Mechanisms of geochemical and geophysical variations along the western Galápagos Spreading Center

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[1] Improved insights into the processes of hot spot–ridge interaction along the Galápagos Spreading Center (GSC) are revealed by geochemical data between ~91°W and 98°W. Principal components analysis reveals that >87% of the total isotopic variability can be explained with only two mantle source components. The “depleted” component has lower ratios of highly to moderately incompatible elements, higher Nd isotopic ratios, and lower Sr and Pb isotopic ratios. The second component is relatively enriched in incompatible elements, has more radiogenic Pb and Sr and less radiogenic Nd, and is comparable to the C or “common” mantle component observed at many locations in the Pacific. The enriched component’s signature is strongest nearest the hot spot at ~92°W and diminishes with distance from the hot spot to 95.5°W. Near 95.5°W, lava compositions change sharply, becoming dominated by the depleted component and remaining so farther west, to 98°W. Thus, the Galápagos hot spot most clearly influences the composition of the GSC between 91°W and 95.5°W. The depleted component between 91°W and 98°W differs from that evident at the Galápagos Archipelago, along the GSC east of 91°W, and along the East Pacific Rise. This suggests some form of compositional zoning in the regional mantle. If the depleted materials are intrinsic to the Galápagos mantle plume, then the plume is bilaterally zoned and feeds a depleted component to the GSC at 91°W–98°W that is distinct from the depleted material elsewhere in the region. This possibility is supported by melting models in which the Galápagos plume is a uniform mixture of a depleted matrix and fine-scale enriched veins. The expression of the more fusible veins is predicted to be enhanced nearest the hot spot (~92°W), where plume-like upwelling drives rapid flow and melting deeper in the melting zone (where the veins are melting). With increasing westward distance from the hot spot, the deep, plume-driven flow is predicted to decrease, as does the expression of the enriched veins in lava compositions. The model therefore adequately explains the compositional and crustal variations from 92°W to 95.5°W. The average model composition of the plume in this region does not differ significantly from that of the ambient mantle beneath other ridges not influenced by hot spots.
1. Introduction

[2] Many hot spots are located near mid-ocean ridges. Hot spots are commonly thought to derive from volatile-rich [e.g., Bonatti, 1990; Asimow and Langmuir, 2003], chemically fertile [e.g., Sobolev et al., 2007], and/or unusually hot material rising as a “plume” from the deep mantle [e.g., Morgan, 1971]. Plumes have been inferred to feed material laterally to nearby ridges [e.g., Morgan, 1978] and deliver material along the ridge axis over distances as great as 1000 km [e.g., Vogt, 1976; Schilling et al., 1976]. The surface expression of plume-ridge interaction is recognized by a bathymetric swell and thickened crust, which may be attributable to unusually hot or fertile mantle from the plume [e.g., Ito and Lin, 1995; Canales et al., 2002]. The geochemistry of rocks formed along plume-affected ridge segments also differs from those of “normal” ridge segments (i.e., those that are far from and not influenced by hot spots). Elemental and radiogenic isotope ratios vary considerably along axis, with compositions ranging from those of normal mid-ocean ridge basalts (MORB) to those more common among ocean island basalts (OIB) [e.g., Schilling, 1991; Blichtert-Toft et al., 2005; Nicolaysen et al., 2007]. Although many factors can control these geochemical and geophysical variations, the relative roles of mantle flow, magma flux, and mantle heterogeneity remain poorly understood [e.g., Ito et al., 2003; Hall and Kincaid, 2003, 2004].

[3] The Galápagos Spreading Center (GSC), in the equatorial eastern Pacific Ocean, is a classic example of off-axis plume-ridge interaction. In this paper, we present trace element and Sr, Nd and Pb isotope data for GSC rocks collected during the G-PRIME experiment in 2000 [Detrick et al., 2002]. These data are combined with previously published major element data [Cushman et al., 2004] and geophysical observations. G-PRIME sample spacing was 7–10 km, allowing us to resolve shorter-wavelength fluctuations than previously detected. We use the data and observations to examine variations in extents of melting and fractional crystallization as well as the mantle composition along the ridge. Finally, we investigate the possible upper mantle dynamics that can explain the geochemical and geophysical variations.

2. Geological Setting

[4] The east-west trending GSC separates the Cocos Plate (north) from the Nazca Plate (south) and intersects the East Pacific Rise to the west (Figure 1). The Galápagos hot spot is inferred to be centered at the western edge of the Galápagos Archipelago, some ~200 km to the south of the GSC near ~92°W [e.g., White et al., 1993]. The GSC initiated around 23 Ma with the breakup of the Farallon Plate along a preexisting fracture zone near the Galápagos hot spot [e.g., Hey, 1977; Barckhausen et al., 2001; Werner et al., 2003]. For the past several million years, the GSC has been migrating away from the Galápagos hot spot [e.g., Hey et al., 1980; Wilson and Hey, 1995]. The hot spot has impacted the geometry of the plate boundary [Hey et al., 1989], and is associated with several propagating ridge segments, including two on the eastern GSC (EGSC, east of the 91°W transform fault) and two on the western GSC (WGSC, west of 91°W) with propagating rift tips currently at 93.3°W and 95.5°W. The hot spot has also influenced the morphology of the axial topography [Detrick et al., 2002; Sinton et al., 2003; Werner et al., 2003; White et al., 2008], the thickness of the crust [Canales et al., 2002], and the geochemistry of the crust along the GSC [Schilling et al., 1982; Verma et al., 1983; Detrick et al., 1982; White et al., 2008].
et al., 2002; Schilling et al., 2003; Cushman et al., 2004; Christie et al., 2005]. Together, the above indicators reveal that the hot spot influence is largest near the hot spot (C24/C176W) and decreases with distance east and west away from the hot spot.

3. Previous Geochemical Studies

The foundational geochemical work in the area is by Schilling et al. [1976, 1980, 1982], Verma and Schilling [1982], and Verma et al. [1983] and, e.g., Christie and Sinton [1981, 1986]. Recently, Schilling et al. [2003] presented radiogenic isotope data for the entire length of the GSC (83°W–101°W). Cushman et al. [2004] published major element and volatile (H2O and CO2) data for a detailed sampling of the WGSC between ~91°W and 98°W, and Christie et al. [2005] published major and trace element data for the EGSC (85.5°W–91°W). Rotella et al. [2009] used elemental and isotopic data to investigate magmatic processes near the 93.3°W overlapping spreading center. Clear geographic trends emerge from these studies. Nearest the hot spot (~92°W–90.5°W), concentrations of the most incompatible trace elements and ratios of these elements to moderately incompatible elements are highest, 87Sr/86Sr and 206Pb/204Pb are elevated, and 143Nd/144Nd and 176Hf/177Hf are lowest. These findings indicate that near the hot spot, where the crust is thickest, the GSC most extensively samples one or more mantle sources with a long-term enrichment in the more incompatible elements. The elemental concentrations in this area were thought to reveal relatively low (not high) extents of melting due to enhanced hydrous melting in this near–hot spot area [Asimow and Langmuir, 2003; Cushman et al., 2004]. The above expressions of the enriched source material and hydrous melting decrease with crustal thickness and distance, east and west, away from the hot spot. The origin of the along-axis variations,
constrained by additional trace element and isotopic data, is a central topic of this study.

4. Methods

[6] Detrick et al. [2002] and Cushman et al. [2004] reported the first geochemical results of the G-PRIME experiment, which sampled most of the WGSC at a higher spatial resolution than in previous studies. Major and minor element analyses of glasses were obtained by electron microprobe and H2O and CO2 analyses by Fourier transform infrared spectroscopy (FTIR). Using the same samples, we present new data obtained at the University of Hawaii for a suite of trace elements, and Sr, Nd, and Pb isotope compositions.

4.1. Elemental Analyses

[7] Chlorine, fluorine, and sulfur data (Table 1) were acquired on glass chips using a Cameca SX50 electron microprobe following Thordarson et al. [1996]. Average reproducibility is ±2% for S, ±10% for Cl and ±20% for F.

