MANTLE DYNAMICS: MAGMA GENERATION AND DELIVERY

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Key Problems:

1. What is the overall pattern of mantle flow beneath spreading centers? Is the flow largely passive, following the form that would be expected to be induced simply by the motion of the separating plates, or is there a dynamic effect of buoyancy that alters and intensifies the flow in certain areas? Magma is generated by the pressure release melting of upwelling mantle, so the overall pattern of flow controls the spatial distribution of magma production.

2. Is there a narrow conduit for mantle flow beneath slow spreading ridges? There are two primary explanations that have been advanced for the dynamic origin of median valley topography: crustal and lithospheric stretching, or restricted flow of an upwelling, viscous fluid within a narrow conduit. The later model requires flow to be confined within a zone about the width of the median valley to a depth of tens of kms and has profound implications for the rheological properties of the mantle.

3. What is the role of absolute plate motion, or motion relative to the deep mantle beneath the ridge? If the plates are moving relative to the deeper, upwelling flow, then heterogeneities in the flow may produce characteristic signatures in magma generation that may migrate along the axis of the spreading center. In addition, absolute plate motion may be responsible for asymmetric spreading.

4. To what extent do hotspots, on– and off–ridge, supply or contaminate the material that undergoes partial melting beneath the ridge? Is there significant along–axis flow from a few primary zones of upwelling? There is geochemical and bathymetric data that suggest that in at least some areas, along–axis flow may be dominant.

5. Is a small fraction of partial melt broadly distributed within the oceanic asthenosphere, or has it all been swept into the vicinity of the spreading centers? Theoretical models of porous flow suggest the asthenosphere should be drained of any melt fraction, but low seismic velocities and high electrical conductivities indicate that there may be a small liquid component remaining.

6. To what extent do local chemical inhomogeneities within the upwelling mantle affect the temporal variability in the volume of melt production along individual ridge segments? There clearly are variations in volume of melt produced as a function of time and there are also clearly variations in chemistry

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of MORBs that are not simply fractionation effects. Is there a link between the two?

7. What is the horizontal and vertical extent of the zone of melt production? What are the minimum and maximum depths of magma generation? The distribution of the melt production zone should be controlled by the pattern of mantle flow, and the temperature and composition of the upwelling material. Magma may be generated in a zone extending as much as 200 km away from the ridge axis in the direction of spreading.

8. What is the effect of segmentation of the ridge by fracture zones and overlapping spreading centers on the form of mantle flow and the distribution of melt production? Some models of mantle flow predict that melt will be produced at depth beyond the end of a spreading center segment beneath the sea floor on the older side of a fracture zone.

9. Is there a melt accumulation zone within the mantle, either at the base of the crust or where the magma is actually generated? If the melt were not separated during the generation process, partial melt concentrations would reach 20% or more. It is thought that melt segregation is efficient and would not allow such concentrations at depth, but there could be zones where melt is accumulated before intermittent transportation to crustal magma chambers.

10. Is there a narrow, low density root beneath the axis of spreading centers? Gravity data suggests that, in addition to the low densities associated with thermal expansion of hot mantle and crustal rocks, there may be some low density body at the base of the crust or deeper. A melt accumulation zone is possible, but may be inconsistent with petrologic data that indicates that primary melts are in equilibrium with residual mantle at depths of 30 km or more.

11. How is magma transported laterally from a presumed wide zone of melt generation to a narrow magma chamber? The primary driving force for melt segregation and transportation, the buoyancy of the magma, is directed vertically upward. What drives the horizontal flow? The two primary suggestions are shear deformation of the matrix setting up a nonhydrostatic pressure gradient directed toward the ridge axis, and dynamic forcing of the magma (by suction) toward the eruption center.

12. Is there a narrow vertical conduit in which melt ascends beneath a spreading center segment? Are these conduits regularly spaced? Are they quasi–permanent? The conduit for melt may be distinct from the hypothesized conduit for mantle flow. Melt may ascend repeatedly along the same conduit, but it also could ascend by dynamic crack propagation or diapiric uprise, in effect creating

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a new conduit for each ascending blob.

