

Chapter 2

Plumes and Hotspots

2.1 Introduction

The plate tectonic processes adequately explain two principal types of basaltic volcanism on the earth's surface, the mid-oceanic ridge and island arc volcanism. Apart from these, another important type of basaltic volcanism occurs away from the plate margins. The most volumetric forms of this type are the continental flood basalts, oceanic plateaus and aseismic ridges, collectively refereed as large igneous provinces. These are explained by the plume hypothesis, initially put forward by Wilson (1963) and Morgan (1971, 1972) to explain the age-progressive chains of volcanic islands like Hawaii that stretch across the ocean basins. The plume hypothesis attributes flood basalts and giant oceanic plateaus to the melting of the head of a plume (Richards et al. 1989; Campbell and Griffiths 1990) and aseismic ridges to the melting of a plume tail (Wilson 1963; Morgan 1971).

A mantle plume is a buoyant upwelling of abnormally hot rock within the Earth's mantle. As the plume rises to shallow depths below the lithosphere, the head may melt partially to give rise to volcanic provinces. Hotspots are observed surface features on the crust characterized by excess volcanisms that are thought to be fed by the underlying anomalously hot mantle.

Mantle plumes are interpreted to be the cause of many or most of the hotspots. However, in recent literature the terms are often used interchangeably. Thus, Anderson and Natland (2005) wrote *“Unfortunately, the terms hotspot and plume have become confused. In recent literature the terms are used interchangeably. A plume is a hypothetical mantle feature. A hotspot is a region of magmatism or elevation that has been deemed to be anomalous in some respect because of its volume or location. In the plume hypothesis, a hotspot is the surface manifestation of a plume, but the concepts are different; one is the presumed effect, and the other is the cause.”*

2.2 The Plume Hypothesis

Fluid flow from thermal boundary layers occurs due to buoyancy, as a direct result of density differences due to temperature between materials on either side of the layer. The earth interior has two such thermal boundary layers—the crust-mantle boundary, where heat is transferred by conduction, and the lower mantle-outer core boundary, where heat conduction takes place by convecting plumes of hot material rising from the liquid outer core.

Laboratory experiments (e.g., Griffiths and Campbell 1990) suggest that Rayleigh-Taylor instability occurring in the D'' layer in outer core leads to the formation of upwellings that are focused into narrow conduits, which rise through the mantle, driven by their thermal buoyancy. These plume materials are hotter than the surrounding mantle by about 100–300 °C (Campbell and Davies 2006), indicating two orders of lower viscosity, and hence are lighter. However, to gain

Fig. 2.1 Photograph of a laboratory model of a starting thermal plume showing the structure caused by heat conduction and consequent entrainment of surrounding fluid (Campbell and Griffiths 1990). The dark area is hot material from the plume source and the lighter material is cooler entrained material

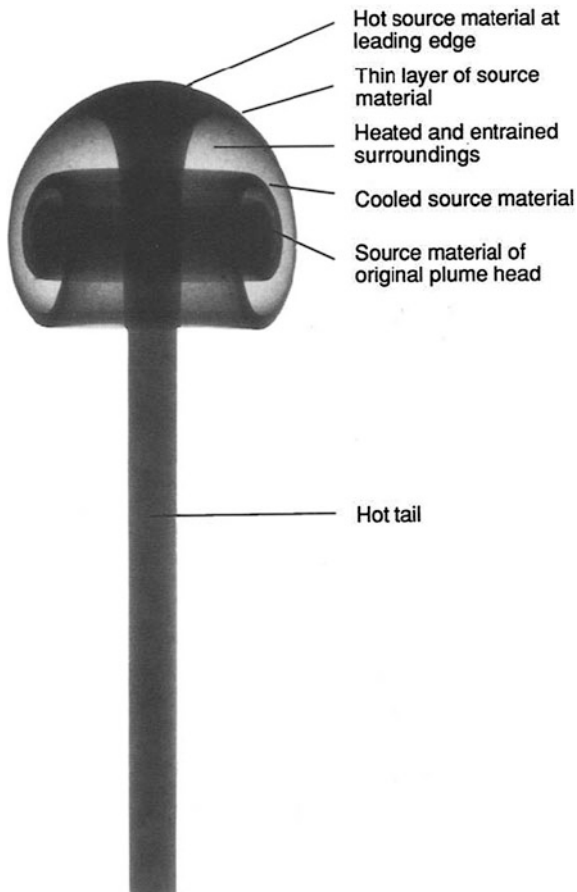
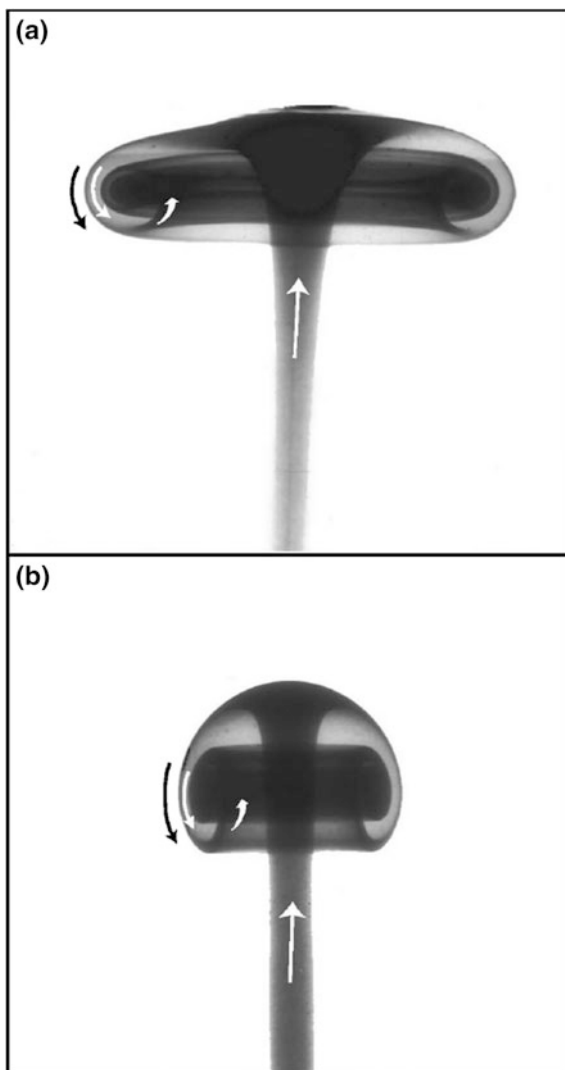


Fig. 2.2 Photograph of a laboratory model of a starting thermal plume (a) *mid-way* through its ascent and (b) at the *top* of its ascent (Campbell and Davies 2006). The *white arrows* show motion within the plume and the *black arrows* the direction of motion in the boundary layer adjacent to the plume



sufficient buoyancy, a large amount of material needs to accumulate, connected by a relatively narrow tail or feeder conduit, which connects the plume head with the reservoir. The tail is hotter and less viscous than the head, which cools by heat exchange with the surrounding, entraining partial melt from the mantle during its ascent (Figs. 2.1 and 2.2). As the plume rises through the mantle, the head grows by entrainment and a constant flux of materials by the tail (Griffith and Campbell 1990), which ascents faster. When a stagnation point is reached, material flows radially with a spiraling motion, giving the head a characteristic mushroom shape.

