Oceanic spreading center–hotspot interactions: Constraints from along-isochron bathymetric and gravity anomalies

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ABSTRACT

We analyzed bathymetric and gravity anomalies along present and paleo-axes of oceanic spreading centers influenced by the Iceland, Azores, Galápagos, Tristan, and Easter hotspots. Residual bathymetry (up to 4.7 km) and mantle Bouguer gravity (up to $-340$ km) anomalies are maximum at on-axis hotspots and decrease with increasing ridge-hotspot separation distance ($D$), until becoming insignificant at $D \sim 500$ km. Along-isochron widths of bathymetric anomalies (up to 2700 km) depend inversely on paleo–spreading rate, reflecting the extent to which plume material will flow along axis before being swept away by the spreading lithosphere. Flux balance arguments suggest that the five hotspots feed material to ridges with comparable fluxes of $\sim 2.2 \times 10^6$ km$^3$/m.y. Assuming that the amplitudes of these geophysical anomalies reflect temperature-dependent crustal thickness and mantle density variations, we suggest that ridge temperature anomalies are maximum (150–225 °C) when plumes are ridge centered and decrease with increasing ridge-hotspot distance due to cooling of the ridgeward-migrating plume material.

INTRODUCTION

When mantle plumes rise near oceanic spreading centers, they generate not only near-ridge hotspots, but also melt anomalies at the axis of the nearby ridges (e.g., Morgan, 1978). Direct evidence that near-ridge plumes divert toward and feed ridges is the ocean-island basalt (OIB) geochemical signature in ridge basalts (e.g., Hart et al., 1973). Furthermore, along-axis gradients in the strength of OIB signatures and in topography (e.g., Vogt, 1976; Schilling, 1991) indicate that once a plume reaches a ridge, it spreads laterally along axis.

Previous studies of ridge-plume interactions have focused primarily on present-day spreading centers. Ito and Lin (1995), however, demonstrated that 70%–75% of off-axis bathymetric and gravity anomalies of the Cocos plate can be attributed to the anomalous crustal thicknesses generated at the paleo-Galápagos ridge axis. We attributed long-wavelength (>200 km) variations in bathymetry and gravity along crustal isochrons to temperature conditions beneath the hotspot-influenced ridge axis at the time the crust was created.

In this study we investigated the evolution of five prominent plume-ridge systems over wide ranges in ridge-hotspot separation distance and spreading rate. The results of this study provide observational constraints on the amplitudes of along-isochron bathymetric and gravity anomalies as they depend on ridge-hotspot separation distance, and along-isochron widths of bathymetric anomalies as they depend on ridge spreading rate.

ALONG-ISOCRON BATHYMETRIC AND GRAVITY ANOMALIES

Iceland, Azores, Tristan, Galápagos, and Easter (Fig. 1) are the five hotspots that impose the most prominent bathymetric and geo-
chemical anomalies observed at nearby oceanic spreading centers (Hart et al., 1973; Hamelin et al., 1984; Schilling, 1985). Encompassing each of the five systems, we obtained shipboard bathymetric data from the National Geophysical Data Center (NGDC) and Lamont-Doherty Earth Observatory (LDEO), and gridded bathymetry from NGDC. To derive residual bathymetry, we first corrected the raw data for isostatic effects of sediment loading and then subtracted predicted depths of a cooling mantle half space (Carlson and Johnson, 1994). Sediment thicknesses were obtained from the LDEO database (A. Cazenave, Centre National d’Etudes Spatiales, Toulouse, France), and density contrasts between the sediments and mantle, and mantle and water were assumed to be 1600 kg/m³ and 2300 kg/m³, respectively.

