10 Volcanoes and Plate Tectonics

10.1 Introduction

This chapter introduces the third part of the book, presenting the regional tectonic frame of volcanoes. In this third part the previously described volcano-tectonic processes are considered at a wider scale, highlighting the interaction of the volcanoes with the regional tectonic context. The focus here is on how the tectonic setting may affect the location, distribution, style, type and frequency of volcanic activity, on both the longer-term (i.e., thousands to millions of years) and, as suggested by recent studies, the shorter-term (i.e., years or less, as in the co- and post-seismic cycles of regional earthquakes). The change in scale in this part of the book permits to appreciate first-order magmatic processes related to plate-tectonics mechanisms, in an exciting journey around key regions of our planet. This journey allows explaining the major differences among the volcanic provinces, as well as showing how magma plays a leading role not only in promoting intra-plate processes but, quite unexpectedly, also in shaping plate boundaries.

In particular, this chapter introduces the main tectono-magmatic processes occurring in the frame of plate tectonics at the planetary scale, along divergent and convergent plate boundaries, and at hot spots; transform plate boundaries, as usually not generating magma, are not considered. In the following chapters, key regional examples are illustrated for an in-depth understanding. In more detail, Chap. 11 describes regional examples of divergent plate boundaries, including continental, transitional and oceanic magmatic rifts, providing a final discussion. Chapter 12 describes regional examples of subducting convergent plate boundaries, experiencing extensional, strike-slip and contractional regimes along the volcanic arc, with a final discussion. Finally, Chap. 13 describes the regional intraplate volcanism of oceanic and, subordinately, continental hot spots, with a final discussion.

The main aims of this chapter are to:

- provide an overview of plate tectonics, presenting the main tectonic settings where volcanic activity occurs;
- describe the main tectono-magmatic processes responsible for the observed differences in volcanic activity in each setting;
- discuss the general causal relationships between magmatism and plate tectonics.

10.2 The Plate Tectonics Frame

Plate tectonics is the paradigm that aims to describe most of the solid Earth processes. The theory has been formulated in the late 1960s to explain in a coherent frame several geological

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and geophysical data acquired in the previous decades, building on the concepts of continental drift and seafloor spreading (Wegener 1912; Vine and Matthews 1963; Vine 1966). The theory postulates that the outer portion of our planet consists of seven major rigid plates (North American, Pacific, Eurasian, African, Indo-Australian, South American and Antarctic) and several minor plates which are slowly moving with regard to each other (Fig. 10.1; e.g., McKenzie and Parker 1967; Isacks et al. 1968; Le Pichon 1968; Morgan 1968; Wilson 1968).

These mobile plates involve not only the Earth’s crust, but also the underlying upper mantle, with both crust and upper mantle forming the lithosphere, which is generally defined as the outer portion of the Earth with homogeneous elastic behaviour on time scales of thousands of years at least. However, depending on the considered problem, there are different definitions of lithosphere. The **seismic lithosphere** is defined by a zone of higher seismic velocity, contrasting with the underlying low velocity zone (or asthenosphere, see below). In the ocean basins its thickness increases from less than 20 km at oceanic ridges to more than 60 km approaching the continents. The temperature at the base of the seismic lithosphere is estimated as 600–650 °C. The **elastic lithosphere** is derived from the flexural behaviour of the Earth's surface when subjected to loading and unloading (such as by huge volcanic edifices, as in Hawaii, USA). Its thickness in ocean basins is similar to that of the seismic lithosphere, although thinner (30–40 km) under continents. In the ocean basins, the base of the elastic lithosphere roughly corresponds to a temperature of approximately 500 °C for dry olivine rheology. The **thermal lithosphere** is determined by a conductive gradient with a basal temperature of about 1280 °C. Therefore, the thermal lithosphere is about twice as thick as the seismic and elastic lithosphere. In the ocean basins, its thickness increases from a few