Numerical and analogue modelings suggest that the plume shape and mobility is controlled by the magnitude of the viscosity contrast with the surrounding mantle (Kellogg and King 1997; Lowman et al. 2004; Lin and van Keken 2006). As the contrast increases, the plume conduit becomes narrower and the head becomes broader, mushroom shaped, allowing the plume to rise more efficiently through the mantle. Nevertheless, a great variety of shapes are also possible if the density contrasts due to chemical variations are taken into account (Farnetani and Samuel 2005; Lin and van Keken 2006).

2.3 Characteristics of Plumes

The plumes can be characterized as follows (Campbell 2006, 2007; Campbell and Davies 2006):

1. New plumes consist of a large head followed by a narrow tail. Laboratory experiments on thermal plumes (e.g., Griffiths and Campbell 1990) suggest that the shape is controlled by the viscosity contrast between the plume material and the surrounding mantle. Flood basalts/oceanic plateaus and hotspots tracks are considered the eruptive products from the plume head and tail, respectively.
2. Plumes originate from a thermal boundary layer, probably the core-mantle boundary. Most of the plumes are supposed to originate from the D'' layer in the lower mantle. However, using seismic methods to trace plume tails to their origin is difficult because of their small diameter, and the method had met with little success.
3. Flattened plume heads should be 2000–2500 km in diameter at the base of the lithosphere, while the narrow tails have a diameter of 100–200 km in the upper mantle. The plume-head diameter depends on the thermal anomaly of the plume with the adjacent mantle, its buoyancy flux, kinematic viscosity of the lower mantle, and the height of ascent from the origin, with the last one being the most important factor.
4. Plume heads stalled at the base of the lithosphere should be seismically detectable for ca. 100 my, and remain hot for about 300 my. It depends on the rate of heat conduction by the lithosphere, which itself is a slow process. This explains the low heatflow anomaly near some hotspots, as in Hawaii. As the plume loses heat, it cools down, and depending upon its thickness, the thermal anomaly disappears within 300 my, as predicted in the case of the Ontong-Java plateau.
5. The temperature excess of a plume head is predicted to be $\sim 300^\circ\text{C}$ at the centre of the head and to taper to $\sim 100^\circ\text{C}$ towards its margins. These temperature anomalies are predicted based on anomalous crustal thicknesses produced by the plume (McKenzie and Bickle 1988). The highest temperature in the centre is due to the rising tail connecting the head with the mantle reservoir. At the margins, the plume entrains mantle material and becomes cooler.

6. Plume heads can erupt millions of cubic kilometres of magma. The melting rate of a mantle plume depends on the depth of emplacement of the hot plume head at the base of the lithosphere, and the presence of recycled oceanic crusts in the form of eclogite in the plume source. Modelling indicates a melting rate of 1–10 km³/a, and a total melt volume of 1–20 million km³, depending on the assumed conditions (Leitch and Davies 2001). These results cover the range of observed values for most of the large igneous provinces.
7. Rifting over a plume head produces a narrow zone of thickened oceanic crust, typically 100–200 km wide. As long as lithospheric extension pulls up the hot plume material towards the rift, anomalous melting and crustal thickening continues. The 150–200 km thick plume head is completely pulled up after only 150–200 km of extension, limiting the anomalous crust to a narrow zone at the centre of the newly formed ocean basin.
8. Both plume heads and tails erupt high temperature picrites. The temperature of erupted magmas are estimated from their MgO content, which shows a linear relationship of 4 wt% increase in MgO with 100 °C rise in magma temperature. The maximum MgO content of plume-derived picrate from Reunion-Deccan, Iceland-North Atlantic province, Hawaii, Siberian Traps, Emeishan, etc. ranges from 18 to 22 wt%, indicating a thermal anomaly of 150–250 °C for both the plume head and the tail.
9. Picrites erupt early during flood volcanism and are most abundant near the centre of the plume head and less abundant towards its margins. This occurs because only the hottest materials in the centre of the plume head ascends to shallow levels, and undergo adiabatic decompression melting to produce picritic lava, which decreases towards the periphery. These picritic flows are often covered by later lava, and are rarely seen in flood basalt provinces. However, they have been documented for the Parana-Etendeka, Deccan, Emeishan, North Atlantic, Siberian Traps and Karoo.
10. Flood basalt volcanism is preceded by a domal uplift of 500–1000 m due to arrival of the plume head in the upper mantle. The uplift is caused by the impact of the plume head, buoyancy due to ponding of the plume materials at the base of the lithosphere, thermal and magmatic underplating. The best documented example of domal uplift occurs in the Emeishan flood basalt province in south-west China (He et al. 2003), where the timing of the uplift is documented to begin 3–5 my before volcanism.
11. The position of hot-spots produced over plumes is fixed with respect to the mantle. The lower mantle has an average viscosity about 30 times higher than that of the upper mantle, indicating a ten times slower convective velocity. This slow convective motion fixes the lower mantle plumes within the upper mantle, so that they move ten times slower than the plates. This makes the plume effectively stationary with respect to the lithospheric plates, so that the hotspot track follows the direction of the plate motions.

2.4 Classification of Plumes

Increasing research focused on the evolution of hotspot magmatism led to the development of a number of ideas for their genesis, apart from variations in the mantle plume hypothesis itself. Turcotte and Oxburgh (1973) proposed tensional cracking of the lithosphere and subsequent volcanism to explain the age progressive chains of volcanism. Morgan (1978) had proposed a different type of hotspot island apart from those produced by plumes. More recent studies (Sleep 1990, 2002; Clouard and Bonneville 2001) have debated with the ideas of primary versus secondary hotspots, as discussed later. Anderson (1998, 2000) have rejected the plume hypothesis altogether, explaining the presence of mid-plate volcanism by plate movement-induced stresses fracturing the lithosphere and edge-driven convection in the upper mantle by localized thermal perturbations. Similar views have also been mentioned by Ballmer et al. (2007). The presence of such diverse views led Courlliot et al. (2003) to attempt a classification of the surface hotspots in terms of their originating plumes, based on criteria that look for deep-seated mantle sources:

1. the presence of a linear chain of volcanoes with continuous age progression, indicating a long-lived plume source;
2. a flood basalt at the origin of the track, corresponding to the impingement of the plume head below the lithosphere;
3. a buoyancy flux of greater than 10^3 kg s^{-1} , for generating a topographic anomaly
4. a consistently high $^4\text{He}/^3\text{He}$ or $^{21}\text{Ne}/^{22}\text{Ne}$ ratio, indicating a deep-seated, primitive mantle reservoir; and
5. a significantly low shear wave velocity in the underlying mantle, indicating the presence of hot, partially melted material below the hotspots.

