3D Magma Chamber Structure beneath the 9°03’N Overlapping Spreading Centre

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Abstract—Overlapping spreading centres (OSC) play a key role in models of magma supply at fast spreading ridges. To investigate the relationship between ridge-axis discontinuities and magmatic segmentation, we conducted a three-dimensional seismic reflection and tomography experiment at the 9°03’N OSC along the East Pacific Rise (EPR). Shadow zones in the refraction record sections indicate that a low velocity zone (LVZ) lies beneath parts of the overlap and its basin. Subsequent three-dimensional tomographic analysis imaged a broad LVZ, 8-10 km wide, with negative velocity anomalies as large as 1 km/s, suggesting that the OSC is magmatically robust. The complimentary seismic datasets reveal that the mush zone and melt sill at either end of the overlap basin are not simply centred beneath the rise crest but are skewed inwards as a result of misalignment between the ridge and mantle supply. Thus, although the magma supply seems to be concentrated at the northern and southern extremities of the overlap basin, the OSC is locally rather than distantly fed. The subsequent focussing of the mush zone and sill beneath the axis of the eastern limb appears to be due to melt migration towards the tip. The western limb of the OSC is less magmatically robust than the eastern limb and the absence of the LVZ from the tip may indicate that it is in the process of dying.

Introduction

Between major transform faults, the fast spreading EPR is offset by smaller discontinuities in which the two segments overlap significantly but are not separated by a transform fault. The average segment length bounded by these small offsets is 50 km while transform faults occur every ∼400 km (Macdonald et al., 1991). Macdonald et al. (1988) identified eight of these offsets between the Rivera and the Siqueiros transform faults and named them overlapping spreading centres. Lonsdale (1985) identified four offsets south of the Siqueiros transform fault and called them non-transform offsets.

Interpretation of these features varies, depending upon the underlying crustal accretion model for the EPR. In one model, the supply of magma is essentially two-dimensional with the crustal magma chamber being locally fed along an entire ridge segment. Perhaps there is some along-axis focussing of the magma supply, which is reflected in the finer scale segmentation of the ridge. In an alternative model, the mantle upwelling is highly focussed into mantle diapirs with a single diapir supplying melt to segments between major discontinuities (Macdonald et al., 1988). In the context of the diapiric model, a significant along-axis transport of melt in the crustal magma chamber is necessary, to redistribute melt away from regions of localised upwelling toward axial discontinuities (Cormier, 1997). The subduced along-axis gravity and bathymetric variability at the rise crest suggests that the EPR may experience focused upwelling without having focussed accretion. The diapiric model presents OSCs as magmatically starved regions of the ridge axis far from the upwelling diapirs, where the limbs are underlain by small and intermittent magma chambers (Macdonald et al., 1984). The two-dimensional model suggests that the OSC will be locally fed. Lonsdale (1983) argues that OSCs are magmatically robust and are fed by a large magma chamber that spans the OSC basin. In order to test these different models of magma supply at the fast-spreading EPR, the ARAD 3-D (Anatomy of a Ridge Axis Discontinuity) seismic survey was undertaken at the 9°03’N OSC. The experiment was designed to test whether magma was remotely or locally supplied to the OSC and to examine whether each limb of the OSC has
its own melt supply or whether they share a single supply.

There is no seismic method that is by itself capable of imaging the complete velocity structure of the oceanic crust. Refraction and reflection methods have different inherent strengths and weaknesses related to their geometry. A combined three-dimensional reflection and refraction seismic experiment such as the ARAD 3-D survey is needed to provide the data necessary to image accurately the internal structure of an OSC.

**Geological setting**

The 9°N segment and its flanks have been extensively mapped with SeaBeam and SeaMARC II systems (Macdonald et al., 1992). The average full spreading rate for the Brunhes/Matuyama boundary is 117 mm/yr (Carbotte and Macdonald, 1992), ranking this segment as fast even within the fast spreading centres category. The ridge segment is bounded by two major transform faults, the Clipperton at 10°10’N and the Siqueiros at 8°22’N (see inset in Figure 1).