[8] Data for several trace elements were acquired on a Siemens 303AS XRF (X-ray fluorescence) spectrometry system using a Rh target, end window X-ray tube. Whole rocks were crushed in an alumina swing mill, and powders were analyzed for major elements on fused disks following methods similar to those of Norrish and Chappell [1977]. Trace elements were analyzed on pressed powder pellets (Table 1). Peak intensities for the trace elements were corrected for backgrounds, line interferences and matrix absorption using methods similar to those of Chappell [1991]. Corrected intensities were calibrated against a wide range of natural rock standards. Accuracy and precision data for this system are reported by Sinton et al. [2005].

[9] When available, unaltered glass chips were selected for analysis of additional incompatible trace elements and data were obtained using a VG PQ-2S inductively coupled plasma source mass spectrometer (ICP-MS) following procedures of Neal [2001]. Reproducibility is generally better than 5% at levels above 0.5 ppm (Table 1). An indication of accuracy is given by measured (an average of 5 analyses) and recommended [see Govindaraju, 1994] values for BHVO-1 analyzed as an unknown.

4.2. Radiogenic Isotopic Compositions (Sr, Nd, and Pb) and Pb Concentrations

[10] Preparation and mass spectrometry for Sr, Nd, and Pb isotope analyses followed Mahoney et al. [1998] and Sheeh et al. [2003]. Measurements were made on a multicollector VG Sector thermal ionization mass spectrometer (TIMS). Fractionation corrections are 0.1194 for 87Sr/86Sr and 0.242436 for 142Nd/144NdO (i.e., Nd was run as an oxide). Ratios are reported relative to measured 87Sr/86Sr = 0.710238 ± 0.000014 (2σ, n = 29) for NBS987 Sr and to 143Nd/144Nd = 0.511850 ± 0.000008 (measured at 0.511843; 2σ, n = 25; this uncertainty corresponds to ±0.2 εNd units) for La Jolla Nd. Lead isotope ratios for most samples were measured with a double-spike method [Galer, 1999]. Most of these measurements were made in multistatic mode on 5–10 ng loads of sample; mean values (±2σ) obtained for 33 analyses on 5 to 10 ng loads of NBS 981 Pb were 16.938 ± 0.003, 15.494 ± 0.004, and 36.712 ± 0.008 for 206Pb/204Pb, 207Pb/204Pb, and 208Pb/204Pb, respectively. The remainder of the double-spike measurements were made in multidynamic mode on larger sample loads (15–40 ng); mean values (±2σ) obtained for 22 analyses on 30–40 ng loads of NBS 981 Pb were 16.9406 ± 0.0008, 15.4974 ± 0.0010, and 36.7216 ± 0.0022, for 206Pb/204Pb, 207Pb/204Pb, and 208Pb/204Pb, respectively. For a set of samples from the 95.5°W propagating rift, Pb isotope ratios are conventional (non-double-spike), multistatic measurements made on 5–10 ng loads. These conventional data are reported relative to the NBS 981 Pb values of Todt et al. [1996]; the 2σ range for conventional measurements of NBS 981 Pb (n = 31) was ± 0.010 for 206Pb/204Pb, 0.011 for 207Pb/204Pb, and 0.032 for 208Pb/204Pb. Within-run errors on the isotopic data in Table 2 are less than or equal to the external uncertainties on these standards, except for the dynamic mode Pb isotope analyses; for those analyses, within-run errors are listed adjacent to the corresponding isotope ratios in Table 2. Concentrations of Pb were determined by isotope dilution; uncertainties are less than 0.5%. Total procedural blanks are <40 pg for Pb, <35 pg for Sr, and <12 pg for Nd. Note that in Figures 5, 6, 12, 13, and 16, 143Nd/144Nd is presented as εNd as defined in Table 2.

5. Results

[11] Cushman et al. [2004] defined three main MORB groups, based on K/Ti, along the WGSC:
N- (normal) MORB have K/Ti < 0.09, T- (transitional) MORB have K/Ti between 0.09 and 0.15, and E- (enriched) MORB have K/Ti > 0.15. We use these definitions in our study as well. The three main geochemical groups roughly coincide with the three main physiographic provinces of the WGSC: E-MORB in the eastern axial high province (91°W–92.7°W), T-MORB in the middle section...
Table 2. Radiogenic Isotope Compositions for Samples From the Western GSC

<table>
<thead>
<tr>
<th>Sample</th>
<th>Longitude (°W)</th>
<th>$\varepsilon_{\text{Nd}}^a$</th>
<th>$^{143}\text{Nd}/^{144}\text{Nd}$</th>
<th>$^{87}\text{Sr}/^{86}\text{Sr}$</th>
<th>$^{206}\text{Pb}/^{204}\text{Pb}^b$</th>
<th>$^{207}\text{Pb}/^{204}\text{Pb}^b$</th>
<th>$^{208}\text{Pb}/^{204}\text{Pb}^b$</th>
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</thead>
<tbody>
<tr>
<td>7D-4</td>
<td>91.28</td>
<td>7.3</td>
<td>0.513014</td>
<td>0.70287</td>
<td>18.996</td>
<td>15.574</td>
<td>38.697</td>
</tr>
<tr>
<td>10D-1</td>
<td>91.36</td>
<td>7.1</td>
<td>0.513003</td>
<td>0.70293</td>
<td>19.0678 (20)</td>
<td>15.5877 (24)</td>
<td>38.7839 (42)</td>
</tr>
<tr>
<td>11D-1</td>
<td>91.41</td>
<td>6.9</td>
<td>0.512991</td>
<td>0.70295</td>
<td>19.1119 (47)</td>
<td>15.5919 (54)</td>
<td>38.8357 (96)</td>
</tr>
<tr>
<td>12D-5</td>
<td>91.49</td>
<td>7.2</td>
<td>0.513009</td>
<td>0.70294</td>
<td>19.0729 (30)</td>
<td>15.5655 (36)</td>
<td>38.8326 (62)</td>
</tr>
<tr>
<td>16D-2</td>
<td>91.75</td>
<td>7.1</td>
<td>0.513002</td>
<td>0.70292</td>
<td>19.005</td>
<td>15.590</td>
<td>38.805</td>
</tr>
<tr>
<td>17D-4</td>
<td>91.80</td>
<td>6.8</td>
<td>0.512988</td>
<td>0.70296</td>
<td>19.0225 (23)</td>
<td>15.5921 (28)</td>
<td>38.7966 (47)</td>
</tr>
<tr>
<td>19D-1</td>
<td>91.96</td>
<td>6.5</td>
<td>0.512971</td>
<td>0.70300</td>
<td>18.9766 (30)</td>
<td>15.5911 (37)</td>
<td>38.8332 (62)</td>
</tr>
<tr>
<td>20D-1</td>
<td>92.01</td>
<td>6.5</td>
<td>0.512971</td>
<td>0.70302</td>
<td>19.125</td>
<td>15.608</td>
<td>38.938</td>
</tr>
<tr>
<td>25D-1</td>
<td>92.32</td>
<td>7.0</td>
<td>0.512997</td>
<td>0.70301</td>
<td>18.9594 (16)</td>
<td>15.5852 (20)</td>
<td>38.7772 (34)</td>
</tr>
<tr>
<td>25D-1c</td>
<td>92.32</td>
<td>6.8</td>
<td>0.512987</td>
<td>0.70301</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>28D-1</td>
<td>92.52</td>
<td>6.9</td>
<td>0.512992</td>
<td>0.70299</td>
<td>18.987</td>
<td>15.589</td>
<td>38.820</td>
</tr>
<tr>
<td>28D-1c</td>
<td>92.52</td>
<td>6.8</td>
<td>0.512990</td>
<td>0.70305</td>
<td>18.987</td>
<td>15.589</td>
<td>38.820</td>
</tr>
</tbody>
</table>

$^a\varepsilon_{\text{Nd}}$ is calculated as the difference, in parts per 10,000, of the measured sample $^{143}\text{Nd}/^{144}\text{Nd}$ from a present-day chondritic average (chond) value of 0.51264; that is, $\varepsilon_{\text{Nd}} = (^{143}\text{Nd}/^{144}\text{Nd}_{\text{sample}} - ^{143}\text{Nd}/^{144}\text{Nd}_{\text{chond}}) \times 10^3$.

$^b$For Pb isotope compositions determined by the double spike method, values listed to the third decimal place were determined in static mode on 5–10 ng loads; those listed to the fourth decimal place were determined in dynamic mode on 15–40 ng loads (individual within-run errors are given in parentheses adjacent to each measurement and apply to the last two digits).