Applying these criteria to a selection of hotspots (Table 2.1) that were active in the last one million years (Davies 1988; Sleep 1990; Steinberger 2000) led to a threefold classification of their possible sources (Courlliot et al. 2003; Fig. 2.3):

1. Seven hotspots, including Hawaii, Easter, Louisville, Iceland, Afar, Reunion and Tristan, satisfy all five criteria. They were possibly originated by plumes from the lower mantle, due to chemical heterogeneity and instability in the D'' layer (Olson et al. 1987; Bercovici and Kelly 1997), termed “Morganian” by Courlliot et al. (2003). They can also be called as primary plumes.
2. About twenty hotspots may be associated with plumes from the bottom of the transition zone, at the top of large domes that correspond to the superswells, also known as superplumes (Courlliot et al. 2003). Examples include Caroline, McDonald, Pitcairn, Samoa and Tahiti. These can be called as secondary plumes.
3. The remainder of about twenty hotspots could originate from thermal perturbations in the upper mantle. They are called as “Andersonian” by Courlliot et al. (2003), also known as tertiary plumes. These hotspots may be linked to

Table 2.1 Main characteristics of selected hotspots (Courtillot et al. 2003)

Scores for 49 hotspots with respect to five criteria used to diagnose a potentially deep origin										
Hotspot	Lat	Lon (°E)	Track	Flood/plateau	Age (Ma)	Buoy.	Reliab.	³ He/ ⁴ He	Tomo (500)	Count
<i>Afar</i>	<i>10N</i>	<i>43</i>	<i>No</i>	<i>Ethiopia</i>	<i>30</i>	<i>1</i>	<i>Good</i>	<i>High</i>	<i>Slow</i>	<i>4</i>
Ascension	8S	346	No	No	–	NA	NA	NA	0	0+?
Australia E	38S	143	Yes	No	–	0.9	Fair	NA	0	1+?
Azores	39N	332	No?	No	–	1.1	Fair	High?	0	1+?
Baja/Guadalupe	27N	247	Yes?	No	–	0.3	Poor	Low	0	0+?
Balleny	67S	163	No	No	–	NA	NA	NA	0	0+?
Bermuda	33N	293	No	No?	–	1.1	Good	NA	0	0+?
Bouvet	54S	2	No	No	–	0.4	Fair	High	0	1+?
Bowie	53N	225	Yes	No	–	0.3	Poor	NA	slow	2+?
Cameroon	4N	9	Yes?	No	–	NA	NA	NA	0	0+?
Canary	28N	340	No	No	–	1	Fair	Low	Slow	2
Cape Verde	14N	340	No	No	–	1.6	Poor	High	0	2
<i>Caroline</i>	<i>5N</i>	<i>164</i>	<i>Yes</i>	<i>No</i>	<i>–</i>	<i>2</i>	<i>Poor</i>	<i>High</i>	<i>0</i>	<i>3</i>
Comores	12S	43	No	No	–	NA	NA	NA	0	0+?
Crozet/Pr. Edward	45S	50	Yes?	Karoo?	183	0.5	Good	NA	0	0+?
Darfur	13N	24	Yes?	No	–	NA	Poor	NA	0	0+?
Discovery	42S	0	No?	No	–	0.5	Poor	High	0	1+?
Easter	27S	250	Yes	<i>mid-Pac mnt?</i>	<i>100?</i>	<i>3</i>	<i>Fair</i>	<i>High</i>	<i>Slow</i>	<i>4+?</i>
Eifel	50N	7	Yes?	No	–	NA	NA	NA	0	0+?
Fernando	4S	328	Yes?	CAMP?	201?	0.5	Poor	NA	0	0+?
Galapagos	0	268	Yes?	Caribbean?	90	1	Fair	High	0	2+?
Great Meteor/New England	28N	328	Yes?	No?	–	0.5	Poor	NA	0	0+?

(continued)

Table 2.1 (continued)

Scores for 49 hotspots with respect to five criteria used to diagnose a potentially deep origin										
Hotspot	Lat	Lon (°E)	Track	Flood/plateau	Age (Ma)	Buoy.	Reliab.	³ He/ ⁴ He	Tomo (500)	Count
<i>Hawaii</i>	20N	204	Yes	<i>Subducted?</i>	>80?	8.7	Good	High	Slow	4+?
Hoggar	23N	6	No	No	–	0.9	Poor	NA	Slow	1
<i>Iceland</i>	65N	340	Yes?	<i>Greenland</i>	61	1.4	Good	High	Slow	4+?
Jan Mayen	71N	352	No?	Yes?	–	NA	Poor	NA	Slow	1+?
Juan de Fuca/Cobb	46N	230	Yes	No	–	0.3	Fair	NA	Slow	2+?
Juan Fernandez	34S	277	Yes?	No	–	1.6	Poor	High	0	2+?
Kerguelen (Herald)	49S	69	Yes	Rajmahal?	118	0.5	Poor	high	0	2+?
<i>Louisville</i>	51S	219	Yes	<i>Ontong-Java</i>	122	0.9	Poor	NA	Slow	3+?
Lord Howe (Tasman East)	33S	159	Yes?	No	–	0.9	Poor	NA	Slow	1+?
Macdonald (Cook-Austral)	30S	220	Yes?	Yes?	–	3.3	Fair	High?	Slow	2+?
Marion	47S	38	Yes	Madagascar?	88	NA	NA	NA	0	1+?
Marqueses	10S	222	Yes	Shatski?	???	3.3	NA	Low	0	2+?
Martin/Trinidad	20S	331	Yes?	No	–	0.5	Poor	NA	Fast	0+?
Meteor	52S	1	Yes?	No	–	0.5	Poor	NA	0	0+?
Pitcairn	26S	230	Yes	No	–	3.3	Fair	High?	0	2+?
Raton	37N	256	Yes?	No	–	NA	NA	NA	Slow	1+?
<i>Reunion</i>	21S	56	Yes	<i>Deccan</i>	65	1.9	Poor	High	0	4
St. Helena	17S	340	Yes	No	–	0.5	Poor	Low	0	1
<i>Samoa</i>	14S	190	Yes	<i>No?</i>	14?	1.6	Poor	High	Slow	4
San Felix	26S	280	Yes?	No	–	1.6	Poor	NA	0	1+?
Socorro	19N	249	No	No	–	NA	Poor	NA	Slow	1+?
Tahiti/Society	18S	210	Yes	No	–	3.3	Fair	High?	0	2+?

(continued)

Table 2.1 (continued)

Scores for 49 hotspots with respect to five criteria used to diagnose a potentially deep origin										
Hotspot	Lat	Lon (°E)	Track	Flood/plateau	Age (Ma)	Buoy.	Reliab.	³ He/ ⁴ He	Tomo (500)	Count
Tasmanid (Tasman Central)	39S	156	Yes	No	–	0.9	Poor	NA	Slow	2
Tibesti	21N	17	Yes?	No	–	NA	Poor	NA	0	0+?
<i>Tristan</i>	37S	348	Yes	<i>Parana</i>	<i>I33</i>	<i>1.7</i>	<i>Poor</i>	<i>Low</i>	<i>0</i>	3
Vema	33S	4	Yes?	Yes? (Orange R.)	–	NA	Poor	NA	0	0+?
Yellowstone	44N	249	Yes?	Columbia?	16	1.5	Fair	High	0	2+?