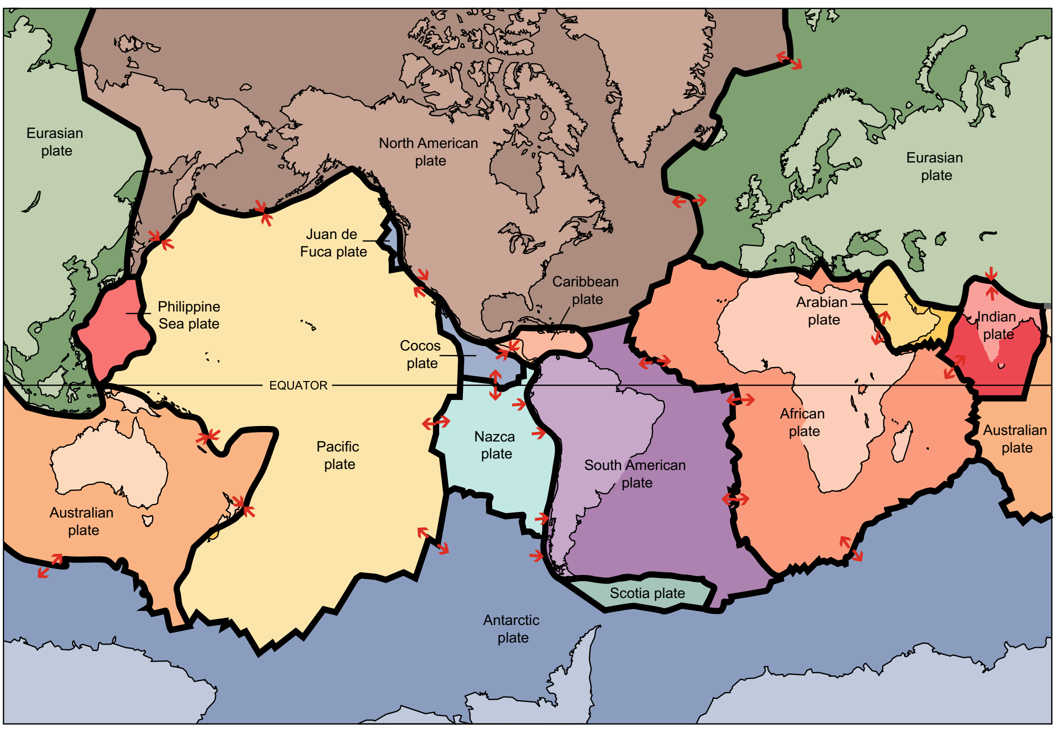

Fig. 10.1 Overview of the tectonic plates composing the Earth lithosphere and relative motion along their boundaries (red arrows; from USGS; https://commons.wikimedia.org/w/index.php?curid=535201)

kilometres at oceanic ridges to approximately 100 km on older oceanic lithosphere, thickening to approximately 280 km beneath the continents (Parsons and McKenzie 1979). Based on their composition, the lithospheric plates may be oceanic or continental. The **oceanic lithosphere** includes, in its upper part, the basaltic crust generated along oceanic ridges. The thickness of the oceanic lithosphere varies between a few kilometres at the ridge axis to 50–140 km away from the ridge, when the lithosphere becomes older, colder and denser. The **continental lithosphere** is formed through magmatism and accretion of continental terranes (which are fault-bounded regions with distinctive geological history), and its crust has a bulk granitic composition. The thickness of the continental lithosphere varies between ~40 and ~280 km. The basaltic oceanic crust is denser (average density of ~2900 kg/m^3) than the granitic continental crust (~2700 kg/m^3), thus determining the higher density of the oceanic lithosphere with regard to the continental lithosphere.

Below the lithosphere lies the **asthenosphere**, which is the highly viscous portion of upper mantle with variable portion of partial melt (up to ~10%). Lithospheric plates move relative to or with the asthenosphere, driven by body forces or traction forces, respectively. Body forces acting within the lithosphere are generated by gravity and density contrasts between the plates and the mantle. The two main body forces are ridge push and slab pull (Forsyth and Uyeda 1975; McKenzie 1977; Turcotte and Schubert 1982; and references therein; Stuwe 2007, and references therein). **Ridge push** is the result of gravitational forces acting on the young, raised oceanic lithosphere around ocean ridges, causing it to slide down the similarly raised but weaker asthenosphere and push on lithospheric material farther from the ridges. **Slab pull** is the pulling force exerted by a cold, denser oceanic plate plunging into the mantle due to its own weight: in fact, the subducting plate is also named slab. Because the oceanic plate is denser than the hotter mantle beneath it, the density contrast causes the plate to sink into the mantle. Traction forces are imparted by the convecting asthenosphere along the base of the lithosphere and along the slab at depth. The convecting asthenosphere is the shallow manifestation of a larger convection occurring within the upper mantle, driven by the density contrast between its hotter and lighter rising portions and the colder and denser sinking portions. Plate tectonics is thus driven by thermal processes acting not only in the outer portion of the Earth, but also within. The latter include internal cooling related to primordial heat associated with initial planetary formation and heat generated by radioactive decay. Magmatic and, in particular, volcanic activity are the final result of these thermal processes and their interaction with the lithospheric plates. The relative motion between the lithospheric plates is on the order of a few cm/year, with the fastest rates of ~15 cm/year observed along portions of the ridge in the Pacific Ocean (East Pacific Rise; Gordon and Jurdy 1986; Gripp and Gordon 2002; Steinberger et al. 2004; Kreemer et al. 2014).