$^c$Second dissolution of a separate glass split (i.e., full procedural duplicates) for 25D-1, 28D-1, and 73D-1 and, for 80D-2, a second dissolution of a holocrystalline part of the same sample.

$^d$Pb isotopes on these samples were determined by conventional (non-double-spike) multistatic measurements on 5–10 ng loads.
(92.7°W–95.5°W) with transitional axial morphology, and N-MORB in the western axial valley province (95.5°W–98°W) [Detrick et al., 2002; Sinton et al., 2003]. There are, however, a few exceptions. For example, N-MORB are sporadically present east of 95.5°W, T-MORB are present near 91°W, and one E-MORB was recovered near 97°W [Cushman et al., 2004].

5.1. Trace Elements

Concentrations of the most incompatible trace elements (e.g., Rb, U, Nb, La) in the E-MORB are generally elevated above those in T- and N-MORB (Figure 2). The middle rare earth elements (REE) are also slightly more abundant in E-MORB, although there is some overlap with the T-MORB. In contrast, heavy REE concentrations are similar for E- and T-MORB, and are only slightly elevated over those of the N-MORB. These characteristics give the E-MORB patterns a negative slope (e.g., normalized Sm/Yb ≥ 1) from the middle to heavy REE in primitive mantle–normalized plots. In summary, from east to west, abundances of incompatible trace elements generally decrease, with some overlap among the three MORB groups; however, the amount of overlap increases with decreasing elemental incompatibility.

Concentrations of H₂O, Cl, and F are also elevated in the E-MORB relative to the T- and N-MORB, but S does not vary as systematically (Figures 3 and 4). The element best correlated with H₂O and F is Ce (r² = 0.988 and r² = 0.983, respectively); Cl and S are weakly correlated with La and Dy (r² = 0.768 and r² = 0.684, respectively).

5.2. Radiogenic Isotopes

The radiogenic ratios correlate strongly with each other (r² values generally >0.8), forming simple linear arrays in isotope space (Figure 5). Also, the three MORB groups defined by K/Ti generally occupy distinct portions of the isotope arrays (Figure 5). A few exceptions include sample D2-1, dredged from the transform fault at 90.9°W, and D73-1 from near the offset at 95.5°W. Sample D2-1 is an N-MORB in terms of K/Ti but has isotopic ratios more similar to those of the E-MORB sampled on the axis nearby. Sample D73-1 is anomalous in many respects. It is a T-MORB in terms of K/Ti, but its incompatible trace element pattern lies well within the N-MORB field. It is one of several high-Al₂O₃ MORB from the GSC that Eason and Sinton [2006] interpret to have been affected by high-pressure fractionation in the upper mantle. It has the most radiogenic ⁸⁷Sr/⁸⁶Sr and least radiogenic εNd of all samples from the entire WGSC. It also has the lowest ²⁰⁶Pb/²⁰⁴Pb of any of the WGSC samples, similar to values of N-MORB nearby, whereas its ²⁰⁷Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb lie above the trend formed by the N-MORB in Figures 5c and 5d.

Along axis, ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb become less radiogenic westward, away from the hot spot (Figure 6). εNd and ⁸⁷Sr/⁸⁶Sr also vary between 91°W and the propagating ridge segment at 95.5°W, but more subtly than the Pb isotopes do. Between the propagating rift tip near 95.5°W and ~96.5°W, all isotopic ratios change suddenly: εNd rises and Sr and Pb isotopic ratios drop by 30%–50% of the total variation seen in our data. Considerable variation in trace element and isotopic ratios also characterizes samples near the 93.3°W ridge offset [Rotella et al., 2009], but the variation is smaller than that near 95.5°W and a stepwise change in composition is not evident. Between 98°W and 96.5°W, isotopic compositions remain nearly uniform.

6. Factors Determining Trace Element and Isotope Compositions in GSC Rocks

As is typical in studies of oceanic basalts, it is essential to first address the influences of crustal processes on lava compositions, including extents of fractional crystallization, assimilation of wall rock by magmas, and post-eruptive alteration of lavas by seawater, before we can gain the understanding we seek on mantle processes, namely, extent and depth of partial melting and composition of the mantle sources.

6.1. Crustal Processes

6.1.1. Assimilation and Alteration

Both assimilation and alteration can mask or overprint primary geochemical signatures of oceanic rocks. Alteration effects have been minimized by analyzing only fresh glass chips (see section 4) for mobile elements and isotopes. Some trace elements are useful indicators of both processes, as interactions between magmas or lavas and the surrounding oceanic crust and seawater create variations in mobile elements that do not covary regularly with primary magmatic indicators. For example, assimilation of hydrated minerals tends to increase concentrations of Cl relative to elements
having similar incompatibility, such as La (Figure 7 [e.g., Simons et al., 2002]).

[18] E-MORB data scatter widely around a regression line fit to the T- and N-MORB Cl and La data (Figure 7a); D11-1 and D12-5 have particularly high Cl contents relative to their La concentrations. When Cl contents are normalized to La and plotted against Mg # (Mg # = 100 x Mg$^{2+}$/[Mg$^{2+}$ + Fe$^{2+}$]), data for samples D11-1 and D12-5 lie on the extreme end of the plot, at the lowest Mg # and highest Cl/La (Figure 7b). Thus, assimilation of Cl-rich material is likely to have affected these two samples the most. In contrast,

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**Figure 2.** Concentrations of incompatible trace elements in WGSC lavas normalized to estimated primitive mantle concentrations of McDonough and Sun [1995]. Only data for samples with Mg # > 50 are plotted. (a) E-MORB, (b) T-MORB, and (c) N-MORB. Fields for the other subgroups are shown in each plot; the T-MORB field excludes sample D73-1, and the N-MORB field excludes sample D2-1. The inset shows that some N-MORB have primitive mantle–normalized concentrations of <1 for some highly incompatible elements.
the low Cl content of sample D10-1 suggests it may have experienced extensive degassing [e.g., Dixon and Clague, 2001]. Removing D11-1 and D12-5 from consideration yields a statistically in-
significant trend ($r^2 = 0.07$). The remaining weak
correlation is mainly controlled by T-MORB, with
moderate Cl/La and moderate Mg #, and the N-MORB,
with generally lower Cl/La and higher Mg # (corre-
lation coefficient of 0.4). There are no trends within
individual MORB groups. Thus, compositional
differences between the T- and N-MORB cannot
be due to crustal assimilation and instead likely
reflect changes in crustal processes such as crys-
tallization [Detrick et al., 2002; Cushman et al.,
2004; Asimow and Langmuir, 2003].

Figure 3. Primitive mantle–normalized concentrations of volatile and selected trace elements in approximate order
of solid-liquid incompatibility (most to least incompatible from left to right) in WGSC glasses. (a) E-MORB,
(b) T-MORB, and (c) N-MORB data are shown along with the fields for the other subgroups. Only data for samples
having Mg # > 50 are plotted or included in the fields; the T-MORB field excludes sample D73-1, and the N-MORB
field excludes sample D2-1. Primitive mantle concentrations are from McDonough and Sun [1995], except for H$_2$O
(335 ppm, assuming H$_2$O/Ce = 200 for the primitive mantle), which is from Workman and Hart [2005].
6.1.2. Fractional Crystallization

Fractional crystallization also modifies melt composition during ascent and storage. The E-MORB span a wide range of Mg # (from <30 to ~60) that largely overlaps the smaller ranges of the other two groups (Figure 8). The lowest–Mg # E-MORB have pronounced negative Sr anomalies (Sr* = Sr_PM – (Pr_PM + Nd_PM)/2; where subscript “PM” indicates normalization of the measured value to that of the primitive mantle estimate of McDonough and Sun [1995]), indicating removal of substantial amounts of plagioclase from the melt (Figure 8a). High concentrations of highly incompatible elements are present in the same samples: for example, Nb concentrations are highest in four E-MORB with Mg # < 40 (Figure 8b). Although the concentrations of many major and trace elements in some E-MORB, particularly samples D10-1, D11-1, D11-3, D12-5, and D19-1, are likely to have been influenced significantly by fractional crystallization, the variation among the three MORB groups in element abundances at the same Mg # (e.g., Al_2O_3 and Nb at Mg # of 50–60 (Figures 8b and 8c)) cannot be explained by fractionation. Furthermore, ratios of incompatible elements that are
relatively unaffected by fractionation (e.g., Nb/Zr (Figure 8d)) must reflect different parental magma compositions, arising from variable extents of melting or in mantle source composition, especially for samples having Mg # > 50. Weak correlations of Nb/Zr (and Cl/La) with Mg # within groups are consistent with fractionation effects, but the ~3 × variation in Nb/Zr at Mg # = 55 between groups must reflect mantle processes, as supported by the strong correlations among incompatible trace element and isotopic ratios in GSC magmas, as we discuss below.