Columns from left to right are: (1) hotspot name; (2, 3) hotspot latitude and longitude; (4) existence of a linear track or chain of dated seamounts extending from the presently active hotspot site; (5, 6) existence and age of a trap (or flood basalt, oceanic plateau); (7, 8) buoyancy flux (in 10³ kg s^{−1}) and its reliability; (9) existence of consistently high ³He/⁴He ratios for the hotspot; (10) existence of a slow shear velocity (V_s) anomaly at 500 km depth below the hotspot surface trace, based on the tomographic model of Ritsema et al. (1999); (11) count of positive responses to the five characteristics listed in the text (columns 4, 5, 7, 9 and 10). Hotspots with a total count of at least two (out of five) are shown in bold type, those with a count of at least three are in bold italics

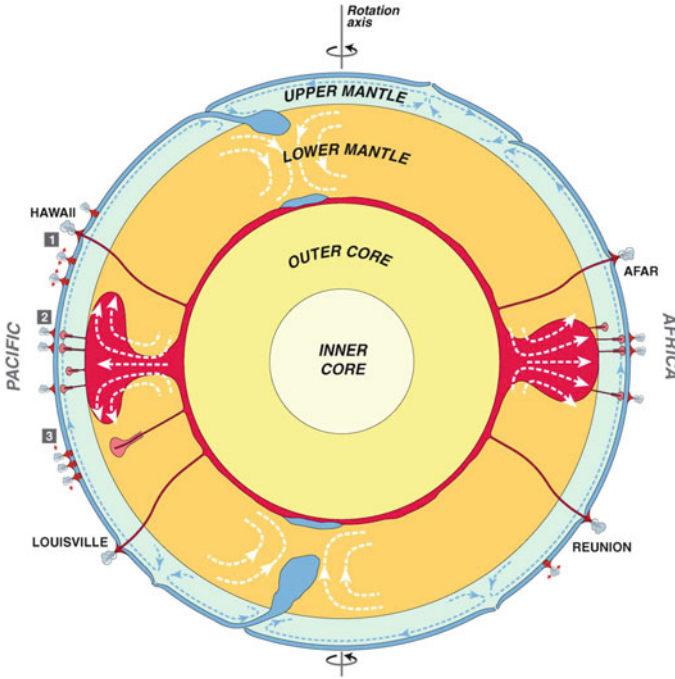


Fig. 2.3 A schematic cross-section of the dynamic Earth, outlining the sources of the three types of plumes/hotspots identified in Courliot et al. (2003): 1—the primary, deeper plumes possibly coming from the lowermost mantle boundary layer (D'' in the broad sense), 2—the secondary plumes possibly coming from the transition zone at the locations of the superswells, while 3—the tertiary hotspots may have a superficial origin, linked to tensile stresses in the lithosphere and decompression melting (Courliot et al. 2003)

perturbations in the asthenosphere and/or be a passive response to forms of lithospheric breakup. They are the subject matter of current research (e.g., Anderson 1998, 2000).

2.4.1 Superplumes

The geographic distribution of the world's major hotspots falls on two hemispheres antipodal to each other, and corresponds to crustal uplifts in East Africa and Central Pacific ocean (Courliot et al. 2003). The same pattern is visible in seismic tomography images of shear wave velocity extending from the transition zone to the core-mantle boundary (Masters et al. 1982, 1996; Dziewonski and Woodhouse 1987; Su et al. 1994; Li and Romanowicz 1996; Ritesma et al. 1999; Gu et al. 2001) (Fig. 2.4). They are associated with superplumes, whose origin appears to be

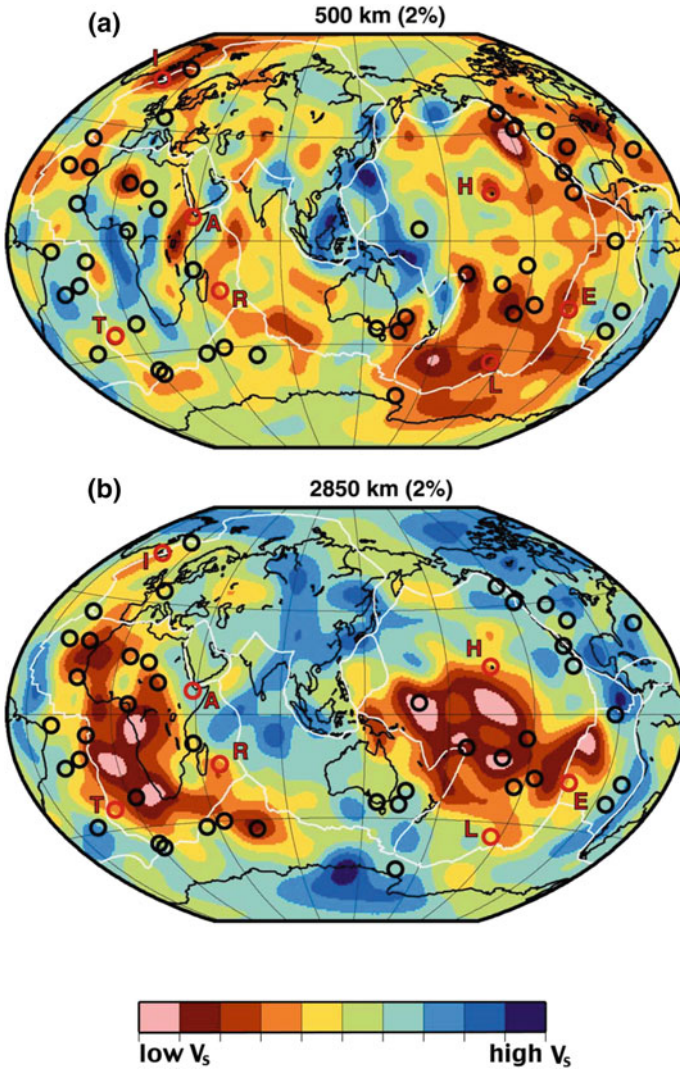


Fig. 2.4 Distribution of the 49 hotspots (black circles) from Table 2.1 superimposed on a section at (a) 500 km and (b) 2850 km depths through Ritsema et al.'s (1999) tomographic model for shear wave velocity (V_s) (Courillot et al. 2003). The primary hotspots are shown as red circles with the *first letter* of their name indicated for reference

related directly to the mantle circulation. The present day convection in the lower mantle appears to be quadrupolar (Busse 1983; see Fig. 2.3), where cold, dense material subducts and sinks in the mantle, while hot, less dense, partially melted material ascends slowly below two antipodal positions below Africa and Central Pacific, forming the superplumes, with many hotspots situated over them (Chase