13. What controls the apparent episodicity of the seafloor spreading process? There may be characteristic time periods for replenishment of magma chambers or melt accumulation zones in the mantle necessary to build up magma supply to a level at which eruption or transport can occur. Variations in mantle composition on a small scale may contribute to episodicity in melt production rates. Stretching or necking of the crust and release of stress by faulting may also be important, particularly at slow–spreading ridges.

State of our knowledge:

At present, we have very few direct observations of the deep structure underlying mid–ocean ridges. Petrological and geochemical observations give us constraints on the composition of the source rocks for the magma and the depth and temperature at which the magma equilibrated with the mantle. There is general agreement that most mid–ocean ridge basalts (MORBs) are produced at temperatures of 1150°C to 1400°C and pressures of 5 to 20 kb. Topography reflects the combined effects of thermal expansion of the crust and mantle, dynamic uplift, faulting, and crustal thickness variations. The decay of sea floor topography in proportion to the square root of sea floor age and the decrease of heat flow with age of the sea floor clearly establish the importance of the gradual cooling of the upper mantle. In a gross sense, we understand the thermal structure of the upper mantle and its physical manifestations from measurements of the flexure of oceanic plates under loads, the depth and mechanism of intraplate earthquakes, the distribution and frequency of intraplate volcanism, geoid and gravity anomalies across mid–ocean ridges and fracture zones, the delay of seismic body waves traveling through the oceanic mantle, the apparent resistivity of the mantle as revealed in magnetotelluric experiments, and the dispersion of seismic surface waves.

Our understanding of the details of the deep structure, mantle flow, and the magma generation and delivery system beneath spreading centers stems largely from the construction of theoretical models based on simple physical principles and our knowledge of the gross structure and evolution of the oceanic upper mantle. There are many aspects of the models that should be tested by observation, and, in many cases, there are alternative models that satisfy the existing constraints. In the following section, one possible description of the processes involved in magma generation and delivery is briefly summarized, but this is far from being an exhaustive review of such models. It may serve as a starting point for designing an observational program for understanding mantle

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dynamics.

Models of mantle dynamics, magma generation and delivery

Much of the flow pattern in the mantle can be predicted from the motions of the surface plates. Immediately beneath the plates, the flow is viscously coupled to the lithosphere and must have the direction and velocity of the overlying plate. On a global scale, downwelling at subduction zones must be balanced by upwelling beneath spreading centers to conserve mass, and the general pattern of return flow beneath the plates can be predicted (figure 1). Although the vertical extent of the flow is a matter of debate, it is clear that beneath some ridges, such as the Nazca–Pacific plate boundary in figure 1, the flow may be primarily upwelling to great depth while beneath others, upwelling may give way to predominantly horizontal flow at relatively shallow depths. Small scale convection beneath the plates is ignore in global circulation models such as these.

The same principles that are used to predict global mantle circulation can be applied on a smaller scale to predict the pattern of mantle flow beneath a ridge–transform–ridge system. A simple example of the three–dimensional flow that might be expected is illustrated in figure 2. This example assumes uniform viscosity in the mantle and ignores the thickening of the plates with age, but serves to illustrate some of the complexities that may arise in a three–dimensional system, such as upwelling beyond the end of a ridge segment and horizontal transport along the ridge axis. Deviations from uniform viscosity will have very important dynamic consequences. For example, although it is not predicted by models of the thermal structure, it is possible that upwelling is largely concentrated within a narrow channel or conduit (figure 3). This would provide an explanation for median valley topography of slow spreading ridges and would severely restrict the lateral dimensions of the melt production region.

Relative motion between the surface plates and the underlying deeper mantle directed along the axis of the ridge may play an important role in the temporal variability in magma generation at any one ridge segment. Evidence for along–axis motion comes from V–shaped topographic ridges or seamount chains that are not aligned in the direction of sea–floor spreading. These could be generated either by variable, along–axis flow from hotspots, as observed along the Reykjanes ridge south of Iceland (figure 4a), or by absolute motion of the plates relative to the mesosphere tapping chemical heterogeneities imbedded in the mantle (figure 4b). Along–axis flow may also be revealed by compositional gradients away from hotspots.