Free-air gravity data were taken from the ship surveys and the satellite altimetry-derived gravity grid of Sandwell and Smith (1992). To isolate the effects of sub–sea-floor density structure, we generated mantle Bouguer anomalies by subtracting from the free-air gravity the attractions of the sea-floor–water (density contrast, $\Delta \rho = 1800$ kg/m³) and crust-mantle ($\Delta \rho = 500$ kg/m³) interfaces using raw bathymetry, and assuming a crust of uniform thickness (6.5 km) (e.g., Kuo and Forsyth, 1988).

Coordinates of present-day ridge axes and crustal isochrons were defined by using plate boundary and age data of Müller et al. (1993a). Because our focus was on anomalies generated at the axes of spreading centers, we considered only data from sea floor unaffected by off-axis volcanism, as detailed in the Figure 1 caption. From our residual bathymetry and mantle Bouguer grids, we then extracted along-isochron profiles (Fig. 2).

**ANOMALY AMPLITUDES VS. PALEORIDGE-HOTSPOT DISTANCE**

To determine hotspot locations relative to paleo–spreading centers, we assumed that the hotspots were stationary with respect to each other and used plate-reconstruction poles (Lonsdale 1988; Müller et al., 1993b) to rotate isochrons with respect to the hotspots back to their positions at the time of accretion. We then measured distances between the paleo–ridge axes and hotspot centers, which we took to be the locations of most recent volcanism.

The along-isochron variations in residual bathymetric ($\Delta RB$) and mantle Bouguer anomalies ($\Delta MBA$) display a decrease with increasing paleoridge-hotspot distance ($D$, Fig. 3). The on-ridge hot spot cases ($D \leq 50$ km) for the Tristan system (80–90 Ma isochrons) and the Iceland system (0–30 Ma isochrons) display the highest $\Delta RB$ (3.5–4.7 km) and most negative $\Delta MBA$ (−250 to −340 mgal), which are approximately twice those of ridge-centered cases for the Galápagos and Azores systems. At $D \sim 500$ km, the hotspot signals become very weak and in the case of Tristan, become indistinguishable from normal ridge-segmentation-related variations. The individual Galápagos and Tristan systems show a decrease in $\Delta RB$ and $\Delta MBA$ with increasing $D$, whereas the Azores system is more complex and the Easter trend is very weak. The predominant decrease of $\Delta RB$ and $\Delta MBA$ with increasing $D$ is consistent with Schilling’s (1985) study of present-day ridge-axis bathymetry.

**ANOMALY WIDTHS VS. PALEO–SPREADING RATE**

Whereas amplitudes of $\Delta RB$ and $\Delta MBA$ are functions of ridge-hotspot distance, along-isochron widths ($W$) of the bathymetric anomalies (see Fig. 2 caption) depend primarily on the full spreading rate ($U$) at the time of crustal accretion. The maximum values of $W$ are found along the slowest-spreading Mid-Atlantic Ridge near Iceland (2700 km, Fig. 4A); these values are comparable to the along-axis extent of the helium isotope anomaly, but are a factor of two greater than the widths of rare-earth-element anomalies (Schilling, 1986). Values of $W$ decrease with increasing $U$ to a minimum along the fast-spreading East Pacific Rise.

The observed dependence of $W$ on spreading rate lends strong support to previous notions of along-axis plume material flow (Vogt, 1976; Schilling, 1985). Similarly to Schilling (1991), we estimate that the net flux of plume material feeding the ridge ($Q$) is eventually carried away by the spreading lithospheric plates (Fig. 4A, inset), such that

$$Q = \int_{-W/2}^{W/2} P(y) h y dy = \frac{h U W}{2},$$

where $y$ is the along-axis coordinate, $h$ is the thickness of the fully developed lithosphere (assumed to be 80 km), and $P(y)$ is the percentage of accreted lithosphere derived from the plume material assumed to decrease linearly from 1 at $y = 0$ to 0 at $y = \pm W/2$.

