Modelling the seismic properties of fast-spreading ridge crustal Low-Velocity Zones: insights from Oman gabbro textures

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Abstract

Although considerable progress has been made in the study of fast-spreading, mid-ocean ridge magma chambers over the past fifteen years, the fraction of melt present in the chamber remains poorly constrained and controversial. We present new constraints obtained by modelling the seismic properties of partially molten gabbros at the ridge axis. P-wave velocities at low frequencies are calculated in the foliation/lineation reference frame using a differential effective medium technique. The model takes into account the lattice preferred orientation of the crystalline phase and the average shape of the melt phase. The structural parameters are obtained from the Oman ophiolite. The structural reference frame is given by the general trend of the gabbro foliation and the melt fraction and shape are estimated using the textures of nine upper gabbro samples. The estimated melt fraction and shape depend on the assumptions regarding which part of the observed textures represent the melt in the gabbroic mush of the magma chamber. However, we can put limits on the reasonable values for the melt fraction and shape. Our results are consistent with a melt fraction of the order of 10 to 20% in the Low-Velocity Zone (i.e. the magma chamber), which is anisotropically distributed with the melt pockets preferentially aligned parallel to the foliation and approximated by oblate ellipsoids with approximate dimensions of 4:4:1. These results are also consistent with the seismic structure of the East Pacific rise at 9°30'N. The anisotropic melt distribution can, at least partially, explain the vertical velocity gradient described in the LVZ. © 1999 Elsevier Science B.V. All rights reserved.

Keywords: mid-ocean ridges; magma chambers; gabbro; anisotropy; seismic properties; Oman ophiolite

1. Introduction

During the last fifteen years, our understanding of fast-spreading oceanic ridges has improved considerably, due to (1) the accumulation of geophysical data at the fast-spreading East Pacific Rise (EPR) (e.g. Hale et al., 1982; Vera et al., 1990; Solomon and Toomey, 1992; Caress et al., 1992; Sinton and Detrick, 1992; Evans et al., 1994; Crawford et al., 1999), and (2) the progress of the work in ophiolites inferred to have been derived from a fast-spreading environment such as the Oman ophiolite (e.g. Nicolas et al., 1988; Nicolas and Boudier, 1995; Boudier et al., 1997). However, these two approaches lead to some discrepancies in the ridge/magma chamber models. One of the differences relates to the amount of melt present in the crystallising mush at the ridge axis, and consequently to the size of the magma chamber. We provide herein some new constraints,
obtained by modelling the seismic properties of the partially molten gabbros at the ridge axis using constraints derived from Oman gabbro microstructures, and by comparing our results to the seismic data at the EPR. The microstructure of upper gabbros is used to estimate the shape distribution and amount of the liquid phase in the magma chamber, and to constrain how these parameters influence seismic properties. From these data, we calculate the seismic velocities, using the tensorial method of Mainprice (1997).

2. What is the magma chamber?

Seismic data and tomography at the EPR image a magma lens at the base of the sheeted dike complex (approximately 1–4 km wide, 50–200 m thick), overlying a triangular Low-Velocity Zone (LVZ) approximately 6–8 km wide and 2–4 km thick (Vera et al., 1990; Hale et al., 1982; Detrick et al., 1987; Solomon and Toomey, 1992; Wilcock et al., 1992; Barth and Mutter, 1996). From the seismic point of view, the magma chamber is usually restricted to the magma lens (e.g. Caress et al., 1992) or to the magma lens and a narrow mush zone immediately below (e.g. Wilcock et al., 1992). The LVZ is interpreted as a zone that contains a hot, mostly solid material, with at most 5% of melt (e.g. Caress et al., 1992; Wilcock et al., 1992, 1995). This estimate is based on the comparison between the EPR seismic data and published experimental data on ultra-sonic (high-frequency) velocities in partially molten basalts (e.g. Murase and McBirney, 1973; Manghani et al., 1986) and peridotites (e.g. Sato et al., 1989), or seismic (low-frequency) attenuation in gabbros at sub-solidus temperatures (Kampfmann and Berckhemer, 1985).