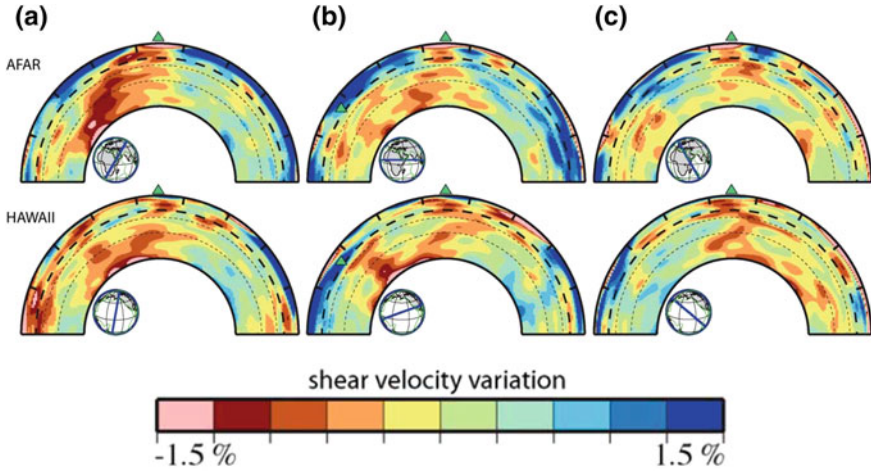


Fig. 2.5 Shear velocity anomalies derived from model S20RTS in 180° wide cross-sections through the mantle. The *triangles* indicate the location of the Afar and Hawaii hotspots. The *red* regions have lower shear velocity than the global average at the same depth, while the *blue* regions have higher velocity than the global average, ranging between $\pm 1.5\%$. The 670-km discontinuity is marked by the *thick dashed line*, while the *thin dashed lines* indicate 1000 and 1700 km depth (modified from Ritsema and Allen 2003)

1979). These massive mantle upwellings, when studied in detail, appear to be of complex structure and compositionally heterogeneous (Breger and Romanowicz 1998). The shape of the African superplume at its base is complex, and it does not ascend vertically into the mantle (Fig. 2.5). A lowering of the shear wave velocity between 670 and 1000 km indicates an obstruction in the upwelling from the core-mantle boundary (Ritesma et al. 1999). However, its sharp margins suggest a stable structure in the deep mantle, and compositional heterogeneity with the surrounding mantle (Ritesma and Allen 2003).

2.4.1.1 Origin of Primary Plumes

Some deep-seated plumes can be traced up to the transition zone in the upper mantle (Morganian plumes of Courlliot et al. 2003). They must, therefore, generate from instabilities in a thermal boundary layer, either at the transition zone itself or at the core-mantle boundary. Seismic tomography is not yet able to resolve plume tails in the lower mantle, but certain indirect evidences point to their origin in the lower mantle, at the D'' layer. These plumes have large flood basalt provinces associated with them, that requires melting of more than 10^8 km^3 of mantle material (Richards et al. 1989; Campbell and Griffiths 1990).

Laboratory fluid mechanic experiments suggest that it is easier to produce thermal instabilities of such large magnitude in the core-mantle boundary than at the

transition zone (Whitehead and Luther 1975; Stacey and Loper 1983; Olson et al. 1987; Bercovivi and Kelly 1997). Many of these flood basalt provinces are enriched in Fe–Si with respect to Mg (Courlliot et al. 2003), which again favors a depleted lower mantle origin for the originating plumes (Javoy 1999). The hotspot tracks associated with these plumes tend to last long, up to 130 my. For example, hotspots like Louisville and Tristan are becoming inactive, while the younger ones, like Reunion and Iceland, are still active. The observation is supported by fluid mechanic experiments, which indicate that plumes with a large head and a long enduring tail can only generate at lower mantle depths (Whitehead and Luther 1975; Stacey and Loper 1983; Olson et al. 1987; Bercovivi and Kelly 1997). Recent experiment by Farnetani and Samuel (2005), however, indicates that deep-seated plumes without a distinct head-tail structure and laterally heterogeneous tail can also occur due to shearing in the mantle.

It seems probable that the lower mantle generates two different modes of upwelling, superplumes and primary plumes, both of which develop from buoyancy anomalies. Thermo-chemical convection in a heterogeneous mantle is able to produce both these modes, depending on the local buoyancy ratio, representing a ratio of chemical density anomaly to a ratio of thermal density anomaly. For a large amount of thermal buoyancy anomaly, producing a low buoyancy ratio, the superplumes are generated. For a high chemical density anomaly, producing a high buoyancy ratio, long lived thermo-chemical primary plumes are developed (Davaille 1999, Davaille et al. 2002). Density anomalies of chemical origin in the mantle, as inferred from mineral physics and seismic studies, would be sufficient to produce both these modes depending on local conditions (Bina 1998).

2.4.2 *Secondary and Tertiary Plumes*

Secondary plumes are those generated in the upper mantle, either from a primary plume/superplume or from thermal perturbations in the transition zone. The 670 km discontinuity in the mantle transition zone presents a suitable stagnation zone for plumes starting from the D'' layer. Here a separation takes place between heavier and lighter fractions in a plume containing a high proportion of denser components, such as subducted former basaltic crust (Campbell and Davies 2006). This leads to the formation of a new thermo-chemical plume head starting from the transition zone (Farnetani and Samuel 2005). Secondary thermal plumes are also generated from the thermal boundary layer between a superplume and the rest of the mantle, when the superplume get stalled at the transition zone, as seems to be the case under Polynesia (Courlliot et al. 2003). Such secondary plumes produce short-lived tracks without any flood basalts, such as Tahilti and Pitcarin. However, their motions are consistent with the primary plumes from which they are derived. Another possible origin of plumes from the 670 km discontinuity in the transition zone is by local thermal instability (White and Mckenzie 1989). Such plumes are likely to be

composed of heated upper mantle material with an entrained lower mantle axial zone (Fitton et al. 1997).

Anderson (1998, 2000) had suggested that long wavelength temperature variations of the asthenosphere depart from the mean by ± 200 °C. This ‘normal’ temperature variation at the top of the convecting mantle due to slab cooling, cratonic roots, continental insulations and edge-driven convection is sufficient to induce pseudo-plume-like upwellings leading to excess intra-plate magmatism. The process can be augmented by shears driven by moving plates and the rise of the asthenosphere between cratons. Ballmer et al. (2007) have pointed out that small-scale sub-lithospheric convection can give rise to chains of volcanic islands, which can lack a definite age progression, without requiring the presence of any plume underneath.