6.2. Mantle Processes

6.2.1. Melting Variations Along the WGSC

Along-axis geochemical gradients can result from variations in extent and depth of melting, although isolating these effects from variations in source composition is challenging, especially for heterogeneous mantle. For example, if the mantle contains lithologies with different solidi, a low extent of melting will tend to favor sampling of the more fusible material(s), which may also have higher concentrations of incompatible trace elements and more radiogenic Sr and Pb. Nonetheless, some qualitative conclusions about partial melting can be drawn from ratios like La/Sm and Nb/Zr. These ratios involve elements with sufficiently different partition coefficients that the ratios are somewhat influenced by partial melting, particularly at small amounts of melting. These ratios tend to increase toward the Galápagos hot spot and from the N- to T- to E-MORB (Figures 9 and 10), which is consistent with the extent of melting decreasing from the N-MORB-dominated western province of the GSC toward the hot spot. Decreasing extent of melting coinciding with increasing crustal thickness toward the hot spot is perhaps counterintuitive; however, this result can be reconciled with an increasing influence of water in the mantle which causes an enhanced flux of hydrous melts from greater depths near the Galápagos hot spot [Asimow and Langmuir, 2003; Cushman et

Figure 5. Radiogenic isotope compositions grouped by MORB type: (a) εNd, (b) 87Sr/86Sr, (c) 207Pb/204Pb, and (d) 208Pb/204Pb versus 206Pb/204Pb. MORB types defined previously on the basis of K/Ti are also largely distinguishable in isotope composition. This is especially true for the Pb isotopes, whereas Sr and Nd isotope ratios show more overlap, particularly between E- and T-MORB. Outlying samples are labeled in Figure 5d. All errors (at 2σ) are smaller than the symbols. Pb isotope ratios analyzed by the double-spike method are distinguished by filled symbols; open symbols denote conventional Pb isotope measurements (not double spiked) and have slightly larger errors (see section 4).
source composition are explored more fully in section 7.2.

[21] Constraints on depths of melting are provided by ratios involving the heavy REE, which are sensitive indicators of garnet in the residue, although source composition can also play a role. For example, the N- and T-MORB have similar Sm/Yb<sub>PM</sub> and Hf/Lu<sub>PM</sub>, whereas the E-MORB as a group have higher values (Figures 9 and 10). These findings suggest that if depth of melting changes along the GSC, it is likely to be deepest...
beneath the E-MORB-dominated portion of the ridge. That said, this does not preclude the possibility that the elevated Sm/Yb\textsubscript{PM} and Hf/Lu\textsubscript{PM} of the E-MORB are source characteristics.

### 6.2.2. Principal Components Analysis of the Isotope Arrays

To quantitatively assess the number of mantle end-member components involved in generating the WGSC rocks, we performed a principal components analysis (PCA) using \(^{87}\text{Sr}/^{86}\text{Sr}, \text{\varepsilon}_{\text{Nd}}, 206\text{Pb}/204\text{Pb}, 207\text{Pb}/204\text{Pb},\) and \(^{208}\text{Pb}/204\text{Pb}\). PCA does not yield the isotopic compositions of the mantle end-member components but resolves the variation in the data into the fewest number of eigenvectors that can explain that variation (see Albarède [1995] for a more thorough description of PCA). The first principal component (PC) for WGSC lavas accounts for more than 87% of the isotopic variation observed (Figure 11). The second PC explains a further 9%, and the next three PCs explain the remaining 4%. The systematic change in lava compositions along the GSC is thus primarily along the first PC. The simplest interpretation is that the first PC reflects binary mixing between two main mantle end-member components, as indeed can be inferred from inspection of the isotopic arrays in Figures 5 and 12, especially the linear trend in Pb-Pb isotope space.

The second PC may reflect a third mantle end-member component, but we prefer an interpretation by which it is related to small-amplitude heterogeneity within one (or both) of the two dominant end-member components. This interpretation is favored by the alignment of the data along the second PC.

**Figure 8.** Elemental and elemental ratio variations against Mg #. (a) Sr* versus Mg # (Sr* = Sr\textsubscript{PM} - (Pr\textsubscript{PM} + Nd\textsubscript{PM})/2, where PM indicates normalization to estimated primitive mantle values): strongly negative values of Sr* in E-MORB with Mg # < 45 indicate extensive plagioclase fractionation. (b) Nb versus Mg #. Note the progressively higher concentrations of Nb at Mg # of 50–60 from N-MORB to T-MORB to E-MORB. Increasing Nb contents of parental magmas between these groups must be inherited from mantle processes, e.g., progressively smaller extents of partial melting or a source enriched in Nb. (c) Al\textsubscript{2}O\textsubscript{3} versus Mg #. Elevated Al\textsubscript{2}O\textsubscript{3} for a given Mg #, as recorded in the E-MORB, is consistent with delay in plagioclase fractionation likely resulting from increased magmatic H\textsubscript{2}O [Sinton and Fryer, 1987]. (d) Nb/Zr\textsubscript{PM} versus Mg # shows a wide variation in Nb/Zr\textsubscript{PM} at a given Mg #, with the exception of the four highly fractionated E-MORB. Nb/Zr\textsubscript{PM} is affected by extent of partial melting (lower extents raise the ratio) and by mantle source composition.
by the small value of the second PC, and because its significance (score) does not change systematically along the length of the WGSC (Figure 11). The next three PCs have only minor significance; their scores are most variable at 95.5°W, for which several of the Pb isotope analyses were not done by the double-spike method. These least significant PCs are possibly related to small-amplitude heterogeneity in the two main mantle components but are also comparable to levels of analytical uncertainty.

When plotted versus longitude, variations of the first PC reveal the pattern seen in the isotopic data themselves: a gradual change from 91°W to 95.5°W, followed by a sudden change near 95.5°W, and mostly constant compositions between 96.5°W and 98°W (Figures 6 and 11). A similar abrupt change is also apparent in ΔNb (Figure 9e), a measurement of the deficiency or excess of Nb relative to Zr that is attributed to long-term source variations at hot spots; positive ΔNb has been associated with plume mantle sources and negative values with depleted upper mantle sources [Fitton et al., 1997]. The stepwise change in compositions near 95.5°W is consistent with, but not adequately resolved in Schilling et al.’s [2003] isotope data because of their wider sample spacing. Schilling et al. [2003] emphasized two samples collected between 98°W and 100°W, which are even less radiogenic in Sr and Pb and more radiogenic in Nd than those taken between 96.5° and 98°W. Schilling et al. [2003] therefore interpreted their average gradients in isotopic composition as revealing a steady decline in plume-ridge interaction over the entire distance between 91°W and 101°W. Our geochemical results indicate that plume effects decrease dramatically west of 95.5°W, further supported by reduced gradients in crustal thickness [Canales et al., 2002] and axial depth [Ito and Lin, 1995; Sinton et al., 2003] between 95.5°W and 98°W. We suspect that the region west of 98°W is beyond the Galápagos plume’s influence on the ridge, as discussed further below.

6.2.3. Enriched Component Beneath the WGSC

One of the two main source components associated with the first PC is characterized by radiogenic Pb and Sr, unradiogenic Nd, and enrichment in the highly incompatible elements. A logical question is whether this component also is evident in lavas of the Galápagos Archipelago. A HIMU-like [e.g., Zindler and Hart, 1986; Stracke et al., 2005] enriched end-member for the Galápagos

Figure 9.