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If upwelling does occur in a broad region as predicted by models of the passive flow induced by plate motion, then melt production can be expected to extend to a distance of 100 to 200 km from the ridge axis (figure 5). Both the contours of melt fraction shown here and the models of melt fraction as a function of temperature and pressure assume none of the melt is removed. It is generally believed that any melt above some threshold concentration (2%?) will separate from the matrix and migrate upwards. Removal of the melt may substantially alter the melting relationships, although there is little or no experimental data on the effects of melt segregation in realistic mantle assemblages. Melt production rates should be proportional to the dot product of the mantle flow velocity and the gradient in melt fraction. This is illustrated in figure 6 for a cross-section along the ridge axis and across a fracture zone into the mantle beneath old sea floor, assuming the passive flow model of figure 2.

The pattern of melt migration must be substantially different from the pattern of mantle flow. The melt must migrate horizontally and vertically in order to be concentrated in magma chambers within the crust that are at most a few kms wide. The mechanism for melt segregation is not well established, but it is generally thought to be aided by compaction of the crystalline matrix driving out the fluid. One possibility is that magma in small blebs begins its ascent vertically, but is driven horizontally by nonhydrostatic pressure gradients generated by the deforming mantle matrix. Melt might then be expected to migrate along the paths shown in figure 7. Another possibility is that melt accumulates within the melt production zone forming a gravitationally unstable layer beneath a depleted mantle layer. Instabilities would develop with characteristic spacing of the order of the depth of the melt layer. With continued growth, the instabilities would develop into diapirs that transport the melt to near surface magma chambers at the ridge axis (figure 8).

The gradual growth or replenishment of melt accumulation zones in the mantle, followed by rapid segregation of the fluid phase and diapiric removal of the magma could explain some of the episodic behavior of the seafloor spreading process (figure 9). In this model, the characteristic period would be the time necessary for mantle upwelling to generate a new layer of partial melt that is large enough for instabilities to develop quickly. There could also be other characteristic periods associated with the depletion of crustal magma chambers, the stress-release cycle of earthquakes, or the development of stretching instabilities within the crust beneath the median valley.

Key Experiments:

There are two basic types of field experiments that are required to solve the key questions about mantle dynamics and magma generation and delivery. Much of

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our present knowledge is inferred from incomplete knowledge of regional variations in crustal structure and composition, both along the axis of the ridge and in the direction of spreading. Thus, the first set of observations needed comprises thorough surveys that would systematically elucidate the temporal and spatial variability that provides the indirect clues to deeper processes. The second set of observations would directly test models of the melt formation and migration process by geophysical and petrological measurements of the physical properties of the source region.

Surveys along axis. Continuous surveys along the ridge crest for distances of the order of 1000 km are needed to establish the long–wavelength framework of mantle heterogeneity and flow. These surveys should thoroughly cover a swath or corridor extending to approximately 50 km either side of the ridge axis and should include as a minimum one slow–spreading and one fast spreading ridge section. Within this swath, there should be systematic, closely spaced dredging along the spreading center and along one off–axis isochron, essentially continuous multi–narrow beam echo–sounding (SeaBeam), magnetic and gravity anomaly measurements, a multichannel seismic reflectionsurvey along the same lines as the dredging, and bottom–water sampling for evidence of hydrothermal activity.

Surveys along a flow line. Continuous surveys in the direction of spreading out to sea floor perhaps 40 m.y. old are needed to establish the pattern of temporal variations in magma production and detailed spreading center geometry. These swaths should be wide enough to include 3 or 4 ridge segments (200 to 300 km) so that it will be clear which variations are uniform changes and which represent events that propagated along the ridge axis at the time of formation of the sea floor. The measurements required are similar to those for along axis surveys with the addition of heat flow, exploratory photography, and passive seismic and electromagnetic studies to monitor earthquake activity and to study upper mantle structure.

