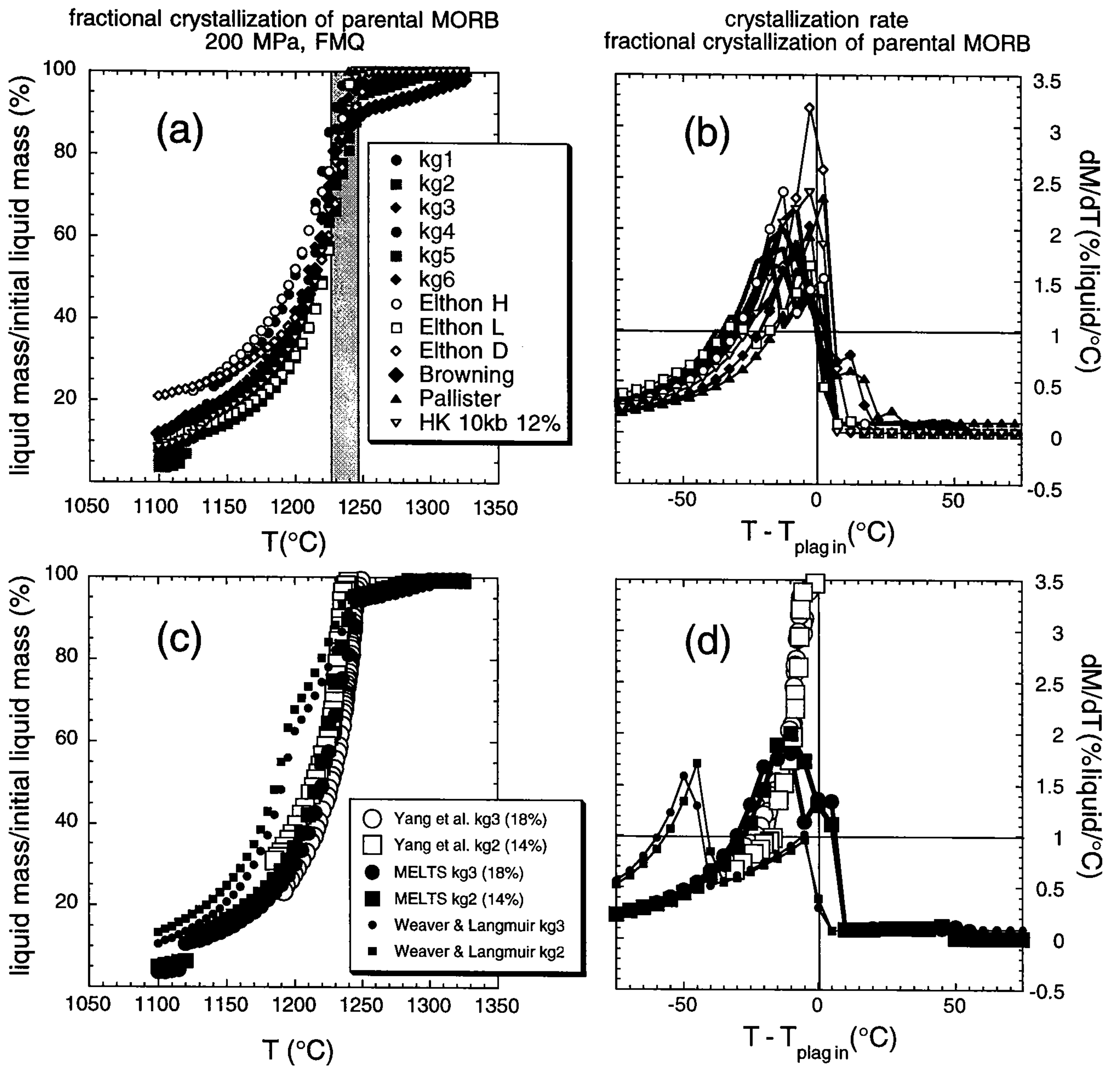

Figure 1. Modeling results for fractional crystallization of proposed, parental mid-ocean ridge basalt (MORB) liquids, H_2O-free at 200 MPa along the Fayalite-Magnetite-Quartz oxygen fugacity buffer. The main result is that the crystallization rate, in % crystallized/°C (labelled dM/dT, %/°C), increases 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.