Variations in incompatible trace elements along axis. (a) La/Sm<sub>PM</sub>, (b) Sm/Yb<sub>PM</sub>, (c) Nb/Zr<sub>PM</sub>, (d) Nb/La<sub>PM</sub>, and (e) ΔNb (= 1.74 + log<sub>10</sub>(Nb/Y) - 1.92 * log<sub>10</sub>(Zr/Y)) versus longitude [Fitton et al., 1997]. Major physiographic features are shown by dotted (ridge offsets) and solid (transform fault) lines. The ratios in Figures 9a–9c can be affected by both extents (and depths) of partial melting and by mantle source composition, whereas the values in Figures 9d and 9e are primarily influenced by source composition. La/Sm<sub>PM</sub> and Nb/Zr<sub>PM</sub> are different among the three groups: E-MORB have the highest values followed by T- and N-MORB, Sm/Yb<sub>PM</sub> is only slightly elevated in some E-MORB samples, and Nb/La<sub>PM</sub> and ΔNb are comparable in the E- and T-MORB but distinguishably lower in the N-MORB. Data labeled “other WGSC” are from Rotella et al. [2009].
Archipelago is apparent in the southernmost Galápagos island, Floreana [Harpp and White, 2001], but it has $^{206}\text{Pb}/^{204}\text{Pb}$ values that are too high relative to its $\varepsilon_{\text{Nd}}$ for it to be the enriched end-member along the WGSC (Figure 12b). A second enriched component invoked by Harpp and White [2001] to explain variations in the archipelago and previous data from the WGSC [Schilling et al., 2003] is the Wolf-Darwin (WD) component; however, this component is offset to higher $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ relative to $^{206}\text{Pb}/^{204}\text{Pb}$ than the trend of the WGSC arrays shown in Figures 12a and 13a. Last, Harpp and White’s [2001] “PLUME” component lies closest to our data; however, it cannot be an end-member because WGSC data extend to more extreme Pb isotope ratios (Figure 12). Although Schilling et al. [2003] argued that the WGSC and the GSC east of 91°W (EGSC) required distinct enriched components, the data shown in Figures 12 and 13 suggests that these two portions of the GSC share a common enriched component. In this sense, our new samples and data reduce the complexity inferred by Schilling et

Figure 10. Covariations in incompatible trace element ratios with isotopic compositions. (a) $\text{La}/\text{Sm}_{PM}$, (b) $\text{Hf}/\text{Lu}_{PM}$, and (c) $\Delta\text{Nb}$ versus $\varepsilon_{\text{Nd}}$ and (d) $\text{Nb}/\text{Zr}_{PM}$, (e) $\text{Sm}/\text{Yb}_{PM}$, and (f) $\text{Nb}/\text{La}_{PM}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$. $\text{La}/\text{Sm}_{PM}$ and $\text{Nb}/\text{La}_{PM}$ correlate well with isotopic composition and again are distinct among the three MORB groups. In contrast, $\text{Hf}/\text{Lu}_{PM}$ and $\text{Sm}/\text{Yb}_{PM}$ show little correlation with isotope composition. $\Delta\text{Nb}$ and $\text{Nb}/\text{La}_{PM}$ vary considerably with isotope ratios but are similar between the E- and T-MORB and distinctly lower in the N-MORB.
al. [2003], although the origin of the enriched component is still unclear.

[26] A plausible enriched end-member is Hanan and Graham’s [1996] “C,” which is similar to Hart et al.’s [1992] “FOZO.” Both “C” and “FOZO” are recognized as being common to many OIB and MORB worldwide. One characteristic that is often linked with C and FOZO is high $^{3}$He/$^{4}$He [e.g., Hanan and Graham, 1996; Hart et al., 1992; Dixon et al., 2002], and indeed some lavas of the Galápagos Archipelago with C/FOZO-like Sr-Nd-Pb isotopic signatures have high $^{3}$He/$^{4}$He (as high as 27 times the atmospheric ratio [Graham et al., 1993; Kurz and Geist, 1999]). In contrast, all GSC lavas studied thus far have low, MORB-like $^{3}$He/$^{4}$He (<8.5 times the atmospheric ratio) [e.g., Detrick et al., 2002; Graham et al., 2006]. This may suggest that the C-like end-member is the same in both settings, but that most of its He was extracted by volcanism in the islands [Kurz and Geist, 1999; Graham et al., 2006] prior to lateral flow of plume mantle toward and then remelting beneath the ridge.

[27] It is notable that although the same mantle (plume) material involved in forming the Galápagos Archipelago is probably now interacting with the WGSC, none of the archipelago’s enriched end-member compositions, as identified by Harpp and White [2001], are apparent along the WGSC, at least not in pure enough form to be recognized individually. Possible explanations include the following:

[28] 1. The archipelago components WD, FLO, and PLUME are thoroughly mixed beneath the WGSC so as to behave as a single enriched component in the ridge basalts, as in pseudobinary mixing [Douglass and Schilling, 2000; Schilling et al., 2003].

[29] 2. One or more of Harpp and White’s [2001] end-member components are not present beneath the WGSC because they were efficiently extracted from the plume beneath the archipelago (as suggested for He above).

[30] 3. One or more of the end-member components that feed the archipelago are not present beneath the WGSC because the mantle flow paths containing these particular components in the plume simply fail to reach the WGSC.

[31] 4. Still another possibility is that the volumes of these end-members are so small that their compositions are swamped beneath the WGSC by the extensive melting of the two more voluminous
components (our enriched and depleted WGSC components). In this case, our enriched WGSC (possibly C) component would be voluminous in the plume but have a composition that is intermediate to Harpp and White’s [2001] end-members.

[32] Resolving which possibility is most correct will likely require further investigations of volcano compositions between the archipelago and the GSC, coupled with modeling of 3-D mantle flow and melting.

6.2.4. Depleted Component(s) Beneath the GSC

[33] In Figures 12 and 13, the depleted component of the WGSC lies at the opposite end of the isotope arrays, having relatively high \( \varepsilon_{Nd} \) and less radiogenic Pb and Sr. A depleted component with compositions similar to those along the East Pacific Rise (EPR) was proposed by Harpp and White [2001] to explain data from the archipelago, and, based on two EPR-like samples west of 98°W, Schilling et al. [2003] interpreted the GSC array in terms of mixing between the (enriched) plume and an EPR-like depleted component in the ambient upper mantle. However, these authors’ and our data from the WGSC east of 98°W reveal a conspicuously shallower trend in Pb-Pb isotope space and extrapolate to higher \( ^{207}\text{Pb} / ^{204}\text{Pb} \) and \( ^{208}\text{Pb} / ^{204}\text{Pb} \) at low \( ^{206}\text{Pb} / ^{204}\text{Pb} \) values than data for either the

![Figure 12.](image)
EGSC or the EPR (Figure 13a). This finding requires the WGSC (east of 98°W) to be mixing with a depleted end-member that is distinct from that beneath the archipelago and the EGSC. The two compositions found west of 98°W appear to lie off of the trend of the WGSC east of 98°W, suggesting that an EPR-like depleted component is present west of 98°W.

[34] Thus we confirm Schilling et al.’s [2003] conclusion that the depleted component beneath most of the WGSC differs from that along the EGSC. A common interpretation is that depleted components are intrinsic to the ambient upper mantle [e.g., Schilling et al., 2003] whereas enriched components are contained in, or come from, plumes. For this to be true along the GSC, the ambient (depleted) upper mantle beneath 91°W and 98°W must differ from that beneath the EGSC, the archipelago, the EPR, and possibly beneath the GSC west of 98°W. Such a possibility requires compositional zoning in the ambient upper mantle at scales of hundreds of kilometers.

