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7.09.1 Introduction

The original work by Wilson (1963, 1973), Morgan (1971, 1972), and Crough (1978) established the concept of ‘hot spot’ as a broad swelling of topography capped by volcanism, which, combined with plate motion, generates volcanoes aligned in a chain and with ages that progress monotonically. In some cases, these chains project back to massive volcanic plateaus, or large igneous provinces (LIPs), suggesting that hot-spot activity began with some of the largest magmatic outbursts evident in the geologic record (Morgan, 1972; Richards et al., 1989; Duncan and Richards, 1991). Hot-spot volcanism is dominantly basaltic and therefore largely involves melting of mantle peridotite, a process that also produces mid-oceanic ridge volcanism. Yet mid-ocean ridge basalts (MORBs) and hot-spot basalts typically have distinct radiogenic isotope characteristics (Hart et al., 1973; Schilling, 1973). These differences indicate that the two forms of magmatism come from mantle materials that have preserved distinct chemical identities for hundreds of millions of years.

The above characteristics suggest that hot-spot volcanism has an origin that is at least partly decoupled from plate processes. A straightforward explanation is that hot spots are generated by convective upwellings, or plumes of unusually hot, buoyant mantle, which rise from the lower mantle (Wilson, 1963, 1973; Morgan, 1971, 1972; Whitehead and Luther, 1975) possibly through a chemically stratified mantle (e.g., Richter and McKenzie, 1981). The large mushroom-shaped head of an initiating mantle plume and the trailing, more narrow plume stem has become a popular explanation for the formation of a LIP followed by a hot-spot track (e.g., Richards et al., 1989; Campbell and Griffiths, 1990).

Studies of hot spots have flourished over the past few decades. Recent articles and textbooks have reviewed some of the classic connections between hot spots and mantle plumes (e.g., Jackson, 1998; Davies, 1999; Condie, 2001; Schubert et al., 2001), the role of mantle plumes in deep-mantle convection and chemical transport (Jellinek and Manga, 2004), and oceanic hot spots (e.g., Ito et al., 2003; Hekinian et al., 2004). Alternative mechanisms, which emphasize processes in the asthenosphere and lithosphere, are being re-evaluated and some new ones proposed (Foulger et al., 2005). It has become clear that few hot spots confidently show all of the above characteristics of the classic description. The term hot spot itself implies a
localized region of anomalously high mantle temperature, but some features that were originally called hot spots may involve mantle with little or no excess heat, volcanoes spanning large distances of a chain with similar ages, or both. Thus, the terms ‘magmatic anomaly’ or ‘melting anomaly’ may be more general and appropriate to describe the topic of this chapter.

Progress made in the last decade on studies of hot spots and melting anomalies is emphasized here. We summarize the recent observations and discuss the major dynamical processes that have been explored and evaluate their ability to explain the main characteristics. Mechanisms involving hot mantle plumes have seen the most extensive quantitative testing, but the recent observations compel the exploration and rigorous testing of other mechanisms. We summarize the main observations, outline mechanisms that have been proposed, and pose questions that need quantitative answers.

7.09.2 Characteristics

Guided by the classical description of hot spots, we examine four main characteristics: (1) geographic age progression along volcano chains, (2) initiation by massive flood basalt volcanism, (3) anomalously shallow topography surrounding volcanoes (i.e., a hot-spot swell), and (4) basaltic volcanism with geochemical distinction from MORBs. Given the marked progress in seismic methods over the past decade, we also summarize the findings of mantle seismic structure beneath hot spots and surface melt anomalies. Table 1 summarizes what we have compiled about the above characteristics for 69 hot spots and melting anomalies. Figure 1 shows a global map of their locations with abbreviations and the main large igneous provinces that we will discuss.

7.09.2.1 Volcano Chains and Age Progression

7.09.2.1.1 Long-lived age-progressive volcanism

At least 13 hot-spot chains record volcanism lasting >50 My (Table 1). The Hawaiian–Emperor and the Louisville chains, for example, span thousands of kilometers across the Pacific basin (~6000 and >4000 km, respectively), record volcanism for >75 My (Duncan and Clague, 1985; Watts et al., 1988; Duncan and Keller, 2004; Koppers et al., 2004), and were among the first chains that led to the establishment of the hot-spot concept. As both chains terminate at subduction zones, the existing volcanoes likely record only part of the activities of these hot spots. The Galápagos is the other Pacific hot spot with a similar duration. Its interaction with the Galápagos Spreading Center has produced two chains: the Galápagos Archipelago–Carnegie Ridge on the Nazca Plate (Sinton et al., 1996) and the Cocos Ridge on the Cocos Plate. The Cocos Ridge records oceanic volcanism for ~14.5 My (Werner et al., 1999) and projects toward the Caribbean LIP (Duncan and Hargraves, 1984), which has $^{40}$Ar/$^{39}$Ar dates of 69–139 Ma (e.g., Sinton et al., 1997; Hoernle et al., 2004). The geochemical similarity of these lavas with the Galápagos Archipelago is compelling evidence for a ~139 My life span for the Galápagos hot spot (Hoernle et al., 2002, 2004).

In the Indian Ocean, Müller et al.’s (1993b) compilation of ages associates the Réunion hot spot with volcanism on the Mascarene Plateau at 45 Ma (Duncan et al., 1990), the Cocos–Laccadive Plateau ~60 Ma (Duncan, 1978, 1991), and finally the Deccan flood basalts in India, which are dated at 65–66 Ma (see also Sheth (2005)). The Comoros hot spot can be linked to volcanism around the Seychelles islands dated at 63 Ma (Emerick and Duncan, 1982; Müller et al., 1993b). Volcanism associated with the Marion hot-spot projects from Marion island (~<0.5 Ma (McDougall et al., 2001)) along a volcanic ridge to Madagascar. While geologic dating is sparse, Storey et al. (1997) infer an age progression along this track back to ~88 Ma. The Kerguelen hot spot is linked to Broken Ridge and Ninetyeast Ridge on the Australian Plate, as well as multiple stages of volcanism on the Kerguelen Plateau dating to 114 Ma (Frey et al., 2000; Nicolaysen et al., 2000) (Figure 2).

In the Atlantic Ocean, the Tristan–Gough and St. Helena chains record volcanism on the African Plate for ~80 My (O’Connor and Roex, 1992; O’Connor and Duncan, 1990; O’Connor et al., 1999). The connection of Tristan–Gough to the Paraná flood basalts in South America and the Etendeka basalts in Namibia suggests a duration for Tristan–Gough of ~130 My (see Peate (1997) and references therein). The Trindade–Martin Vaz chain (Fodor and Hanan, 2000) extends eastward from Brazil to where the Alto Paraná and Poxoreu volcanic provinces erupted ~85 Ma (Gibson et al., 1997). In the North Atlantic, the Madiera and Canaries chains have recorded age-progressive volcanism for nearly 70 My (Guillou et al., 2004; Geldmacher et al., 2005). The Canaries are unusual in that single volcanoes often remain active for tens of millions of years (Figure 3) (Geldmacher et al., 2005).
<table>
<thead>
<tr>
<th>Name (abbreviation)</th>
<th>Hot spot E. Long., N. Lat.</th>
<th>Age progression?</th>
<th>Age range</th>
<th>Swell?/width (km)</th>
<th>Connection to LIP?</th>
<th>Geoch. distinct from MORB</th>
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<tr>
<td>Pacific</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Austral (AU)</td>
<td>−140.0, −29.37</td>
<td>No</td>
<td>0–58.1 Ma</td>
<td>Yes/600</td>
<td>No</td>
<td>$^{206}\text{Pb}/^{204}\text{Pb}$</td>
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<tr>
<td>Baja (BAJ)</td>
<td>−113, 27</td>
<td>—</td>
<td>—</td>
<td>No</td>
<td>No</td>
<td></td>
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<tr>
<td>Bowie-Kodiak (BOW)</td>
<td>−130, 49.5</td>
<td>Ok</td>
<td>0.1–23.8 Ma</td>
<td>Yes/250</td>
<td>No</td>
<td>May be $^{206}\text{Pb}/^{204}\text{Pb}$</td>
</tr>
<tr>
<td>Caroline (CAR)</td>
<td>−197, 5.3</td>
<td>Weak</td>
<td>1.4 Ma (east) to 4.7–13.9 Ma (west)</td>
<td>—</td>
<td>No</td>
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<tr>
<td>Cobb (COB)</td>
<td>−128.7, 43.6</td>
<td>Good</td>
<td>1.5–29.2 Ma</td>
<td>Yes/370</td>
<td>No</td>
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<tr>
<td>Cook (CK)</td>
<td>−149.5, −23.5</td>
<td>No</td>
<td>0.2–19.4 Ma</td>
<td>Yes/500</td>
<td>No</td>
<td>$^{206}\text{Pb}/^{204}\text{Pb}$</td>
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<tr>
<td>Easter (EAS)</td>
<td>−109, −27</td>
<td>Good</td>
<td>0–25.6 Ma</td>
<td>Yes/580</td>
<td>May be Tuamotu and Mid-Pacs</td>
<td>$^{206}\text{Pb}/^{204}\text{Pb}$</td>
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<tr>
<td>Foundation (FOU)</td>
<td>−111, −39</td>
<td>Good</td>
<td>2.1–21 Ma</td>
<td>Yes/250</td>
<td>No</td>
<td></td>
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<tr>
<td>Galápagos (GAL)</td>
<td>−91.6, −0.4</td>
<td>Yes</td>
<td>0–14.5 Ma offshore; 69–139 Ma, Caribbean LIP</td>
<td>Yes/300</td>
<td>Caribbean LIP</td>
<td>$^{206}\text{Pb}/^{204}\text{Pb}$</td>
</tr>
<tr>
<td>Geologist (GEO)</td>
<td>−157, 19</td>
<td>No</td>
<td>82.7–84.6 Ma</td>
<td>—</td>
<td>No</td>
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<tr>
<td>Guadalupe (GUA)</td>
<td>−118, 29</td>
<td>—</td>
<td>&lt;3.4 to ~20.3 Ma</td>
<td>May be/?</td>
<td>No</td>
<td></td>
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<tr>
<td>Hawaiian-Emperor (HAW)</td>
<td>−155.3 18.9</td>
<td>Good</td>
<td>0–75.8 Ma</td>
<td>Yes/920</td>
<td>No</td>
<td>$^{3}\text{He}/^{4}\text{He}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ for Islands but not Emperors $^{206}\text{Pb}/^{204}\text{Pb}$</td>
</tr>
<tr>
<td>Japanese-Wake (JWK)</td>
<td>—</td>
<td>No</td>
<td>78.6–119.7 Ma</td>
<td>No</td>
<td>No</td>
<td></td>
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<tr>
<td>Juan Fernandez (JFE)</td>
<td>−79, −34</td>
<td>Weak</td>
<td>1–4 Ma (2 volcanoes dated)</td>
<td>Yes/?</td>
<td>No</td>
<td>$^{3}\text{He}/^{4}\text{He}$ and $^{87}\text{Sr}/^{86}\text{Sr}$</td>
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<tr>
<td>Line Islands (LIN)</td>
<td>—</td>
<td>No</td>
<td>35.5–91.2 My</td>
<td>Partially/?</td>
<td>May be Mid-Pacs</td>
<td>—</td>
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<tr>
<td>Louisville (LOU)</td>
<td>−141.2, −53.55</td>
<td>Good</td>
<td>1.1–77.3 Ma</td>
<td>Yes/540</td>
<td>Doubtfully OJP</td>
<td>$^{206}\text{Pb}/^{204}\text{Pb}$, may be $^{87}\text{Sr}/^{86}\text{Sr}$ $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$</td>
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<tr>
<td>Magellan Seamounts (MAG)</td>
<td>—</td>
<td>No</td>
<td>87–18.6 Ma</td>
<td>No</td>
<td>No</td>
<td></td>
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<tr>
<td>Marquesas (MQS)</td>
<td>−138.5, −11</td>
<td>Ok</td>
<td>0.8–5.5 Ma</td>
<td>Yes/850</td>
<td>May be Shatsky or Hess</td>
<td>$^{87}\text{Sr}/^{86}\text{Sr}$, may be $^{206}\text{Pb}/^{204}\text{Pb}$ $^{206}\text{Pb}/^{204}\text{Pb}$</td>
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<tr>
<td>Marshal Islands (MI)</td>
<td>−153.5, −21.0</td>
<td>No</td>
<td>68–138 Ma</td>
<td>May be/?</td>
<td>No</td>
<td></td>
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<tr>
<td>Mid-Pacific</td>
<td>—</td>
<td>No</td>
<td>73.5–128 Ma</td>
<td>No</td>
<td>It could be a LIP</td>
<td>—</td>
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<tr>
<td>Location</td>
<td>Age</td>
<td>Type</td>
<td>Is LIP?</td>
<td>87Sr/86Sr</td>
<td>206Pb/204Pb</td>
<td>Another Analysis</td>
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<td>---------------</td>
<td>-------------</td>
<td>-----------</td>
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<td>-----------</td>
<td>-------------</td>
<td>-----------------------------------</td>
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<tr>
<td>Mountains (MPM)</td>
<td>—</td>
<td>Ok</td>
<td>65.5–95.8 Ma</td>
<td>No</td>
<td>No</td>
<td></td>
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<tr>
<td>Pitcairn (PIT)</td>
<td>−129.4, −25.2</td>
<td>Good</td>
<td>0–11.1 Ma</td>
<td>Yes/570</td>
<td>No</td>
<td>87Sr/86Sr</td>
</tr>
<tr>
<td>Puka–Puka (PUK)</td>
<td>−165.5, −10.5</td>
<td>Ok</td>
<td>5.6–27.5 Ma</td>
<td>Yes/?</td>
<td>No</td>
<td>206Pb/204Pb</td>
</tr>
<tr>
<td>Samoa (SAM)</td>
<td>−169, −14.3</td>
<td>Weak</td>
<td>0–23 Ma</td>
<td>Yes/396</td>
<td>No</td>
<td>87Sr/86Sr, 3He/4He, and 206Pb/204Pb</td>
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<tr>
<td>San Félix (SF)</td>
<td>−80, −26</td>
<td>—</td>
<td>—</td>
<td>Yes/?</td>
<td>No</td>
<td></td>
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<tr>
<td>Shatsky (SHA)</td>
<td>—</td>
<td>Yes</td>
<td>128–145 Ma</td>
<td>No</td>
<td>It is a LIP</td>
<td></td>
</tr>
<tr>
<td>Society (SOC)</td>
<td>−148, −18</td>
<td>Good</td>
<td>0.01–4.2 Ma</td>
<td>Yes/?</td>
<td>No</td>
<td></td>
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<tr>
<td>Socorro (SCR)</td>
<td>−111, 19</td>
<td>—</td>
<td>—</td>
<td>Yes/?</td>
<td>No</td>
<td></td>
</tr>
<tr>
<td>Tarava (TAR)</td>
<td>173, 3</td>
<td>Weak</td>
<td>35.9 Ma and 43.5 Ma</td>
<td>Yes/?</td>
<td>No</td>
<td></td>
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<tr>
<td>Tuamotu (TUA)</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>Yes/?</td>
<td>It could be a LIP</td>
<td></td>
</tr>
<tr>
<td>North America</td>
<td>—</td>
<td>Yes</td>
<td>16–17 Ma</td>
<td>Yes/600</td>
<td>May be Columbia River Basalts</td>
<td></td>
</tr>
<tr>
<td>Yellowstone (YEL)</td>
<td>−111, 44.8</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>Australia</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>Balleny (BAL)</td>
<td>164.7, −67.4</td>
<td>Weak</td>
<td>—</td>
<td>May be Lord Howe rise</td>
<td>206Pb/204Pb (2 analyses)</td>
<td>206Pb/204Pb</td>
</tr>
<tr>
<td>East Australia (AUS)</td>
<td>143, −38</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
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<tr>
<td>Lord Howe (LHO)</td>
<td>159, −31</td>
<td>—</td>
<td>—</td>
<td>It could be LIP</td>
<td>—</td>
<td></td>
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<tr>
<td>Tasmanian (TAS)</td>
<td>153, −41.2</td>
<td>Yes</td>
<td>—</td>
<td>Yes/290</td>
<td>May be Lord Howe rise</td>
<td>—</td>
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<tr>
<td>Atlantic</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td></td>
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<tr>
<td>Ascension/ Circe (ASC)</td>
<td>−14, −8</td>
<td>—</td>
<td>&lt;1 Ma (Ascension) and 6 Ma (Circe)</td>
<td>820</td>
<td>No</td>
<td>206Pb/204Pb</td>
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<tr>
<td>Azores (AZO)</td>
<td>−28, 38</td>
<td>Seafloor spreading</td>
<td>0–20 Ma, possibly ~85 Ma</td>
<td>Yes/2300</td>
<td>No</td>
<td>87Sr/86Sr, 206Pb/204Pb</td>
</tr>
<tr>
<td>Bermuda (BER)</td>
<td>−65, 32</td>
<td>—</td>
<td>—</td>
<td>Yes/500 × 700 (parallel × perp to plate motion)</td>
<td>No</td>
<td>—</td>
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<tr>
<td>Bouvet (BOU)</td>
<td>3.4, −54.4</td>
<td>—</td>
<td>?</td>
<td>Yes/900</td>
<td>—</td>
<td>206Pb/204Pb, may be 87Sr/86Sr and 3He/4He</td>
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(Continued)
<table>
<thead>
<tr>
<th>Name (abbreviation)</th>
<th>Hot spot E. Long., N. Lat.</th>
<th>Age progression?</th>
<th>Age range</th>
<th>Swell?/width (km)</th>
<th>Connection to LIP?</th>
<th>Geoch. distinct from MORB</th>
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<tr>
<td>Cameroon (CAM)</td>
<td>6, –1</td>
<td>No</td>
<td>1–32 Ma</td>
<td>Yes/500–600</td>
<td>No</td>
<td>$^{206}$Pb/$^{204}$Pb</td>
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<td>Canaries (CAN)</td>
<td>–17, 28</td>
<td>Ok</td>
<td>0–68 Ma</td>
<td>No</td>
<td>No</td>
<td>$^{206}$Pb/$^{204}$Pb</td>
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<td>Cape Verde (CAP)</td>
<td>–24, 15</td>
<td>No</td>
<td>Neogene</td>
<td>Yes/800</td>
<td>No</td>
<td>$^{87}$Sr/$^{86}$Sr and $^{206}$Pb/$^{204}$Pb</td>
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<td>Discovery (DIS)</td>
<td>–6.45, –44.45</td>
<td>—</td>
<td>25 Ma</td>
<td>Yes/600</td>
<td>No</td>
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<td>Fernando Do Norona (FERN)</td>
<td>–32, –4</td>
<td>—</td>
<td>—</td>
<td>Yes/200–300</td>
<td>—</td>
<td>$^{87}$Sr/$^{86}$Sr and $^{206}$Pb/$^{204}$Pb</td>
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<td>Great Meteor (GM)</td>
<td>–28.5, 31</td>
<td>—</td>
<td>—</td>
<td>Yes/800</td>
<td>—</td>
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<td>Iceland (ICE)</td>
<td>–17.58, 64.64</td>
<td>Yes</td>
<td>0–62 Ma</td>
<td>Yes/2700</td>
<td>N. Atlantic LIP</td>
<td>$^{3}$He/$^{4}$He</td>
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<td>Jan Mayen (JM)</td>
<td>–8, 71.17</td>
<td>—</td>
<td>—</td>
<td>Yes</td>
<td>N. Atlantic LIP?</td>
<td>$^{87}$Sr/$^{86}$Sr</td>
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<td>Madeira (MAD)</td>
<td>–17.5, 32.7</td>
<td>Yes</td>
<td>0–67 Ma</td>
<td>No</td>
<td>No</td>
<td>$^{206}$Pb/$^{204}$Pb</td>
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<td>New England (NEW)</td>
<td>–57.5, 35</td>
<td>Yes</td>
<td>81–103, 122–124 Ma</td>
<td>No</td>
<td>No</td>
<td>$^{206}$Pb/$^{204}$Pb, may be $^{87}$Sr/$^{86}$Sr</td>
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<td>Shona (SHO)</td>
<td>–4, –52</td>
<td>—</td>
<td>not dated</td>
<td>Yes/$^{700}$</td>
<td>—</td>
<td>—</td>
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<tr>
<td>Sierra Leone (SL)</td>
<td>–29, 1</td>
<td>—</td>
<td>not dated</td>
<td>—</td>
<td>It could be a LIP</td>
<td>—</td>
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<tr>
<td>St. Helena (SHE)</td>
<td>–10, –17</td>
<td>Yes</td>
<td>3–81 Ma</td>
<td>Yes/720</td>
<td>No</td>
<td>$^{206}$Pb/$^{204}$Pb</td>
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<td>Trindade-Martin Vaz (TRN)</td>
<td>–12.2, –37.5</td>
<td>Probably</td>
<td>&lt;1 Ma to ~85 Ma</td>
<td>Yes/1330</td>
<td>Small eruptions north of Paraná</td>
<td>$^{87}$Sr/$^{86}$Sr and $^{206}$Pb/$^{204}$Pb</td>
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<td>Tristan-Gough (TRI)</td>
<td>–9.9, –40.4</td>
<td>Yes</td>
<td>0.5–80 Ma and 130 Ma</td>
<td>Yes/850</td>
<td>Rio Grande-Walvis and Paraná-Entendeka</td>
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<td>Vema (VEM)</td>
<td>16, –32</td>
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<td>&gt;11 Ma</td>
<td>Yes/200–300</td>
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<td>Indian</td>
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<td>Amsterdam-St. Paul (AMS)</td>
<td>77, –37</td>
<td>No</td>
<td>—</td>
<td>Yes/300–500</td>
<td>May be Kerguelen</td>
<td>may be $^{87}$Sr/$^{86}$Sr</td>
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</tbody>
</table>