2.5 Plume-Related Lithospheric Uplift

Impingement of a hot, ascending mantle plume below the lithosphere results in the formation of a broad swell, 2000–2500 km in diameter, above the flattened plume head (e.g., Griffiths and Campbell 1991; He et al. 2003; Fig. 2.6). The different mechanisms generating the lithospheric uplift can be summarized as:

1. Dynamic uplift is generated when abnormally hot mantle is emplaced beneath the lithosphere. The support mechanism is a combination of impact of the plume, buoyancy due to ponding of the plume material below the lithosphere (Guillou-Frottier et al. 2007) and thermo-mechanical erosion at the base of the lithosphere (Davies 1994; d’Acremont et al. 2003). It is indicated by a positive

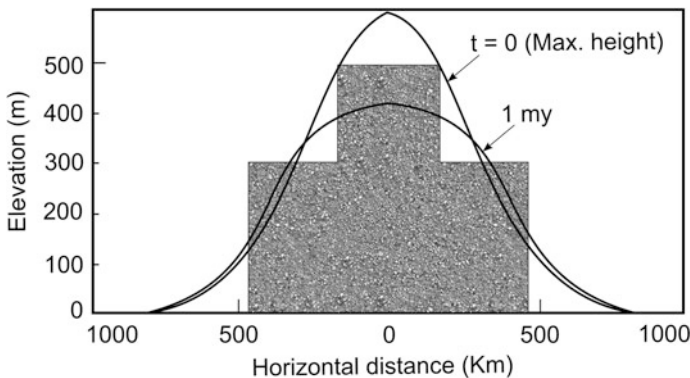


Fig. 2.6 Uplift above a plume head, as predicted by Griffiths and Campbell (1991), compared with the uplift observed at the center of the Emeishan flood basalt by He et al. (2003) (modified from Campbell and Davies 2006). Predicted profiles are given for maximum uplift ($t = 0$) when the top plume is at a depth of ~ 250 km, and 1 my later ($t = 1$ my) when flattening of the head is essentially complete

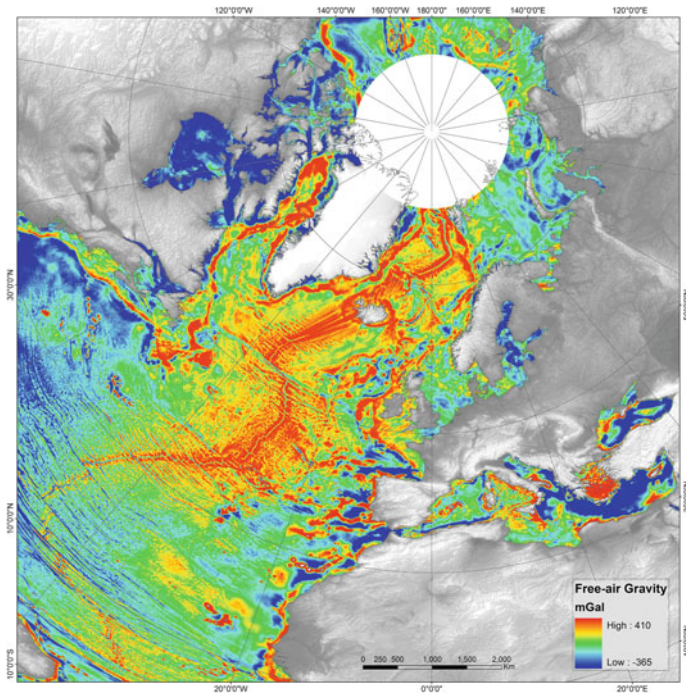


Fig. 2.7 Free-air gravity anomaly over part of the northern hemisphere, displayed using a Polar Lambert azimuthal equal-area projection. Note the long-wavelength gravity high centred on Iceland, and extending from south of the Azores to Spitzbergen and from Baffin Island to Scandinavia. Gravity anomaly data from Sandwell and Smith (2009)

anomaly of wavelength more than 1000 km long in the free-air gravity data (Sclater et al. 1975; McKenzie 1994) (Fig. 2.7). The uplift is a transient phenomenon, which disappears when the thermal anomaly dissipates by conduction and convection (Nadin et al. 1995, 1997; Jones et al. 2002). The Iceland plume in the North Atlantic region serves as a typical example.

2. Permanent uplift occurs due to crustal thickening when plume-related thermal anomalies induce melting and the melt is injected into or just beneath the crust. The ensuing magmatic underplating results in isostatic uplift of the crust (Brodie and White 1994, 1995; Clift 1997; Clift and Turner 1998; White and Lovell 1997). For example, permanent uplift affected Britain and Ireland during Paleocene time, and the regions flanking the line of continental separation at the Paleocene- Eocene boundary (Jones et al. 2002).
3. Isostatic thermal uplift occurs due to heating and decrease of density of the lithosphere by the ascending hot plume material. It is a transient phenomenon, adding to the dynamic topography, which dissipates with the disappearance of the plume head (Sleep 1990; Clift and Turner 1998).

2.6 Morphology of Hotspot Tracks

The features like morphology and volume of eruption of hotspot tracks (Figs. 2.8 and 2.9) strongly depend on the nature the affected lithosphere, particularly its age during the time of eruption. Hotspot ridges fall into two broad categories (Honsho and Tamaki 1995; Fig. 2.10):

1. volcanic ridges formed in intra-plate setting tend to form discrete seamounts or guyots, which originate in a thick plate, regionally supported by lithospheric flexure, and
2. volcanic ridges formed in young lithosphere near sea-floor spreading centers tend to form continuous ridges, locally supported by the Airy isostasy of low density roots.

Example of the former is provided by the Hawaii-Emperor Chain, while the 90°E Ridge provides the second example.

The Hawaii-Emperor Chain extends for about 6000 km across the North Pacific, with about 107 individual volcanoes (Clague and Dalrymple 1987; Fig. 2.11a) being constructed in the last 70 my by the Hawaiian hotspot. The bathymetric profile

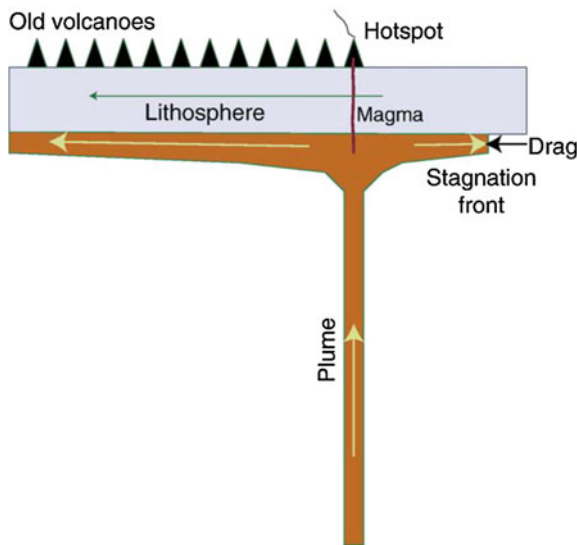


Fig. 2.8 Schematic diagram of a hotspot track produced by plume beneath fast-moving plate (Sleep 2006). A chain of volcanoes forms when the plate passes over the plume. Lateral flow of buoyant plume material complicates the flow pattern beneath the base of the lithosphere. The plate drags the plume material along with it. A stagnation front exists upstream of the hotspot