[35] An alternative interpretation is that two, somewhat different depleted components are intrinsic to the Galápagos plume. One appears in lavas at the archipelago and along the EGSC, and is similar to the depleted component beneath the EPR. The other Galápagos depleted component is unique to the WGSC, 91°W–98°W. The topic of depleted mantle plume components has been discussed and debated for the Hawaiian [e.g., Regelous et al., 2003; Mukhopadhyay et al., 2003; Huang et al., 2005; Frey et al., 2005] and Icelandic [e.g., Kerr et al., 1995; Chauvel and Hémond, 2000; Kempton et

Figure 13. Radiogenic isotope ratios of WGSC lavas compared to proposed mantle end-members. (a) Values of $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for GSC MORB from this study and Schilling et al. [2003] (labeled WGSC* and EGSC* for the western and eastern GSC, respectively) compared to MORB from the southern and northern East Pacific Rise (SEPR and NEPR, respectively) and the Costa Rica Rift (CRR; see Figure 12 caption for data sources), along with end-members proposed specifically for the Galápagos Archipelago and end-members proposed to have a global distribution. Binary mixing (purple line) between our estimated WGSC end-members (purple crosses, denoted GDM* and GEM* for Galápagos depleted and enriched mantle, respectively) is shown. Note that in Pb isotope space, this calculated mixing line is indistinguishable from both the 1st PC and from a linear regression through the double-spike Pb isotope data. WGSC data of Schilling et al. [2003] fall on our proposed mixing line, with the exception of two samples from west of 98°W, whereas most of the EGSC data of Schilling et al. [2003] lie below this line. Our N-MORB do not appear to share a common depleted end-member with SEPR or NEPR MORB; our GDM* end-member is more radiogenic in $^{208}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ relative to its $^{206}\text{Pb}/^{204}\text{Pb}$, similar to compositions found in Indian MORB (IMORB). Our enriched end-member, GEM*, does not coincide with any of the previously proposed Galápagos end-members. It is most similar to a mix of the WD and PLU end-members of Harpp and White [2001]; it is also similar to the proposed “C” component of Hanan and Graham [1996]. Values of (b) $\varepsilon_{\text{Nd}}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for the same rocks and end-members as in Figure 13a and (c) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for the same rocks and end-members as in Figure 13a. The 2σ errors on data are smaller than the symbols.
plume is predicted to push all of the ambient asthenosphere aside, so that only plume material melts beneath the section of ridge it influences [see, e.g., Ito et al., 2003]. If the original large- to intermediate-scale composition of the plume is uniform in terms of the relative amounts of enriched versus depleted material, then numerical models predict an unrealistically constant composition where the plume underlies the ridge, and a sudden change in composition at the edge of plume influence [Ito et al., 1997] (Figure 14b). The failure of geodynamic studies to predict gradients in composition along plume-influenced ridges may reflect shortcomings of the models, misinterpretation of the observations, or both. Resolving this issue is beyond the scope of this paper.

7. Mechanism Responsible for Geochemical Variations Along the WGSC

7.1. Plausible Scenarios and Overall Predictions

[36] Two classes of explanation have been forwarded for the origin of along-axis geochemical variations in studies of hot spot–ridge interaction. The first emphasizes mantle heterogeneity on intermediate scales (tens to hundreds of kilometers), relates lava composition directly to bulk mantle composition, and assumes the depleted components are primarily within the ambient upper mantle and not in plumes. The other class emphasizes heterogeneity at smaller scales (~1 km or smaller), considers depleted components to be intrinsic to plumes, and relates lava compositions to intermediate-scale variations in mantle flow and melt extraction, and not to intermediate-scale variations in the composition of the bulk source. Figure 14 depicts examples of the two classes.

7.1.1. Model A1

[37] The original explanation of GSC geochemistry is that of Schilling and colleagues. The near-ridge plume is more enriched than the depleted ambient upper mantle. Plume mantle flows toward, and then disperses along, the ridge axis where it becomes progressively diluted by ambient asthenosphere [e.g., Schilling, 1991] (Figure 14a). This hypothesis predicts a continuous, steady isotopic mixing trend along axis without any abrupt transitions.

[38] The main difficulty with the above hypothesis is that although geodynamic modeling studies predict plumes to interact with and spread substantially along ridges [e.g., Ribe, 1996; Ito et al., 1997; Sleep, 1996; Albers and Christensen, 2001], none have predicted any appreciable solid-state or magma mixing between plume and ambient mantle. The
enriched veins begin melting. At shallower depths, where the more refractory depleted matrix begins to melt, mantle flow (and melting) is slower, because it is impaired by the rigid lithospheric lid. As a consequence, this model predicts the most prominent expression of the enriched veins to be in lavas nearest the plume where deep mantle flow is strongest. With progressive distance away from the plume, the enriched component signature in lavas steadily decreases as mantle flow becomes less influenced by the plume.

7.2. Quantitative Modeling of Model B1: Veined Plume With Variable Melt Extraction and Upwelling

7.2.1. Methodology

Our goal is to explain the variations in crustal thickness, incompatible element ratios, and isotopic ratios along the WGSC, including the sudden shift in compositions near 95.5°W. We use a model like B1 above as our starting point.

[43] The melting model is explained in detail by Ito and Mahoney [2005a] (see also the online supporting material of Ito and Mahoney [2005a]); here we outline the main assumptions. We simulate non-modal fractional melting of peridotite in the presence of garnet (>3 GPa), spinel (<3 GPa), and after clinopyroxene is exhausted (F > 0.2). Melting of both depleted matrix and enriched peridotite veins occurs as mantle ascends through the melting zone (Figures 15a and 15b). Each melt increment is assumed to be chemically isolated from the surrounding solid prior to pooling with other melt increments (i.e., pooled, accumulated fractional melting). The composition of the pooled melts at each point along the ridge is the weighted average

Figure 14. Cartoons of five different mechanisms for hot spot–ridge interaction, overlain with idealized isotopic profiles that would be generated. Colors represent different materials as indicated in the legend. (a) Model A1. The concept of Schilling et al. [2003] involves a narrow plume that is compositionally distinct from the ambient mantle. The plume progressively mixes with the local depleted ambient mantle causing the plume signature in lava composition to gradually decrease with distance from the plume center. (b) Ito et al.’s [1997] geodynamic model involves a plume that ascends and displaces the ambient mantle while it spreads beneath the lithosphere. No solid-state mixing occurs; thus, a sharp boundary (black curve) separates enriched and ambient, depleted mantle. Geodynamic models predict that over most of the region of plume influence, the plume layer extends deeper than the solidus (dashed line), so lavas do not reflect any mixing between plume and ambient upper mantle. (c) Model A2. Like model A1 but incorporating our new results by including a zone of ambient upper mantle that is distinct from that elsewhere in the region. The shape of the intermediate-scale heterogeneity between 91 and 98°W is shown schematically; that is, the boundaries would be near 91°W and 98°W but not necessarily have the vertical sides as depicted. The question marks indicate uncertainty as to whether there is a compositional boundary at the edge of the plume on the EGSC. (d) Model B1. After Ito and Mahoney [2005a], this model involves a uniformly heterogeneous mantle source beneath the entire ridge that is dominantly depleted but contains small enriched blobs or veins (in red). The rate that mantle flows through the deepest portion of the melting zone (purple arrows) is largest above the plume stem and decreases toward the sides (expressed with arrow length). (e) Model B2. A hybrid model in which the deep upwelling varies along axis; as in Figure 14d, the matrix is depleted, but here the abundance of enriched veins changes slightly near 95.5°W to give rise to the step in composition. Also, the depleted matrix beneath the WGSC is distinct from that elsewhere.
of that needed to create the residual mantle column (RMC; [Plank and Langmuir, 1992; Langmuir et al., 1992]) which is exiting the melting zone perpendicular to the ridge axis. The average is weighted by the extent of melting ($F$) and by the RMC width (approximately proportional to the upwelling rate), both of which vary with depth (discussed below and shown in Figure 15).

[44] In our model, the depleted end-member composes 90% of the mantle and has only ~35 ppm water, or 10% of the water present in the enriched end-member. This end-member thus begins melting at shallow depths. We also assume it has a trace element content equal to that of the average depleted MORB mantle estimate of Workman and Hart [2005] and has high-$\varepsilon_{Nd}$, low-$^{206}\text{Pb}/^{204}\text{Pb}$ and low-$^{87}\text{Sr}/^{86}\text{Sr}$ values. The enriched end-member is assumed to be peridotite (not pyroxenite) veins or blobs, making up the remaining 10% of the mantle, and crossing the solidus ~60 km deeper than the depleted component does (as predicted for a water content of ~350 ppm; Asimow et al. [2004]). This component is assumed to have an incompatible element composition like the primitive mantle (using McDonough and Sun’s [1995] estimate) and isotopic compositions like C. Estimated compositions for the two end-members (see GDM and GEM in Figure 13) were determined by trying different possible isotopic compositions with values similar (in isotope space) to those of either the depleted mantle or C. Thus, multiple iterations of our model were carried out to determine the likeliest end-member isotopic compositions assuming melting conditions as discussed above. Importantly, the binary mixing curve in Pb isotope space (Figure 13a) produced using these conditions and end-members coincide with a line formed by a linear regression through the double-spike Pb isotope data. Table 3 specifies the values of all model parameters as well as our derived end-member isotope compositions.