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<table>
<thead>
<tr>
<th>Location</th>
<th>Latitude</th>
<th>Longitude</th>
<th>ESR/86Sr</th>
<th>Age (Ma)</th>
<th>Notes</th>
</tr>
</thead>
</table>
| Comores (COM) | 44, –12  | —         | Yes      | 0–5.4 Ma on island chain and  
~50 Ma (Seychelles)           | Yes/700–800 —                                                        |
| Conrad (CON)  | 48, –54  | —         | Not dated | Half-width 400 south of seamounts | It could be a LIP —                                                  |
| Crozet (CRO)  | 50, –46  | —         | Not dated | Yes/1120                      | May be Madagascar —                                                 |
| Kerguelen (KER)| 63, –49  | Yes       | 0.1–114 Ma (Kerg) and 38–82 Ma (Ninety east-Broken Ridge) | Yes/1310 | It is a LIP — |
| Marion (MAR)  | 37.75, –46.75 | Weak | <0.5 Ma (Marion) and 88Ma (Madagascar) | Half-width 500 or Along-axis >1700 | Madagascar Plateau and Madagascar Island flood basalts — |
| Réunion (REU) | 55.5, –21 | Yes       | 0–66 Ma  | Yes/1380                      | Mascarene, Chagos-Lacc. and Deccan 30–70 My — —                        |
| Africa        | —        | —         | —        | —                            | —                                                                      |
| Afar (AF)     | 42, 12   | No        | —        | —                            | —                                                                      |
| East Africa/Lake Victoria (EAF) | 34, 6 | No | — | — | — |
| Darfur (DAR)  | 24, 13   | No        | —        | —                            | —                                                                      |
| Ahaggar (AHA) | 6, 23    | No        | —        | —                            | —                                                                      |
| Tibesti (TIB) | 17, 21   | No        | —        | —                            | —                                                                      |
| Eurasia       | —        | —         | —        | —                            | —                                                                      |
| Eifel (EIF)   | 7, 50    | No        | —        | —                            | —                                                                      |

Horizontal bars indicate that there are no data, where we did not find any data, or where available data are inconclusive.
This is significantly longer than, for example, the activity of Hawaiian volcanoes which have a main stage lasting ~1 My (e.g., Clague and Dalrymple, 1987; Ozawa et al., 2005). The long life span of some of the Canary volcanoes has contributed to uncertainty in defining a geographic age progression, but is consistent with a slow propagation rate of age-progressive volcanism (Figure 3).

Iceland is often cited as a classic hot spot (Figure 4). The thickest magmatic crust occurs along the Greenland– and Faeroe–Iceland volcanic ridges extending NW and SE from Iceland. Anomalously thick oceanic crust immediately adjacent to these ridges shows datable magnetic lineations (Macnab et al., 1995; White, 1997; Jones et al., 2002b). Extrapolating the ages of these lineations onto the Greenland– and Faeroe–Iceland Ridges reveals age-progressive volcanism with seafloor spreading, which is most easily explained by the Iceland hot spot causing excess magmatism very near to or at the Mid-Atlantic Ridge (MAR) since the time of continental breakup (Wilson, 1973; White, 1988, 1997). Earlier volcanism occurs as flood basalts along the continental margins of Greenland, the British Isles, and Norway ~56 Ma, and even further away from Iceland in Baffin Island, West and East Greenland, and the British Islands beginning ~62 Ma (e.g., Lawver and Mueller, 1994; Saunders et al., 1997). The latter date provides a minimum estimate for the age of the Icelandic hot spot.

A few volcano chains fail to record volcanism for longer than a few tens of millions of years but the initiation of volcanism or the connection to older

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**Figure 1** (Continued)
volcanic provinces is somewhat unclear. For example, the New England seamount chain records ~20 My of oceanic volcanism (Duncan, 1984), but an extrapolation to the volcanic provinces in New England could extend the duration another 20 My (see O’Neill et al. (2005)). The duration of activity at the Azores hot spot is not clear. Gente et al. (2003) hypothesize that the Azores hot spot formed the Great Meteor and Corner seamounts as conjugate features ~85 Ma. Yet, age constraints of these edifices are poor and a geochemical association with the Azores group is yet to be tested. The most robust feature of this hot spot is its sudden influence on the MAR starting ~20 Ma as seen in geophysical surveys and dated using interpolations of seafloor isochrons (Cannat et al., 1999; Gente et al., 2003).

Figure 1  (a) Pacific region. Elevation on the continents and residual topography in the oceans. Residual topography is the predicted bathymetry grid of Smith and Sandwell (1997) corrected for sediment loading and thicknesses (Laske and Masters, 1997), and for seafloor subsidence with age (Stein and Stein, 1992) using seafloor ages, updated from Müller et al. (1993a) (areas without ages are interpolated using cubic splines). Grid processing and display was done using GMT (Wessel and Smith, 1995). Color change from blue to turquoise is at 300 m and delineates the approximate boundaries of anomalously shallow seafloor. Circles mark estimated locations of most recent (hot spot) volcanism. Pairs of lines are used to measure the widths of some of the hot-spot swells (Table 1). Flood basalt provinces on the continents and continental margins are red; abbreviations are identified in Section 7.09.2.3.2. Axes are in degrees latitude and east longitude. (b). Atlantic and Indian oceans.
On the Pacific Plate (e.g., see Clouard and Bonneville (2005) and references therein), the Cobb (1.5–29.2 Ma (Turner et al., 1980; Desonie and Duncan, 1990) and Bowie–Kodiak chains (0.2–23.8 Ma (Turner et al., 1980)) terminate at the subduction zone south of Alaska. The Easter chain on the Nazca Plate extends to 25.6 Ma (Clouard and Bonneville, 2005) but the record on the Pacific Plate may extend further into the past, perhaps to the Tuamotu Plateau (Ito et al., 1995; Clouard and Bonneville, 2001). This tentative connection certainly warrants more age dating.
7.09.2.1.2 Short-lived age-progressive volcanism

At least eight volcanic provinces show age-progressive volcanism lasting <22 My (e.g., Clouard and Bonneville, 2005). The Society and Marquesas (Caroff et al., 1999) islands represent voluminous volcanism but for geologically brief durations of 4.2 and 5.5 My, respectively. Clouard and Bonneville (2001) use geometrical considerations to argue that the Marquesas hot spot could have formed the Line Islands and Hess Rise. If this interpretation is correct, the large gap in volcanism between the three provinces suggests a strongly time-varying mechanism. Durations of 10–20 My occur along the Pitcairn (0–11 Ma), Caroline (1.4–13.9 Ma), Foundation (2.1–21 Ma), Tarava (35.9–43.5 Ma), and Pukapuka (5.6–27.5 Ma) chains in the Pacific, as well as the Tasmandid chain (7–24.3 Ma) (McDougall and Duncan, 1988; Müller et al., 1993b) near Australia.

An intriguing, yet enigmatic form of age-progressive volcanism is represented by the Pukapuka and Sojourn Ridges, which extend NW away from the East Pacific Rise. With respect to its geographic trend and duration, the Pukapuka Ridge resembles some of the other volcano chains in the region, such as the Foundation chain (Figure 5) (O’Connor et al., 2002, 2004). The Pukapuka Ridge, however, stands out because of its smaller volcano volumes and the more rapid and variable rate of age progression (Janney et al., 2000). It is thus unclear at this point whether these...
differences indicate deviant behaviors of a common mechanism or a distinct mechanism entirely. Batiza (1982) proposed a distinction between hot-spot volcanoes and smaller (and more numerous) ‘non-hot-spot’ volcanoes. The association of such seamounts with near-axis, mid-ocean ridge volcanism is good reason to consider a volcano group non-hot-spot, but such a characterization is less straightforward for Pukapuka.

While it projects to the region of the young Rano–Rani seamounts near the East Pacific Rise, most of Pukapuka formed on older seafloor (Figure 5). The above ambiguities blur the distinction between hot-spot and non-hot-spot volcanism.

### 7.09.2.1.3 No age-progressive volcanism

An important form of oceanic volcanism does not involve simple geographic age progressions. Amsterdam–St. Paul and Cape Verde are two examples that represent opposite extremes in terms of size and duration. Amsterdam–St. Paul is on the Southeast Indian Ridge, just NE of Kerguelen. Geochemical distinctions between lavas at Amsterdam–St. Paul and on Kerguelen suggest that they come from separate sources in the mantle (Doucet et al., 2004; Graham et al., 1999). With this interpretation, the Amsterdam–St. Paul hot spot represents a relatively small and short-lived (≤5 My) melting anomaly. Its small size and duration, as well as its location on a mid-ocean ridge likely contribute to the lack of an identifiable age progression. The Cape Verde volcanoes, on the other hand, are larger and likely to have existed since early Neogene (e.g., McNutt, 1988). The islands do not show a monotonic age progression, but this is reasonably well explained with a hot spot occurring very close to the Euler pole that describes the motion of the African Plate relative to a hot-spot reference frame (McNutt, 1988).

Other examples of volcanoes with complex age-space relations are not well explained within the hot-spot framework. These include older provinces in the Pacific such as the Geologist seamounts, the Japanese and Markus–Wake seamounts, the Marshal Islands,
the Mid-Pacific Mountains, and the Magellan Rise (e.g., Clouard and Bonneville (2005) and references therein). The lack of modern dating methods applied to samples of many of these provinces leads to significant age uncertainties, and the large number of volcanoes scattered throughout the western Pacific makes it difficult to even define volcano groups. Three notable examples show nonprogressive volcanism with ages confirmed with modern dating methods. The Line Islands and Cook–Austral groups are oceanic chains that both involve volcanism over tens of millions of years with synchronous or near-synchronous events spanning distances >2000 km (Figure 6) (Schlanger et al., 1984; McNutt et al., 1997; Davis et al., 2002; Devey and Haase, 2004; Bonneville et al., 2006). The Cameroon line, which extends from Africa to the SW on to the Atlantic seafloor (Figure 7) (Marzoli et al., 2000), may represent a continental analog to the former oceanic cases.

The remaining oceanic hot spots in Table 1 do not have sufficient chronological data to test for time–space relations. This list includes cases that are geographically localized (e.g., Discovery, Ascension/Circe, Sierra Leone, Conrad, Bermuda), have spatial distributions due to interactions with spreading centers (e.g., Bouvet, Shona, Balleny), or have elongated geographic trends but have simply not been adequately dated (Socorro, Tuamotu, Great Meteor).

### 7.09.2.1.4 Continental hot spots

In addition to the intraplate oceanic volcanism there are a number of volcanic regions in the continents that are not directly associated with present-day subduction or continental rifting. In Figure 1 we included a number of these areas such as the Yellowstone (YEL)–Snake River volcanic progression, the European Eifel hot spot (EIF), and African hot spots such as expressed by the volcanic
Mountains of Jebel Mara (DAR), Tibesti (TIB), and Ahaggar (AHA) (Burke, 1996). The Yellowstone hot spot is the only one with a clear age progression. The lack of age progression of the other hot spots could indicate very slow motion of the African and European Plates.

The Yellowstone hot spot is centered on the caldera in Yellowstone National Park. Its signatures include a topographic bulge that is 600 m high and approximately 600 km wide, high heat flow, extensive hydrothermal activity, and a 10–12 m positive geoid anomaly (Smith and Braile, 1994). The trace of the Yellowstone hot spot is recorded by silicic-caldera-forming events starting at 16–17 My at the Oregon–Nevada border, ~700 km WSW of the hot spot (Figure 8). The original rhyolitic volcanism is followed by long-lived basaltic volcanism that now forms the Snake River Plain. The effective speed of the hot-spot track is 4.5 cm yr\(^{-1}\), which is interpreted to include a component of the present-day plate motion (2.5 cm yr\(^{-1}\)) and a component caused by the Basin and Range extension. The excess topography of the Snake River Plain decays systematically and is consistent to that of a cooling and thermally contracting lithosphere following the progression of the American Plate over a hot spot (Smith and Braile, 1994).

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**Figure 7** Sketch map of the Cameroon Line. Reported ages refer to the volcanism on the continent and to the onset of basaltic volcanism on the ocean islands. Inset, top left: sketch map of Western Cameroon Highlands showing location of basaltic samples used for \(^{40}\)Ar/\(^{39}\)Ar dating. Inset, bottom right: West African Craton (WAC), Congo Craton (CC) and Kalahari Craton (KC). Reproduced from Marzoli A, Piccirillo EM, Renne PR, Bellieni G, Iacumin M, Nyobe JB, and Tongwa AT (2000) The Cameroon volcanic line revisited; petrogenesis of continental basaltic magmas from lithospheric and asthenospheric mantle sources. *Journal of Petrology* 41: 87–109, by permission of Oxford University Press.
7.09.2.1.5 The hot-spot reference frame

The existence of long-lived volcano chains with clear age progression led Morgan (1971, 1972) to suggest that hot spots remain stationary relative to one another and therefore define a global kinematic reference frame separate from the plates (Morgan, 1983; Duncan and Clague, 1985). However, ongoing studies have established that hot spots do move relative to the Earth’s spin-axis and that there is motion between the Pacific and Indo-Atlantic hot spots with speeds comparable to the average plate speed (Molnar and Stock, 1987; Acton and Gordon, 1994; Tarduno and Gee, 1995; DiVenere and Kent, 1999; Raymond et al., 2000; Torsvik et al., 2002). Paleomagnetic evidence (Tarduno et al., 2003; Pares and Moore, 2005) suggests rapid southward motion of the Kerguelen hot spot in the past 100 My and of the Hawaiian hot spot prior to the Hawaiian–Emperor Bend at 50 Ma (Sharpe and Clague, 2006). Southward motion of the Hawaiian hot spot appears to have slowed or nearly ceased since this time (Sager et al., 2005). Combined with a possible shift in plate motion (Norton, 2000; Sharpe and Clague, 2006) this change most likely contributes to the sharpness of the Hawaiian–Emperor Bend (Richards and Lithgow-Bertelloni, 1996). Near present-day (i.e., <4–7 Ma) motion between hot spots globally is unresolvable or at least much slower than it has been in the geologic past (Wang and Wang, 2001; Gripp and Gordon, 2002).

While past rapid motion between groups of hot spots in different oceans is likely, rapid motion between hot spots on single plates is not. Geometric analyses of volcano locations suggest no relative motion, to within error, between some of the larger Pacific hot spots (Harada and Hamano, 2000; Wessel and Kroenke, 1997). Such geometric methods are independent of volcano ages and therefore have the advantage of using complete data sets of known volcano locations, unhindered by sparse age dating of variable quality. Koppers et al. (2001), however, argue that geometric analyses alone are incomplete and that when age constraints are considered as well, motion between the large Hawaiian and Louisville hot spots is required. But Wessel et al. (2006) show that the hot-spot track predicted by Koppers et al. (2001) misses large sections of both chains and present an improved plate-motion model derived independently of age dating. Thus, at this point, there appears to be little or no motion between the prominent Hawaiian and Louisville hot spots. Shorter chains in the Pacific do appear to move relative to larger chains (Koppers et al., 1998;
Koppers and Staudigel, 2005) but as noted above, the existence of age progression along many short-lived chains is questionable. One enigma, in particular, is the apparent motion required between Iceland and both the Pacific and Atlantic hot spots (Norton, 2000; Raymond et al., 2000). For a more complete description of the methods and above issues regarding absolute plate motion, see Chapter 6.02.

7.09.2.2 Topographic Swells

Figure 1 illustrates that most oceanic hot spots and melt anomalies show anomalously shallow topography extending several hundred kilometers beyond the area of excess volcanism. The prominence of hot-spot swells as established in the 1970s was initially attributed to heat anomalies in the mantle (Crough, 1978, 1983; Detrick and Crough, 1978). The evidence for their cause is somewhat ambiguous: for one, heat flow data fail to show evidence for heated or thinned lithosphere (Stein and Stein, 1993, 2003; DeLaughter et al., 2002). However, it is likely that hydrothermal circulation associated with volcanic topography obscures the deep lithospheric signal (McNutt, 2002; Harris and McNutt, 2007). Seismic studies are limited but provide some clues. Rayleigh-wave dispersion shows evidence for a lithosphere of normal thickness beneath the Pitcairn hot spot (Yoshida and Suetsugu, 2004), whereas an S-wave receiver function study argues for substantial lithospheric thinning a few hundred kilometers up the Hawaiian chain from the hot spot (Li et al., 2004).

Shallower-than-normal topography along the axes of hot-spot-influenced ridges indicates another form of hot-spot swell. Gravity and crustal seismic evidence indicate that topography along the ridges interacting with the Galápagos (Canales et al., 2002) and Iceland (White et al., 1995; Hooft et al., 2006) hot spots are largely caused by thickened oceanic crust. In the case of the Galápagos spreading center, a resolvable contribution to topography likely comes from the mantle (Canales et al., 2002), but for Iceland, it appears that crustal thickness alone can explain the observed topography (Hooft et al., 2006).

An example of a much larger-scale swell is the South Pacific Superswell (McNutt and Fischer, 1987) which spans a geographic extent of ~3000 km and encompasses the hot spots in French Polynesia. Other swells of comparable size include the ancient Darwin Rise in the far northwestern Pacific (McNutt et al., 1990) and the African Superswell (e.g., Nyblade and Robinson, 1994), which encompasses the southern portion of Africa and the South Atlantic down to the Bouvet Triple Junction. Both the South Pacific and African Superswells involve clusters of individual hot spots. The hot spots in the South Pacific (French Polynesia) are mostly short lived, including some without simple age progressions (see Section 7.09.2.1) (McNutt, 1998; McNutt et al., 1997). This region is also known to have anomalously low seismic wave speeds in the mantle (e.g., McNutt, 1998). The African Superswell, on the other hand, does not show a seismic anomaly in the upper mantle but rather a broad columnar zone of slow velocities in the lower mantle (e.g., Dziewonski and Woodhouse, 1987; Li and Romanowicz, 1996; Ritsema et al., 1999).