Ridge–axis experiments. Designed to directly detect the physical manifestations of the melt generation region and the form of the accompanying mantle flow, these experiments will need to encompass an area on the order of 250 km square to achieve resolution of structure at depths up to 100 km. The crucial experiments are seismic tomography and monitoring of wave propagation from local and regional earthquakes to study the seismic velocity structure of the upper mantle, magnetotelluric experiments to study the electrical conductivity of mantle matrix and melt, and a detailed sampling program including both

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dredge and drilled samples to study the petrological variations that accompany changes in the deep physical structure revealed by seismological and electromagnetic techniques. To achieve maximum resolution of deeper structure, active electrical and seismic surveys will have to be executed so that the effect of variations in crustal structure on travel time delays and apparent resistivity can be removed. Measurements of the deformation of the sea floor using tiltmeters and precise pressure (depth) sensors could help reveal the deep plumbing of the magma system if discrete transport events were captured.

Theoretical, laboratory and numerical experiments. It is essential that there be a strong program of modelling, theoretical studies and laboratory tests to accompany the field experiments outlined above. Only by simulating the deeper structure in physical laboratory experiments or in computer experiments can the field experiments be properly interpreted. In addition, these artificial experiments can point out the critical measurements needed in field experiments to confirm or disprove a hypothesis about the nature of the magma generation and transport system.

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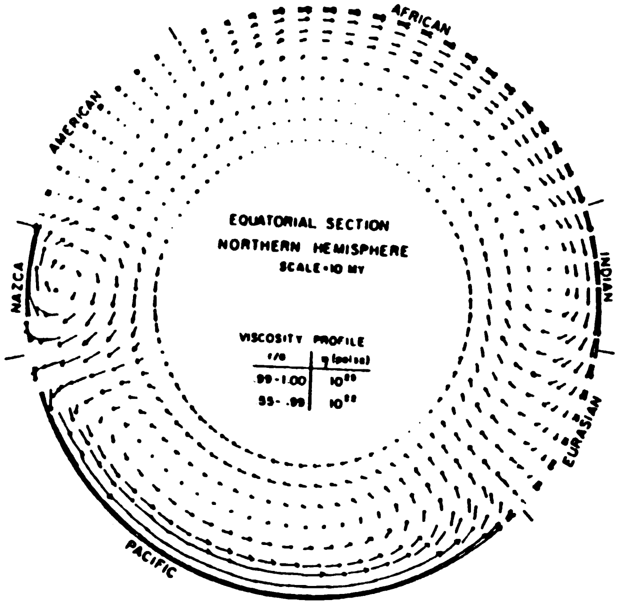
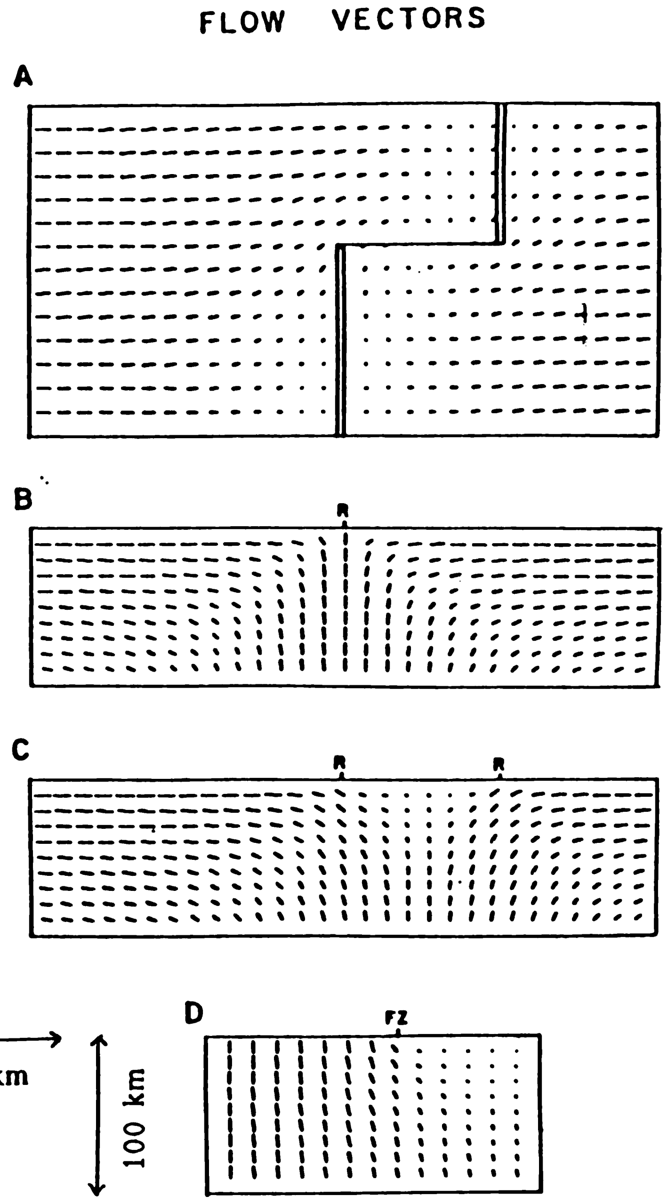