On the other hand, the Oman ophiolite gabbros, which are inferred to be produced at an ancient fast-spreading centre (e.g. Nicolas et al., 1994; Nicolas and Boudier, 1995; Boudier et al., 1997), display dominantly magmatic microstructures (e.g. Nicolas et al., 1988), suggesting that the amount of melt present in the system (i.e., the LVZ) was high enough to allow the deformation of the crystallising gabbros to be controlled by the presence of melt (Nicolas and Ildefonse, 1996; Ildefonse et al. in prep.). The threshold from suspension flow to solid-state flow in a magmatic mush is defined classically around 35–40% of melt (e.g. Arzi, 1978; Van der Molen and Paterson, 1979; Fernandez and Gasquet, 1994). On the basis of simple analogue experiments with suspensions of paraffin wax particles in a mixture of oil and alcohol, it has been suggested that this threshold may be as low as 20% if the particles are aligned (shape preferred orientation) by the flow of the suspension (Nicolas, 1992; Nicolas et al., 1993). This is the lowest published estimate for the transition from suspension flow to solid-state flow and possibly the minimum estimate for the amount of melt present in the crystallising mush at the ridge axis. Thus, there is a large apparent discrepancy in the amount of melt (and consequently in the size of the magma chamber) estimated from the seismic constraints at the EPR (< 5%) and from the field constraints in the Oman ophiolite. However, the latter estimate is not well constrained since it has been shown that deformation in the Oman gabbros occurs by grain boundary sliding accommodated by melt-assisted pressure solution (Nicolas and Ildefonse, 1996). For such deformation, there is no clearly defined lower bound on the amount of melt required. In this paper, we aim to put some new constraints on the melt fraction and its seismic signature in partially molten gabbros, taking into account the spatial distribution of melt estimated from textures of Oman gabbros (we use the term ‘texture’ in the classical geological sense, i.e., to describe the geometry of the mineralogical aggregate at the microstructural scale, characterised by the shape, size and distribution of the different crystal phases).

3. Structural reference frame

In order to compare the results of our modelling with the seismic velocity profiles of the EPR (e.g. Vera et al., 1990), we need to define a common reference frame. At the sample scale which is the scale of our modelling, the reference frame is the magmatic foliation and lineation of the gabbros. P-wave velocities will be calculated in three directions (X, Y, and Z, with XY the foliation plane and X the lineation). In a typical lower crustal section from Oman, the gabbro foliation in the lower two thirds
is nearly parallel to the Moho, while upsection the foliation steepens rapidly and progressively becomes subparallel to the overlying sheeted dikes (Nicolas et al., 1996). The lower gabbros are also layered, but the layering tends to disappear upsection where the gabbros are generally homogeneously foliated, or locally nearly isotropic (e.g. Boudier et al., 1996). Since the magmatic foliation develops in the magma chamber, we assume that the structural pattern observed in the Oman gabbros reflects the structural pattern in the present-day LVZs, an assumption that is also supported by numerical modelling (Chenevez et al., 1998). The direction of the ray paths used to obtain tomographic images of the lower crust beneath the EPR axis is approximately horizontal below the magma lens (Wilcock et al., 1993; Wilcock et al., 1995). They are either nearly perpendicular to the upper gabbro foliation (i.e. parallel to Z) or nearly parallel to the lower gabbro foliation and lineation (i.e. parallel to X). The relationships of these various elements are shown in Fig. 1, superimposed to the Vp profile of the EPR at 9°N (Vera et al., 1990).

4. Calculation of the seismic properties from petrofabrics and average shape of the melt phase

The method we use to model the seismic properties of the gabbroic mush is described in detail by Mainprice (1997). It is a tensorial model based on the standard ‘Differential Effective Medium’ technique (e.g. Bruner, 1976) modified using the poro-elastic method of Gassmann (1951) which was extended to elastically anisotropic media (Brown and Korringa, 1975). Seismic velocities are calculated in the chosen reference frame, at the low frequency limit (i.e., the pore fluid is considered to be everywhere connected and the pressure uniform) which corresponds to seismic frequencies, by taking into account the intrinsic elastic constants, density, and lattice preferred orientation of all constitutive crystalline phases, and the compressibility, density, and spatial distribution of the liquid phase. The volume averaged shape of the melt is represented by a triaxial ellipsoid. To a first approximation, the attenuation is considered to result only from the squirt-flow mechanism (e.g. Mavko, 1980; Schmeling, 1985; see discussion in Mainprice,
Fig. 2. Typical textures in the Oman gabbros. (a) Subdoleritic texture (upper gabbros). (b) Tabular texture (upper gabbros). (c) Partly re-equilibrated lower gabbro texture with lobate grain boundaries. (d) Strongly re-equilibrated mosaic texture in an anorthitic layer. The scale bar is 1 mm.
1997). Calculations made with liquid represented by oblate ‘pancake’ ellipsoids of various aspect ratio demonstrate the strong influence of the distribution of the liquid on the seismic response of a melt mush (Mainprice, 1997). For this reason, we attempt in the following section to constrain the shape of the liquid phase from observations of Oman gabbro textures.