We treat the hotspot to ridge flow as a simple laminar flow problem in which the lithospheric drag opposes the ridgeward flow of plume material. The channel connecting the ridge and hotspot has a characteristic width $w_3$ and thickness $w_2$ (see Fig. 4, A and B, insets). Therefore, the net flux from the hotspot to the ridge is

$$Q = w_3 w_2 V \left( \frac{U}{4} \right),$$

where $V$ is the average ridgeward velocity of plume flow, $w_3 w_2 V$ is the ridgeward flux, and $w_3 w_2 / 4$ is the opposing plate-driven flux. Combining equations 1 and 2 yields the dependence of $W$ on spreading rate,

$$W = \left( \frac{2w_3 w_2}{h U} \right) \left( \frac{V}{4} - \frac{U}{4} \right).$$

The solid curve in Figure 4A is that predicted for assumed values of $V = 70$ km/m.y. and $w_3 w_2 = 3 \times 10^8$ km², which yields a root-mean-square misfit to the data of 500 km. Similar misfits are achieved for $V = 30–100$ km/m.y. and corresponding values of $w_3 w_2$ of $8–2 \times 10^4$ km². These results suggest that the ridgeward fluxes from the five
hotspots are comparable, the average value being $2.2 \times 10^6$ km$^3$/m.y. Increasing or decreasing $w_1w_2V$ by $1 \times 10^6$ km$^3$/m.y. increases the data misfit by a factor of two.

Our theoretical relation between $W$ and $U$ is based on one end-member scenario in which lateral spreading of plume material beneath ridges is strictly ridge parallel. A numerical study that considers both ridge-parallel and ridge-perpendicular spreading of plumes beneath ridges (Feighner et al., 1995) may represent the other end member; it thus predicts that $W$ is proportional to $(Q/U)^{1/2}$ rather than $(Q/U)$, as does our model.

PALEO–RIDGE-AXIS TEMPERATURE ANOMALIES

We show here how the amplitudes of $\Delta RB$ and $\Delta MBA$ may reflect the temperature anomalies beneath the paleo–ridge axes. We assume that $\Delta RB$ and $\Delta MBA$ arise from crustal-thickness and mantle-density anomalies, both of which depend on the ridge-axis temperature anomaly at the time of crustal accretion. $\Delta RB$ and $\Delta MBA$ can be related to a hotspot-induced mantle temperature anomaly ($\Delta T$) using the model of Ito and Lin (1995), which considered changes in mantle density by thermal expansion, and in crustal thickness by increased decompression melting. Assuming passive mantle upwelling, Ito and Lin (1995) imposed temperature anomalies below the melting zone and then combined the effects of crustal-thickness and mantle-density variations to yield theoretical isostatic bathymetric variations and $\Delta MBA$.

Applying this method for ranges of imposed temperature anomalies and model spreading rates, we derive the empirical relations:

$$\Delta T = (0.11U + 35.3)\Delta RB,$$  \hspace{1cm} (4)

and

$$\Delta T = -(0.0017U + 0.45)\Delta MBA.$$  \hspace{1cm} (5)

The dependence of $\Delta T$ on $U$ reflects a subtle dependence of crustal thickness on spreading rate that is consistent with calculations of Su et al. (1994). For $\Delta T = 100$ °C and $U = 20–100$ km/m.y., for example, we predict corresponding values of crustal thickening of 9–4.5 km.

Temperature anomalies derived according to the observed $\Delta RB$ and $\Delta MBA$ are maximum for the on-ridge cases (150–225 °C), and decrease to near zero for $D \simeq 500$ km (Fig. 4B). Such a behavior can be interpreted as the cooling trend of plume material as it migrates from hotspot centers to nearby ridges, the ridge-centered cases reflecting the temperature anomaly of the hotspot itself.