Figure 1. Extrapolated velocity vectors in the mantle projected on a great circle section through the earth for a model with uniform viscosity (from Hager and O'Connell, 1979). This section is an equatorial section viewed toward the northern hemisphere. Vectors are plotted at several depths at intervals of 5° and represent instantaneous velocities extrapolated for 10 m.y.



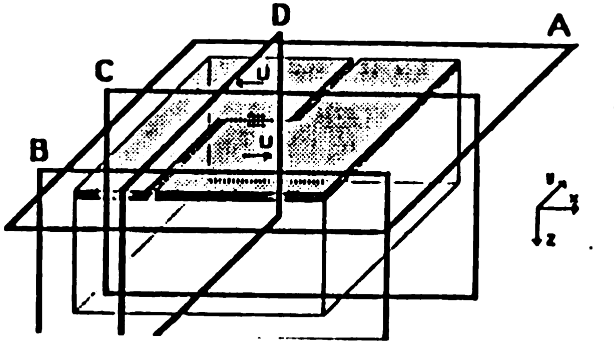


Figure 2. Flow vectors in the mantle in the vicinity of a transform offset between two spreading centers (Phipps Morgan and Forsyth, 1987). The orientation of the four sections on which the vectors are projected is shown in the inset.

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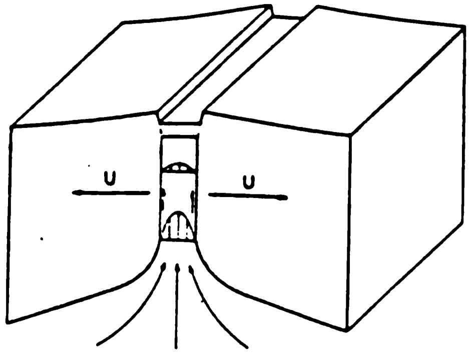

Figure 3. Ridge axis conduit model for the mechanical structure of slow spreading accreting plate boundary. The shaded region represents relatively undeformable or stiff mantle and the vertical extent of the conduit may be tens of km.