For individual hot spots, we identify the presence of a swell if residual topography exceeds an arbitrarily chosen value of 300 m and extends appreciably (>~100 km) away from volcanic topography (Figure 1). We find that such hot-spot swells are very common (Table 1). Swells are even apparent on chains with very small volcanoes such as the Tasmandid tracks and the Pukapuka (Figure 5) and Sojourn Ridges (Harmon et al., 2007).

One characteristic of hot-spot swells is that they are not present around extremely old volcanoes (Table 1). The Hawaiian Swell for example is prominent in the youngest part of the chain but then begins to fade near ~178° W, disappears near the Hawaiian–Emperor Bend (50 Ma), and is absent around the Emperor chain (Figure 1). Swells also appear to fade along the Louisville chain (near a volcano age of ~34 Ma), along the Tristan chain near Walvis Ridge (62–79 Ma), and possibly along the Kerguelen track as evident from the shallow seafloor that extends away from the Kerguelen Plateau on the Antarctic Plate but that is absent around the southermost portion of Broken Ridge (~43 Ma) on the Indian Plate. A swell appears to be present over the whole length of the St Helena chain out to ~80 Ma, but it is not possible with the current data to separate the swell around the oldest portion of this chain with that around the Cameroon line. The SE portion of the Line Islands (ages ranging from 35 to 91 Ma) most likely has a swell but the NE portion (55–128 Ma) may not. Given the diversity of possible times of volcanism in the Line Islands, however, it is not clear which episode is associated with the current
swell. Old chains that lack swells in our analysis include the Japanese–Wake seamounts (>70 Ma), the Magellan Seamounts (>70 Ma), the Mid-Pac (>80 Ma), and the Musician Seamounts (>65.6 Ma). Overall, it thus appears that if a swell forms at a hot spot, it decreases in height with time until it can no longer be detected at an maximum age of 80 Ma, but often even after <50 Ma.

While swells are very prominent even around small or short-lived volcano chains, the Madeira and Canary hot spots are two cases that break the rule. The lack of obvious swells around these large and long-lived volcano chains is indeed very puzzling.

7.09.2.3 Flood Basalt Volcanism

LIPs provide further constraints on the nature of hot spots and mantle dynamics. In this chapter we use the term LIP to represent continental flood basalt provinces, oceanic plateaus, and volcanic passive margins, which are typified by massive outpourings and intrusions of basaltic lava, often occurring within a couple of million years. Reviews of the nature and possible origin of LIPs are provided by Richards et al. (1989); Coffin and Eldholm (1994); Mahoney and Coffin (1997); and Courtillot and Renne (2003). As characterized by Coffin and Eldholm (1994), continental LIPs, such as the Siberian Traps or the Columbia River Basalts, often form by fissure eruptions and horizontal flows of massive tholeiitic basalts. Volcanic passive margins, such as those of the North Atlantic Volcanic Province or Etendeka–Paraná, form generally just before continental rifting. The initial pulse is rapid and can be followed by a longer period of excess oceanic crust production and long-term generation of a seamount chain. Oceanic LIPs form broad, flat-topped features of thickened oceanic crust with some eruptions being subaerial (Kerguelen oceanic plateau) and others seeming to be confined to below sea level (e.g., Ontong Java Plateau (OJP) and Shatsky Rise). This section reviews some basic geophysical and geological observations of LIPs that formed since the Permain. For information on older LIPs we refer the reader to other reviews (Ernst and Buchan, 2001, 2003). Figure 1 shows locations (abbreviations defined in the following text), and Figure 9 summarizes the areas and volumes of the provinces described as follows.

7.09.2.3.1 Continental LIPs

**Columbia River Basalts (CRBs).** The CRBs erupted a volume of ~0.17 Mkm$^3$ between 16.6–15.3 Ma (Courtillot and Renne, 2003). Excellent exposures provide insights into flow structures and relationships to feeder dikes. Individual eruptions have volumes in excess of 2000 km$^3$ and flow over distances up to 600 km (Hooper, 1997). Lack of collapse structures suggests that large amounts of magma were rapidly derived from a deep source without being stored in the crust (Hooper, 1997).

**Emeishan (EM).** The EM province in west China is estimated to have spanned an area of at least 2 Mkm$^2$ and volume of 1 Mkm$^3$ when it first formed (Zhou et al., 2002). Eruption ages are dated at 251–255 Ma and ~258 Ma (e.g., see Courtillot and Renne (2003), and Zhang et al. (2006) and references therein), with a date of ~259 Ma being confirmed by a Zircon U–Pb dating (Ali et al., 2005). A rapid, kilometer scale uplift appears to have preceded the basalt eruption by

![Figure 9](image) **Figure 9** Histograms of estimated areas affected by magmatism and magmatic volumes of the large igneous provinces discussed in the text. The OJP–Manihiki–Hikurangi LIP would have the area shown by white bar and an even larger volume than shown, as indicated by arrow. Asterisks indicate provinces with voluminous magmatism that could have endured for >3 My to a few tens of millions of years; no asterisks indicate cases that mostly likely erupted in <3 My.
Basalts erupted rapidly and were accompanied by high MgO basalts (He et al., 2003; Zhang et al., 2006). Since, the Emeishan traps have been fragmented and eroded, they currently encompass an area of only ~0.3 Mkm$^2$ (Xu et al., 2001).

**Siberia (SIB).** The Siberian Traps are presently exposed over an area of only 0.4 Mkm$^2$ and have an average thickness of 1 km (Sharma, 1997). There are strong indications that the volcanics extend below sedimentary cover and into the West Siberian Basin (Reichow et al., 2005). Additional dikes and kimberlites suggest a maximum extent of 3–4 Mkm$^2$ with a possible extrusive volume of >3 Mkm$^3$. Lack of significant sedimentary rocks or paleosols between flows suggests rapid extrusion (Sharma, 1997). Most of the province probably erupted within ~1 My coinciding with the Permo-Triassic boundary at 250 Ma (Courtillot and Renne, 2003). Individual flows can be as thick as 150 m and can be traced over lengths of hundreds of kilometers. Use of industry seismic and borehole data in the West Siberian Basin indicates that the basin elevation remained high during rifting, suggesting dynamic mantle support (Saunders et al., 2005).

**Yemen/Ethiopia/East Africa Rift System (EAF).** An early volcanic episode in southernmost Ethiopia starting ~45 Ma was followed by widespread flood basalt volcanism in Northwest Ethiopia ~30 Ma and in Yemen starting 31–29 Ma. The 30 Ma Ethiopia event consisted of tholeiites and ignimbrites (Pik et al., 1998) that erupted within 1–2 My (Hofmann et al., 1997; Ayalew et al., 2002). In Yemen, 0.35–1.2 Mkm$^2$ of mafic magmas were produced, followed by less voluminous silicic volcanism starting ~29 Ma (Menzies et al., 1997). Flood volcanism appears to have occurred several million years prior to the onset of extension along the EAR ~23 Ma (Morley et al., 1992; Hendrie et al., 1994) and in the Gulf of Aden ~26 Ma (Menzies et al., 1997). The volcanism is bimodal with shield volcanoes forming on top of tholeiitic basalts (Courtillot and Renne, 2003; Kieffer et al., 2004). Currently, the Ethiopian and Kenyan Rift Systems are on an area of elevated topography ~1000 km in diameter. A negative Bouguer gravity anomaly suggests this topography is dynamically supported in the mantle (Ebbing et al., 1989).

**Older continental flood basalts:** These are often more difficult to detect in the geological record due to the effects of surface uplift and erosion. A general characteristic that is attributed to continental LIPs is radiating dike swarms (Mege and Korme, 2004; Mayborn and Lesher, 2004). These dike swarms provide the main pathways for basaltic magmas vertically from the mantle, as well as laterally over distances up to 2500 km, as suggested for the 1270–3 My-old Mackenzie dike swarm in N. America (Lecheminant and Heaman, 1989; Ernst and Baragar, 1992). The longest dikes usually extend well beyond the original boundaries of the main lava field.

### 7.09.2.3.2 LIPs near or on continental margins

**Central Atlantic Magmatic Province (CAMP).** The CAMP is primarily delineated by giant dike swarms and is associated with the early breakup of Gondwana between North Africa, North America, and Central South America. Widely separated eruptions and dike swarms are present over an area of ~7 Mkm$^2$ with an estimated magmatic volume of ~2 Mkm$^3$ (Marzoli et al., 1999), while seismic and magnetic studies on the eastern margin of North America suggest that this offshore portion of the CAMP could have a volume as large as 3 Mkm$^3$ (Holbrook and Kelemen, 1993). These estimates bring the total volume to near 5 Mkm$^3$. $^{40}$Ar/$^{39}$Ar dates spanning 197–202 Ma suggests an emplacement episode lasting ~5 My (Hames et al., 2000; Marzoli et al., 1999; Courtillot and Renne, 2003). Only the offshore portion is mapped in Figure 1.

**Chon Aike (CHON).** In contrast to the other large igneous provinces discussed in the chapter, the Chon Aike Province in Patagonia is primarily silicic with rhyolites dominating over minor mafic and intermediate lavas. The rhyolites may have formed due to intrusion of basalts into crust that was susceptible to melting. The province is relatively small with an area of 0.1 Mkm$^2$ and total volume of 0.235 Mkm$^3$ (Pankhurst et al., 1998). Chon Aike had an extended and punctuated eruptive history from Early Jurassic through Early Cretaceous (184–140 Ma). Pankhurst et al. (2000) recognize episodic eruptions with the first coinciding with the Karoo and Ferrar LIPs. The province potentially extends into present-day West Antarctica.

**Deccan (DEC).** The Deccan Traps provide one of the most impressive examples of continental flood basalts. It formed by primarily tholeiitic magmatism over Archean crust interspersed over an area of ~1.5 Mkm$^2$, with an estimated volume of 8.2 Mkm$^3$ (Coffin and Eldholm, 1993). Eruptions straddle the magnetic chron C30n, C29r, C29n within 1 My.
around the K–T boundary, as confirmed by 40Ar–39Ar and Re–Os dating (Allègre et al., 1999; Courtillot et al., 2000; Hofmann et al., 2000). An iridium anomaly embedded between flows suggests that the Chicxulub impact happened while the Deccan Traps were active (Courtillot and Renne, 2003). Seafloor spreading between India and the Seychelles started a few million years after the major Deccan event, ~63 Ma (Vandamme et al., 1991; Dyment, 1998). Unlike older continental flood basalts associated with the breakup of Gondwana, the Deccan basalts that are least contaminated by continental lithosphere closely resemble hot-spot basalts in oceanic areas and the major element contents agree with predictions for high-temperature melting (Hawkesworth et al., 1999).

**Karoo–Ferrar** (KAR–FER). The Karoo province in Africa and Ferrar basalts in Antarctica record a volume of 2.5 Mkm$^3$ which erupted at ~184 Ma (Encarnacion et al., 1996; Minor and Mukasa, 1997), possibly followed by a minor event at 180 Ma (Courtillot and Renne, 2003). The short (<1 My) duration is questioned by Jourdan et al. (2004, 2005) who obtained ages of ~179 Ma for the northern Okavango dike swarm in Botswana and consequently prefer a longer-lived initial activity that propagated from the south to the north. In Africa, tholeiitic basalts dominate but some picrites and some rhyolites occur (Cox, 1988). The triple-junction pattern of the radiating dike swarm that supplied the Karoo basalts was likely controlled by pre-existing lithospheric discontinuities that include the Kaapvaal and Zimbabwe Craton boundaries and the Limpopo mobile belt (Jourdan et al., 2006). The Ferrar Province spans an area of ~0.35 Mkm$^2$ (Elliot and Fleming, 2004) in a linear belt along the Transantarctic Mountains. The two provinces were split by continental rifting and then seafloor spreading ~156 Ma.

**Madagascar** (MDR). Wide-spread voluminous basaltic flows and dikes occurred near the northwestern and southeastern coasts of Madagascar during its rifting from India around 88 Ma (Storey et al., 1997). The oldest seafloor magnetic anomaly to form is chron 34 (~84 Ma). Flood volcanism was probably prolonged as it continued to form the Madagascar Plateau to the south, perhaps 10–20 My later as inferred from the reconstructed positions of Marion hot spot.

**North Atlantic Volcanic Province** (NAVP). The NAVP covers ~1.3 Mkm$^2$ (Saunders et al., 1997) with an estimated volume of 6.6 Mkm$^3$ (Coffin and Eldholm, 1993) and is closely linked to continental rifting and oceanic spreading (e.g., Nielsen et al., 2002) (Figure 2). Prior to the main pulse of flood volcanism, seafloor spreading was active south of the Charlie–Gibbs fracture zone at 94 Ma and propagated northward into the Rockall Trough, which stopped in the late Cretaceous near or prior to the earliest eruptions of the NAVP. The early NAVP eruptions occurred as large picritic lavas in West Greenland and Baffin Island (Gill et al., 1992, 1995; Holm et al., 1993; Kent et al., 2004) soon followed by massive tholeiitic eruptions in West and Southeast Greenland, British Isles, and Baffin Island at 61 Ma (2 Mkm$^3$) and in East Greenland and the Faeroes at 56 Ma (>2 Mkm$^3$) (Courtillot and Renne, 2003). The initial episodes were followed by rifting between Greenland and Europe recorded by Chron 24, 56–52 Ma, continental margin volcanisms, and ocean crust formation, which included the formation of thick seaward-dipping seismic reflection sequences. Spreading slowed in Labrador Sea ~50 Ma, stopped altogether at 36 Ma, but continued further to the west on the Aegir Ridge and eventually along the Kolbeinsey Ridge at ~25 Ma, where it has persisted since. This provides an intriguing suggestion that the presence of hot spots can guide the location of seafloor spreading following continental breakup.

Many of the volcanic margin sequences erupted subaerially or at shallow depths, suggesting widespread regional uplift during emplacement (Clift and Turner, 1995; Hopper et al., 2003). Uplift in the Early Tertiary is documented by extensive erosion and changes in the depositional environments as far as the North Sea Basin (e.g., see Nadin et al. (1997) and Mackay et al. (2005) and references therein). Reconstructions from drill cores show that uplift was rapid and synchronous and preceded the earliest volcanism by >1 My (Clift et al., 1998).

**Paraná–Etendeka** (PAR–ET). Paraná and Etendeka are conjugate volcanic fields split by the breakup of South America and Africa. The Paraná field in South America covers 1.2 Mkm$^2$ with estimated average thickness of 0.7 km (Peate, 1997). Extensive dike swarms surrounding the provinces suggest the original extent could have been even larger (Trumbull et al., 2004). Volcanism is bimodal with dominating tholeiitic lavas and rhyolites. 40Ar–39Ar dates suggest a peak of eruption ~133–130 Ma (Turner and al., 1994; Renne et al., 1996; Courtillot and Renne, 2003), preceded by minor eruptions in the northwest of the Paraná basin at 135–138 Ma (Stewart et al., 1996). Younger magmatism persisted along the coast.
(128–120 Ma) and into the Atlantic Ocean, subsequently forming the Rio Grande (RIO) and Walvis (WAL) oceanic plateaus.

The Etendeka province covers 0.08 Mkm$^2$ and is very similar to the Paraná flood basalts in terms of eruptive history, petrology, and geochemistry (Renne et al., 1996; Ewart et al., 2004). Seafloor spreading in the South Atlantic progressed northward, with the oldest magnetic anomalies near Cape Town (137 or 130 Ma). Earliest magnetic anomaly near Paraná is ≈127 Ma. The formation of onshore and offshore basins suggests a protracted period of rifting well in advance of the formation of oceanic crust and the emplacement of the Paraná basalts (Chang et al., 1992).

### 7.09.2.3.3 Oceanic LIPs

**Caribbean (CBN).** The Caribbean LIP is a Late Cretaceous plateau, which is now partly accreted in Colombia and Ecuador. Its present area is 0.6 Mkm$^2$ with thickness of oceanic crust ranging from 8–20 km. A volume of 4 Mkm$^3$ of extrusives erupted in discrete events from 91–88 Ma (Courtillot and Renne, 2003). The full range of $^{40}$Ar/$^{39}$Ar dates of 69–139 Ma (e.g., Sinton et al., 1997; Hoernle et al., 2004), however, suggests a protracted volcanic history that is poorly understood. The cause volcanism has been attributed to the Galápagos hot spot, which is currently in the Eastern Pacific.

**Kerguelen (KER).** The Kerguelen hot spot has a fascinating history of continental and oceanic flood eruptions and rifting, as well as prolonged volcanism (Figure 2). The breakup of India, Australia, and Antarctica coincided closely in time with the eruption of the Bunbury basalts in southwest Australia, dated at 123 and 132 Ma (Coffin et al., 2002). The first massive volcanic episode formed the Southern Kerguelen Plateau at 119 Ma, the Rajmahal Traps in India at 117–118 Ma followed by lamprophyres in India and Antarctica at 114–115 Ma (Coffin et al., 2002; Kent et al., 2002). The Central Kerguelen Plateau formed by ~110 Ma and Broken Ridge formed by 95 Ma (Coffin et al., 2002; Frey et al., 2000; Duncan, 2002). The above edifices represent the most active period of volcanism with a volcanic area and volume of 2.3 Mkm$^2$ and 15–24 Mkm$^3$, respectively (Coffin and Eldholm, 1993), but unlike many other flood basalt provinces, it spanned tens of millions of years. Another unusual aspect is the presence of continental blocks as suggested from wide-angle seismics (Operto and Charvis, 1996) and more directly from trace-element and isotopic data of xenoliths and basalts from Southern Kerguelen Plateau and Broken Ridge (Mahoney et al., 1995; Neal et al., 2002; Frey et al., 2002). During 82 to 43 Ma northward motion of the Indian Plate formed the Ninetyeast Ridge on young oceanic lithosphere, suggesting the Kerguelen hot spot stayed close to the Indian–Antarctic spreading center (Kent et al., 1997). At ~40 Ma, the Southeast Indian Ridge formed and separated Kerguelen and Broken Ridge. Volcanism has since persisted on the Northern Kerguelen Plateau until 0.1 Ma (Nicolas et al., 2000). Dynamic uplift of the plateau in the Cretaceous is indicated by evidence for subaerial environment, but subsidence since then is not much different than normal oceanic subsidence (Coffin, 1992).

**Ontong Java Plateau (OJP).** A recent set of overview papers on the origin and evolution of the OJP is provided by Fitton et al. (2004) and Neal et al. (1997). This plateau extends across ~2 Mkm$^2$, has crust as thick as 36 km, and has volumes estimated at 44 Mkm$^3$ and 57 Mkm$^3$ for accretion off, and on a mid-ocean ridge, respectively (Gladczenko et al., 1997). The majority of the basalts were erupted in a relatively short time (~1–2 My) near 122 Ma as revealed by $^{40}$Ar/$^{39}$Ar (Tejada et al., 1996), Re–Os isotope (Parkinson et al., 2002), and paleomagnetic (Tarduno et al., 1991) studies. Basalts recovered from Site 803, and the Santa Isabel and Ramos Islands indicate a smaller episode of volcanism at 90 Ma (Neal et al., 1997). Age contrasts between the surrounding seafloor and OJP suggest it erupted near a mid-ocean ridge. Furthermore, rifting is evident on some of its boundaries and Taylor (2006) suggests that the OJP is only a fragment of what was originally an even larger edifice that included Manihiki (MAN) and Hikurangi (HIK) Plateaus. The inferred original size of the edifice makes it, by far, the largest of any flood basin province in the geologic record, approaching an order of magnitude more voluminous than any continental flood basalt.