5. Estimate of the liquid phase geometry from upper gabbro textures

5.1. Oman gabbro textures

Microscopically, the textures in the upper gabbros are dominantly subdoleritic or tabular, with euhedral to subhedral plagioclase laths (Fig. 2a,b). Down-section, the plagioclase tends to become anhedral, with either lobate or curvilinear grain boundaries, and mosaic-like textures are frequently observed in plagioclase-rich layers (e.g. Boudier et al., 1996) (Fig. 2c,d). The subdoleritic textures are typically encountered at the top of the igneous section, close to the root of the sheeted dike complex; they reflect faster cooling of the mush (i.e., there is not enough time for the crystal grain boundaries to migrate toward a minimum surface energy pattern, the late minerals crystallise in the interstices). Faster cooling of the upper gabbros is also supported by the frequent compositional zoning of the euhedral plagioclases, which is never seen in the layered gabbros (e.g. Pallister and Hopson, 1981). Tabular textures (e.g. Boudier et al., 1996) are also present in the upper gabbros; they are defined by tabular, nearly euhedral plagioclases with average dimensions of 1.0 × 0.2 mm which are often strongly aligned, and clinopyroxene and olivine appearing either as large oikocrysts or small interstitial crystals.

5.2. Image analysis, separation of mineral phases

We assume that the interstitial phases in subdoleritic or tabular textures, when present, represent to a first approximation the melt phase at a given instant; and is thus the crystallised melt phase present in the steady-state magma chamber (the LVZ), at the structural level considered (upper gabbros). A similar approach was used by Bryon et al. (1996) to determine the occlusion of porosity in crystallising granites, and by Jousselin and Mainprice (1999) to estimate the shape of the melt phase in impregnated peridotites of the Moho transition zone in the Oman ophiolite. These interstitial minerals, crystallised from trapped melt, are dominantly clinopyroxene and olivine when present, more or less replaced by secondary minerals (amphibole, serpentine). Thus, the first step in our analysis required us to separate the plagioclases from the other mineralogical phases. We obtained a binary image, with the plagioclase in white, and the EBP (‘everything but plagioclase’) phase in black using three different techniques:

(1) Automatic image analysis of a sample section after oxidation by heating in a furnace at 900°C. Oxidation eventually changes the colours of the minerals (olivine red, clinopyroxene green, plagioclase white, serpentinised olivine yellow, and amphibole brown), which can then be selected easily by a classical image analysis device, if the grains are big enough. We used this technique only once in this work, since its resolution was not sufficient.

(2) Automatic image analysis of a thin section. The natural light image is acquired using a slide scanner, a video camera, or a photo enlarger. This technique is fast and allows easy separation of the plagioclase (uncoloured) from the other phases.

(3) Drawing the thin section using a microfilm reader. This technique is the most accurate, but also the most time-consuming. Every grain and grain boundary can be recognised with no ambiguity and the grains manually assigned a colour.

Once we have a binary image, the average shape of the selected phase is evaluated using an autocorrelation technique (Panozzo Heilbronner, 1992), which gives an oriented average ellipse as a function of size.

5.3. Estimates of the melt fraction

The distribution of the mineral phases was determined on three orthogonal sections from each of nine upper gabbro samples. The volume fraction of each phase was then estimated by averaging the fraction measured on the three orthogonal sections (Table 1). Six samples display subdoleritic textures, and three samples display tabular textures. For each of these
<table>
<thead>
<tr>
<th>Texture:</th>
<th>88OA6a subdoleritic</th>
<th>88OA6b subdoleritic</th>
<th>92OA83 subdoleritic</th>
<th>95OA129 subdoleritic</th>
<th>95OA133 subdoleritic</th>
<th>94OA36 subdoleritic</th>
<th>92OA39a tabular</th>
<th>92OA39b tabular</th>
<th>94OB28 tabular</th>
<th>average</th>
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<td>Image acquisition technique (see text for number)</td>
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<td>3</td>
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<td>A. plagioclase (%)</td>
<td>47.7</td>
<td>45.1</td>
<td>40.5</td>
<td>81.2</td>
<td>61.1</td>
<td>70.6</td>
<td>64.3</td>
<td>60.7</td>
<td>67.8</td>
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<td>B. clinopyroxene + olivine (%)</td>
<td>41.3</td>
<td>41.7</td>
<td>52.8</td>
<td>18.8</td>
<td>38.9</td>
<td>29.4</td>
<td>28.2</td>
<td>33.7</td>
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<td>C. opaque (%)</td>
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<td>0.4</td>
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<td>D. grain boundaries (%)</td>
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<td>6.1</td>
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<td>E. EBP (B + C + D + E) (%)</td>
<td>46.7</td>
<td>46.4</td>
<td>56.7</td>
<td>18.8</td>
<td>38.9</td>
<td>29.4</td>
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<td>F. rims of zoned plagioclases (%)</td>
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<td>G. cores of zoned plagioclases (%)</td>
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<td>33.4</td>
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<td>H. rims of Cpx + Ol (arbitrary) (%)</td>
<td>32</td>
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<td>40.9</td>
<td>15.9</td>
<td>17.9</td>
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<td>48.4</td>
<td>15.9</td>
<td>17.9</td>
<td>39.2</td>
<td>48.4</td>
<td>15.9</td>
<td>17.9</td>
<td>39.2</td>
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<td>49.4</td>
<td>35.2</td>
<td>54.5</td>
<td>21.7</td>
<td>31.1</td>
<td>49.4</td>
<td>54.5</td>
<td>21.7</td>
<td>31.1</td>
<td>49.4</td>
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<tr>
<td>M. empirical melt fraction 3 (%)</td>
<td>17.3</td>
<td>9.8</td>
<td>17.3</td>
<td>9.8</td>
<td>17.3</td>
<td>9.8</td>
<td>17.3</td>
<td>9.8</td>
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<td>9.8</td>
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<td>N. calculated minimum melt fraction (%)</td>
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<td>6</td>
<td>14.2</td>
<td>12.3</td>
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<td>14.2</td>
<td>12.3</td>
<td>6</td>
<td>14.2</td>
<td>12.3</td>
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</tbody>
</table>