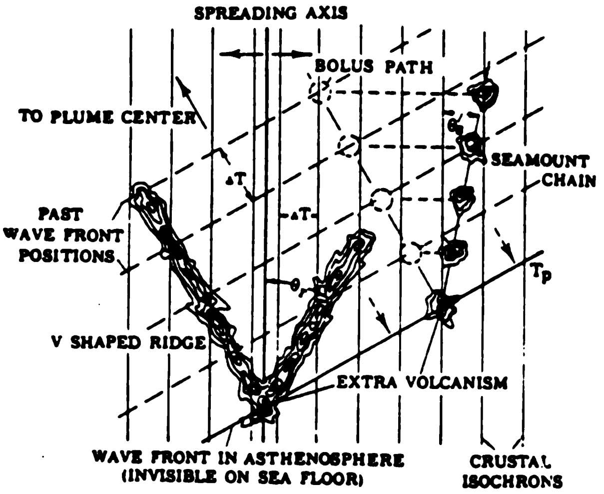

Figure 4a. Schematic model for the origin of secondary V-shaped topographic ridges on the sea floor south of Iceland (Vogt, 1971). The ridges and the seamount chain are caused by the passage of a basalt-rich front or bolus moving from upper left to lower right in the asthenosphere. South of Iceland, the fronts are thought to be generated by pulses of activity from the Iceland hot spot. The current position of the front is marked by the solid line labeled Tp. The angles θ_r and θ_s depend on the relative velocity of the front and the spreading velocity.

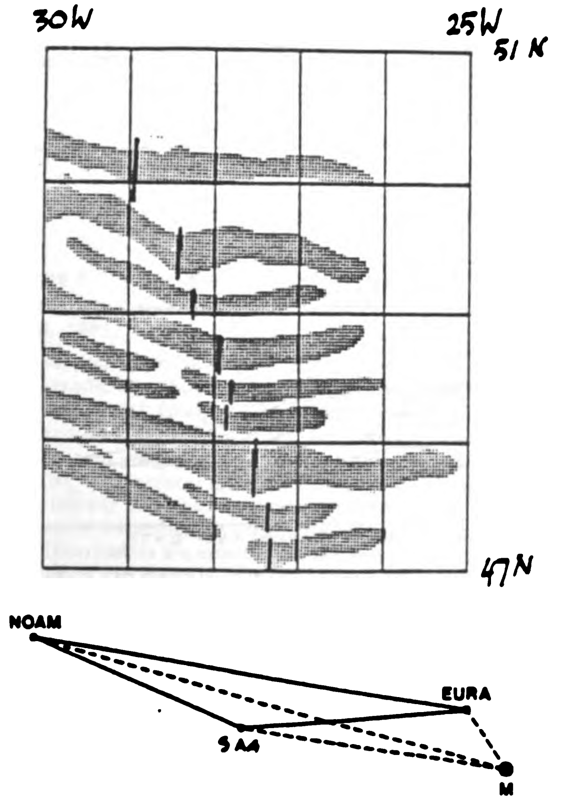

Figure 4b. Stippled areas show interpreted trends of topographic belts which clearly form broad V's about the spreading axis in the North Atlantic (after Johnson and Vogt, 1973). The belts can be traced to the present day segmentation of the ridge axis as indicated by the short, heavy lines. The trends can be predicted from the motions of the North American and European plates relative to the mesosphere (M) and the component of absolute motion of this two-plate system along the strike of the ridge (SAA) (Schouten et al. 1987)

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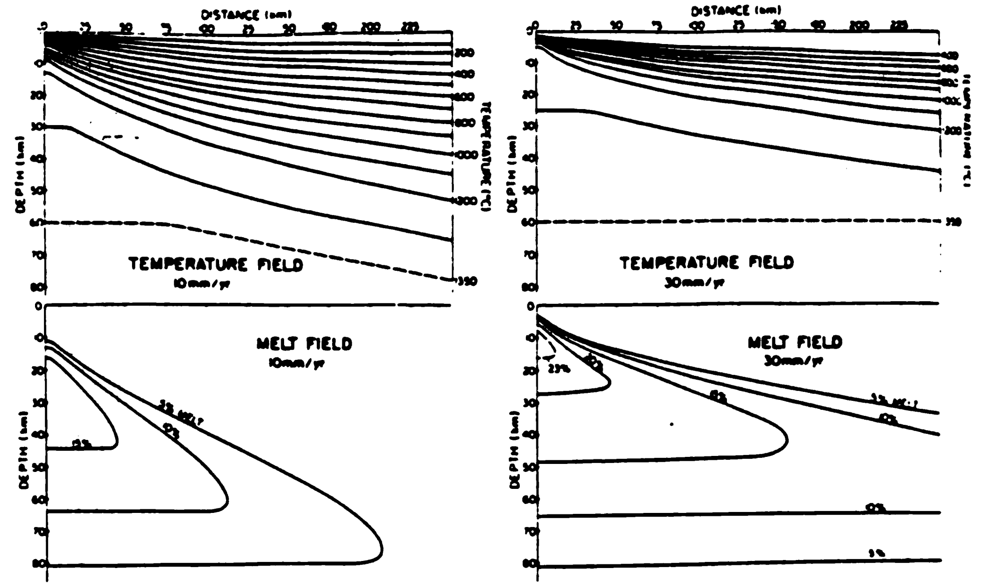