A combination of various seismic experiments carried out along the 9°N segment has led to the definition of a new structural model of the axial magma chamber at fast-spreading mid-ocean ridges (Kent et al., 1990; Sinton and Detrick, 1992). A thin, narrow, lens-like melt reservoir (50 to hundreds of meters thick, ~1 km wide) is nearly continuous along the axis of the segment (Detrick et al., 1987; Kent et al., 1990, 1993; Collier and Singh, 1997). A broad LVZ underlies the melt lens and is believed to be a region of hot material and possibly molten rocks (Harding et al., 1989; Vera et al., 1990). Velocity anomalies as low as -0.5 to -2.7 km/s have been reported in the axial LVZ (Vera et al., 1990; Toomey et al., 1990, 1994; Dunn et al., 2000) and extend 3 to 10 km off axis. The melt fraction in the axial LVZ could be as high as 16 to 38% (Dunn et al., 2000).

**3-D tomographic inversions**

The observation of geometric shadow zones on refraction lines crossing the OSC (Figure 1) is indicative of a LVZ beneath the overlap basin. The seismic arrivals that travelled at depth beneath the OSC basin are delayed and attenuated, resulting in a reduced signal-to-noise ratio and reduced apparent trace-to-trace coherence in the data. We infer the LVZ to be a region of elevated temperature and possibly partially molten rock beneath the ridge-axis discontinuity. We performed a tomographic analysis of 20,000 Pg arrivals recorded by 19 ocean bottom instruments to assess the extent of the LVZ. The shots came from the 7 ridge parallel and 4 ridge perpendicular refraction lines (28 to 72-km length profiles) and the shot-receiver ranges of the Pg arrivals varied from 2 and 42 km. Further details of the experimental procedure may be found in Bazin et al. (submitted to J. Geophy. Res.).
We start the tomographic inversion with a one-dimensional velocity profile derived by forward modelling; no LVZ is explicitly built into this starting velocity model (Van Avendonk et al., 1998). The starting RMS travel-time residual is 121 ms and is reduced to 25 ms after 15 iterations (variance reduction of 96%). Nominally this slightly underfits the data, as the total predicted RMS errors, which includes errors from bathymetry, travel-time picks, instrument and shot locations, instrument clock drift, and travel-time computation, is 17 ms. The main source of error is picking errors which averages 12 ms but can reach 24 ms. A superior data fit would be feasible if a weaker smoothing constraint was applied in the inversion and finer grid cells were chosen.

A map view of the 3-D velocity model solution at mid-crustal depth, 2.3 km below seafloor, is displayed in Figure 2. The velocities are referenced to 6.8 km/s to allow direct comparison with the LVZ imaged at 9°30'N by Toomey et al. (1994). At this depth, a LVZ is present beneath both the western and eastern limbs, but does not cover the entire OSC: the western ridge tip north of 9°02'N does not manifest any LVZ. The southern portion of the LVZ underlies the western limb; however, the northern portion of the LVZ is not centred beneath the eastern axial high but shows a westward shift toward the basin. The truncation of the low velocity body north of Line L and south of line P is an artefact of the ray coverage.

The tomographic model shows individual low velocity zones beneath the eastern and western limbs of the OSC that are shifted inwards from the ridge axis at the northern and southern extremities of the basin. We present two resolution tests, the first demonstrates that the experimental geometry and tomographic inversion is capable of recovering a structure similar to that observed with a LVZ beneath each limb. The second demonstrates that the inversion would not artificially fragment a single LVZ body beneath the basin into separate LVZ domains. Two elongated low velocity perturbations (maximum amplitude of -30% with gaussian shape) are placed at 1.5 km depth below seafloor beneath each OCS limb (Figure 3). The model solution fit the synthetic data to the same level as the actual data. The amplitude and width of the anomalies are recovered and there is no lateral smearing. The line of instruments through the basin prevents the two LVZs from linking up into a single body. However, the resolving power is not as good at the centre of the eastern limb as elsewhere due to instrument failures in the region.