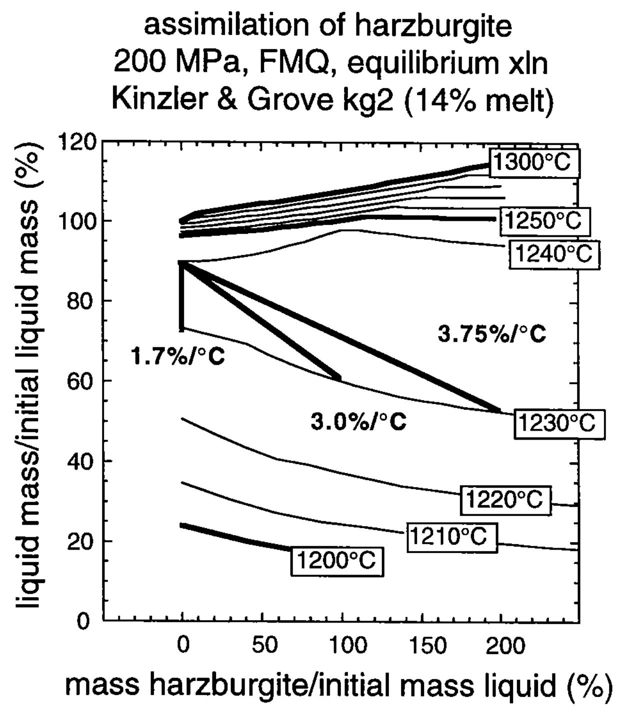

Figure 2. Modeling results for equilibrium crystallization of parental MORB liquid composition kg2 plus varying proportions depleted mantle harzburgite, calculated using the MELTS model, H2O-free at 200 MPa along the Fayalite-Magnetite-Quartz oxygen fugacity buffer (FMQ). The kg2 liquid composition is 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 (± 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 ~ 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