[45] The effects of mantle flow are simulated by considering the rate at which the residue exits the melting zone as a function of depth normalized to the half-spreading rate, yielding a dimensionless, relative rate of flow (Figure 15c). Thus, the RMC effectively varies in width (per increment of time) as a function of depth and position along the axis. Far from the plume, the RMC is assumed to have a uniform width, as predicted for passive upwelling and corner flow driven purely by plate spreading (a relative flow rate of 1) [Ito and Mahoney, 2005a]. In the plume-influenced areas, we consider an additional component of buoyancy-driven flow that starts near where the depleted matrix begins to melt, or more specifically when melt extent ($F$) is 0.02, at ~65 km (Figure 15a), and extends to a depth of 125 km [Ito et al., 1997]. This flow profile is designed to simulate the effects of water on mantle rheology, in which the mantle begins with a moderate water content and low viscosity so that buoyancy drives rapid mantle flow at depth [Ito et al., 1999]. As water is extracted from the matrix at low degrees of partial melting, the mantle becomes too viscous [Hirth and Kohlstedt, 1996] for buoyancy to influence the flow (i.e., at ~60 km (Figure 15c)) and upwelling slows to the rate driven by the strong plates. The result is rapid, buoyancy-driven flow and melting where the enriched peridotite veins are melting (Figures 15a and 15b), followed by more sluggish, plate-driven flow and melting where the depleted matrix is primarily melting.

Figure 15. Profiles of (a) extent of fractional melting, or $F$; (b) melt productivity, or the rate change in melt fraction per increment of pressure release, $dF/dP$; and (c) relative rate of flow (flow out of the melting zone normalized by the half-spreading rate) versus depth for conditions listed in case I (Table 3 and section 7.2.2) for model B1 (see section 7.2). For Figure 15c, cases II, IV, and V produce no change, but in case III the maximum relative flow rate is 7.5 (or reduced by half from the 15 shown for case I).
Table 3. Model Parameters Used for Five Different Test Cases

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<td>$^{87}$Sr/$^{86}$Sr</td>
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Notes:
- Case I is the combination of parameters that is the best fit to our data and the other four reflect one or more deviations from this preferred model.
- Case II considers the effect of a diminished temperature differential. In case III, relative mantle flow is decreased to half the rate of case I. Case IV examines the effect of the amount of enriched component available. Case V shows the impact of a change in Kd values for three of the modeled elements. Note that Nd isotope compositions are represented in e units as defined in Table 2.
- $^{208}$Pb/$^{204}$Pb is the change in relative effective mantle velocity through the melt zone.
- $^{87}$Sr/$^{86}$Sr is the change in temperature along axis.
- $^{87}$Sr/$^{86}$Sr is the height of the plume layer.
- DM stands for our “depleted end-member” and assumes the elemental abundances of “average depleted mantle” from Workman and Hart [2005] (W&H ADM); elements used in the model and their concentrations (in ppm) are: Th = 0.0079, U = 0.0032, Nb = 0.1485, La = 0.192, Ce = 0.550, Pb = 0.018, Sr = 7.664, Nd = 0.581, Zr = 13.54, Y = 0.338, Yb = 0.365, H2O = 35.
- EM stands for our “enriched end-member,” and we use the elemental abundances of “primitive mantle” from McDonough and Sun [1995] (M&S PM) except for water, which is from Workman and Hart [2005]; elements used in the model and their concentrations (in ppm) are as follows: Th = 0.085, U = 0.211, Nb = 0.713, La = 0.687, Ce = 1.775, Pb = 0.185, Sr = 21.1, Nd = 1.354, Zr = 11.2, Sm = 0.444, Y = 4.55, Yb = 0.493, and H2O = 335.
- Kds, or partition coefficients, for the first three cases are from Salters and Stracke [2004] for all elements except Nb (clinopyroxene and garnet only [Johnson, 1998]), La (clinopyroxene and garnet only [Johnson, 1998]), Zr (olivine and orthopyroxene [Salters et al., 2002]) or clinopyroxene and garnet [Johnson, 1998]), Y (Green, 1994), and Yb (clinopyroxene and garnet only [Johnson, 1998]). For the fourth case, all Kds are from Salters and Stracke [2004] except for Y and Zr, which are as above.

7.2.2. Reference Model: Case I

Case I represents our reference model. The excess flow due to plume buoyancy at the base of the melting zone decreases along axis as an exponential function of distance, from a high of 15 times the spreading rate at 92°W to 1 times the spreading rate at 96°W, at which point the RMC has a uniform thickness created entirely by plate-driven flow (Figure 15c). The excess mantle temperature also decreases exponentially over the same distance, with a peak of 30°C at 92°W.

The models are run in the forward sense, predicting crustal thickness and lava composition along the ridge axis. To demonstrate the sensitivity of the model predictions to differences in mantle temperature, flow rate, end-member proportions, and bulk distribution coefficients, we present five different cases. Also, by comparing the predictions of the different cases with the observations, we will constrain the values of the model parameters (see Table 3 and Figures 16 and 17), which we will associate with plume-ridge interaction along the WGSC.
Figure 16
melting zone, essentially the same effect attributed to a variation in bulk water content by Asimow and Langmuir [2003], Asimow et al. [2004], and Cushman et al. [2004]. However, in this case bulk water content is the same everywhere. The enhanced flux of low-degree melts from deep in the melting zone is consistent with observations of higher Nb/ZrPM, La/SmPM, and Sm/YbPM in the E-MORB (Figure 17).

The variations in isotopic ratios and most incompatible element ratios are reproduced relatively well in the 95.5°W–92°W region; mixing arrays in isotope space are likewise reproduced well (not shown). Case I also predicts crustal thickness to increase toward the hot spot, but overpredicts crustal thickness nearest the hot spot. Overall, this model case provides a reasonable match with the geochemical and geophysical observations in the region of the WGSC most influenced by the hot spot 95.5°W–92°W. Case I does not explain the data west of 95.5°W (nor do cases II–V), an issue discussed in section 7.3.

7.2.3. Lower Excess Temperature: Case II

A second set of calculations examines the effects of mantle temperature by simulating a smaller temperature variation along axis (ΔT = 10°C). Compared to case I, this case slightly underpredicts crustal thickness farther from the hot spot and better predicts crustal thickness near 92°W. Case II predicts the variations in trace element and isotope ratios equally as well as case I. The smaller temperature change of case II causes average extents of melting to decrease more toward the hot spot than case I. The mean F of ~5.5% for the E-MORB, as predicted at 92°W, is similar to the 5% estimated by Cushman et al. [2004] (3%–5%) and Asimow and Langmuir [2003]. Overall, cases I and II show that ΔT is probably 10°C–30°C. This finding is consistent with prior estimates by Cushman et al. [2004] (ΔT ~ 20°C) and Canales et al. [2002] (ΔT ~ 30°C).

7.2.4. Slower Deep Upwelling: Case III

Case III is identical to case I but with a factor of two reduction in maximum relative flow rate (i.e., 7.5 times the half-spreading rate). Compared to case I, case III predicts slightly thinner crust, and a much more subtle change in mean extent of melting along the GSC (from ~9% at 96°W to ~8% at 92°W). The predictions for observed variations are comparable to those of case I in that some fits are slightly better (e.g., 206Pb/204Pb, crustal thickness) whereas others are poorer (e.g., εNd and La/SmPM).

7.2.5. More Enriched Veins (Case IV) and Different Partition Coefficients (Case V)

Case IV differs from case I in having a larger proportion (20%) of the enriched component in the starting source. Obviously, varying the isotopic compositions of the end-members also would have a strong effect on the model’s predictions; however, there is a direct trade-off between isotopic compositions and the amount of enriched material present, so we focus on the latter. A doubling in the amount of enriched component leads to thicker crust compared to case I, because of the larger fraction of water-rich veins that melt where buoyant upwelling is the most rapid. The resulting Pb isotope compositions, however, are too radiogenic. A similar misfit results for Nd isotopic ratios and the modeled trace element ratios are generally overpredicted. This case suggests that the amount of enriched veins is less than 20%.

Case V is the same as case I but uses the distribution coefficients of Salters and Stracke [2004] rather than those used by Ito and Mahoney [2005a]. Changing the distribution coefficients, while keeping all other parameters the same, has a relatively small effect on isotopic and crustal thickness predictions but, as expected, modifies the relevant elemental concentration ratios.