Next, a large elongated low velocity perturbation underlying the entire OSC is placed at 2 km depth below seafloor (Figure 4). Within the overlap basin, the entire anomaly shape and maximum amplitude are recovered and there is no evidence of fragmentation into two distinct LVZs at either end of the OSC basin (Figure 4). Therefore, the two focused zones of low velocity, as resolved by our method, are probably not an artefact of the geometry.
Interpretation and discussion

Our models indicate the presence of a low-velocity zone beneath the 9°03’N OSC, but this body is not continuous under the entire discontinuity. Two zones of low velocities are focused at the two extremities of the basin, and seem to shift toward the two distinct melt lenses at shallow depths and possibly connect at the propagating ridge tip (Figure 2). Resolution tests indicate that our method is able to resolve this LVZ structure. We conclude that neither a broad magma chamber spanning the entire overlap basin, nor distinct bodies centred beneath the two limbs, resides at the 9°03’N OSC.

We compare the extent of the axial LVZ with the melt lens reflections picked in the three-dimensional reflectivity volume (Kent et al., 2000). The MCS survey reveals that the melt sill underlies both limbs and the northern part of the overlap basin. The melt sill beneath the eastern limb is unusually wide (up to 4 km), and shifted toward the west with respect to the axial high, while the melt sill underlying the western limb is close to typical width (∼1 km) (Kent et al., 2000). The eastern velocity anomaly shows the highest amplitude (∼1 km/s compared to a fixed velocity of 6.8 km/s) beneath the widest part of the melt lens and fades toward the south (∼0.8 km/s). As the eastern melt sill narrows to ∼1 km width, the LVZ jumps toward the east and becomes realigned with the topographic high. At the latitude where the eastern melt lens vanishes and starts appearing on the western limb, the low velocity anomaly increases again but is now focused beneath the western limb. Two-way travel-times computed by vertical integration through our velocity model reveal that the melt sill lies on average 1705 m below seafloor. Within estimation error, this depth estimate is consistent with the 1680 m of Kent et al. (1993) measured at the same location from the 1985 MCS seismic data with the rise crest velocity function of ESP 5 (Vera et al., 1990). There is a depth difference of ∼220 m between the eastern and western melt sills. This could indicate a different heat budget between the two OSC limbs (Phipps Morgan and Chen, 1993). In addition we observe an increase in two-way travel-times of the melt sill reflector along its narrow portion toward the south (Figure 5), while no systematic deepening is measured over the segment scale (Kent et al., 1993). This reflector seems to coincide with that of isovelocity contour 5.8

Figure 3. Resolution test investigating a previously proposed magma chamber model for the 9°03’N OSC (Macdonald et al., 1984). The left panel shows a map view of the low velocity perturbation applied for the resolution test: two elongated anomalies were placed 1.5 km below the seafloor. The right panel displays the recovered anomaly using the same shot-receiver geometry as in the tomographic inversion. The velocity sections are mapped at depth of 1.5 km below seafloor. Red/orange colours indicate slow velocity anomalies. Thin black contours are -1.5, -1, -0.5 and 0 km/s velocity anomalies. The triangles show the locations of the 19 instruments used. White 2700 m isobath is overlain for reference and contours the ridge axis.

The two resolution tests demonstrate that the truncation of the LVZ north of 9°03’N (Line N, Figure 1) and south of 8°57’N (Line P) in Figures 3 and 4 is an artefact of the ray coverage.

Figure 4. Resolution test investigating a previously proposed magma chamber model for the 5°30’N non-transform offset (Lonsdale, 1983). The left panel shows a horizontal section of the low velocity perturbation applied for the resolution test: a broad low velocity anomaly was placed 2 km below the seafloor. The right panel displays the recovered anomaly. The horizontal sections are mapped at depth of 2 km below seafloor.
km/s, which also deepens by a few hundreds of meters over this last section of the tip. Our observation is in agreement with waveform modelling results showing velocity at the sill is between 5.6 and 6 km/s at various locations along the EPR (Hussenoeder et al., 1996; Collier et al., 1997).