The plateau has been sampled on the tectonically uplifted portions in the Solomon Islands (e.g., Tejada et al., 1996, 2002; Petterson, 2004) and with eight ocean drill holes (most recently during Ocean Drilling Program Leg 192; Mahoney et al. 2001). The volcanics are dominated by massive flows of low-K tholeiitic basalts. Petrological modeling is consistent with the primary magmas formed by 30% melting of a peridotitic source (Fitton and Godard, 2004; Herzberg, 2004b; Chazey III and Neal, 2004), which would require a hot (1500°C) mantle under thin lithosphere. The low volatile
content of volcanic glasses (Roberge et al., 2004, 2005), as well as the range and limited variability of Pb, Sr, Hf, and Nd isotope compositions (Tejada et al., 2004) resemble characteristics of many MORB.

Besides its gigantic volume, the other aspect that makes OJP extremely enigmatic is that it appears to have erupted below sea level with little evidence for hot-spot-like uplift (Roberge et al., 2005; Korenaga, 2005b; Ingle and Coffin, 2004; Coffin, 1992; Ito and Clift, 1998). These aspects must be explained by any successful model for the origin of this flood basalt province.

Shatsky–Hess Rise (SHA–HES). Shatsky Rise is one of the large Pacific oceanic plateaus with an area of 0.48 Mkm² and volume of 4.3 Mkm³. The initial eruption is associated with the jump of a triple junction between the Pacific, Izanagi, and Farallon Plates toward the plateau (Nakanishi et al., 1999; Sager, 2005) near 145 Ma (Mahoney et al., 2005). Subsequently, volcanism progressed northeast together with the triple junction (which migrated with repeated jumps as indicated by seafloor magnetic lineation) until ~128 Ma. The most voluminous portion of the plateau is the central-southwest portion. Magmatism appears to have diminished toward the northeast (Figure 1). Thus Shatsky appears to show a short-lived age-progressive volcanism on timescales much like many smaller volcano groups (e.g., Cook–Austral and Pukapuka). Volcanism, however, may have continued with a renewed pulse starting some 10–20 My later with the formation of the Hess Rise, which is comparable in area to Shatsky. Age constraints on Hess Rise are poor because of the lack of sampling and its location on Cretaceous Quite Zone seafloor. The possible coincidence of both plateaus at a mid-ocean ridge suggests a dynamic linkage between their formation and seafloor spreading. Another notable aspect is that Pb and Nd isotope compositions for Shatsky Rise are indistinguishable from those of the present-day East Pacific Rise (Mahoney et al., 2005).

7.09.2.3.4 Connections to hot spots

The possible links between hot spots and LIPs are important for testing the origin of both phenomena, with particular regard to the concept of a starting mantle plume head and trailing, narrower plume stem. While linkages are clear for some cases, a number of proposed connections are obscured by ridge migrations or breakup of the original LIP. Below we list the connections of hot spots to LIPs, in approximate order of decreasing reliability.

At least six examples have strong geographical, geochronological, and geochemical connections between hot-spot volcanism and flood basalt provinces. These are: (1) Iceland and the North Atlantic Volcanic Province, including Greenland, Baffin Island, Great Britain volcanics, Greenland–UK (Faeroe) Ridge (Saunders et al., 1997; Smallwood and White, 2002); (2) Kerguelen, and Bunbury, Naturaliste, Rajmahal (E. India), Broken Ridge, and Ninetyeast Ridge (Kent et al., 1997); (3) Réunion and Deccan (Roy, 2003), W. Indian, Chagos–Laccadive, Mascarenas, Mauritius; (4) Marion and Madagascar (Storey et al., 1997); (5) Tristan da Cunha and Paraná, Etendeka, Rio Grande, Walvis Ridge (Peate, 1997); (6) Galapagos and Caribbean (Feigenson et al., 2004; Hoernle et al., 2004).

In addition to these six examples, a tentative link exists between Yellowstone and the early eruptive sequence of the CRBs based on geochemistry (Dodson et al., 1997), but the paleogeographical connection is somewhat indirect since the main CRB eruption occurred up to 500 km north of the hot-spot track. This may suggest a mechanism for the formation of the flood basalts that is independent of the Yellowstone hot spot (Hales et al., 2005). Yet, the earliest manifestation of the CRBs is possibly the Steens Mountain basalt (Oregon) which is located well to the south of the main eruptive sequences (Hooper, 1997; Hooper et al., 2002) and closer to the proposed location of the Yellowstone hot spot. This suggests a south-to-northward propagation of the basalts and supports the suggestion that flood basalts may have been forced sideways from their mantle source by more competent continental lithosphere toward a weaker ‘thin spot’ in the lithosphere (Thompson and Gibson, 1991).

The links are less clear—in part due to lack of data—for other flood basalt provinces. The Bouvet hot spot has been linked to the Karoo–Ferrar; the Balleny hot spot may be connected to the Tasmanian Province or to Lord Howe Rise (Lanyon et al., 1993); and the Fernando hot spot has been linked to the Central Atlantic Magmatic Province. Only three Pacific hot spots possibly link back to LIPs: Louisville–OJP, Easter–Mid-Pac, and Marquesas–Hess-Shatsky (Clouard and Bonneville, 2001). The Louisville–OJP connection is doubtful at best: kinematic arguments against such a link have been made (Anttreter et al., 2004) and geochemical distinctions between the oldest Louisville seamounts and OJP would require distinct geochemistry between plume head and tail or a difference in melting conditions (Mahoney et al., 2005b, 2005c, 2005d, 2005e; Ingle and Coffin, 2004; Coffin, 1992; Ito and Clift, 1998). These aspects must be explained by any successful model for the origin of this flood basalt province.

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island basalts (OIBs) have much larger variability, ranges for all three isotope ratios. In contrast, ocean-systems vary individually, each fall within the above while the median values from the three major spreading spots and melting anomalies has played a key role in the origin of many flood basalt provinces (Anderson, 1994b).

7.09.2.4 Geochemical Heterogeneity and Distinctions from MORB

The geochemistry of MORBs and basalts from hot spots and melting anomalies has played a key role in our understanding of mantle dynamics. Isotope ratios of key trace elements reflect long-term (10^2–10^3 My) concentration ratios between parent and daughter elements and therefore have been used to fingerprint the mantle from which different lavas arise. We will focus on three commonly used ratios: 87Sr/86Sr, 206Pb/204Pb, and 3He/4He. High (or low) 87Sr/86Sr is associated with mantle material that is enriched (or depleted) in highly incompatible elements (i.e., those that partition mostly into melt when in equilibrium with the solid during partial melting) relative to moderately incompatible elements (e.g., Rb/Sr). High (or low) 206Pb/204Pb ratios are associated with mantle with a long-term high (or low) U/Pb. The 3He/4He ratio is a measure of the amount of primordial helium (3He, which was present in the presolar nebula and is only lost from the Earth to space via degassing) relative to 4He, which is primarily generated by radioactive decay of U and Th.

MORB is well characterized by low and small variability in the above ratios. Figure 10 shows frequency distributions of basalts from the three major oceans: the data are from http://www.petdb.org/ and absolutely no data (e.g., Iceland) are excluded. One standard deviation about the median for all three datasets combined defines the MORB range for 87Sr/86Sr of 0.7025–0.7033, for 206Pb/204Pb of 17.98–18.89, and for 3He/4He of 7.08–10.21 (where 4He/3He is given in multiples of the atmosphere ratio, R_a). Significantly, while the median values from the three major spreading systems vary individually, each fall within the above ranges for all three isotope ratios. In contrast, ocean-island basalts (OIBs) have much larger variability, extending from MORB values to much higher values (Figure 10). By comparing the median hot-spot compositions (solid bar) with the MORB ranges, it becomes clear that, with few exceptions, the hot spots and melting anomalies have compositions that are distinguishable from MORB by at least one of the three isotope ratios (see also Table 1). This result appears to be independent of the duration of age-progression (e.g., Tristan vs. Marquesas), the presence or absence of a swell (Hawaii vs. Canaries), or even volcano size (Kerguelen vs. Pukapuka (Janney et al. (2000)) and Foundation (Maia et al. (2000))).

There are at least four possible exceptions: Cobb, Bowie-Kodiak, the Caroline seamounts, and the Shatsky Rise. Helium isotopes are not yet available at these locations but 87Sr/86Sr and 206Pb/204Pb compositions for each case fall within or very near to the MORB range. These examples span a wide range of forms, from a small, short-lived seamount chain (Caroline), to longer-lived, age-progressive volcanism (Cobb, Bowie-Kodiak), to an oceanic LIP (Shatsky). The possibility that these cases are geochemically indistinguishable from MORB has far-reaching implications about mantle processes and chemical structure; it clearly needs testing with further sampling.

7.09.2.5 Mantle Seismic Anomalies

7.09.2.5.1 Global seismic studies

Seismic wave propagation is generally slowed by elevated temperature, volatile content, the presence of melt, and mafic (primarily garnet) content of the mantle (e.g., Anderson, 1989). Seismology is therefore the primary geophysical tool for probing the mantle signature of hot spots and melting anomalies. Seismic tomography has become a popular method of ‘imaging’ the mantle. While it has provided important insights into the deep transport of subducting slabs (e.g., Grand et al., 1997; Bijwaard et al., 1998), seismic tomography has yet to produce robust images with sufficient spatial resolution in the deep mantle beneath hot spots (Nataf, 2000; Ritsema and Allen, 2003). The lack of methods that can probe the lower mantle with sufficient resolution makes it particularly difficult to address questions regarding the putative lower-mantle source for hot spots.

Global seismic models provide a first attempt to trace the surface expressions of hot spots to seismic anomalies into the mantle (Niu et al., 2002; Zhao, 2001; DePaolo and Manga, 2003; Montelli et al., 2004). The most robust features are the voluminous low-velocity
Figure 10  Frequency distributions (dark gray, normalized by maximum frequency for each case, so the peaks are at 1.0) of isotope measurements taken from the shown oceanic hot spots and mid-ocean ridges (lower left). Solid lines mark median values and dashed lines encompass 68% (i.e., one standard deviation of a normal distribution) of the samples. Light gray bars denote the range of values encompassing 68% of all of the MORB measurements (sum of the three ridges shown). Most of these data are from the GEOROC database with key references for $^{3}$He/$^{4}$He data given in Ito and Mahoney (2006). Data for the Puka–Puka are from Janney et al. (2000) and for the Foundation chain from Maia et al. (2000).
anomalies in the lowermost mantle below the South Pacific and Africa Superswell regions (Breger et al., 2001; Ni et al., 2005; van der Hilst and Karason, 1999; Trampert et al., 2004). The sharp edges of these anomalies (Ni et al., 2002; To et al., 2005) and their reproduction in dynamical models (Tan and Gurnis, 2005; McNamara and Zhong, 2005) suggest both a compositional and thermal origin of these anomalies. It is more difficult to identify low-velocity anomalies at smaller spatial scales and shallower depths. For example, Ritsema and Allen (2003) investigated the correlation between seismic low-velocity regions in the global S-wave model S20RTS (Ritsema et al., 1999) and hot spots from a comprehensive list (Sleep, 1990). They confidently detected anomalously low seismic wave speeds in the upper mantle for only a small numbers of hot spots. Stronger correlations between a number of hot spots and deep-mantle anomalies were found using finite-frequency P- and S-wave tomography (Montelli et al., 2004), but the ability of this method to actually improve the resolution with the available data is debated (de Hoop and van der Hilst, 2005; van der Hilst and de Hoop, 2005; Boschi et al., 2006).

The topography of seismic discontinuities in the transition zone may also reflect heterogeneity. A hotter anomaly causes the exothermic phase change at 410 km to occur deeper and the endothermic phase change at 660 km to occur shallower, assuming the olivine system dominates the phase changes (Ito and Takahashi, 1989; Helffrich, 2000). Global observations of the transition zone thickness provide some support for hotter-than-normal mantle below some hot spots (e.g., Helffrich, 2002; Li et al., 2003a, 2003b), although other observations suggest that global seismic data can resolve strong correlations between topography of the 410 km discontinuity and seismic velocities only at wavelengths larger than those of individual hot spots (Chambers et al., 2005).

7.09.2.5.2 Local seismic studies of major hot spots

Improved insights into the upper mantle beneath hot spots can be obtained using regional or array studies, including the use of surface waves (e.g., Pilidou et al., 2005).

Iceland. Regional seismic studies have confidently imaged a body of anomalously slow seismic wave speeds in the upper mantle beneath Iceland. Conventional ray theory was used to first image the anomaly (Allen et al., 1999a, 2002; Foulger et al., 2001; Wolfe et al., 1997) but improved finite-frequency techniques (Allen and Tromp, 2005) resolve the feature to be roughly columnar with lateral dimension of 250–300 km and peak P- and S-wave anomalies of $-2.1\%$ and $-4.2\%$, respectively (Hung et al., 2004) (Figure 11). Recent studies using Rayleigh waves and local earthquakes confirm these high amplitudes, which most likely require a combination of excess temperature and melt (Yang and Shen, 2005; Li and Detrick, 2006).

In addition, studies of surface waves and shear-wave splitting reveal significant seismic anisotropy in the Icelandic upper mantle (Li and Detrick, 2003; Bjarnason et al., 2002; Xue and Allen, 2005). Overall, they find that the fast S-wave propagation directions are mostly NNE–SSW in central Iceland with stronger E–W components to the west and east. The anisotropy in west and eastern Iceland deviates significantly from the directions of motion of the two plates and thus could indicate a large-scale mantle flow in the region (Bjarnason et al., 2002). However, in central Iceland, the strong rift-parallel anisotropy near the active rift zones is interpreted to indicate ridge-parallel flow associated with a mantle plume (Li and Detrick, 2003; Xue and Allen, 2005).

The anomalous seismic structure extends well below 410 km as evidenced by the thinning of the transition zone beneath Iceland (Shen et al., 1998) (Figure 12). While Shen et al. (1998) show evidence for both a deepening 410- and shoaling 660-discontinuity, Du et al. (2006) argue that the discontinuity at 660 km is instead flat. The precise nature of both discontinuities is important in determining whether the Iceland anomaly initiates in the upper mantle or is present below 660 km (Shen et al. 1998). One global tomography model suggests that the anomaly extends into the lower mantle (Bijwaard and Spakman, 1999) and another study identifies an ultralow-velocity zone near the core–mantle boundary (CMB) below Iceland (Helmberger et al., 1998). The available seismic data, however, leave a lower mantle origin open to debate (e.g., Foulger et al., 2001) and a robust test awaits improved regional seismic experiments.

Other Atlantic hotspots. Slow surface-wave speed anomalies extend to 200 km below the Azores hot spot as part of an along-strike perturbation of the velocity structure beneath the MAR (Pilidou et al., 2004). Receiver function analysis at Cape Verde (Lodge and Helffrich, 2006) indicates a thickened crust ($\sim15$ km) and a high-velocity, low-density zone to a depth of $\sim90$ km. The oldest volcanoes sit on top of the thickest parts of the crust and the high-velocity layer. Such high velocities in the shallow mantle beneath active hot spots are unusual. They
suggest major-element heterogeneity (e.g., due to melt depletion) dominate over other effects such as the presence of volatiles, melt, or excess temperature.

**Hawaii.** Anomalously low seismic wave speeds beneath the Hawaiian hot spot have been found from preliminary surface wave (Laske et al., 1999) and tomographic (Wolfe et al., 2002) studies. The anomaly imaged by tomography appears broader and higher in amplitude for S-waves (200 km diameter and up to −1.8%) than it does for P-waves (100 km diameter and −0.7%). A significant low S-wave speed (<4 km s⁻¹) is observed below 130 km suggesting the presence of partial melt below this depth (Li et al., 2000). Additional evidence for an upper-mantle melt anomaly is provided by a seafloor magnetotelluric study (Constable and Heinson, 2004) which suggested a columnar zone of 5–10% partial melting with a radius <100 km and a depth extent of 150 km. Insights into the deeper structure using seismic tomography are currently hindered by poor coverage of stations and earthquake sources, but resolution tests indicate that the anomaly is unlikely to be restricted to the lithosphere (Wolfe et al., 2002). A deep origin is suggested by evidence for a thinning of the transition zone by 40–50 km (Li et al., 2000, 2004; Collins et al., 2002). These studies, as well as a
combined seismic and electromagnetic inversion show that the transition-zone anomaly is consistent with an excess temperature of 200–300 K (Fukao et al., 2001). A still deeper origin, possibly to the CMB, is suggested by a recent tomographic study that incorporates core phases (Lei and Zhao, 2006).

**Galápagos.** The Galápagos hot spot is part of a broad region in the Nazca and Cocos Basins with significantly reduced long-period Love and Rayleigh wave speeds (Vdovin et al., 1999; Heintz et al., 2005). The transition zone surrounding the Galápagos has similar thickness to that of the Pacific Basin except for a narrow region of ~100 km in radius slightly to the west of the Galápagos archipelago where it is thinned by ~18 km. This amount of thinning suggests an excess temperature of 130 K (Hooft et al., 2003). Preliminary results of a regional tomography study indicate a low-velocity feature of comparable dimension, extending above the transition zone into the shallow upper mantle (Toomey et al., 2001). The western edge of the archipelago shows shear-wave splitting of up to 1 s with a direction consistent with E–W plate direction. The anisotropy disappears beneath the archipelago where the upper-mantle wave speeds are anomalously low, suggesting that melt or complex flow beneath the hot spot destroy the plate-motion derived anisotropy (Fontaine et al., 2005).

**Yellowstone.** Early tomographic studies revealed a complex velocity structure in the upper mantle beneath the Snake River Plain, southwest of the Yellowstone hot spot. This structure was interpreted to represent compositional variability restricted to the upper mantle associated with melting (Saltzer and Humphreys, 1997). More recent work suggests that a narrow, low-velocity feature extends from the upper mantle into the top of the transition zone (Waite et al., 2006; Yuan and Dueker, 2005). The shallow upper-mantle anomaly is present over a distance of more than 400 km, spanning from the northeastern extent of the Snake River Plain to Yellowstone National Park, including a short segment to the northeast of the Yellowstone hot spot. The anomaly is strongest at depths 50–200 km with peak anomalies of ~2% for Vp and ~5.5% for Vs (Waite et al., 2006). The velocity reductions are interpreted to represent 1% partial melt at a temperature of 200 K above normal (Schutt and Humphreys, 2004). Initial transition-zone studies showed significant topography of the 410 discontinuity throughout the region (Dueker and Sheehan, 1997). More recent studies show that the 410 discontinuity deepens by 12 km near the intersection of the low-velocity anomaly identified by Waite et al. (2006) and Yuan and Dueker (2005), but interestingly, the 660 km discontinuity appears flat in this area (Fee and Dueker, 2004). Shear-wave splitting measurements around the Yellowstone–Snake River Plain show fast S-wave speeds primarily aligning with apparent plate motion, except for two stations in the Yellowstone caldera, perhaps due to local melt effects (Waite et al., 2005).

**Eifel.** The Eifel region in Western Germany is characterized by numerous but small volcanic eruptions with contemporaneous uplift by 250 m in the last 1 My. Tomographic imaging indicates a mantle low-velocity anomaly extending to depths of at least 200 km (Passier and Snieder, 1996; Pilidou et al., 2005). Inversions using a high-resolution local array study indicate a fairly narrow (100 km) P-wave anomaly of ~2% that possibly extends to a depth of 400 km (Ritter et al., 2001; Keyser et al., 2002). The connection with the deeper mantle is unclear but has been suggested to include the low-velocity structure in the lower mantle below central Europe (Goes et al., 1999) (Figure 13). Shear-wave splitting measurements show the largest split times for S-waves polarized in the direction of absolute plate motion, but the pattern is overprinted by complex orientations, suggestive of parabolic mantle flow around the hot spot (Walker et al., 2005). A comparison of the seismic anomaly structure below the Eifel, Iceland, and Yellowstone is provided in Figure 14 (from Waite et al., 2006).