Figure 16. Model B1 (see section 7.2) predictions compared to our data. Prescribed variations in model variables (dashed lines) and predicted results (solid lines) shown are (a) relative mantle flow (i.e., flow exiting the melting zone normalized by spreading rate, left axis) and mean extent of melt (total quantity of melt divided by the height of the melting zone, right axis), (b) potential temperature, (c) crustal thickness compared to that measured seismicly by Canales et al. [2002], (d) εNd, and (e) 206Pb/204Pb versus longitude. Case I is our reference case (parameters given in Table 3). Maximum relative flow rate (Figure 16a) is the same for cases I, II, IV, and V and is 50% lower for case III; therefore, case III predicts a thinner crust than case I. For a reduced temperature anomaly (case II), average extents of melting decrease more rapidly toward the east, and less crust is produced than in case I. For larger proportions of enriched mantle material (case IV), average extents of melting are generally less and crust is thicker than in case I. All 2σ errors on measured data are smaller than the symbols.
In summary, the geochemical and geophysical data between 95.5°W and 92°W are generally matched with maximum temperature anomalies of ΔT = 10°C–30°C, excess deep mantle flow of 7.5–15 times the spreading rate, and a proportion of enriched veins <20% and probably close to 10%.

7.2.6. Buoyancy Flux of Model B1

The reasonableness of a relative mantle flow rate (ΔV) of 7.5–15 times the spreading rate may be assessed by estimating the buoyancy flux B,

\[ B = Q \Delta \rho, \]

where Q is the volume flux of the excess buoyant upwelling and Δρ the density anomaly of the plume. Volume flux Q can be estimated as,

\[ Q = U H W \left[ 1 + 2/3(\Delta V/2)/2 \right], \]

where U is full spreading rate (1.6 × 10^{-9} m/s), H is the total thickness of the plume layer (1.2 × 10^{5} m, about twice as thick as the zone of rapid upwelling), W is total along-axis width of plume influence (i.e., both west and east of 91°, or ~1 × 10^{6} m), and the maximum excess flow rate has been halved to account for the decrease to zero along the ridge axis (ΔV/2 = 7.5, the factor of 2/3 comes from integrating the quadratic function of plume-driven flow over depth). This gives Q = 670m^{3}/s. Assuming Δρ is purely thermal, it is the product of upper mantle density (3300 kg/m^{3}), the coefficient of thermal expansion (3.5 × 10^{-5} °C^{-1}), and the upper bound for excess temperature (ΔT = 30°C). Thus, the estimated buoyancy flux implied by our reference case I is \( B = 2.3 \times 10^{3} \) kg/s. This estimate is comparable to the buoyancy flux estimate of ~2 × 10^{3} kg/s by Ito et al. [1997]. Our lower bound estimate for the excess flow of 7.5 times the spreading rate, for ΔT = 10°C–30°C, brings our estimate for B down to 0.5–1.5 × 10^{3} kg/s, or within the estimates of Sleep [1990] and Canales et al. [2002], which are based on the flux needed to support the Galápagos hot spot swell.

7.3. Hybrid Model B2 and Implications

Model B1 shows that variations in buoyancy-driven mantle flow, combined with the effects of dehydration to retard upwelling where the depleted matrix is melting, can explain most of the geochemical and geophysical variations within the zone of greatest plume influence (91°W–95.5°W). However, model B1 fails to account for the sharp
transition in the compositions between 95.5°W–96.5°W. In the context of model B1, this transition would require a sharp and large decrease in buoyant upwelling, for which we have no simple explanation. A more likely explanation is that the abrupt change near 95.5°W signifies a geochemical boundary associated with intermediate-scale zoning in mantle composition. Our finding that the same depleted component is present east and west of 95.5°W could indicate the mantle between 95.5°W and 98°W is ambient upper mantle similar in composition to the depleted component in the plume, or that the mantle west of 95.5° is still part of the plume. The latter possibility would imply the presence of an outer plume, a sheath containing fewer enriched veins than the inner region it surrounds. It is the inner region that appears to be hottest, most actively upwelling, and therefore generating the largest geophysical and geochemical anomalies (91°W–95.5°W). This hybrid model B2 is illustrated in Figure 14e. As discussed in section 6, this model would require another compositional boundary near 91°W, the approximate longitude that marks the appearance on the EPR of the other depleted component, which is also evident at the Galápagos Archipelago. The WGSC data can therefore be interpreted as revealing an east-west bilateral, compositional zoning of the Galápagos mantle plume. The zoning pertains particularly to depleted components that are intrinsic to the plume.

An important implication is that even the inner, more enriched, region of the Galápagos plume influencing the WGSC between 91°W and 95.5°W is mostly depleted mantle. Its bulk composition is approximately expressed by the model compositions at 95.5°W (Figures 16 and 17) and does not clearly differ from that of the mantle expressed at other ocean ridge segments unaffected by hot spots. The inference is that the source region of the plume material feeding the GSC (e.g., the deeper upper mantle, lower mantle, or core-mantle boundary) may be similar to the source of many normal ridge basalts in terms of the abundance of depleted material. It has yet to be tested whether this material could represent the bulk composition of Galápagos plume at its center, beneath the archipelago.

Last, the discovery that the inferred extent of strong plume influence occurs near the tip of a propagating rift near 95.5°W, as emphasized by both PCA results (Figure 11b) and by a change from positive (plume-like) ΔNb values (Figure 9e), could reveal some form of mechanical interaction between the asthenosphere and lithosphere. Geochemical boundaries at propagating rift tips have been observed at a few other locations, including 20.7°S along the EPR (Sinton et al., 1991; Mahoney et al., 1994). The 95.5°W propagator is one of several, often smaller-offset, propagators that are or were recently active along the GSC [Hey, 1977; Wilson and Hey, 1995]; therefore, not all propagating rifts mark the edge of strong Galápagos plume influence. Indeed, we can see from the PCA scores and analysis of geochemical data [Rotella et al., 2009] that the propagator at ~93.3°W displays no such boundary. Bathymetric gradients and lateral asthenospheric flow, such as those associated with plume-ridge interaction, have been shown to drive rift propagation [Phipps Morgan and Parmentier, 1985; West and Lin, 1999], but a causal link between propagators and the edge of buoyant plume material has not yet been established. Also an analogous compositional boundary along the EPR, for example at the 87°W OSC, has yet to be found [Christie et al., 2005]. Detailed sampling across the Inca (85.5°W) and Ecuador (84.5°W) transforms are probably needed for this test.

8. Conclusions

Geochemical data for WGSC basalts provide new information about the causes of geochemical and geophysical variations along axis. Our major conclusions are as follows.

1. From 91°W to 95.5°W, lava compositions show an overall steady decline in abundances of incompatible trace elements, ratios of highly over moderately incompatible elements, and radiogenic Sr and Pb, and increasingly radiogenic Nd. To the west of 95.5°W, compositions change by 30%–50% of the total variation over an axial distance of ~100 km, to background levels of less radiogenic Sr and Pb and lower abundances of highly incompatible elements. Between 96.5°W and 98°W, compositions show little further systematic variation.

2. Two main mantle source end-members explain >83% of the isotopic variability along the WGSC. The enriched end-member is comparable to the C or “common” mantle component sampled at many ridges and hot spots and is dissimilar to any of the individual end-members of the Galápagos Archipelago. This finding could be revealing differences in sampling and/or mixing between the two...
locations, or spatial variability in the composition and flow within the Galápagos mantle plume.

[62] 3. The depleted WGST end-member is distinct from that beneath the Galápagos Archipelago, the eastern GSC, and the nearby EPR. Thus, there is some form of compositional zoning either in the ambient upper mantle or in the plume.

[63] 4. To explain geochemical and geophysical variations between 91°W and 95.5°W, we propose a model in which the mantle is a uniform mixture of a depleted matrix (90%) and incompatible element–enriched veins (10%). Variations in melting caused by deep, buoyancy-driven flow explain the geochemical and crustal thickness variations along the WGSC between 92° and 95.5°W.

[64] 5. The geochemical boundary near 95.5°W marks a mantle compositional boundary. It separates a region with slightly less enriched material to the west compared to the east.

[65] 6. The bulk composition of our model plume between 91°W and 95.5°W beneath the GSC is comparable to compositions of other ridges far from hot spots.

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References