The dense MCS experiment was designed to quantify the magma distribution in the axial melt lens (Kent et al., 2000; Figure 2). In order to map the completely molten regions of the lens, we applied a Amplitude Versus Offset (AVO) technique developed by Singh et al. (1998), which predicts that reflection coefficients vary differently for melt or mush bodies with respect to the incidence angle. Therefore, differences in reflection amplitudes between partially stacked reflectivity volumes may detect variations in the melt lens composition. A preliminary AVO stack attempt shows that the 4 km wide axial magma chamber (AMC) reflector under lying the eastern limb shows amplitude variations as a function of the angle of incidence: both near- and wide-angle reflections are detected beneath the axial high while only near-vertical reflections are observed at the lens portion that spans the OSC basin. These observations may suggest that the western and deeper part of the lens is more molten than the shallower section which underlies the rise crest.

We suggest the following geometry for the magma supply of the propagating limb: melt arises from beneath the northern extremity of the overlap basin at 9°09’N and is trapped at the base of the dike complex. The molten signature at the western portion of the AMC reflector (Figure 5) suggests that the upwelling liquid has been emplaced at this location more recently than that underlying the rise crest. Around 9°07’N, we observe a right lateral step of the LVZ (Figure 2) and below this point, we do not image a deep root to the low velocity body. Therefore, we propose that the melt supply of the eastern limb emanates from focused upwelling at 9°09’N. A melt migration toward the southern tip following mid-crustal fractures or cracks is suggested by the presence of small right-lateral en echelon steps along the narrow section of the melt lens (Kent et al., 2000). Lavas from the propagating limb are generally more fractionated (higher Fe content and lower Mg number as Mg/(Mg+Fe2+); Langmuir et al., 1986; Natland et al., 1986; Sempéré et al., 1988) than those from the western limb or from the adjoining southern and northern ridge crest. The high level of fractional crystallisation is likely a result of the diminishing and deepening melt lens south of 9°07’N.

Conclusions

Figure 5. Seismic volume showing the difference in reflectivity amplitudes between two partial stacks in a 7.5 x 16 km box covering the eastern limb of the OSC. Short offset reflections (stacked data from 0 to 1.55 km source-receiver offset, coloured in yellow/red) and wide-angle reflections (stacked data from 0.55 to 3.1 km offset, coloured in blue/green) behave differently for the three main interfaces. Seafloor reflections show contributions from all offsets, the sharp velocity gradient at the base of Layer 2A creates mainly wide-angle reflections, while the AMC reflector creates near- and wide-angle reflections beneath the axial high but only near-vertical reflections beneath the OSC basin where the reflector is 4 km wide.
We have successfully applied three-dimensional seismic tomographic inversions to an extensive refraction data set at the 9°03′N OSC along the EPR. A low-velocity body underlies most of the axial discontinuity. At shallow depth, the low velocity zone underlies the melt sill reflection as imaged by Kent et al. (2000). The melt sill lies at approximately 1.7 km below the seafloor and its eastern limb deepens as it narrows towards the ridge tip. Fractionated lavas (low Mg numbers and high Fe content) were found at the ridge tip and are probably associated with the diminishing magma chamber.

The LVZ is not centred below the two rise crests but is skewed toward the OSC basin. The amplitude of the northern low velocity anomaly and the large width of the melt lens would imply that the melt supply is probably greater in the northern part of the 9°03′N OSC basin. We postulate that rather than being fed by along axis flow from robust regions of the 9° segment, the propagating ridge tip is supplied by a local source, which itself is misaligned compared to the neovolcanic zone. Melt arises from beneath the northern part of the overlap basin and is injected into the propagator.

References