**East Africa.** Body and surface-wave studies indicate a strong regional low-velocity anomaly in the mantle
below the East Africa Rift System. Below Afar, surface-wave studies image anomalous velocities (≤−6% in vertically polarized S-waves) extending to a depth of ~200 km (Sebai et al., 2006). In the North Ethiopian Rift, a narrow (75–100 km) tabular feature extends to depths >300 km and is slightly broader in the northern portion which trends toward Afar. Between the flanks the rift zones, P-wave speeds are reduced by 2.5% and S-wave speeds by 5.5% (Bastow et al., 2005). Independent P-wave tomography confirms the general trend and amplitude of the anomaly with a tentative suggestion that it trends to the south and west (Benoit et al., 2006) toward the broad lower-mantle seismic anomaly below Africa (Ritsema et al., 1999; Grand, 2002). The Tanzania Craton, to the northwest, is imaged as having high P- and S-wave speeds to depths of at least 200 km (Ritsema et al., 1998), is thinner than other African cratons, and is surrounded by the slow seismic-wave speeds associated with the East African Rift System (Sebai et al., 2006). The transition zone shows complicated variations but is generally thinner below the Eastern Rift by 30–40 km compared to the more normal thickness under areas of the Tanzania Craton (Owens et al., 2000; Nyblade et al., 2000). Combined, these results suggest that the mantle below the rifts is hotter than normal mantle by 200–300 K with partial melt in the shallow upper mantle.

Seismic anisotropy is dominantly parallel to the main Ethiopian Rift, in an area that extends to nearly 500 km away from the ridge axis, which likely rules out simple extension-driven asthenospheric flow (Gashawbeza et al., 2004). The regional anisotropy is most likely caused by pre-existing features in the late Proterozoic Mozambique Belt but may be locally enhanced by aligned melt in the Ethiopian and Kenyan Rifts (Gashawbeza et al., 2004; Walker et al., 2004; Kendall et al., 2005).

South Pacific Superswell. Recordings of French nuclear explosions in French Polynesia provide evidence for seismically fast velocities in the shallow mantle which suggest compositional heterogeneity without evidence for excess temperature (Rost and Williams, 2003). Rayleigh-wave dispersion measurements across the Pitcairn hot-spot trail suggest an absence of lithospheric thinning (Yoshida and Suetsugu, 2004). Underside reflections of S-waves at the 410- and 660-discontinuities show normal thickness of the mantle transition zone, except in a 500-km-wide area beneath the Society hot spot (Niu et al., 2002). Seismic anisotropy in French Polynesia generally aligns with apparent plate motion, although
the absence of anisotropy beneath Tahiti and minor deviations beneath other islands could indicate the presence of local flow or magma (Russo and Okal, 1998).

**Bowie.** The presence of a narrow, low-velocity zone near the base of the transition zone below the Bowie hot spot is based on delays in seismic records of Alaskan earthquakes measured in the NW United States (Nataf and VanDecar, 1993). The delays are consistent with a zone of 150 km diameter with an excess temperature of 200 K.

**LIPs.** Beneath the Deccan Traps, seismic speeds in the shallow upper-mantle are anomalously low to a depth of at least 200 km (Kennett and Widiyantoro, 1999). The anomaly appears to be absent at depths near the transition zone (Kumar and Mohan, 2005). The Paraná Province is underlain by a distinct region of low seismic-wave speeds in the upper mantle (VanDecar et al., 1995; Schimmel et al., 2003). The OJP has an upper-mantle velocity anomaly of −5% with respect to Preliminary Reference Earth Model (PREM) (Dziewonski and Anderson, 1981) with a maximum depth extent of 300 km (Richardson et al., 2000), whereas ScS reverberations show that this region is less attenuating than ‘normal’ Pacific asthenosphere (Gomer and Okal, 2003). Seismic anisotropy beneath the OJP is weak, which is interpreted to indicate that the residual mantle root has remained largely undeformed since it formed ~120 Ma (Klosko et al., 2001). The presence of anomalously slow mantle beneath such old flood basalts is enigmatic in that any thermal anomaly is expected to have diffused away.

**7.09.2.6 Summary of Observations**

Long-lived (>50 My) age-progressive volcanism occurs in 13 hot spots. At present day, these hot spots define a kinematic reference frame that is deforming at rates lower than average plate velocities. Over geologic time, however, there has been significant motion between the Indo-Atlantic hot spots, the Pacific hot spots, and Iceland. Short-lived (<22 My) age progressions occur in at least eight volcano chains. The directions and rates of age progression in the short-lived chains suggest relative motion between these hot spots, even on the same plate. Finally, a number of volcano groups, which sometimes align in chains (e.g., Cook–Austral, Cameroon), fail to show simple age–distance relations but instead show episodic volcanism over tens of millions of years.

Anomalously shallow topographic swells are very common among hot spots and melting anomalies. These swells are centered by the volcanoes and span geographic widths of hundreds of kilometers to >1000 km. Swells appear to diminish with time; they are usually present around volcanoes with ages <50 Ma and are typically absent around volcanoes older than ~70 Ma. Conspicuously, the Madeira and Canary hot spots are two active volcano chains without substantial swells.
Large igneous provinces represent the largest outpourings of magma but also show a huge range in magmatic volumes and durations. They can be as large as 50 Mkm$^3$ (OJP) to <2 Mkm$^3$ (CRB) (see Figure 9). Voluminous magmatism can occur in dramatic short bursts, lasting 1–2 My (CAMP, CRB, EM, OJP, SIB) or can be prolonged over tens of millions of years (e.g., CHON, CBN, KER, NAVP, SHA). Main eruptive products are tholeiitic basalts which, on the continents, typically are transported through radiating dike swarms. High-MgO basalts or picrites are found in a number of provinces (NAVP, OJP, CAR) which indicate high degrees of melting. Smaller rhyolitic eruptions in continental flood basalts (KAR, PAR, CHON, EARS) indicate melting of continental crust. Dynamic topographic uplift is evident around the main eruptive stages of some LIPs (EM, NAVP, SIB, KER) but may not have occurred at some oceanic plateaus (OJP, SHA). While an appreciable amount of the geologic record is lost to subduction, a half-dozen recently active hot-spot chains are confidently linked to LIPs with the remainder having tenuous or non-existent links. Those volcano chains that clearly backtrack to LIPs involve those at the centers or margins of continents; Kerguelen is one of the possible exception. Also, most of the LIPs we have examined are associated with rifting, either between continents (PAR, KAR, CAMP, NAVP, DEC, MDR) or at mid-ocean ridges (OJP–MAN–HIK, KER, SHA–HES). The above characteristics compel substantial revisions and/or alternatives to the hypothesis of an isolated head of a starting mantle plume as the only origin of LIPs.

Basalts from hot spots and other melting anomalies, for the most part, are more heavily influenced by mantle materials that are distinct in terms of $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, and/or $^{3}\text{He}/^{4}\text{He}$ ratios from the MORB source. Four possible exceptions, which show MORB-like $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ compositions (but lack constraints from $^{3}\text{He}/^{4}\text{He}$) are the Shatsky Rise and the Bowie–Kodiak, Cobb, and Caroline chains.

Most hot spots are associated with anomalously low seismic-wave speeds below the lithosphere and in the upper mantle. The transition zone below hot spots is often thinned by tens of kilometers. The above findings are consistent with elevated mantle temperature by 150–200 K and with excess partial melt in the shallow upper mantle. Improved understanding of mineral physics at appropriately high pressure and temperature are needed to better constrain the magnitude of the possible temperature anomalies and to quantify the potential contribution of compositional heterogeneity. Finally, while there are hints of seismic anomalies extending into the lower mantle and even to the CMB, definitive tests of a deep origin for some melting anomalies require more extensive regional seismic experiments and modern methods of interpretation.

The key characteristics described above provide information needed to test various proposed dynamical mechanisms for the formation of hot spots and melting anomalies. Some trends and generalities are apparent but substantial deviations likely reflect a range of interacting processes. In other words, it is very unlikely that a single overarching mechanism applies to all cases.

### 7.09.3 Dynamical Mechanisms

This section reviews the mechanisms proposed to generate hot spots and melting anomalies. We begin with a summary of methods used to quantitatively explore the mechanisms (Section 7.09.3.1). We then discuss the shallower processes of melting (Section 7.09.3.2) and swell formation (Section 7.09.3.3) before addressing the possible links to the deeper mantle, with specific focus on the extensive literature on mantle plumes (Section 7.09.3.4). In the context of whole mantle convection, we discuss possible causes of volcano age progressions and the inferred approximately coherent motion among hot spots on the same (Indo–Atlantic and the Pacific) plates (Section 7.09.3.5). Proposed mechanisms for generating LIPs and their possible connection to hot spots are explored in Section 7.09.3.6. The diversity of observations of hot spots and LIPs requires important modifications to the thermal plume hypothesis, as well as alternative possibilities as presented in Section 7.09.3.7. In light of these possibilities we discuss possible causes for the differences in geochemistry between hot-spot basalts and MORB in Section 7.09.3.8.

#### 7.09.3.1 Methods

The origin and evolution of hot spots and melting anomalies can be constrained by studying the transport of energy, mass, and momentum in the solid and partially molten mantle (see Chapter 7.02). Key parts of the above processes can be described mathematically by the governing equations and solved with analytical or numerical approaches, or can be studied by simulation in laboratory experiments using analog materials.

Analytical approaches provide approximate solutions and scaling laws that reveal the main
relationships between phenomena and key parameters (see Chapter 7.04). Relevant examples of applications include boundary layer analysis, which is important to quantifying the time and length scales of convection, and lubrication theory, which has been instrumental in understanding hot-spot swell formation. A separate approach uses experimental methods (see Chapter 7.03), involving analog materials such as corn syrup and viscous oils to simulate mantle dynamics at laboratory timescales. Analog methods were instrumental in inspiring the now-classic images of plumes, with voluminous spherical heads followed by narrow columnar tails. Numerical techniques are necessary to solve the coupling between, and non-linearities within the equations such as those caused by strongly varying (e.g., temperature, pressure, and composition-dependent) rheology (see Chapter 7.05), phase transitions, and chemical reactions. Numerical models are therefore well suited for simulating more-or-less realistic conditions and for allowing quantitative comparisons between predictions and observations. In this way, numerical techniques provide a means to directly test conceptual ideas against the basic laws of physics and to delineate the conditions under which a proposed mechanism is likely to work or has to be rejected based on observations.

Unfortunately, fully consistent modeling is difficult to achieve and is hampered by multiple factors. First, the constitution of the Earth’s mantle can only be approximated. Information about material properties such as density, rheology, and thermal conductivity, are essential for quantitative modeling but is not precisely known and becomes increasingly imprecise with depth. Second, the transition from the brittle lithosphere to the viscous asthenosphere involves a rapid temperature increase; the common assumption that the deformation of the Earth’s mantle can be approximated as creeping viscous flow is only correct at high temperatures and the details of the lithosphere–asthenosphere interaction depend on poorly known processes that are difficult to model self-consistently. Finally, the problem is multiscale, involving processes occurring from scales as small as individual grains (e.g., fluid–solid interaction, chemical transport), to as large as the whole mantle (see Chapters 7.02, 2.06, and 2.14). Addressing these challenges will require careful comparisons between the different techniques, adjustments according to improved insights from experimental and observational work, and smart use of increasing computing technology.

### 7.09.3.2 Generating the Melt

Understanding of the causes for excess melt generation is essential for our understanding of the dynamics of hot spots and melting anomalies. The rate that an infinitesimal bulk quantity of mantle melts to a fraction \( F \) can be described by

\[
\frac{DF}{Dt} = \left( \frac{\partial F}{\partial T} \right)_{P,c} (DT/Dt) + \left( \frac{\partial F}{\partial P} \right)_{P,T} (DC/Dt) + \left( -\frac{\partial F}{\partial P} \right)_{S} (-DP/Dt) \tag{1}
\]

The first two terms on the right-hand side describe nonisentropic processes. The first term describes the melt produced by heating and is proportional to the change in \( F \) with temperature \( T \) at constant pressure \( P \) (i.e., isobaric productivity); melting by this mechanism may occur in a variety of settings but is likely to be comparatively small and thus has not been a focus of study. The second term describes melt generated by the open-system change in composition. This term may be important behind subduction zones where the addition of fluids into the mantle wedge causes ‘flux’ melting and the formation of arc and back-arc volcanism. Finally, the third term describes isentropic decompression melting, which is perhaps the most dominant process of melt generation at mid-ocean ridges, hot spots, and other melting anomalies.

For decompression melting, the rate of melt generation is controlled by the melt productivity \( \partial F/\partial P \), which is positive if temperature exceeds the solidus. Both the solidus and value of \( \partial F/\partial P \) (e.g., McKenzie, 1984; Hirschmann et al., 1999; Phipps Morgan, 2001) depend on the equilibrium composition of the solid and liquid at a given pressure. The rate of decompression \( -DP/Dt \) is controlled by mantle dynamics and is primarily proportional to the rate of mantle upwelling. The total volumetric rate of melt generation is approximately proportional to \( DF/Dt \) integrated over the volume \( V \) of the melting zone,

\[
\dot{M} = \frac{\rho_m}{\rho_c} \int_V (\partial F/\partial P)_S (DP/Dt) \, dV \tag{2}
\]

where \( \rho_m \) is mantle density and \( \rho_c \) is igneous crustal density. Melting anomalies thus require one or more of the following conditions: excess temperature, presence of more fusible or fertile material, and mantle upwelling. Higher temperatures increase \( V \) by increasing the pressure at which the solidus is intersected, more fusible mantle can change both \( -\partial F/\partial P \)\(_S\) and increase \( V \), and both factors may influence \( DP/Dt \) through their effects on mantle buoyancy.
7.09.3.2.1 Temperature
Melting caused by an increase in temperature has been a major focus of previous studies. One way to estimate mantle temperature is based on comparisons between predicted and observed melt-production rates at different settings. Another way involves using compositions of primitive lavas. Given an assumed starting source composition and a model for melt–solid interaction as the melt migrates to the surface (e.g., ‘batch’, ‘fractional’, ‘continuous’ melting), the liquid concentrations of key oxides (e.g., Na$_2$O, CaO, Al$_2$O$_3$, SiO$_2$, FeO, MgO) as well as incompatible trace elements can be predicted based on their dependence on $P$, $T$, and $F$, which are related through solid mineralogy and liquid composition. The above relationships can be established by thermodynamic theory and constrained with laboratory experiments.

Estimates of temperature variations at mid-ocean ridges are aided by the relative simplicity of lava compositions, relatively straightforward measurements of magma production rate (e.g., crustal thickness times spreading rate), and the ability to correlate variations in the above observational parameters in space (i.e., along mid-ocean ridges). In addition, spreading rate provides a constraint on the rate of mantle upwelling, for example, if one assumes, to first order, that mantle upwelling is a passive (kinematic) response to plate spreading. Based on the conditions needed for a lherzolitic mantle to yield observed crustal thicknesses and major element variations near Iceland, excess mantle temperatures beneath Iceland relative to normal mid-ocean ridges are estimated to range from about 100°C to >250°C (e.g., Klein and Langmuir, 1987; McKenzie and Bickle, 1988; Langmuir et al., 1992; Shen and Forsyth, 1995; Presnall et al., 2002; Herzberg and O’Hara, 2002). Excess temperature estimates based on inversions of crustal thickness and incompatible trace-element compositions also fall within the above range for Iceland (White et al., 1995; Maclellan et al., 2001) and other hot spots (White et al., 1992).

Beneath Hawaii, the maximum mantle potential temperatures (i.e., temperature at zero pressure after removing the effects of adiabatic decompression) is estimated at 1500–1600°C based on predicted melt production rates from numerical models of mantle upwelling, driven by thermal buoyancy (Ribe and Christensen, 1999; Watson and McKenzie, 1991). This temperature range is 200–300°C higher than the estimated potential temperature of 1280°C beneath normal mid-ocean ridges using the same melting model (McKenzie and Bickle, 1988).

Another method of estimating mantle temperatures is based on Fe–Mg content of primary melts and the olivine phenocrysts with which they equilibrate. This method depends on experimentally constrained partitioning of Fe and Mg between liquid and olivine, measured forsterite content of olivine crystals, and estimated Fe and Mg content of primary magmas (i.e., magmas that segregated from the mantle melting zone and have not been further modified by shallow processes such as crystal fractionation or accumulation). One group suggests that the mantle is no hotter beneath Hawaii than beneath many mid-ocean ridges (Green et al., 2001). Other groups, however, suggest elevated temperatures of 50–100°C (Herzberg and O’Hara, 2002; Herzberg, 2004a) and 100–200°C (Putirka, 2005) beneath Hawaii and Iceland. As essentially all sampled lavas have evolved to varying degrees after they left the mantle source, an important uncertainty is the MgO content of the primary liquids. Putirka (2005) argues, for example, that the lower MgO contents derived by Green et al. (2001) for Hawaii could lead to an underestimate of temperature. A recent critical evaluation of the criteria for determining mantle potential temperatures below ridges and hot spots is provided by Herzberg et al. (in press).

7.09.3.2.2 Composition
A major source of uncertainty for all of the above temperature estimates is the composition of the mantle source. Water and CO$_2$, for example, can dramatically reduce melting temperatures even in the small proportions (i.e., well below saturation) likely to be present in the MORB source (Asimow and Langmuir, 2003; Dasgupta and Hirschmann, 2006). While such small concentrations of volatiles are not likely to increase the total extent of melting significantly, they can enhance the amount of melt produced for a given temperature by appreciably expanding the volume of the melting zone. As the mantle beneath hot spots is likely to be more volatile rich, temperature estimates based on dry peridotite may be too high. For example, excess temperatures beneath the hot-spot-influenced Galápagos spreading center may be reduced from ~50°C for anhydrous melting models to <40°C when water is considered (Cushman et al., 2004; Asimow and Langmuir, 2003). Similarly, estimates for the mantle excess temperature beneath Azores have been revised from ~75°C to ~55°C (Asimow et al., 2001).
Hydrous melting models have yet to be explored in detail for the larger Iceland and Hawaii hot spots. The mantle beneath hot spots may also contain more fusible, mafic lithologies, such as those generated by the recycling of subducted oceanic crust. The presence of such ‘fertile’ mantle has been suggested for hot spots such as Hawaii (Hauri, 1996; Takahashi, 2002), Iceland (Korenaga and Kelemen, 2000), the CRRs (Takahashi et al., 1998), Galápagos (Sallares et al., 2005), and others (Hofmann, 1997). Pyroxenite lithologies have both a lower solidus and greater productivity ((\partial F/\partial P) than peridotite (Pertermann and Hirschmann, 2003) and therefore require much lower temperatures to produce the same volume of melt as peridotite.

Some have argued that fertile mantle melting could generate many melting anomalies with very small or zero excess temperatures (Korenaga, 2005b). An important difficulty with this hypothesis is that mafic materials will tend to form eclogite, which is significantly denser than lherzolite in the upper mantle (Irifune et al., 1986). This material must therefore produce substantial melt at depths where upwelling (i.e., −DP/DT, eqn [2]) is not appreciably impeded by negative buoyancy. It has been suggested that eclogite becomes neutrally buoyant near the base of the upper mantle (∼660 km) (Hirose et al., 1999; Ringwood and Irifune, 1988) (see Chapter 2.18). To reach the solidus and initiate melting at this depth most likely requires temperatures >300 K higher than normal (e.g., Hirose and Fei, 2002). Alternatively, it has been suggested that rapid upwelling driven by shallow thermal convection (Korenaga, 2004) (Figure 15) or fast seafloor spreading (Korenaga, 2005b) could entrain eclogite upward from a neutrally buoyant layer at 660 km and cause melting in the upper mantle without usually hot mantle. More recent experiments, however, suggest that subducted basalt is actually denser than peridotite throughout the upper mantle (Aoki and Takahashi, 2004) and thus would unlikely accumulate near 660 km. It is clear that a more complete understanding of the properties and phase relations of different lithologies at a range of mantle pressures and temperatures is needed to test the importance of fertile mantle melting.