Figure 5. Isotherms and equilibriums isopleths (contours of equal melt fraction) for a cross-section of a spreading center assuming a flow model similar to that in cross-section B of figure 2 (Reid and Jackson, 1981). Spreading rates are assumed to be 10 mm/yr and 30 mm/yr in the two examples. Even though melt could exist at distances greater than 200 km from the axis in these models, very little melt is produced because the flow is predominantly horizontal at that distance.

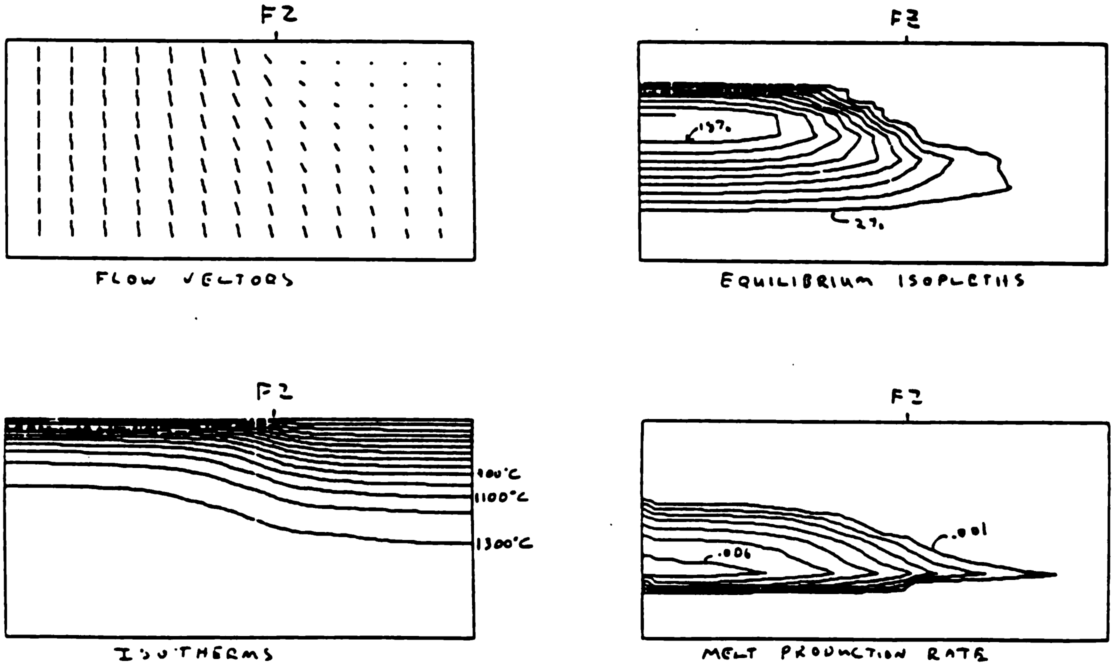

Figure 6. Flow vectors, isotherms, isopleths, and melt production rates for a cross-section along a ridge crest and across a fracture zone (section D in figure 2).