Average shape of F (X : Y : Z): 2.1 : 1.5 : 1 : 2.3 : 1.1 : 1 : 2.05 : 1.4 : 1 : 2.15 : 1.33 : 1

Modes are given for the different phases considered (average values from three perpendicular sections). EBP = 'Everything But Plagioclase'), together with estimates of melt fraction and shape. See text for further details.
samples, we obtained different approximations of the steady-state melt volume fraction, depending on what we assume to be liquid in the gabbroic mush of the LVZ (Figs. 3–5). First we measured the amount of the EBP phase in the nine samples (Fig. 3b, Fig. 4b). It ranges from 18 to 60% (Table 1, Fig. 5) and averages 38%. However, the EBP phase is not a valid approximation of the melt fraction, because the plagioclase is not the only phase to crystallise early, and late-stage melts can also crystallise plagioclase (Fig. 6).

Some of the samples studied (92OA39b, 92OA83a, 88OA6b, 88OA6a) display plagioclase zoning, easily visible in a polarising optical microscope or on back-scattered images from a scanning electron microscope (Fig. 7). Zoning of the plagioclases results from relatively rapid crystallisation in a close system; the composition of the cores is around An80 and the rims average An60 (Pallister and Hopson, 1981). It also demonstrates that the last melts crystallise plagioclase. Thus, when present in the samples (Table 1), we incorporate the plagioclase-rim phase into the melt phase (Fig. 3c–f, Fig. 4c–f).

On the other hand, clinopyroxene (or olivine in samples 92OA39b and 95OA129) starts to crystallise above the eutectic temperature (Fig. 6). It forms large crystals, often enclosing small plagioclases, which can be distinguished easily from smaller interstitial pyroxenes. The cores of these early clinopyroxenes (and olivines) must be part of the crystallised phase in the steady-state mush. Unfortunately, we have no information on either the fraction or the geometry of the parts of these crystals which were removed from the liquid phase. Chemical zoning of clinopyroxenes from the upper gabbros is very weak and does not appear to be consistent (Pallister and Hopson, 1981); we observed no zoning on SEM back-scattered images (Fig. 7). Moreover, clinopyroxene and olivine in the upper gabbros are frequently hydrothermally altered and replaced, at least partially, by secondary amphibole and serpentine, which makes recognition of subtle primary zoning impossible. Consequently, the area of the core region that must be subtracted from the liquid phase is unconstrained (Fig. 3c,d, Fig. 4c,d). The last part of the texture which must be considered when estimating the melt phase is the grain-boundary phase, which ranges from 5 to 10% on the drawings (Table 1). In Fig. 3c and Fig. 4c (empirical estimate 1), the system is closed and melt pockets are isolated; in Fig. 3d and Fig. 4d (empirical estimate 2), all the grain-boundaries are wetted and the melt phase is connected. The empirical melt fraction 1 ranges from 16 to 48% and averages 25%; the empirical melt fraction 2 ranges from 22 to 55% and averages 38% (Table 1, Fig. 5). These values probably overestimate the steady-state melt fraction in the gabbroic mush of the LVZ, but they are clearly incompatible with values of a few percent suggested by geophysicists for the LVZ at the EPR. Indeed, they are of the same order of magnitude as the melt fraction inferred to be present in the upper melt lens by Hussenoeder et al. (1996) who compare their seismic data with the experimental values of Murase and McBirney (1973). We shall consider these estimates as upper bounds.