Figure 15 Two forms of small-scale convection in the upper mantle. (a) Solutions of 2-D numerical models in which sublithospheric thermal instabilities drive convection (arrows show mantle flow) in the upper mantle at two time steps as labeled. Colors show temperature with blue being cold (lithosphere) and light yellow being hottest. Green tracers of subducted crustal fragments first form a layer at 660 km; some eventually sink into the lower mantle and some are drawn upward where they could melt. (b) Predictions of 2-D numerical models in which small-scale convection is driven by the edge of thick lithosphere (gray). Colors show temperature contrast from the mantle adiabat. (a) From Korenaga J (2004) Mantle mixing and continental breakup magmatism. Earth and Planetary Science Letters 218: 463–473. (b) From King S and Anderson DL (1998) Edge-driven convection. Earth and Planetary Science Letters 160: 289–296.
7.09.3.2.3 Mantle flow

The final major factor that can lead to melting anomalies is enhanced mantle upwelling \((-\Delta P/\Delta t)\). Thermal buoyancy can cause rapid upwelling and dramatically enhance melt production (e.g., Ito et al., 1996; Ribe et al., 1995). Compositional buoyancy could also enhance upwellings (Green et al., 2001). However, compositionally lighter material such as those with less iron and dense minerals (i.e., garnet), perhaps due to prior melt depletion (Oxburgh and Parmentier, 1977; Jordan, 1979), are likely to be less fusible than undepleted or fertile mantle. Behaving in complimentary fashion to fertile mantle, depleted mantle must be light enough such that the associated increase in upwelling rate \((-\Delta P/\Delta t)\) overcomes the reduction in fusability \((\nabla F/\partial P)_S\) in eqn [2]).

Buoyancy-driven upwelling, however, only requires enhanced lateral variations in density. Besides hot, plume-like upwellings from the deep mantle, large density variations can occur near the lithospheric thermal boundary layer (TBL). Sublithospheric boundary layer instabilities can drive small-scale convection in the upper mantle (e.g., Richter, 1973; Korenaga and Jordan, 2003; Huang et al., 2003; Buck and Parmentier, 1986) (Figure 15(a)). A related form of small-scale convection can occur where there are large variations in lithospheric thicknesses such as that near rifted continental margins (Buck, 1986; King and Anderson, 1998) (Figure 15(b)). While the physics of convection in such situations have been explored to some degree, the volumes, timescales, and compositions of magmatism that could be produced have not.

The process of thinning the continental lithosphere can also cause rapid passive upwelling in the underlying asthenosphere. Thinning could occur by the foundering and delamination of the lower lithosphere or by continental rifting. Both mechanisms have been proposed to form flood volcanism on continents or continental margins without elevated mantle temperatures (Hales et al., 2005; van Wijk et al., 2001) (see Section 7.09.3.6).

Another form of enhanced mantle upwelling can occur in response to melting itself. Partial melting reduces the density of the solid residue (discussed above) and generates intergranular melt. Both factors can reduce the bulk density of the partially molten mantle and drive buoyant decompression melting (Figure 16). Buoyant decompression melting has been shown to be possible beneath mid-ocean ridges but could also occur away from mid-ocean ridges (Tackley and Stevenson, 1993; Raddick et al., 2002). The key requirements for buoyant decompression melting to occur mid-plate is the presence of an appreciable thickness of mantle to be very near its

\[ F = \frac{\partial F}{\partial P} S, \]

\[ \Delta P/\Delta t, \]

\[ \nabla F/\partial P, \]

\[ \text{Time} = 0.0021 \text{ My}^{-1}, \]

\[ \text{Time} = 2.4 \text{ My}, \]

\[ \text{Time} = 6.2 \text{ My}, \]

\[ \text{Figure 16} \]

Predictions of 2-D numerical models which simulate buoyant decompression melting for three time steps as labeled. Left column shows fractional melting rate \((\text{red} = 0.0021 \text{ My}^{-1}, \text{blue} = 0)\), middle column shows melt fraction retained in the mantle \(\phi\) \((\text{red} = 0.02)\), and right column shows volume fraction of melt extracted \(\xi\) \((\text{red} = 0.108)\). Density decreases as linear functions of both \(\phi\) and \(\xi\), and lateral density variations are what drive upwelling. Melting is therefore limited by the accumulated layer low-density residue. Reproduced from Raddick MJ, Parmentier EM, and Scheirer DS (2002) Buoyant decompression melting: A possible mechanisms for intra plate volcanism. Journal of Geophysical Research 107: 2228 (doi: 1029/2001JB000617), with permission from AGU.
solidus and a perturbation to initiate upwelling and melting. Perturbations could be caused by rising small bodies of hotter or more fusible mantle, the flow of mantle from beneath the old and thick side of a fracture zone to the young and thin side (Raddick et al., 2002), or even sublithospheric convection.

Lastly, mantle upwelling can be driven by vertical motion of the lithosphere, such as that due to lithospheric flexure. The flexural arch around large intraplate volcanoes, for example, is caused by the growth of volcanoes which not only pushes the underlying lithosphere downward but causes upward flexing in a donut-shaped zone around the volcano. Flexural arching also occurs on the seaward side of subduction zones. Decompression melting can occur beneath flexural arches, again, if an appreciable thickness of asthenosphere is near or at its solidus (Bianco et al., 2005).

### 7.09.3.3 Swells

One of the more prominent characteristics of hot spots and melting anomalies is the presence of broad seafloor swells. One indication of the possible origin is an apparent dependence of swell size with the rate of plate motion. We measure swell widths using the maps of residual topography (Figure 1) with the criterion that the swell is defined by the area that exceeds a height of +300 m in the direction perpendicular to volcano chains. This direction usually corresponds to the direction in which the swell is least wide (except for Trindade). The parallel bars in Figure 1 mark the widths we measure for different hot spots. Figure 17(a) shows how swell width W varies with the plate speed \( U_p \) at the hot spot, relative to the hot-spot reference frame (compiled by Kerr and Mériaux (2004)). For plate speeds <80 km My\(^{-1} \), \( W \) appears to decrease with increasing plate speed. The prominent Pacific hot spots (Hawaii, Marquesas, Easter, Pitcairn, and Louisville) break the trend and have widths comparable to many of the swells in the Atlantic. We now test whether the observations can be explained by buoyant, asthenospheric material ponding beneath the lithosphere.

#### 7.09.3.3.1 Generating swells: Lubrication theory

Lubrication theory is a simplifying approach of solving the equations governing fluid flow against a solid interface. The method eliminates partial derivatives with respect to depth by assuming that fluid layers are thin compared to their lateral dimension, and therefore describes fluid-layer thickness in map view. This theory was first applied to the formation of hot-spot swells by Sleep (1987) and Olson (1990) (see Chapter 7.04). Here, buoyant asthenosphere is introduced at the base of a moving, rigid (lithospheric) plate. The buoyant material is dragged laterally by plate motion and expands by self-gravitational spreading away from the source such that the extent \( W \) perpendicular to plate motion increases with distance \( x \) from the source (Figure 18(a)). Without plate motion the material would expand axisymmetrically like a pancake.

As confirmed with laboratory experiments and numerical models, the width of the buoyant material and swell far from the source can be approximated by the dimensionless equation (Figure 18(a)):

\[
W/L_0 = C_1 (x/L_0)^{1/5}
\]  \[3\]

where \( C_1 \) is a constant (~3.70) and \( L_0 \) is the characteristic length scale of the problem defined as (Ribe and Christensen, 1994)

\[
L_0 = \left( \frac{B^2 \gamma}{96 \pi^3 \Delta \rho^2 \eta U_p^4} \right)^{1/4}
\]  \[4\]

This length scale contains information about the key parameters controlling the width of the flow: buoyancy flux \( B \) (kg s\(^{-1} \)), acceleration of gravity \( g \), density contrast between the buoyant and normal mantle \( \Delta \rho \), viscosity of the buoyant mantle \( \eta \) and plate speed \( U_p \). Alternatively, if the buoyant material is introduced along a semi-infinite line, then \( W \) will increase more rapidly with distance. Using the same reasoning as Kerr and Mériaux (2004), this case can be directly compared to [3], using the length scale \( L_0 \)

\[
W/L_0 = C_2 (x/L_0)^{4/5}
\]  \[5\]

where \( C_2 \) is a constant (in Figure 18(a) \( C_2 L_0^{4/5} = 1.23 \)).

In either case, \( W \) is predicted to be of the same order as \( L_0 \propto B^{0.4}/U_p \). This relation predicts \( L_0 \) and thus \( W \) to be inversely proportional to \( U_p \) because faster plates allow less time for the layer to expand while it is dragged a distance \( x \). The above relations also show an important dependence on \( B \), which is not considered in our initial plot in Figure 17(a). Larger buoyancy fluxes (proportional to \( Q \)) lead to larger \( W \) primarily by enhancing the rate of gravitational expansion.
Buoyancy flux can be estimated based on the volumetric rate of swell creation, (Davies, 1988; Sleep, 1990)

\[ B = \bar{b} \bar{W} U_p (\rho_m - \rho_w) \]  

where \( \bar{b} \) and \( \bar{W} \) are averages of swell height and width, respectively, and \( \rho_m - \rho_w \) is the density contrast between the mantle and water.

Using estimated values for \( B \) and \( U_p \) (Kerr and Meriaux, 2004), a plot of \( \bar{W} \) versus \( B^{3/4}/U_p \) indeed shows a positive correlation (Figure 17(b)). Some of the scatter could be due to errors in \( B \), perhaps due to uncertainties in swell height (Cserepes et al., 2000) as well as the oversimplifying assumption in [6] that the buoyant material flows at the same speed as the plate (Ribe and Christensen, 1999). Other sources of scatter could be differences in \( \rho_m \) and \( \rho_w \) between hot spots.

For hot spots interacting with mid-ocean ridges, swell widths are the extent that positive residual topography extends in the direction parallel to ridges (Ito et al., 2003). Swell widths along hot-spot-influenced ridges and nearby seafloor isochrons appear to depend on the full spreading rate \( U \) as well as hot-spot–ridge separation (Figures 17(c) and 17(d)). Lubrication theory predicts that along-axis ‘waist’ widths \( \bar{W} \) will reach a steady state when the volume flux of buoyant mantle expands beneath the lithosphere predicts a swell width scale \( L_0 \propto B^{-3/4}/U_p \), where \( B \) is buoyancy flux. Line shows best fitting regression to the data, excluding Cape Verde. (c) Widths \( \bar{W} \) for hot-spot–ridge interaction are the total along-isochron span of positive residual topography: diamonds, Iceland; white triangles, Azores; white squares, Galápagos; inverted triangles, Tristan; circles, Easter. Plate rate \( U \) is the half spreading rate of the ridge during times corresponding to isochron ages (Ito and Lin, 1995). Black symbols mark present-day ridge-axis anomalies. Bold error bars show along-axis mantle-Bouguer gravity anomaly widths along the Southwest Indian Ridge near the Bouvet (black) and Marion (purple) hot spots (Georgen et al., 2001). Curves show predictions of scaling laws based on lubrication theory for a range of volume fluxes of buoyant mantle \( Q \). (d) Along-isochron widths of residual bathymetric anomalies vs plume-ridge separation distance at times corresponding to isochron ages for the Tristan–MAR system. Curve is best fitting elliptical function \( E(x/W_0^2) \) in eqn [7]. (c, d) Reproduced from Ito G, Lin J, and Graham D (2003) Observational and theoretical studies of the dynamics of mantle plume–mid-ocean ridge interaction. Review of Geophysics 41: 1017 (doi:10.1029/2002RG000117), with permission from AGU.
material at the source $Q$ is balanced by the sinks associated with lithospheric accretion near the mid-ocean ridge (Figure 18(b)). Results of numerical models are well explained by the scaling law (Feighner and Richards, 1995; Ribe, 1996; Ito et al., 1997; Albers and Christensen, 2001; Ribe et al., 2007):

$$\bar{W}/W_0 = C_s \left( \frac{Q \Delta \rho g}{48 \pi \sigma U^2} \right)^{1/2} E(x_r/W_0)$$  \[7\]

where $C_s \approx 2$, $c \approx 0.7$, and $E$ is an equation for an ellipse in terms of the normalized distance $x_r/W_0$ between the source and ridge axis. The characteristic width scale is

$$W_0 = \left( \frac{Q}{U} \right)^{1/2}$$  \[8\]

The curves in Figure 17(c) show widths predicting by [7] and [8] for seven cases of hot-spot-ridge interaction. The general inverse relationship between $W$ and $U^{1/2}$ explains the data reasonably well. Dispersion of $W$ at a given $U$ can be caused by differences in $Q$ and in hot-spot-ridge separation $x_r$ (Figure 17(d)).

Overall, the apparent correlations between hot-spot widths, fluxes, and plate rates can be well explained by buoyant material being introduced at the base of the lithosphere. Compositonally or thermally buoyant upwellings rising from below the asthenosphere are possible sources and have been widely explored in context of the mantle-plume hypothesis (see also below). Buoyant mantle could also be generated near the base of the lithosphere, perhaps due to buoyant decompression melting. Such a mechanism for swell generation may be an alternative to deep-seated thermal upwellings.

### 7.09.3.3.2 Generating swells: Thermal upwellings and intraplate hot spots

Hot mantle plumes provide a straightforward mechanism to explain both the swells and excess volcanism associated with some hot spots. Three dimensional (3-D) numerical models that solve the governing equations of mass, energy, and momentum equilibrium of a viscous fluid have quantified the physics of plume-generated swells (Ribe and Christensen, 1994; Zhong and Watts, 2002; van Hunen and Zhong, 2003). They have, for example, successfully predicted the shape and uplift history of the Hawaiian swell. They also predict the eventual waning of swell topography to occur as a result of the thinning (see also Figure 18(a)) and cooling plume material beneath the lithosphere. Such a prediction provides a simple explanation for the disappearance of hot-spot swells along the Hawaiian and Louisville chain, as well as the lack of swells around very old portions of other volcano chains. Ribe and Christensen (1994) also predict minimal thinning of the lithosphere; therefore, the predicted elevation in heat flow is smaller than the variability that can be caused by local crustal or topographic effects (DeLaughter et al., 2005; Harris and McNutt, 2007).

A similar model but with melting calculations defined the range of lithospheric thicknesses, potential temperatures, and buoyancy fluxes needed to generate the Hawaiian magma fluxes, swell width, and swell height (Ribe and Christensen, 1999).
For reference, a plume composed of anhydrous lherzolite requires high potential temperatures of 1500–1600°C to roughly match the observations. In addition to a main (shield) melting phase, this model also predicts a secondary zone of upwelling and melting substantially down the mantle ‘wind’ of the plume. The location of this melting zone away from the main zone is consistent with it contributing to part of the rejuvenated stages on some Hawaiian Islands.

7.09.3.3 Generating swells: Thermal upwellings and hot-spot–ridge interaction

Another series of numerical modeling studies help define the conditions for hot upwelling plumes to explain swells and melting anomalies along hot-spot influenced ridges. Initially, studies showed that the swell width and the crustal thickness along the MAR near Iceland required a very broad upwelling (radius ~300 km) of only modest excess temperature (<100°C above an ambient of 1350°C) (Ribe et al., 2005).
1995; Ito et al., 1996). These characteristics appeared to be inconsistent with the evidence from seismic tomography for a much narrower and hotter body (Wolfe et al., 1997; Allen et al., 1999b, 2002). However, calculations involving such a narrow and higher temperature upwelling predict crustal thicknesses that are several times that measured on Iceland – even assuming dry lherzolite as the source material.

One solution is provided by the likelihood that water that was initially dissolved in the unmelted plume material is extracted at the onset of partial melting, and this dramatically increases the viscosity of the residue (Hirth and Kohlstedt, 1996). Numerical models that simulate this effect predict that the lateral expansion of the plume material occurs beneath the dry solidus and generates the observed swell width along the MAR (Ito et al., 1999). Above the dry solidus, where most of the melt is produced, the mantle rises slowly enough to generate crustal thicknesses comparable to those at Iceland and along the MAR (Figure 20). A similar model, but with a variable flux of material rising in the plume produces

![Figure 20](image)

**Figure 20** (a)–(c) Comparisons between observations and predictions of a 3-D numerical model of a hot mantle plume rising into the upper mantle, interacting with two spreading plates, and melting. (a) Observed (light blue from gridded bathymetry and dots from refraction experiments) residual topography along the MAR, compared to the predictions, which assumes isostatic topography due to the thickened crust and low-density mantle. Dots show height above the seismically determined crustal thicknesses that are shown in (b). Dashed curve in (b) is predicted crustal thickness. (c) Perspective view of potential temperatures (white > ~1500°C, orange = 1350°C) within the 3-D model. The vertical cross-sections are along (left) and perpendicular (left) to the ridge. Viscosity decreases with temperature and increases at the dry solidus by $10^2$ because water is extracted from the solid with partial melting (Ito et al., 1999). (a, b) Reproduced from Hooft EEE, Brandsdottir B, Mjelde R, Shimamura H, and Murai Y (2006) Asymmetric plume-ridge interaction around Iceland: The Kolbeinsey Ridge Iceland Seismic Experiment. *Geophysics, Geochemistry, and Geosystems* 7: Q05015 (doi:10.1029/2005GC001123), from AGU. (c) Reproduced from Ito G. Shen Y, Hirth G, and Wolfe CJ (1999) Mantle flow, melting, and dehydration of the Iceland mantle plume. *Earth and Planetary Science Letters* 165: 81–96.
fluctuations in crustal production at the ridge that propagates away from the plume source along the ridge axis (Ito, 2001). This behavior was shown to explain V-shaped ridges that straddle the MAR north and south of Iceland for hundreds of kilometers (Vogt, 1971; Jones et al., 2002b). A north–south asymmetry in crustal thickness documented by Hooft et al. (2006) is not predicted by the above models and could hold clues to larger-scale mantle flow, heterogeneity, or both.

7.09.3.4 Dynamics of Buoyant Upwellings

Our discussion of hot spots and melting anomalies is set amidst the background of larger-scale processes of plate tectonics and mantle convection. The cooling of the oceanic lithosphere is the main driving force for plate motion and the associated convection in the mantle. While the volume of mantle that participates in the plate-tectonic cycle is still debated, a consensus model has emerged of moderated whole-mantle convection, with significant material exchange between the upper mantle with relatively low viscosity and the more sluggishly convection lower mantle. Observational evidence for whole-mantle convection is based on geoid and topography (e.g., Richards and Hager, 1984; Davies, 1998), geodynamic inversions (Mitrovica and Forte, 1997), seismological observations of slab extensions in to the lower mantle (Creager and Jordan, 1986), and seismic tomography (Grand, 1994; Su et al., 1994, Grand et al., 1997). Geodynamic models of whole-mantle convection show reasonable agreement with seismic tomography (Lithgow-Bertelloni and Richards, 1998; McNamara and Zhong, 2005), seismic anisotropy in the lower mantle (McNamara et al., 2002), as well as surface heat flux and plate motions (e.g., van Keken and Ballentine, 1998). Hot spots and melting anomalies represent smaller-scale processes that likely involve mantle plumes.

The lack of detailed knowledge about lower-mantle properties, such as rheology, and thermal conductivity and expansivity, provides speculative opportunities about dynamical behavior; but numerical calculations can be used to map out the likely range of outcomes. For example, whole-mantle convection models with reasonable degrees of internal heating can satisfactorily explain both the average surface heat flow and plate velocities, but only when a higher-viscosity mantle is assumed (e.g., van Keken and Ballentine, 1998). Mineral physics (see Chapter 206) provide strong suggestions for a reduction in thermal expansivity and increase in thermal conductivity with pressure. The combined effects will reduce convective vigor, but models that incorporate reasonable depth variations of these properties predict that this does not render the lower mantle immobile (e.g., van Keken and Ballentine, 1999; van den Berg et al., 2005; Matyska and Yuen, 2006a). Dynamical theory provides an essential stimulus for the mantle-plume hypothesis, since thermal plumes form naturally from hot boundary layers in a convecting system. The CMB is the main candidate to have a significant TBL (e.g., Boehler, 2000), but other boundary layers may exist at locations where sharp transitions in material properties or composition occur, such as the bottom of the transition zone at 670 km depth and the proposed thermochemical layer at the base of the mantle. We will first summarize the fluid dynamics of plumes rising from a TBL before addressing the consequences of chemical buoyancy forces and depth-dependent mantle properties.