An assumption of plate tectonics is that lithospheric plates behave in a rigid way. This implies that the motion at any portion within the same plate is consistent and that the deformation resulting from the relative motion of nearby plates mainly focuses along the edge of the plates, or plate boundaries (Fig. 10.2). **Plate boundaries** thus consist of a zone of deformation resulting from the relative motion of the plates. This zone may have an extremely variable width, in some cases reaching up a thousand of kilometres, as for example in the Himalayan–Tibetan collisional zone. Depending upon the relative motion between neighbouring plates, three main types of plate boundaries are distinguished: divergent, convergent and transform (Fig. 10.3; e.g., van der Pluijm and Marshak 2004; Fossen 2010, and references therein).

Divergent plate boundaries develop along the contact between plates that move away from each other (Fig. 10.3a). This motion determines lithospheric thinning at the plates contact; thinning in turn promotes decompression and magma generation. The rate of divergence is proportional to the flux of magma through the plate boundary, determining its structure, magmatic and volcanic

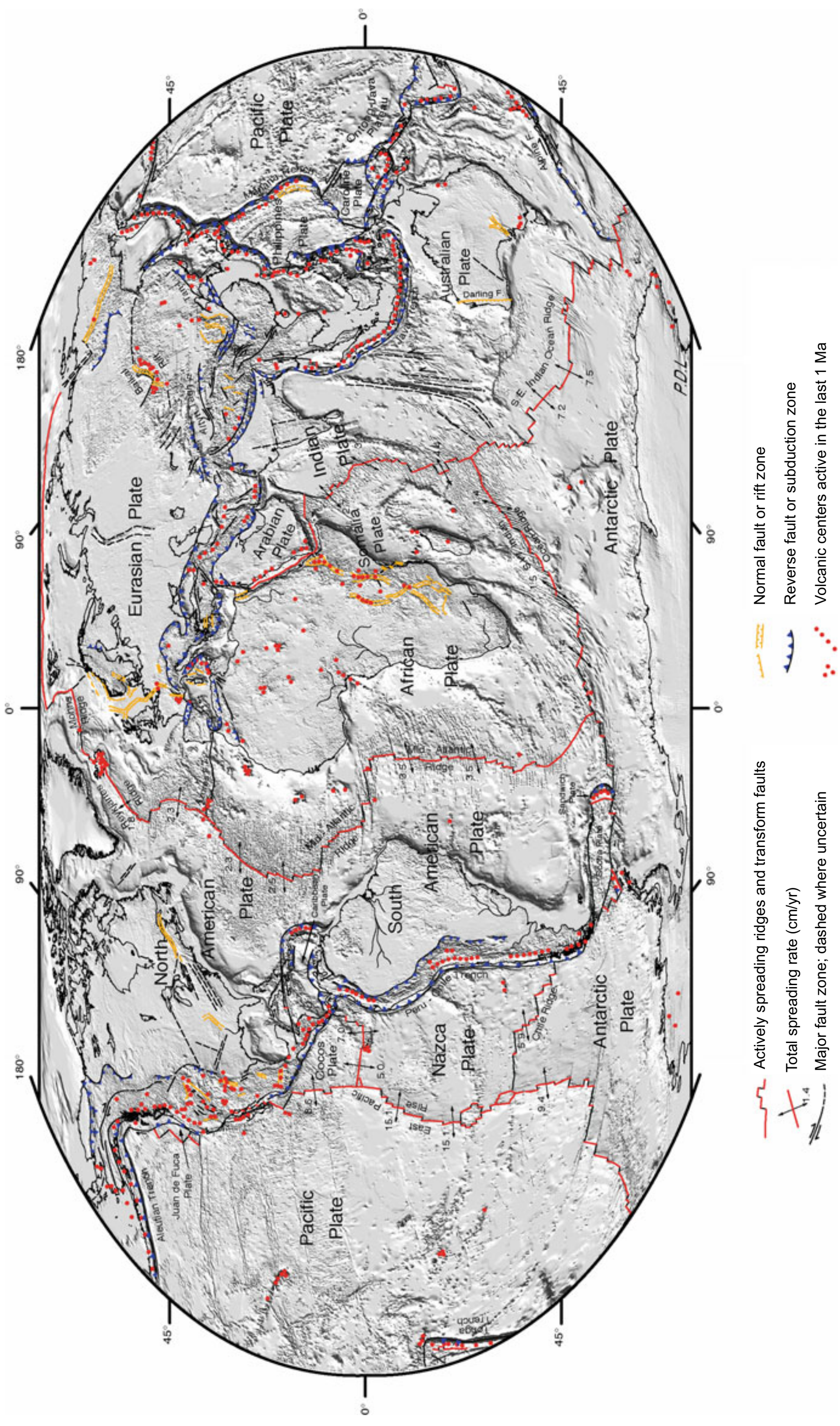

Fig. 10.2 Tectonic and volcanic activity of the Earth in the last one million years, highlighting the overall structure of the plate boundaries and the distribution of volcanoes (from NASA; https://commons.wikimedia.org/w/index.php?curid=61568)

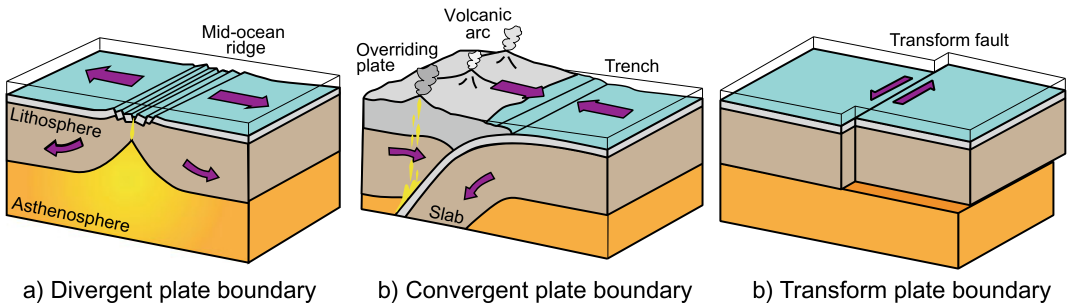

Fig. 10.3 Sketches of a **a** divergent, **b** convergent and **c** transform plate boundary

features. Divergent plate boundaries are found in continental lithosphere (as in the East African Rift System), oceanic lithosphere (as along the Mid-Atlantic Ridge or the East Pacific Rise) and any transitional lithosphere in between (as in Afar, Ethiopia). Continental divergent plate boundaries mark the onset of the divergence between two plates, and are characterized by extension rates of a few mm/yr. Oceanic divergent plate boundaries are the expression of the mature stage of divergence, with extension rates ranging from several mm/year to ~15 cm/year. These oceanic boundaries, coinciding with the oceanic ridges, are by far the most frequent type of divergent plate boundary. The general processes occurring at divergent plate boundaries are introduced in Sect. 10.3, whereas their regional features are described in Chap. 11.