Lower bounds for the melt fraction can be estimated by limiting the melt phase to parts of the texture that can be identified as late trapped liquid. When present, the plagioclase zoning is considered to be part of this trapped melt phase (Table 1, Fig. 3e, Fig. 4e). The other obvious candidates to belong to
Fig. 5. Comparison of the EBP phase and different melt fraction estimates for the nine upper gabbro samples studied. See text for further discussion.

The minimum melt fraction are the interstitial, small clinopyroxenes (and olivines when present). We can empirically estimate this contribution by simply removing all pyroxenes (and olivines) which are bigger than an arbitrary chosen size (Fig. 3f, Fig. 4f). This threshold size depends on the sample and ranges from 0.5 to 1 mm. The corresponding empirical melt fraction 3 (plagioclase zoning + small Cpx/Ol) ranges from 8 to 17% and averages 12% (Table 1, Fig. 5). Another way to estimate the contribution of the Cpx/Ol phase is to calculate from the appropriate phase diagram (An67–Di–Fo, Biggar and Humphries, 1981) the fraction of these minerals which is consistent with the rims of zoned plagioclase. The respective phase proportions (Pl 54%–Cpx 40.5%–Ol 5.5%, or Pl 53.4%–Cpx 46.6%) are estimated from the position of the corresponding eutectic points in the An67–Di–Fo diagram. The minimum melt fraction calculated this way ranges from 1.6 to 14.2% and averages 8.5% (Table 1, Fig. 5).

5.4. Distribution of the melt fraction (average shape)

From the ellipses obtained by the autocorrelation method, we construct ellipsoids representing the average shape of the selected phase (Fig. 8). The shape anisotropy (aspect ratios of the ellipsoid; $X : Y : Z$) of the EBP phase is generally very weak. It is slightly stronger in tabular samples, especially in 94OB28, reflecting the stronger shape fabrics in these samples (compare Fig. 2b with Fig. 2a).

The average shapes of the empirical melt fractions 1 and 2 in the tabular samples also show...
stronger anisotropy. The ellipsoids are slightly more elongated and flatter when the grain boundaries are incorporated into the melt phase (compare empirical melt fraction 2 and 1 in Table 1 and Fig. 8). Similar aspect ratios are also obtained by considering only the rims of the zoned plagioclase. Thus, the average shape of the estimated melt fraction is insensitive to uncertainties regarding which phase represent melt. The shape is generally prolate, intermediate between a plane ellipsoid and a pancake shape, the anisotropy is always weak (averaging about 2.5 : 1.5 : 1), except for the strongly aligned tabular texture of 94OB28 (Fig. 2b) for which the ellipsoid aspect ratio is about 4 : 4 : 1 (Table 1, Fig. 8). The mean shape ellipsoid is always parallel to the foliation of the sample, indicating that the distribution of the melt is directly related to the fabric of the solid component of the partially molten rock. Similarly, the longest axis of the ellipsoids is parallel to the lineation in the sample.

The anisotropy of melt distribution is weaker in the subdoleritic gabbros, which were sampled in the uppermost part of the gabbro section, near the root of the sheeted dike complex. In this work, we favoured this type of texture because their geometry is easier to interpret in terms of liquid distribution. However, they are not typical of the upper gabbro section, which is usually more foliated. Therefore, we believe that the mean measured distribution (2.5 : 1.5 : 1) is a lower bound; the general melt distribution is more likely to be similar to the one obtained in sample 94OB28 (i.e., of the order of 4 : 4 : 1).

5.5. Lower gabbros: a lower bound from the distribution of opaque minerals

In the layered gabbro section (about two lower thirds of the igneous crust), the textures tend to equilibrate, with significant grain-boundary migration after primary crystallisation of the constituent minerals, and the approximations we use in the upper gabbros are no longer valid. We did attempt to perform the same type of analysis on one layered gabbro sample (95OA183, Fig. 9a), by using the exceptionally abundant, late-crystallising opaque (oxides and sulphides) phases as an approximation to the trapped liquid phase. The opaque phase represents 4.5% of the sample volume, and its mean shape is 4 : 4 : 1 (Fig. 9b). This shape is remarkably similar to the most elongated ellipsoid obtained above for the upper gabbros.
6. Calculated seismic velocities