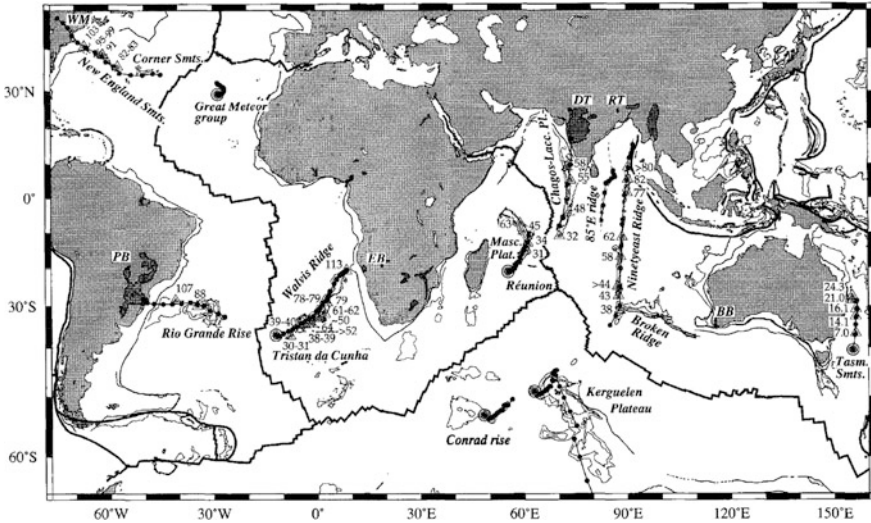


Fig. 2.9 Major hotspot tracks in Atlantic and Indian oceans (Muller et al. 1993). Large shaded circles are locations of present-day hotspots. WM—younger White Mountains, DT—Deccan Traps, RT—Rajmahal Traps, PB—Paraná flood basalts, EB—Etendeka flood basalts, BB—Bunbury basalts. Triangles with numbers indicate radiometric ages of hotspot tracks

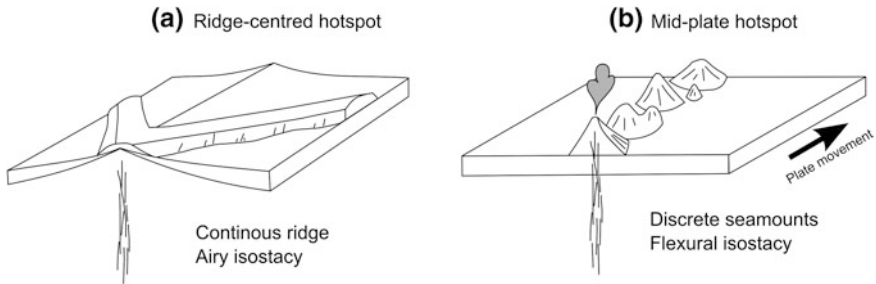


Fig. 2.10 Diagram showing morphological differences between two broad categories of hotspot tracks, (a) hotspots formed in young lithosphere near spreading centers, and (b) hotspots in intra-plate setting (modified from Honsho and Tamaki 1995)

across the chain shows a broad, symmetrical depression, the Hawaiian Deep, surrounding the ridge, and the broad Hawaiian Arch surrounding the Deep. The isostatic equilibrium can be explained by the elastic plate model, with a best fitting elastic thickness of 20–30 km for the Hawaiian and about 20 km for the Emperor Chain. The age of the crust during eruption is between 80 and 90 my for the Hawaiian chain, and then decreases from about 80 my at the bend to 20 my towards the northern end (Honsho and Tamaki 1995).

The 90°E Ridge has resulted from the volcanism of the Kerguelen hotspot (Royer et al. 1991) and extends for about 4000 km along the 90°E meridian

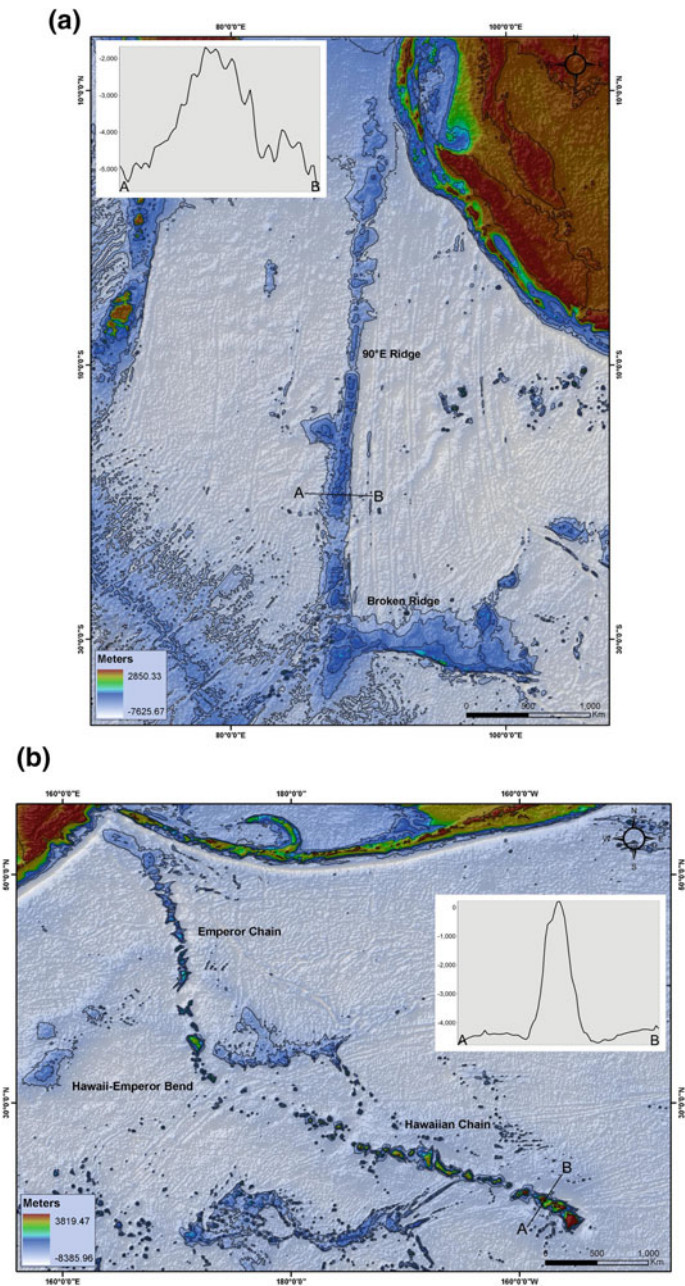


Fig. 2.11 Bathymetry for (a) Hawaiian-Emperor seamount chain, and (b) Ninetyeast ridge, contoured at 500 m intervals (redrawn after Honsho and Tamaki 1995). Lines with annotations represent location of profiles shown in inset

(Fig. 2.11b). It has the continuous shape of an elongated plateau without any break. The profile across it shows no depressions around the ridge, except those associated with fracture zones to the east. The ridge has a steep eastern flank indicative of block faulting (Detrick and Watts 1979). Its isostatic adjustments are explained by the Airy model, with a best fitting elastic thickness of 15–25 km (Honsho and Tamaki 1995). Drilling results from DSDP legs 21 and 26, and ODP leg 121, show a close correspondence between the age of the ridge and the surrounding oceanic crust, indicating its eruption in a young lithosphere near a spreading centre.

2.7 Moat and Arch

The impact of a mantle plume beneath the lithosphere leads to a broad swell around the erupting centers, within which voluminous magma eruption takes place. The load of the volcanoes bends the underlying elastic lithosphere, creating a trough-like depression around the volcano called flexural moat. Away from the moat, isostatic upwelling causes the formation of a peripheral bulge, also termed as flexural arch (Rees et al. 1993). Examples come from Hawaii, Marquesas and Canary Islands (Figs. 2.12 and 2.13).