7.09.3.4.1 TBL instabilities

In its simplest form, the growth of an upwelling instability from a hot boundary layer can be approximated as a Rayleigh–Taylor instability with the onset time and growth rate controlled by the local (or boundary) Rayleigh number,

$$R_{\text{Ray}} = \frac{\rho g \alpha \Delta T \delta^2}{\mu \kappa}$$

[9]

The instability is enhanced by thermal expansivity $\alpha$ ($\rho$ is density and $g$ is acceleration of gravity), the temperature jump across the boundary layer $\Delta T$, and the layer thickness $\delta$, and is hampered by viscosity $\mu$, and thermal diffusivity $\kappa$ (including potential radiative effects in the deep mantle). Large-scale mantle flow tends to suppress the growth of instabilities but the temperature dependence of rheology will enhance its growth. For more specifics on governing equations for boundary layer instabilities and examples of their modeling with laboratory and numerical techniques see chapters 6 and 11 of Schubert et al. (2001). Indeed, analytical methods provide important insights to the rate of formation of the instability and the dependence on ambient conditions (see e.g., Whitehead and Luther, 1975; Ribe and de Valpine, 1994). The growth of a diapir to a full plume can be understood with nonlinear theory; for example, Bercovici and Kelly (1997) show that growth is retarded by draining of the
source layer and the diapir can temporarily stall. Experimental and numerical investigations confirm and expand these predictions and, quite importantly, provide direct verification of model predictions made by independent approaches (e.g., Olson et al., 1988; Ribe et al., 2007). In general, most studies find that for reasonable lower-mantle conditions, the boundary layer instabilities will grow on the order of 10–100 My (e.g., Christensen, 1984; Olson et al., 1987; Ribe and de Valpine, 1994).

As the diapir rises it will generally be followed by a tail of hot material that traces back to the boundary layer. The rise speed of the diapir is proportional to its buoyancy, the square of its radius, and is inversely proportional to viscosity. The morphology of the plume head and tail are controlled by the viscosity contrast between the hot and ambient fluid. A more viscous plume will tend to form a head approximately the same width as the tail (a ‘spout’ morphology) whereas a lower-viscosity plume will tend to form a voluminous plume head much wider than the tail (a ‘mushroom’ or ‘balloon’ geometry following terminology by Kellogg and King (1997)). Since mantle viscosity is a strong function of temperature it is generally expected that the mushroom/balloon geometry should dominate, but Korenaga (2005a) proposes an interesting counter argument for grain-size controlled, high-viscosity plumes. As we will discuss below, chemical buoyancy may have significant control on the shape of the plume as well.

Dynamical experiments without large-scale flow generally demonstrate that a boundary layer will become unstable with many simultaneous plumes that interact with each other as they rise through the fluid (e.g., Whitehead and Luther, 1975; Olson et al., 1987; Kelly and Bercovici, 1997; Manga, 1997; Lithgow-Bertelloni et al., 2001). To study the dynamics of a single plume, it has become common to use a more narrow or point source of heat, which in laboratory experiments can be achieved by inserting a small patch heater at the base of the tank (e.g., Kaminski and Jaupart, 2003; Davaille and Vatteville, 2005) or to inject hot fluid through a small hole in the base of the tank (Griffiths and Campbell, 1990). The latter work showed that with strongly temperature-dependent viscosity the plume head entrains ambient fluid, forming a characteristic mushroom-shaped head. Interestingly, this same shape was observed also by Whitehead and Luther (1975) but for mixing of fluids with similar viscosity (their figure 9). Van Keken’s (1997) replication of Griffiths and Campbell’s (1990) laboratory experiment also showed that this form of plume is retained when it originates from a TBL or when olivine, rather than corn syrup rheology, was assumed. Other relevant numerical experiments are provided by Davies (1995) and Kellogg and King (1997).

7.09.3.4.2 Thermochemical instabilities

Studies of thermal plumes originating from TBLs have guided much of the classic descriptions of mantle upwellings and represent a logical starting point for understanding them. The Earth, however, is more complex since density is likely to be controlled by composition, as well as temperature. The seismic structure in the deep mantle beneath the African and South Pacific Superswell regions provides evidence for such deep compositional heterogeneity. Mantle convection models suggest that dense layers are likely to form distinct large blobs or piles that are away from areas of active downwellings (Tackley, 2002; McNamara and Zhong, 2005). Due to the spatial and temporal interaction between chemical and thermal buoyancy forces, the upwellings that form from a thermochemical boundary layer can be dramatically different from the classical thermal plume and interaction with the lower-viscosity upper mantle can significantly alter their shape (Farnetani and Samuel, 2005). The stable topography of high-density layers could provide an anchoring point above which thermal plumes can rise and thus define a fixed reference frame for different hot-spot groups (Davaille et al., 2002; Jellinek and Manga, 2002).

A compelling cause for compositional heterogeneity is the recycling of oceanic crust in subduction zones (e.g., Christensen and Hofmann, 1994). The density of the mafic (eclogitic) crust likely remains higher than that of the ambient mantle through most of the lower mantle (Ono et al., 2001). A layer generated by oceanic crust recycling is likely to remain stable if its density is in the range of 1–6% greater than that of the ambient mantle (Sleep, 1988; Montague and Kellogg, 2000; Zhong and Hager, 2003; Brandenburg and van Keken, in press). Entrainment of this layer by plumes provides a straightforward explanation for the geochemically observed oceanic crust component in OIBs (Shirey and Walker, 1998; Eiler et al., 2000). The potential for entrainment of a deep chemical boundary was studied systematically by Lin and van Keken (2006a, 2006b) who found that with strongly temperature-dependent viscosity the entrainment would become
episodic under a large range of conditions. The style of entrainment ranged from nearly stagnant large plumes in the lower mantle to fast episodic pulsations traveling up the pre-existing plume conduit, which provides an explanation for the pulses of LIP volcanism (Lin and van Keken, 2005). Oscillatory instabilities in starting plumes can be caused by the competing effects of thermal and chemical buoyancy with particularly interesting effects where the effective buoyancy is close to zero (Samuel and Bercovici, 2006).

7.09.3.4.3 Effects of variable mantle properties

Large variations in material properties can lead to complex forms and time dependence of buoyant upwellings. For example, the combination of increasing ambient viscosity, thermal conductivity, and decreasing thermal expansivity with depth will likely cause plumes to be relatively broad in the deep mantle but become thinner when migrating upward (Albers and Christensen, 1996). Changes in rheology fundamentally alter plume dynamics. The sharp decrease of ambient viscosity for a plume rising from the lower, to the upper mantle will cause a rapid increase in speed and resulting drop in plume width (van Keken and Gable, 1995). Such necking may also cause the formation of a second boundary layer with episodic diapirism with timescales on the order of 1–10 My (van Keken et al., 1992, 1993). The viscosity change can also completely break apart a starting mantle-plume head into multiple diapirs, perhaps contributing to multiple flood basalt episodes (Bercovici and Mahoney, 1994). An important aspect of mantle rheology is the non-Newtonian behavior, which is characterized by a viscosity that has a strong, nonlinear function of stress in addition to temperature and pressure. The strong stress dependence can dramatically enhance the deformation rate of boundary layer instabilities and lead to much higher rise speeds than is observed in Newtonian fluids (where strain rate and stress are linearly related) (Larsen and Yuen, 1997; Van Keken, 1997; Larsen et al., 1999). Such behavior can cause starting plume heads to rise sufficiently fast to almost completely separate from the smaller tail, thus providing an alternative explanation for the observed LIP episodicity (Van Keken, 1997).

The transition zone is also characterized by major phase changes in the upper-mantle mineral assemblages, dominated by the exothermic 400 km discontinuity and the endothermic 670 km discontinuity. The phase changes provide a dynamical influence that can strongly modify plume flow, with predictions for more episodic or faster plume flow in the upper mantle (Nakakuki et al., 1997; Brunet and Yuen, 2000). One model provides a source for plumes even just below the 660 km discontinuity (Cserepes and Yuen, 2000), an intriguing possibility for hot spots with sources above the CMB.

The sluggish nature of the lower mantle may be enhanced if heat radiation becomes efficient at high temperature. This has been explored for mantle plumes by Matyska et al. (1994) who suggest that the radiative components will strongly enhance the stability and size of large ‘super’plume regions in the lower mantle, even without chemical stabilization. The plume stability may be enhanced by the possible post-perovskite transition at the base of the lower mantle (Matyska and Yuen, 2005, 2006a).

Thus, the concept of a cylindrical plume, rising vertically from the CMB to the lithosphere, is probably far too simple. Upwellings are likely to take on complex shapes, have a wide range of sizes, be strongly time dependent, and originate from different depths in the mantle. We will revisit such issues in Section 7.09.3.7.

7.09.3.4.4 Plume buoyancy flux and excess temperature

A long-standing question concerns the efficiency of heat transport in mantle plumes. If hot spots are dynamically supported by plumes that rise from the CMB, then we can use surface observations to estimate the heat from the core. The topography of hot-spot swells provides a fundamental constraint on the buoyancy flux of plumes. Davies (1988) estimated, from swell heights of 26 hot spots provided by Crough (1983), that new topography is being generated at a rate $S = 17.5 \text{ m s}^{-1}$ (in comparison to 300 m s$^{-1}$ for the total mid-oceanic ridge system). Iceland was excluded from this compilation and the overall flux was dominated by the Pacific hot spots and in particular by Hawaii, Society, and the Marquesas. The total buoyancy flux follows by multiplication with the density difference between mantle and sea water, $B = (\rho_m - \rho_w)S = 40 \text{ Mg s}^{-1}$, assuming $\rho_w = 2300 \text{ kg m}^{-3}$. The estimated heat flux carried by the plumes $q_p$ follows from $q_p = \rho_w C_p S/\alpha$, which is around 2 TW, assuming $C_p = 1000 \text{ J kg}^{-1} \text{ K}^{-1}$ and $\alpha = 3 \times 10^{-5} \text{ K}^{-1}$. Sleep (1990) provided a similar analysis for 34 hot spots, including Iceland, and found a slightly larger value of 55 Mg s$^{-1}$ for the total buoyancy flux, implying a
core heat loss of \(\sim 2.7\) TW. The plume heat flux is therefore significantly smaller than the total heat flux of the Earth of \(44\) TW (Pollack et al., 1993).

The low estimates for plume heat flux are suggestive of only a minor contribution of plumes to the cooling of the Earth, and potentially a small contribution of core cooling, under the assumption that hot-spot heat is related to the heat from the core. However, the role of compressible convection, internal heating modes, and interaction with the large-scale flow and plate tectonics may be important in masking heat rising from the core. For example, a large-scale mantle circulation associated with plate tectonics could be drawing heat from the CMB toward mid-ocean ridges rather than allowing it to rise in vertical plumes to form mid-plate hot spots (Gonnermann et al., 2004). Statistical arguments for the power-law distribution of plumes have been used to debate that many small plumes are entrained in the large-scale flow and do not express themselves as hot spots (Malamud and Turcotte, 1999). This idea has been supported by mantle convection models that simulate a wide range of heating modes (Labrosse, 2002; Mittelstaedt and Tackley, 2006).

The temperature increase across the TBL at the CMB can be estimated based on the adiabatic extrapolation of upper-mantle temperatures and mineral physics constraints on the temperature of the core. An estimate exceeding 1000 K (Boehler, 2000) is much greater than the temperature anomalies expected in the upper mantle beneath the major hot spots of only 200–300 K (see Section 7.09.3.2.1). If plumes rise from the CMB, what then is the mechanism for reducing the plume temperature by the time it reaches the base of the lithosphere? While entrainment of, and diffusive heat loss to the surrounding cooler mantle will reduce the plume excess temperature, most calculations based on the classical plume model suggest only modest reductions (e.g., Leitch et al., 1996; Van Keken, 1997).

In our view, the likely important role of thermo-chemical convection and the variable properties of the mantle (see Sections 7.09.3.4.2 and 7.09.3.4.3) can provide a self-consistent resolution to the above discrepancy. For example, Farnetani (1997) showed that if a compositionally dense layer stabilizes at the base of D\(^n\), plumes will tend to rise from only the top of the TBL, which is substantially less hot than the CMB. In addition, decompression and adiabaticity can enhance the cooling of plumes as they rise through the mantle and further reduce their surface expression (Zhong, 2006).

### 7.09.3.5 Chains, Age Progressions, and the Hot-spot Reference Frame

Thermal plumes rising from below the upper mantle to the lithosphere provide a reasonably straightforward explanation for a source of at least some hot-spot swells and age-progressive volcano chains. An origin residing below the asthenosphere allows for the possibility of a kinematic reference frame that is distinct from the plates. On the other hand, if thermal plumes rise from the deep, convecting mantle, it should be intuitive that hot spots are not stationary (Section 7.09.2).

A series of studies initiated by Steinberger (Steinberger and O’Connell, 1998; Steinberger, 2000) have used numerical models to simulate plumes rising in a convecting mantle. Their calculations of a spherical Earth assume that mantle flow is driven kinematically by the motion of the plates with realistic geometries, and dynamically by internal density variations estimated from different seismic tomography models. The viscosity structure includes high viscosities in the lower mantle (\(\sim 10^{22} – 10^{23}\) Pa s) and lower viscosities in the upper mantle (\(\sim 10^{20} – 10^{21}\) Pa s). A plume is simulated by inserting a vertical conduit in the mantle at a specified time in the past. Velocities at each point along the conduit are the vector sum of the ambient mantle velocity and the buoyant rise speed of the conduit, which is computed based on scaling laws derived from theory and laboratory experiments. For simplicity, the ambient mantle flow is not influenced by the plumes. Plume conduits therefore deform with time and where they intersect the base of the lithosphere defines the location of the hot spots (Figure 21).

This method was applied to examine the evolution of the Hawaiian, Louisville, and Easter hot spots in the Pacific ocean (Steinberger, 2002; Koppers et al., 2004). A mantle flow model was found to optimize fits between predicted and observed age progressions along the whole lengths of all three chains. For the Hawaiian hot spot, the models predict absolute southward motion that was rapid (average \(\sim 40\) km My\(^{-1}\)) during 50–80 Ma (prior to the Hawaiian–Emperor Bend) and slower (<20 km My\(^{-1}\)) since 50 Ma, thereby providing an explanation for the paleomagnetic evidence (Tarduno et al., 2003; Pares and Moore, 2005; Sager et al., 2005). The models also predict slow eastward motion of Louisville, consistent with the observed nonlinear age progression (Koppers et al., 2004), as well as WSW motion of the Easter hot spot at rates of \(\sim 20\) km My\(^{-1}\).
The above studies addressed one group of hot spots but a key challenge is to explain the age progressions of both the Pacific and Indo-Atlantic hot spots in a single whole-mantle flow model. Steinberger et al. (2004) started this by considering two kinematic circuits to define the relative motions between the Pacific and African Plates: (1) through Antarctica, south of New Zealand and (2) through the Lord Howe Rise, north of New Zealand, Australia, and then Antarctica. The reference case of fixed hot spots predicts hot-spot tracks that deviate substantially from observed locations (dotted curves in Figure 22) for both plate circuits. Considering moving hot spots derived from mantle flow simulations with plate circuit (1) yields reasonable matches to the tracks for ages <50 Ma but predicts a track too far west of the Emperor chain for ages >50 Ma. Finally, predicted tracks with moving hot spots using plate circuit (2) provide the closest match to the observed tracks. This model successfully predicts the geographic age progressions along most of the Tristan, Réunion, and Louisville tracks, and for the Hawaiian track since 50 Ma, including a bend between the Hawaiian and Emperor seamounts. But the bend is not sharp enough: the models still predict a trajectory for the Emperor seamounts too far west.

The above studies illustrate that models of plumes rising in a geophysically constrained, mantle flow field can explain many key aspects of apparent hot-spot motion. The studies, however, underscore the importance of uncertainties in defining relative plate motions, particularly in the presence of diffuse plate boundaries – for example, that near Lord Howe Rise. Still more uncertainties are associated with the locations of volcanism in time and in paleomagnetic latitudes. The models are sensitive to a number of properties such as mantle viscosity structure, the choice of seismic tomography model, the mapping between seismic velocities and density, as well as the buoyancy and dimensions (which control the rise speed) of the mantle plumes. A recent study has just begun to quantify the observational uncertainties and to use them to define statistically robust mantle flow solutions (O'Neill et al., 2003). But many observations remain poorly understood, including the location and trend of the older portion of the Emperor chain.

7.09.3.6 Large Igneous Provinces

The rapid and massive magmatic production of many LIPs, combined with their strong connection to continental breakup, but inconsistent connection to
present-day hot-spot volcanism are challenging to understand. Moreover, the wide range of eruptive volumes (Figure 8) and durations suggest that there may not be just one overarching mechanism.

The observation of large plume heads followed by thin tails in fluid dynamical experiments has traditionally been used to explain the LIP-hotspot connection (Richards et al., 1989) and remains, because of its simplicity and plausibility, an attractive base model for the formation of many LIPs. Its strengths include that: (1) it is supported by fluid dynamics for increasingly realistic assumptions about mantle composition and rheology (see Section 7.09.3.4), in fact these modifications to the base model allow for an explanation of some of the diversity seen in the geological record (Figure 23); (2) it offers a dynamical cause for the common disconnect between LIP and hotspot volcanism (Farnetani and Samuel, 2005) (i.e., some plume heads rise to the surface without plume tails and some upwellings form narrow plume tails without heads, Figure 23). (3) It predicts the hottest material of rising plume heads will erupt first (Farnetani and Richards, 1995) which explains high MgO basalts early in the LIP record; and (4) it

Figure 22 Computed hot-spot motion (rainbow colored bands with color indicating position in time according to scale on the lower right) and tracks overlain on gravity maps (left color scale). Tracks are plotted on all plates regardless of whether a hot spot was actually on the plates during those times. Ticks along tracks are every 10 My. Dotted (red and green) lines are solutions assuming fixed hot spots. Solid red lines (plate circuit (1) shown for Hawaii and Louisville) and purple lines (plate circuit (2) shown for Hawaii) are for moving hot spots in mantle flowing in response to absolute plate motions that optimize fits to only the Tristan and Réunion hot-spot tracks. Green lines (shown for Réunion and Tristan) are best fit solutions to only the Hawaii and Louisville tracks using plate circuit (1). Black (plate circuit (1)) and blue (plate circuit (2)) are solutions that optimize fits to all four tracks. A least-squares method is used to optimize the fit to locations and radiometric ages of seamounts. Reprinted by permission from Macmillan Publishers Ltd: (Nature) Steinberger B, Sutherland R, and O’Connell RJ (2004) Prediction of Emperor-Hawaii seamount locations from a revised model of global plate motion and mantle flow. Nature 430: 167–173, copyright (2004).
predicts that the arrival of the plume at the surface leads to uplift and extension which is observed in the geological record of many LIPs (see Section 7.09.2.3). Plume-based models, however, have yet to adequately explain the strong correlation between LIPs and continental breakup and the lack of uplift during the OJP eruptions (d’Acremont et al., 2003).

There are a number of alternative mechanisms that address the above issues, which include shallow, sublithospheric processes or meteorite impacts. While the plume model has received significant attention and quantitative hypothesis testing, the majority of the alternatives are currently still in rather qualitative form.