As plume material migrates from a hotspot center to a ridge, it conducts heat to the surrounding mantle (see Fig. 4, A and B, insets). Assuming that the amount of heat conducted in the direction of plume flow is negligible, the heat balance equation is...
\[
\frac{\partial T}{\partial t} + V \cdot \nabla T = - \left( \frac{\partial q_p}{\partial y} + \frac{\partial q_l}{\partial z} \right),
\]

where \( T \) is the average temperature of the plume conduit, \( \rho \) and \( c_p \) are the density and heat capacity of the plume material, respectively, and \( q_p \) and \( q_l \) are the components of conductive heat flow out of the conduit walls. If we assume that heat loss occurs through a thermal boundary layer surrounding the plume channel with thickness \( \delta \),

\[
q_p \sim q_l \sim \frac{k \alpha T}{\delta},
\]

where \( k \) is the mantle thermal conductivity (3 W m\(^{-1}\) °C\(^{-1}\) and \( \delta \) is defined such that \( q = 30-100 \) mW/m\(^3\), comparable to heat-flux values on the sea floor. If it is assumed that the gradients \( \alpha \partial q_{p}/\partial y \) and \( \alpha \partial q_{l}/\partial z \) are proportional to 1/\( w_{p} \) and 1/\( w_{l} \), respectively, and \( Q = w_{p}^{2}V \) (i.e., \( V \ll U/4 \)), the combination of equations 6 and 7 yields

\[
\frac{\partial T}{\partial x} = -\frac{2k}{Q_{6}} (w_{1} + w_{2}) \Delta T,
\]

where \( \kappa = k/\rho c_{p} \) is thermal diffusivity (10\(^{-6}\) m\(^2\)/s). Integrating with respect to \( x \) from 0 to \( D \) yields

\[
\Delta T = \Delta T_{0} \exp \left[ \frac{2k}{Q_{6}} (w_{1} + w_{2}) D \right],
\]

where \( \Delta T_{0} \) is the temperature anomaly at the hotspot center.

Taking \( \Delta T_{0} = 100 \) °C, \( (w_{1} + w_{2}) = 400 \) km, and \( Q = 2.2 \times 10^{6} \) km\(^3\)/m.y., as observed with the consistent \( W \) vs. \( U \) trend above, we produce a theoretical curve (Fig. 4B) that effectively matches the inferred temperature anomalies for \( D > 50 \) km. For \( D \ll 100 \) km, the Iceland and Tristan points lie significantly higher than the theoretical curve. This mismatch may be because (1) the Iceland and early Tristan plumes are hotter than the other hotspots and/or (2) latent heat loss due to melting at the hotspot centers rapidly cools the plume before it migrates to nearby ridges in the off-axis cases. For \( D \sim 500 \) km, \( \Delta T \) is small enough that its effects on \( \Delta RB \) and \( \Delta MBA \) are negligible, even though the plume may still be feeding the ridge. Consequently, the geochemical signal may persist to a ridge-hotspot distance of up to 850 km (Schilling et al., 1985), long after the signals in \( \Delta RB \) and \( \Delta MBA \) have disappeared.

CONCLUSIONS

Along-isochron variations in residual bathymetry and mantle Bouger gravity reflect the influence of hotspots on paleoaxes of nearby spreading centers. The amplitudes of along-isochron anomalies for the five prominent plume–ridge systems reach a maximum of 47 km for \( \Delta RB \) and \( \Delta MBA \), with 340 mgal for \( \Delta MBA \) and decrease with increasing paleo-ridge-hotspot distance. The along-isochron widths (0–2700 km), however, depend inversely on paleo-spread rate. Whereas the widths of \( \Delta RB \) reflect the balance between ridgeward plume flux and lithospheric accretion, the amplitudes of \( \Delta RB \) and \( \Delta MBA \) reflect paleoaxial temperature anomalies that decrease as the plume material cools along its lateral migration to nearby ridges. The five hotspots appear to deliver material to ridges with comparable fluxes of \( \sim 2.2 \times 10^{6} \) km\(^3\)/m.y. and produce excess mantle temperature anomalies of 50 to 225 °C that influence ridge-axis structure to a maximum ridge-hotspot distance of \( \sim 500 \) km.

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