Convergent plate boundaries develop when two nearby lithospheric plates move toward each other, determining overall contraction along the boundary (Fig. 10.3b). If an oceanic plate is involved in the convergence, its higher density determines its subduction into the mantle below the lighter overriding plate; an oceanic **trench** marks the submarine contact between the subducting and overriding plates. The subduction may occur with a variable dip of the slab, which depends upon its density, itself a function of its temperature and age, and the general tectonic setting. For example, younger and/or warmer subducting lithosphere creates a lighter slab gently dipping in the mantle, conversely to older and colder lithosphere, associated with a denser and steeper slab. At a depth of 80–150 km, the slab experiences the partial melting of its least stable components, which mainly consist of hydrated minerals. The fluids associated with these minerals rise in the overlying mantle wedge, decreasing the melting point of the rocks and promoting the generation of mafic magma. This magma rises and interacts with the overlying crust, becoming more evolved and felsic, eventually erupting along volcanic arcs. **Volcanic arcs** are the surface expression of magmatism along convergent plate boundaries, consisting of chains of aligned and regularly spaced volcanoes, mostly stratovolcanoes, lying on the plate that is overriding the slab, as observed along the circum-Pacific Ring of Fire. Conversely, if no oceanic plate is involved in the convergence process, as in the case of two approaching continental plates, no dense plate sinks and **continental collision** replaces subduction. Collision leads to the thickening of the continental lithosphere and to orogenesis (i.e., mountain building) along the plate boundary, as for example along the Alpine-Himalayan region. In this case, volcanism is negligible or absent. The general processes occurring at convergent plate boundaries are introduced in Sect. 10.4, whereas their regional features are described in Chap. 12.