The best studied example comes from the Hawaiian island, where lithospheric flexure induced by the weight of the volcanoes creates the Hawaiian moat, called the Hawaiian Deep, surrounding the island chain. The moat is bounded by a sea-floor rise known as the Hawaiian Arch, located about 250 km seaward of the island chain (Hamilton 1957; Menard 1964). Both the flexural moat and bulge are superimposed on a bathymetric high of larger wavelength of about 1200 km, known as the Hawaiian Swell (Detrick and Crough 1978). The moat and the bulge are divided into flanking segments in the north, south and front, known respectively as the North Hawaiian Moat/Arch, South Hawaiian Moat/Arch, and Frontal Hawaiian Moat/Arch. Moat to bulge relief in the flanking segments exceeds 1 km compared to about 200–450 m in the frontal region. This indicates less subsidence of the oceanic crust in the front (Leslie et al. 2002). The frontal segments are short-lived in comparison to the flanking segments due to being devoured by the growing Hawaiian ridge as the Pacific plate migrates across the Hawaiian hotspot. Widespread mass-wasting derived sediments fill up the moats that get eventually covered by subsequent volcanic rocks of the Hawaiian ridge (Moore and Clague 1992; Rees et al. 1993; Smith et al. 1999; Naka et al. 2002).

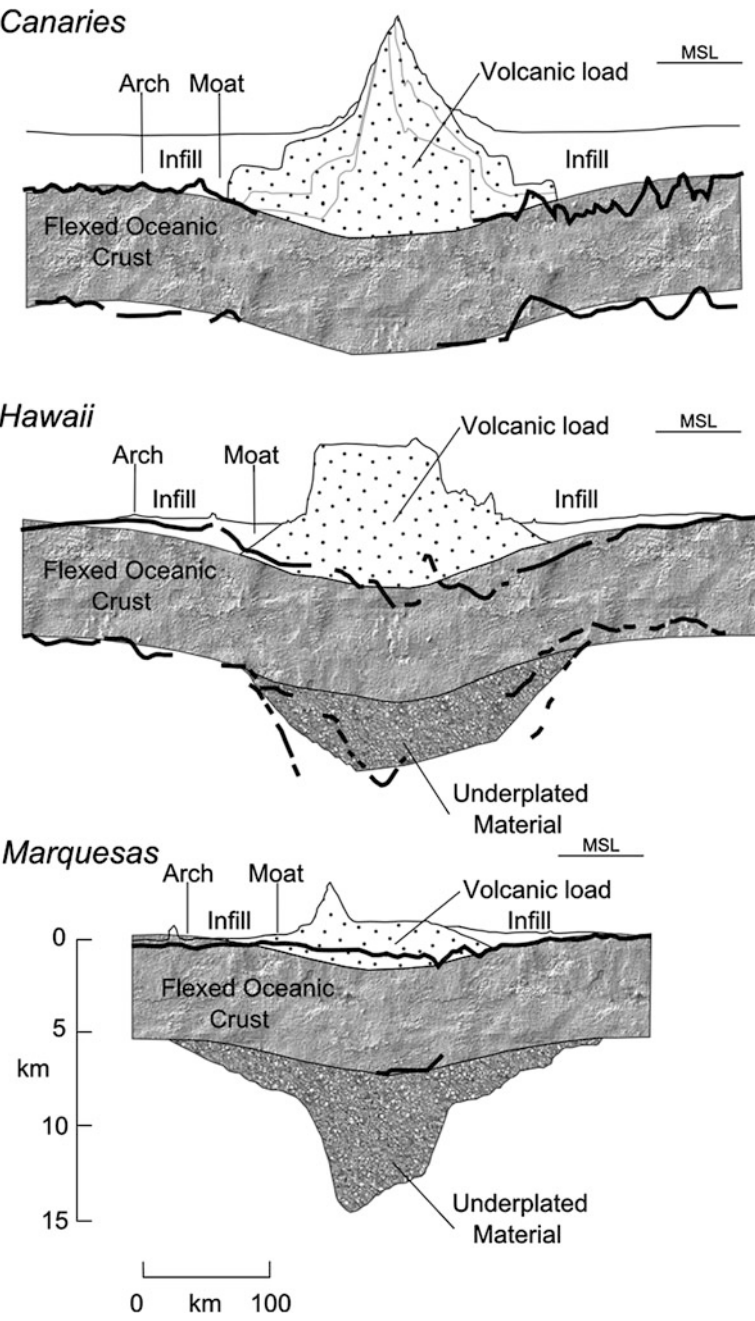


Fig. 2.12 Transects across Canary Islands, Hawaii and Marquesas, showing the development of the flexural moat and arch (modified from Watts et al. 1997)

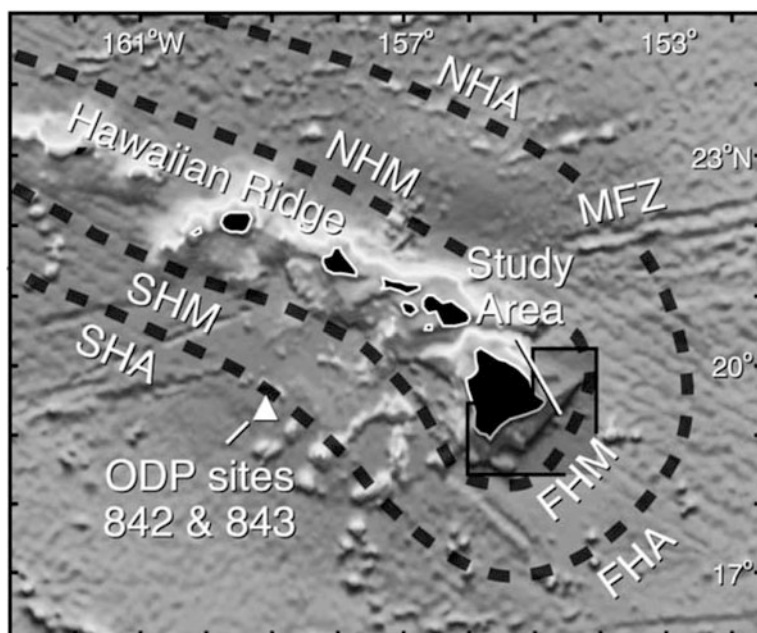


Fig. 2.13 Shaded relief bathymetric map of the Hawaiian Ridge and surrounding flexural moat and arch (Leslie et al. 2002). The moat and arch are subdivided into frontal regions, i.e. the Frontal Hawaiian Moat (*FHM*) and Arch (*FHA*), and flanking segments, i.e. the Northern and Southern Hawaiian Moat, and the Northern and Southern Hawaiian Arch (*NHM*, *SHM*, *NHA* and *SHA*, respectively). *MFZ* denotes Molokai fracture zone. The location of ODP Drill Sites 842 and 843 on the South Arch is indicated by *triangle*, as also the study area of Leslie et al. (2002)

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