P-wave velocities were calculated using the estimates of the melt distribution described above, and the preferred crystal orientations of the crystalline phase. As the crystalline phase is dominantly plagioclase, we approximated its seismic properties by using plagioclase fabrics characteristics of upper and lower gabros (Fig. 10). The fabric of the lower gabbro (sample 90OF11, Fig. 10b) is stronger than the upper gabbro one (sample 95OA129, Fig. 10a). We use the elastic constants for An57 at ambient conditions (Aleksandrov et al., 1974), and the bulk elastic modulus for basalt (14.91 GPa) is calculated from the P-wave velocity measured at 1200°C and ambient pressure; the effect of pressure on the basalt elastic modulus can be neglected, as can the effects of pressure and temperature on the plagioclase elastic constants (Mainprice, 1997; the maximum Vp overestimate for dry plagioclase is 0.6 km s⁻¹). The basalt and An80 densities are 2.7 g cm⁻³ (Mainprice, 1997) and 2.72 g cm⁻³ (Smith and Brown, 1988, p. 290), respectively. Fig. 11 shows the evolution of the compressional velocities along the three principal axes as a function of the melt fraction for the upper gabros. The initial anisotropy (prior to partial melting) is very weak, reflecting the weak plagioclase crystallographic fabric (Fig. 10a). Note that the maximum velocity for a foliated aggregate of plagioclase is perpendicular to the foliation and not parallel as is the case for olivine and pyroxene (Mainprice, 1997). Thus, the anisotropy of the crystalline aggregate is overestimated by considering just the plagioclase fabric. Because of the compensating effect of olivine and pyroxene, the anisotropy of a typical gabbro (i.e., about 60% Pl and 40% Cpx/Ol) is negligible.

6.1. Calculations of seismic velocities for upper gabros

The results of the calculation for upper gabros are shown in Fig. 11 for the two melt distributions estimated above (2.5 : 1.5 : 1 and 4 : 4 : 1). Since the plagioclase fabric used for the crystalline phase is weak (Fig. 10a), the initial anisotropy is also very weak; Vp is slightly faster perpendicular to the foliation (VpZ). Because melt is preferentially aligned with the foliation, VpZ decreases most rapidly with the melt fraction. Consequently, the sense of the
anisotropy reverses, with the fastest velocities parallel to the foliation. If the melt is preferentially aligned in one direction, this direction is the fastest (Fig. 11a). For the estimated range of melt fractions (8.5% to 38%), the melt shape estimate of 2.5 : 1.5 : 1 yields $V_p$ ranging from 4.15 to 6 km s$^{-1}$ (Fig. 11a), and the melt shape estimate of 4 : 4 : 1 yields $V_p$ ranging from 3.7 to 5.7 km s$^{-1}$ (Fig. 11b). These velocities are generally consistent with the upper half of the LVZ, considering the uncertainties in our estimate of the melt fraction and the velocities of the tomographic models (Fig. 1).

6.2. Calculations of seismic velocities for lower gabbros

The only difference between our calculation of lower gabbro velocities and the calculation shown in
Fig. 10. Lattice fabrics of plagioclase for (a) upper gabbro (sample 95OA129, subdoleritic texture), and (b) lower gabbro (sample 90OF11, tabular texture).

Fig. 11b is the stronger plagioclase fabric (Fig. 10b) used for the crystalline phase. This results in a stronger anisotropy at 0% melt (Fig. 12) but this difference is eliminated with a few percent of melt present. As explained above, information on the geometry of the primary igneous texture at the sample/thin section scale in the lower gabbros is lost because of re-equilibration of the texture. It is not possible to estimate the melt distribution from the textures at the thin section scale, and a safer approach is to estimate the melt fraction from the calculated Vp/X assuming a given melt shape. The velocities constrained by the seismic tomography are approximately 5.5 to 7 km s\(^{-1}\) in this region, yielding a melt fraction ranging from 0 to 22% for an average melt shape of 4 : 4 : 1 (Fig. 12).

7. Discussion

The analogy between the Oman ophiolite and fast-spreading oceanic ridges has been presented elsewhere (e.g. Nicolas et al., 1994; Nicolas and Boudier, 1995; Boudier et al., 1997), and we will not discuss it again. The main limitation of our approach is that the present-day microstructures and textures of the Oman gabbros may not correspond to the steady-state partially crystallised mush for two reasons. First, it is not necessarily straightforward to identify any primary igneous texture with the geometry of a trapped liquid (e.g. Means and Park, 1994; McBirney and Hunter, 1995). Second, the textures tend to equilibrate very easily under hypersolidus conditions. In this study, this problem is minimized by choosing samples from the uppermost part of the
igneous section, which crystallised fast enough to preserve apparent primary textures, with euhedral, poikilitic and interstitial phases.