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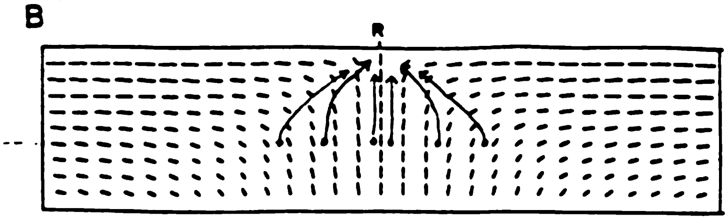


Figure 7. Schematic melt migration paths within the mantle beneath a ridge if the driving force for horizontal motion of the magma towards the ridge axis is the non-hydrostatic stress field generated by deformation of the matrix in the overall mantle flow.

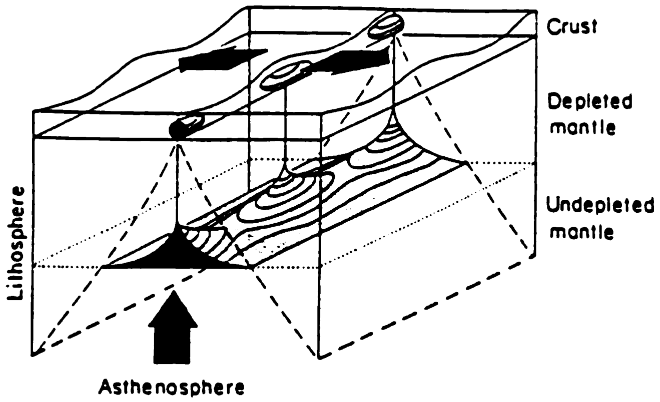
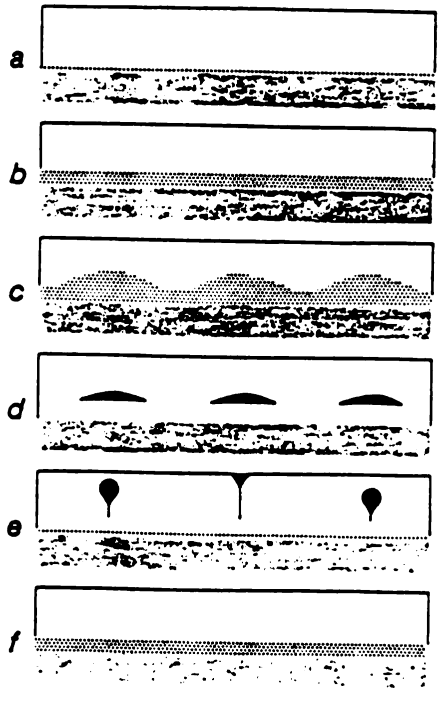


Figure 8. Schematic model for diapiric upwelling of magma from a mantle melt-accumulation zone (Whitehead et al. 1984). Above a certain level (dotted lines), the rising asthenosphere passes through a zone in which partial melt can form and collects in a layer below the lithosphere and depleted mantle layer. Due to its lower viscosity and density, the partial-melt zone develops a gravitational instability which leads to regularly spaced concentration of melt that rise to the surface to from crustal magma chambers.

Figure 9. Along-axis diagram of a mechanism that could be responsible for episodic production and supply of melt to crustal magma chambers(Schouten et al. 1985). Fine stipples, parental upper mantle; coarse stipples, partially molten mantle; no stipples, melt-depleted mantle; solid black, aggregated mantle melt (magma). A region of partially molten mantle begins to collect at the base of the depleted mantle in the asthenosphere beneath the spreading axis (*a*). As the volume of this region grows (*b*), the partially molten mantle, which has a lower density and viscosity than the surrounding asthenosphere, develops a gravitational instability with regularly spaced disturbances (*c*). Once the disturbances are established, the very low viscosity melt (magma), which can migrate by porous flow and moves more rapidly than the partially molten region as a whole, concentrates at the tops of the rising disturbances (*d*), after which it rises diapirically (*e*) towards the magma chambers in the oceanic crust above. The partially molten region is depleted of the melt (*d*) and the process will repeat itself as a new region of partially molten mantle begins to grow again at the base of the freshly depleted mantle (*e*, *f*).



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