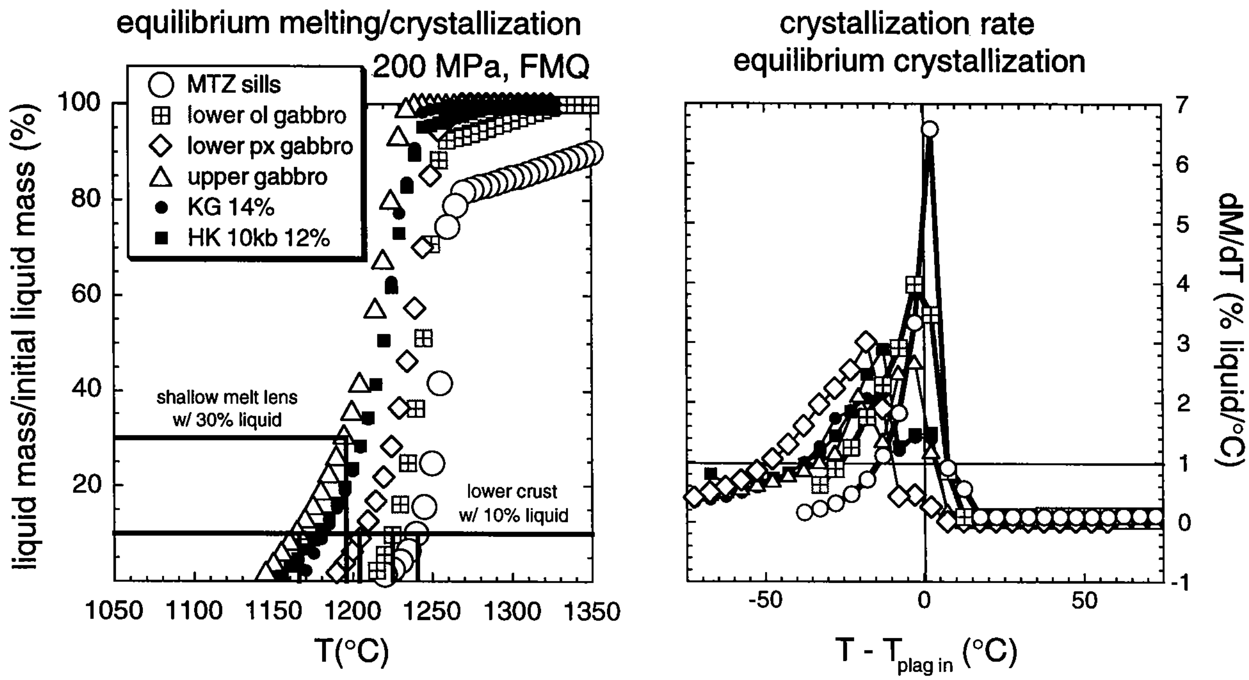

Figure 3. Modeling results for equilibrium crystallization of gabbro compositions from the Oman ophiolite, calculated using the MELTS model, H_2O-free at 200 MPa along FMQ. The composition for sills from the Moho Transition Zone (MTZ) is an average of 26 samples from *Korenaga and Kelemen* [*1997*]. 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 ~ 0.01°C/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, ~ 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°C/m, and rates of crystallization of multiply saturated basalt of 1% to 3%/°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 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*],

$$V_e \propto \frac{l^2}{G} p, \tag{1}$$

where l is the horizontal length of the lens, G is the elastic stiffness of the host rock, and p = P-ρgh 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

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

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

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

The deviation of the horizontal stress from lithostatic stress is $\Delta\sigma_H = \rho g h - \sigma_H$, 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

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

where ω is melt velocity, μ is the melt viscosity, Δρg 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

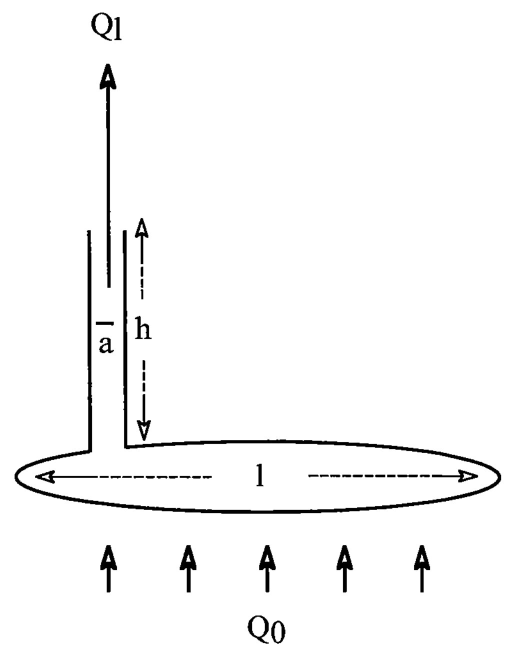

Figure 4. Schematic illustration of a melt lens below a low permeability layer, with an open crack allowing for melt flux out the top of the lens.

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

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

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, $\kappa = 10^{-6} m^2/s$, $c = 730$ J/kg°C, $T_0 = 1230°C$, $1230°C > T_\infty > 1220°C$, 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

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

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 $p \geq p_{max}$

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

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

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

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

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

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),

$$\omega_c a_c^2 \sim \frac{\kappa h}{12}. \qquad (8)$$

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: $p^* = \Delta\rho g h$, $\omega^* = a^{*2} p^*/(12\mu h)$, $t^* = h/\omega^*$, $a^* = hp^*/G$. Finally we define a characteristic Peclet number, $Pe^* = p^* a^{*4}/(\kappa h^2 \mu)$. The non-dimensional equations are then

$$\gamma \dot{p}' = Q'_0 - \delta\omega' a' - Rp'; \qquad (9)$$

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

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

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

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

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

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

where

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