In order to approximate the distribution of the melt phase, we have made a series of hypotheses, leading to various estimates of both the melt content and the melt distribution. Rather than choosing a single value, we favour determining bounds on the melt fraction (see above). The upper limit ranges from 25% to 38%, depending on our assumptions about the distribution of the melt along the grain boundaries. The best estimate for the grain-boundary contribution is somewhere between these two end-members. Experimental studies in various types of rocks show that the distribution of melt at the aggregate scale is anisotropic (e.g. Longhi and Jurewicz, 1995; Laporte et al., 1997; Cmiral et al., 1998).

Melt preferentially wets planar crystalline faces of low solid–melt interfacial energy, and is thus related to crystalline anisotropy. However, there are a few studies on the melt topology in partially molten, deforming gabbroic assemblages, and the results are somewhat contradictory. Longhi and Jurewicz (1995) report a distribution of wetting angles (45° on average) reflecting the crystalline anisotropy in an anorthite aggregate, whereas Dimanov et al. (1998) observed melt pools that are probably unconnected in experiments on partially molten labradorite, with no wetting of the grain boundaries. The relationships between crystal orientations and the distribution of melt are still not well established and cannot be included in this work.

It is also important to consider the scale at which the melt is distributed. We can infer that the melt is distributed homogeneously in the upper gabbros,
since they generally have no layering and display a homogeneous, scale-independent structure. The melt distribution inferred at the sample/thin section scale should be representative of the melt-related seismic signature of the rock at the scale of the seismic wavelength. For the lower gabbros, the situation is more complicated since they are heterogeneous by nature. It has been proposed recently that the lower part of the magma chamber is fed by the episodic intrusion of sills (Boudier et al., 1996; Kelemen et al., 1997). Seismic models in the lower crust have a resolution of 1 km at best, which probably exceeds the vertical dimension of these sills. However, there are no constraints on the timing and the scale of these intrusions. Consequently, their effect on the average melt shape used in the model, and thus on the seismic signature of the lower part of the LVZ is unclear. We also know from the geochemistry that lower gabbros are cumulates (Pallister and Hopson, 1981; Kelemen et al., 1997). A significant amount of melt has to escape from the system (at least 50% according to Kelemen et al., 1997). This implies that a large amount of melt migrates through the gabbroic mush. However, the mechanism for melt migration remains unknown. Short-scale chemical variations apparently preclude pervasive porous flow, and favour a channel model (Korenaga and Kelemen, 1997). However, they are presently no field observations to support this mechanism. We have no good constraints on the melt distribution at a given time in the lower part of the LVZ.

In the upper gabbro, the lower limit on our estimate of the melt content is of order 10% (12.3% for the interstitial minerals only, 8.5% for the calculation using the plagioclase zoning). If we accept 10% as a reasonable estimate for the minimum melt content, the corresponding $V_p$ is $5.7 \text{ km s}^{-1}$ for an average melt shape of $4:4:1$ (Fig. 11b). To explain the lowest velocities in the uppermost part of the LVZ ($4\text{ to }3 \text{ km s}^{-1}$, Fig. 1), one needs a melt content increasing upward toward the melt lens, and/or an increase in the anisotropy of the melt distribution. Fig. 13 shows the decrease of $V_p$ with increasing ellipticity of the average melt shape.

Fig. 13 shows the decrease of $V_p$ with increasing ellipticity of the average melt shape. The largest effect is observed perpendicular to the foliation when the average melt shape changes from $1:1:1$ to $10:10:1$ (Fig. 13a). For all directions, $V_p$ is relatively insensitive to melt shape once the ellipticities exceed 10 (Fig. 13), and $V_p$ will unlikely be lower than $5 \text{ km s}^{-1}$ (Fig. 13a) with a melt fraction of 10%. Thus, the anisotropy of the melt distribution alone is not sufficient to explain the lowest velocities in the uppermost part of the LVZ.

In the lower gabbros, a $V_p$ $5.5\text{ to }6 \text{ km s}^{-1}$, a range typical of the lower part of the LVZ (Fig. 1), corresponds to a melt fraction ranging from 9.5 to

![Fig. 13. Variation in P-wave velocity as a function of melt fraction for various melt shapes. (a) VpZ (perpendicular to the foliation). (b) VpX (parallel to the lineation).](image-url)
22% for a 4 : 4 : 1 melt ellipsoid. Again, we have a lower bound of order 10%. The velocity along the lineation (Vp X), that is parallel to the maximum elongation of the representative melt ellipsoid, is not sensitive to the average shape of the melt fraction (Fig. 13b). Increases in the velocity at the bottom of the LVZ (from 6 to 7.5 km s⁻¹ in Fig. 1) are likely to be related to a progressive increase with depth of the abundance of ultramafic, olivine-rich layers in the basal layered gabbros, as is commonly observed in Oman. Increases in Vp would then be the result of the higher Vp of olivine and to the Vp anisotropy of olivine which counteracts that in plagioclase, since the fast velocity is parallel to the lineation. A decrease in the melt fraction downward is not required, although it may contribute to the trend.