The first alternative to thermal plumes pertains to LIPs formed on continental margins. Anderson (1994b) proposes that excess heat can build below continents during tectonic quiescence and/or supercontinent formation, which then causes the massive eruptions during continental breakup. This hypothesis addresses the correlation between LIPs and continental breakup and the lack of connection of some continental LIPs to hot-spot trails. One aspect not addressed specifically is why the volcanism is typically not margin-wide but, instead, is more restricted in total extent. Nevertheless, the correlation between the LIPs and continental breakup is intriguing and it is quite likely that regional variations in the composition and strength of the lithosphere have an important control on the location of magma eruption.

Second, delamination of continental lithosphere and secondary convection at rifted margins (Figure 15(b)) have been forwarded to generate LIPs near continents (King and Anderson, 1995, 1998; van Wijk et al., 2001; Hales et al., 2005; Anderson, 2005). Like the above concept of subcontinental mantle incubation, these models can explain continental LIPs without hot-spot tracks, but have yet to show how they could form LIPs with hot-spot tracks. Indeed, more quantitative modeling of shallow mantle processes needs to be done.

A third alternative, which could apply to LIPs that form near sites of continental or oceanic rifting, is that compositional, rather than thermal effects cause excess melting, for example, more fertile mantle such as eclogite and/or water in the source (e.g., Anderson, 1994a, 2005; Cordery et al., 1997; Korenaga, 2004; 2005b) (Figure 15(a)). The strengths of this possibility include that some compositional effects are expected and these can strongly enhance melt

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**Figure 23** Temperatures of upwelling instabilities rising from the CMB predicted by 3-D numerical models. The CMB is heated from below and is blanked by a layer initially 200 km in thickness and containing material with an intrinsic, compositional density that is 1% greater than that of the surrounding lower mantle. Upwelling through a depth of 670 km is inhibited by an endothermic phase change with a Clapeyron slope of $-2.5 \text{ MPa} \text{ K}^{-1}$. In the top row a large, roughly spherical plume head rapidly rises and could generate flood volcanism at the surface. The head detaches from the stem and this may delay or prevent the formation of a chain of volcanoes extending away from the flood basalts. In the bottom row, only a narrow upwelling rises into the upper mantle. Reproduced from Farnetani CG and Samuel H (2005) Beyond the thermal plume hypothesis. *Geophysical Research Letters* 32: L07311 (doi:10.1029/2005GL022360), with permission from AGU.
production (see Section 7.09.3.2.2); the lack of uplift of OJP could be explained if dense eclogite occurs in the source; and it can explain the formation of LIPs not connected to long-lived hot-spot tracks. Again, a key dynamic weakness of models invoking eclogite in the source is that eclogite is dense and requires some mechanism to stay near, or be brought back up to the surface.

Fourth, the enigmatic nature of the OJP has led to the suggestion that a meteorite impact could be responsible for the emplacement of LIPs (Rogers, 1982; Jones et al., 2002a; Tejada et al., 2004; Ingle and Coffin, 2004). The strengths are that the decompression of mantle following impact may generate extensive melting (although perhaps with some qualifications (Ivanov and Melosh, 2003)) with less uplift than expected from a hot plume head and without a connection to a hot-spot track. It is rare, however, to find direct evidence for meteorite impact during LIP emplacement. One of the few convincing observations is the iridium anomaly embedded in the Deccan Traps, but if this impact signal is related to the Chixculub impact, it post-dates the start of volcanicism (Courtillot and Renne, 2003). Also, it is statistically unlikely that the majority of Phanerozoic LIPs can be explained by impacts (Ivanov and Melosh, 2003; Elkins-Tanton and Hager, 2005).

Finally, a cursory investigation of planetary impact cratering suggests that many large craters on the Moon, Mars, and Venus form without contemporaneous volcanicism. An interesting recent speculation is the iridium anomaly embedded in the Deccan Traps, but if this impact signal is related to the Chixculub impact, it post-dates the start of volcanicism (Courtillot and Renne, 2003). Also, it is statistically unlikely that the majority of Phanerozoic LIPs can be explained by impacts (Ivanov and Melosh, 2003; Elkins-Tanton and Hager, 2005).

There are a few ways to initiate plumes from boundary layers internal to the mantle. High-resolution 2-D convection simulations by Matyska and Yuen (2006b) predict large-scale ($10^3$–$10^4$ km), superplume-like upwellings as a result of relatively low local Rayleigh numbers caused by high viscosities, low thermal expansivity, and radiative heat transfer in the lower mantle (Figure 24(b) and 24(d)). As superplumes rise through the lower mantle, further rise can be inhibited by the endothermic phase change at 660 km. The hot tops of the superplumes can generate a TBL from which smaller scale ($10^3$ km) upper-mantle upwellings can originate. Another possible surface for a mid-mantle TBL is the top of a chemically dense layer in the lower mantle (see Section 7.09.3.4.2). Laboratory experiments show that when an initially chemically stratified system is heated from below and cooled from above, a variety of forms of upwellings and downwellings occur, depending on the ratio of chemical-to-thermal buoyancy (Davaille, 1999). When the negative chemical buoyancy of the lower layer is $\sim$0.35–0.55 times the positive buoyancy due to the basal heating, the two layers remain separate but the surface between them undulates to form broad downwellings and superplume upwellings. Above the upwellings, smaller instabilities can rise into the upper layer and to the surface (Figure 24(e)). In both of the two examples described above, smaller upper-mantle plumes are shown to rise from the top of broad
superplumes. Such a situation could provide an explanation for the large frequency of short-lived hot spots in the superswell regions of the South Pacific and Africa (e.g., Courtillot et al., 2003; Koppers et al., 2003).

7.09.3.7.2 Forming melting anomalies by upper-mantle processes

Buoyant upwellings have been the focus of a large number of studies but are probably not the only phenomena giving rise to melting anomalies. Upper-mantle processes that are largely decoupled from the lower mantle undoubtedly contribute to the magmatism in various ways.

Small-scale, sublithospheric convection, as introduced in Section 7.09.3.2, is one possible mechanism. Sublithospheric convection could be evidenced by a number of observations: it could limit the maximum thickness of the lithosphere and slow the subsidence of old seafloor (e.g., Huang et al., 2003); it could give rise to the prominent gravity lineations over the Pacific seafloor (Haxby and Weissel, 1986); and it could explain the periodic fluctuations in upper-mantle seismic structure as imaged perpendicular to hot-spot swells in the Pacific (Katzman et al., 1998). Small-scale sublithospheric convection may explain magmatism along lineaments parallel to plate motion (e.g., Richter, 1973; Huang et al., 2003), but perhaps

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**Figure 24** (a) Photographs, the left taken prior to the right, of a laboratory experiment involving sucrose solution heated from the base. Isotherms (white streaks) outline the hot upwellings that detach from the base, as imaged using suspended liquid crystals and glass particles. (b) 2-D numerical simulations of whole-mantle convection in which the mantle is heated internally, at its base, and is cooled at the surface. Viscosity is temperature- and pressure-dependent. Calculations also include an endothermic phase change at 670 km and an exothermic phase change at 2650 km. Coefficient of thermal expansion decreases by >10 times from the surface to the CMB. (b) A thermal conductivity $k = 1$ corresponds to that due to heat diffusion, whereas (c) and (d) include an additional term (dependent on temperature) for radiative heat transfer. (e) Laboratory experiments involving a chemically dense and more viscous fluid underneath a chemically less dense and less viscous fluid. The system is heated from below and cooled from above. Dark and light bands show the interface between the two fluids, which bows upward and resembles the upper surface of a mantle superplume. Smaller plumes are imaged to be rising from this surface into the upper layer. (a) Reproduced from Davaille A and Vatteville (2005) On the transient nature of mantle plumes. Geophysical Research Letters 32, doi:1029/2005GL023029, with permission from AGU. (b)–(d) Reproduced from Matyska C and Yuen DA (2006b) Upper-mantle versus lower-mantle plumes: are they the same? In: Foulier GR and Jurdy DM (eds.) The Origins of Melting Anomalies: Plates, Plumes, and Planetary Processes. GSA, with permission from GSA. (e) Reprinted by permission from Macmillan Publishers Ltd: (Nature) Davaille A (1999) Simultaneous generation of hotspots and superswells by convection in a heterogeneous planetary mantle. Nature 402: 756–760, copyright (1999).
without a systematic geographic age progression. Quantitative studies of melting have yet to be done but are needed to quantify the rates and time dependence of magma production, as well as whether small-scale convection could generate seafloor swells.

Another mechanism that has been recently proposed is fingering instabilities of low-viscosity asthenosphere (Weeraratne et al., 2003; Harmon et al., 2006). When two fluids are contained in a thin layer (Hele–Shaw cell in the laboratory and possibly the asthenosphere in the upper mantle) and one fluid laterally displaces a more viscous fluid, the boundary between the two becomes unstable and undulates with increasing amplitude (Saffman and Taylor, 1958). Fingers of the low-viscosity fluid lengthen and penetrate the high-viscosity fluid. Perhaps hot mantle rising beneath the South Pacific Superswell area is supplying hot, low-viscosity asthenosphere that is fingering beneath the Pacific Plate and generating some of the volcanic lineaments such as Pukapuka (Harmon et al., 2006; Weeraratne et al., in press). Weeraratne and Parmentier (2003) explore this possibility using laboratory experiments that simulate asthenospheric conditions.

Finally, numerous studies have suggested that heterogeneity in lithospheric stresses or structure can allow magma that is already present in the asthenosphere to erupt at the surface (e.g., Anguita and Herman, 2000; Clouard et al., 2003). Lithospheric stress associated with regional (e.g., Sandwell et al., 1995) or local (e.g., Mittelstaedt and Ito, 2005) tectonics, as well as thermal contraction (Gans et al., 2003) could initiate fissures that can propagate and cause volcanic lineaments over a range of scales. Sandwell and Fialko (2004), for example, demonstrate that top-down cooling of the lithosphere generates thermoelastic tensile stress, which is optimally released by local zones of fracturing with spacing comparable to the flexural wavelength of the lithosphere and to the distances between the Pukapuka, Sojourn, and Hotu–Matua Ridges in the South Pacific. Natland and Winterer (2005) propose a lithospheric fissure origin for most or all of the Pacific hot spots, but such a hypothesis has yet to be tested quantitatively.

Another form of stress-influenced magma penetration could redistribute magma from diffuse volumes in the asthenosphere to discrete, localized eruption sites at the surface. The weight of a volcano can draw further magmatism if lithospheric stresses due to loading focuses magma-filled cracks toward the volcano (Muller et al., 2001) or if damage that is related to volcano loading enhances permeability in the lithosphere beneath it (Hieronymus and Bercovici, 1999). Parametrized models of damage-enhanced lithospheric permeability predict volcano chains to form either from a plate moving over a hot-spot-like source of magma (Hieronymus and Bercovici, 1999) or without a hot-spot source, but with nonlithostatic, horizontal tension in the lithosphere related to plate tectonics (Hieronymus and Bercovici, 2000) (Figure 25). The latter result provides another

![Figure 25](image-url)
plausible mechanism for forming lineaments without a monotonic age progression, but instead, with testable progressions of decreasing volcano age in both directions along the lineament. Another test is provided by the relationship between tectonic stress and the direction of volcano propagation. Models based on fracture mechanics predict propagation of volcanism perpendicular to the tensile direction, while Hieronymus and Bercovici (2000) predict volcanism to propagate along the tensile direction by the interaction of point-load flexural and background stresses (Figure 25).

7.09.3.8 Geochemistry of Hotspots and Melting Anomalies Vs MORB

The geochemical differences between basalts erupted at hot spots and melting anomalies versus MORB provide a vital constraint on the causal mechanisms (see also Chapters 7.10 and 2.04). We have focused on $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, $^{3}\text{He}/^{4}\text{He}$ ratios because they are key to tracing at least five flavors of mantle materials with different, time-averaged, chemical histories (Hart et al., 1992; Hanan and Graham, 1996; Zindler and Hart, 1986). Lavas with low $^{3}\text{He}/^{4}\text{He}$ and minimal $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ probably come from depleted mantle material (DM), referring to a long-term ($>10^7$ My) depletion in incompatible elements. Enriched (EM1 or EM2, i.e., high $^{87}\text{Sr}/^{86}\text{Sr}$) and HIMU (i.e., high $^{206}\text{Pb}/^{204}\text{Pb}$) mantle are thought to be influenced by subducted and recycled material, the former by old oceanic sediments or metasomatized lithosphere, and the latter perhaps by oceanic crust that has been hydrothermally altered and then devolatilized during subduction (e.g., Cohen and O’Nions, 1982; Hofmann and White, 1982; Hart et al., 1992; Zindler and Hart, 1986; Hofmann, 1997). High $^{3}\text{He}/^{4}\text{He}$, moderately low $^{87}\text{Sr}/^{86}\text{Sr}$, and intermediate-to-high $^{206}\text{Pb}/^{204}\text{Pb}$ compositions mark the fifth geochemical material; it has been identified with various names and we will refer to it as FOZO (for FOcal ZOne (Hart et al., 1992)). Its origin, however, is not well understood. The ‘standard’ hypothesis is that $^{3}\text{He}/^{4}\text{He}$ measures the primordial nature of the source material with low ratios reflecting material that has experienced substantial degassing of primordial $^{3}\text{He}$ and the high ratios indicating relatively undegassed mantle. But more recent evidence weakens the standard hypothesis and instead suggests that FOZO, in fact, has been depleted in highly incompatible elements. In this scenario, the high $^{3}\text{He}/^{4}\text{He}$ ratio could reflect a low $^{4}\text{He}$ concentration as a result of low U and Th content (Coltice and Ricard, 1999; Stuart et al., 2003; Meibom et al., 2005; Parman et al., 2005).

The key issue is that MORB appears to be heavily influenced by DM and minimally influenced by subducted materials and FOZO, whereas hot spots and melting anomalies appear to be influenced substantially by all five components (albeit to different degrees for different volcano groups). One possibility is that the pressure/temperature dependence of mantle viscosity and mineralogy, as well as density differences between the different mantle materials promotes large-scale layering in mantle geochemistry. DM is likely to be compositionally light and may tend to concentrate in the upper mantle where it is sampled by mid-ocean ridge magmatism. Mantle plumes, which feed hot spots, rise from deeper levels in the mantle and incorporate the other materials in addition to DM.

The formation of the different geochemical components, as well as the possibility of large-scale layering in the presence of vigorous, whole-mantle convection is actively being studied with both computational and laboratory methods (e.g., Christensen and Hofmann, 1994; van Keken and Ballentine, 1999; Davaille, 1999; Ferrachat and Ricard, 2001; Xie and Tackley, 2004). On the one hand, such studies have successfully predicted the formation of deep layers that are concentrated in dense subducted mafic material, which, if entrained in upwelling plumes could explain some of the elevated $^{206}\text{Pb}/^{204}\text{Pb}$ ratios in hot-spot basalts. On the other hand, it remains to be seen how it is possible to generate and physically separate two (or more) different components that may be depleted of mafic components: one with low $^{3}\text{He}/^{4}\text{He}$ that is prominent in MORB and the other with high $^{3}\text{He}/^{4}\text{He}$ that is weakly expressed in MORB and more prominently expressed in some hot-spot lavas. Another challenge is to reconcile the geochemical character of hot spots/melting anomalies with the possibility that some could be caused by plumes originating very deep in the mantle, some by plumes originating from shallower in the mantle, and others from shallow mechanisms completely unrelated to plumes. Finally, the small heat flux of mantle plumes implied by observations of swell buoyancy flux (Section 7.09.3.4.4), as well as constraints on excess temperatures of plumes in the upper mantle (Zhong, 2006) require that incompatible-element-rich materials are present both above and below the source layer of most mantle plumes.
One key process to consider in addressing the above issues is the chemical extraction of the different components by melting (Phipps Morgan, 1999). Geochemical evidence indicates that heterogeneity is likely to be present over a range of spatial scales, including scales much smaller than the size of upper-mantle melting zones (e.g., Niu et al., 1996; Phipps Morgan, 1999; Saal et al., 1998; Reiners, 2002; Salters and Dick, 2002; Stracke et al., 2003; Kogiso et al., 2004; Ellam and Stuart, 2004). The likelihood that different materials begin melting at different depths for a given mantle temperature makes it probable that differences in lithospheric thickness, as well as the rate of mantle flow through the melting zone can influence the relative proportions of incompatible elements that are extracted from the different components.

Mid-ocean ridge magmatism could most substantially melt the refractory component (DM) because the thin lithosphere allows for the greatest amount of decompression melting. Magmatism away from mid-ocean ridges could be less influenced by DM and proportionally more by the other, perhaps less refractory components owing to the thicker lithosphere. Melting of a buoyant upwelling – like a mantle plume – can also emphasize the least refractory components, even beneath relatively thin lithosphere, because the buoyancy pushes mantle through the deepest portions of the melting zone more rapidly than in the shallowest portions (Ito and Mahoney, 2005, 2006). Unraveling the above clues provided by magma geochemistry will thus require integrated geochemical, geophysical, and geodynamic investigations of the character of the mantle source, as well as the mantle convection, melting, and melt extraction.

7.09.4 Conclusions and Outlook

The rich diversity of observations and dynamical behavior makes it likely that a variety of mechanisms cause hot spots and melting anomalies. Our future task is to design observational and theoretical tests of which mechanisms can and cannot explain individual systems. The variety of observational techniques will lead to improved constraints. These observations include volume and durations of volcanism, the nature and depth extent of mantle seismic anomalies, and presence or absence of four key characteristics: swells, age progressions, connections with LIPS, and geochemical distinctions from MORB. We will close this chapter with a short outlook in the form of a wish list for future work.

1. Origin of melting anomalies: We need to explore different mechanisms of mantle flow and melting that include increasingly realistic dynamics (non-Newtonian, time-dependent, and 3-D) and lithologic variability, and we need to test model predictions against observed volumes and durations of volcanism. We need to do this to understand the relative importance of temperature and composition, which are both coupled to the upper-mantle dynamics. Example processes that are relatively less well understood include sublithospheric convection, fertile mantle melting, viscous fingering in the asthenosphere, as well as the role of the lithosphere in controlling magmatism on the surface.

2. Origin of swells: Previous work has shown that plumes can explain many observations of swells but future work is needed to explore whether nonplume mechanisms can cause swells, and in particularly, how they vary with plate speed and buoyancy flux. We are also faced by explaining the presence of melting anomalies without swells, such as the prominent Canaries and Madeira hot spots. Perhaps such systems are dominated by fertile mantle melting.

3. Age progressions and lack thereof: For long-lived age progressions, future challenges involve reducing observational uncertainties with further geophysical studies, more accurate and precise dating, and improving geodynamic models of mantle flow and evolution. The latter will require improved methods of tracking mantle flow further into the past, and in defining the ranges of allowable mantle density and viscosity structures from seismology and mineral physics. For hot spots with short-lived age progressions models will need to consider the possibility of them originating from upwellings from boundary layers above the CMB or nonplume sources. For hot spots without simple age progressions, plume may be unlikely and thus other mechanisms should be explored. In fact some of these mechanisms (e.g., propagating fracture) may predict age progressions and quantitative models are needed to explore these possibilities.

4. LIPs are the most dramatic but potentially the least understood dynamical processes on the planet. It is critical to evaluate how the evolving plume theory can self-consistently address the formation of some LIPs and to which cases alternatives are required.
5. Geochemistry: Differences in isotope geochemistry and in particular the distinctions from MORB of most OIBs require a chemically layered mantle, differences in melting of a nonlayered, heterogeneous mantle, or some combination of above.

6. Seismology: The key challenge is to confidently resolve if any hot spots have seismic anomalies extending into the lower mantle and if so, which ones do and do not. Such information will be critical for evaluating plume versus nonplume mechanisms. Combined with geochemical observations, such information could be the key in addressing the possibility or nature of geochemical layering.

7. Integrated and interdisciplinary work: We need to meet our capabilities of simulating increasingly complex dynamic behaviors with increasing quality of geophysical and geochemical data.

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