The accuracy of the velocity models (Fig. 1) is weak, especially in the lower half of the crust (Dunn et al., 1999). Thus our estimate of the melt fraction for the lower gabbros from Vp is not well constrained. Nevertheless, our results are consistent with melt fraction ranging from 10% to 30% in the LVZ. The upper bound is not well constrained since estimates are highly variable (see Table 1). We can infer that the melt fraction is similar throughout the LVZ, as implied by the constant viscosity in the numerical models of Chenevez et al. (1998). The upper bound can then be lowered to 22%, which is the upper limit for the lower gabbros. With a melt fraction of the order of 10 to 20%, the seismic structure of the LVZ, and in particular its vertical velocity gradient, can be explained by the combined structural effects of the melt being preferentially aligned in the developing foliation, and of the foliation being variably oriented with respect to the seismic ray paths. This melt content is still at least twice as large as independent estimates previously obtained for the LVZ on the basis of seismic (e.g. Caress et al., 1992; Wilcock et al., 1992, 1995) or electromagnetic (Evans et al., 1994) experiments on the EPR. As Wilcock et al. (1995) point out, no laboratory studies completely simulate the conditions of the seismic experiments. We note that our estimate of the melt fraction is consistent with (1) the field-derived evidence (Nicolas and Ildefonse, 1996), (2) compliance measurements on the EPR which yield an estimate of 2–18% for the melt fraction in the LVZ (Crawford et al., 1999), and (3) a recent three-dimensional tomographic model of the EPR at 9°30′N (Dunn et al., 1999) which yields an estimate of 5–22% at 4 km depth and 2–11% at 6 km depth for the melt fraction.

8. Conclusions

We have analysed the texture of nine gabbro samples from the Oman ophiolite, in order to estimate the melt fraction and its average shape in the crystallising mush present in the steady-state magma chamber (LVZ) at fast spreading ridges. We have chosen upper gabbro samples because they display primary textures with interstitial phases that are assumed to represent the approximate geometry of melt in the magmatic mush of the steady-state magma chamber. Our approach does not allow a precise estimate of the melt fraction in the upper half of the LVZ. However, it places bounds on the possible melt fraction and on the spatial distribution of melt at the aggregate scale. The average melt fraction ranges from approximately 10% to 30%. It is important to note that the lower bound (~10%) is significantly higher than previous estimates obtained from most geophysical studies on the EPR, and is in better agreement with estimates derived from the Oman ophiolite and recent geophysical studies on the EPR (Crawford et al., 1999; Dunn et al., 1999). The melt is preferentially aligned with the mineral foliation, and has an average shape which can be approximated by a 2.5 : 1.5 : 1 ellipsoid. We think that this shape probably underestimates ellipticity, and should be taken as a minimum, since most of the samples used for our analysis display weaker foliation than the typical upper gabbros. The 4 : 4 : 1 ellipsoid obtained for the sample with a well foliated tabular texture more likely represents the average melt shape. Direct estimates of the melt fraction and distribution are not possible in the lower gabbros. However, an estimate can be obtained indirectly from our modelling. In the lower part of the LVZ, the influence of the melt shape is weak because the melt lenses are nearly aligned with the seismic ray paths; the melt fraction is estimated from velocities in the tomographic image (Fig. 1) and is about 10 to 20%.

Previous interpretations of the seismic structure of the East Pacific Rise at 9°30′ (Fig. 1) were based on
the assumption that the melt was distributed isotropically and, on the basis of the available experimental data, concluded that the melt fraction was small. We suggest that the same seismic structure is consistent with a larger amount of melt (of the order of 10–20%) which is anisotropically distributed. The anisotropy does not need to be very large (of the order of 4:1 with the long axes parallel to the foliation plane) to account for the amount of melt estimated from the textures of Oman upper gabbros, and the vertical velocity gradient reported in the LVZ of the EPR. The latter observation can be explained by the preferred orientation of the melt, and by the way it is oriented with respect to the seismic ray paths. Lower velocities in the upper half of the LVZ are not exclusively a result of a decrease with depth of the melt content and/or temperature. They may also result from the steepening of the flow pattern in the upper part of the magma chamber beneath the melt lens, as reflected by the steepening of the magmatic foliation in the Oman gabbros.

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