Transform plate boundaries develop when two nearby plates move laterally past each other, without significant convergence or divergence component (Fig. 10.3c). Transform plate boundaries are thus characterized by a dominant component of horizontal motion, accommodated by the development of strike-slip fault systems. Transform boundaries may affect oceanic and continental lithosphere. Oceanic transform faults accommodate motion between active oceanic ridge segments, as along the Mid-Atlantic Ridge. These transform faults are usually found below the sea level, although at times they may appear onland, as in northern Iceland or southern Australia. Continental transform faults accommodate the motion between oceanic ridge segments and collisional regions, as for example observed along the San Andreas Fault (California) or the Alpine Fault (New Zealand), respectively. Conversely to divergent and convergent plate boundaries, the horizontal motion along transform boundaries does not usually involve any important process associated with the generation of magma. Indeed, in the few cases of volcanism along transform zones there is no evidence or model that magma generation results from the activity of the transform. Rather, volcanism appears controlled by other processes, most notably the activity of hot spots (see below), as for example in the Azores Archipelago. As magma is usually not generated along transform plate boundaries, this type of boundary is not considered further, although the effect of a transform setting on magmatism is considered in the Azores hot spot (Sect. 13.7).

At the global scale, mantle convection determines plate motion, with upwelling regions of mantle flux coinciding with divergent plate boundaries and downwelling regions with convergent boundaries, so that the extension created at divergent boundaries is compensated by the contraction at convergent boundaries. As a result, new plates may grow, or be even created, whereas old plates may consume and become ultimately destroyed along subduction or collision zones. These features merge into an imaginary evolutionary cycle, or **Wilson cycle** (Wilson 1972; e.g., Stuwe 2007). This ideally begins with the lithospheric extension of a continent (as along the East African Rift System), evolving in continental break-up and the development of an immature spreading centre producing oceanic lithosphere (as in the Red Sea). This condition may then lead to an ocean with mature ridge, as observed in the Atlantic or in the Pacific. This divergent motion is balanced by plate convergence, which begins with the onset of subduction of an oceanic plate underneath a continental plate, as observed along the west coast of South America. This condition may also involve the subduction of an oceanic ridge underneath the continental plate, as the Juan de Fuca Ridge below North America: in this case, the lower density of the young and hot ridge hinders its sinking into the mantle, generating a flat slab not reaching the partial melting conditions, and thus not generating any volcanism. Alternatively, all the oceanic lithosphere may be consumed in the subduction process, leading to the collision between two lower-density continents and the cessation of subduction, as observed between India and Asia.

Therefore, most of the magmatic activity on our planet develops along divergent and convergent plate boundaries. Nevertheless, a minor though significant portion of magma is erupted away from plate boundaries, in focused areas, or **hot spots**, whose activity is independent of plate motion (Fig. 10.4; Davies 1999). In particular, hot spot volcanism results from the rise through the mantle and the crust of hot and narrow plumes, which often originate at the mantle-core boundary. Being independent of any plate motion, mantle plumes and their surface manifestation hot spots are considered to be stationary and, for this reason, they have been commonly used as a reference frame in an otherwise ever moving Earth's surface. Hot spots are best identified in intraplate domains, that is distant from plate boundaries, as for example the Hawaiian, Reunion or Yellowstone hot spots. However, several major hot spots lie along plate boundaries, especially divergent boundaries, as for example Iceland and Afar (the latter occurring at a triple junction between three divergent boundaries). These less evident hot spots may be recognized by several features, including the locally anomalous production of magma with slightly distinct composition that generates alkali basalts, rather than the tholeiitic basalts commonly produced along divergent plate boundaries. Since in many cases the activity of a hot spot along a divergent plate boundary predates the development of that boundary (as in Afar), it is likely that the thermal weakening induced by the hot spot activity in the lithosphere "captured" the plate boundary. In these cases, the upwelling

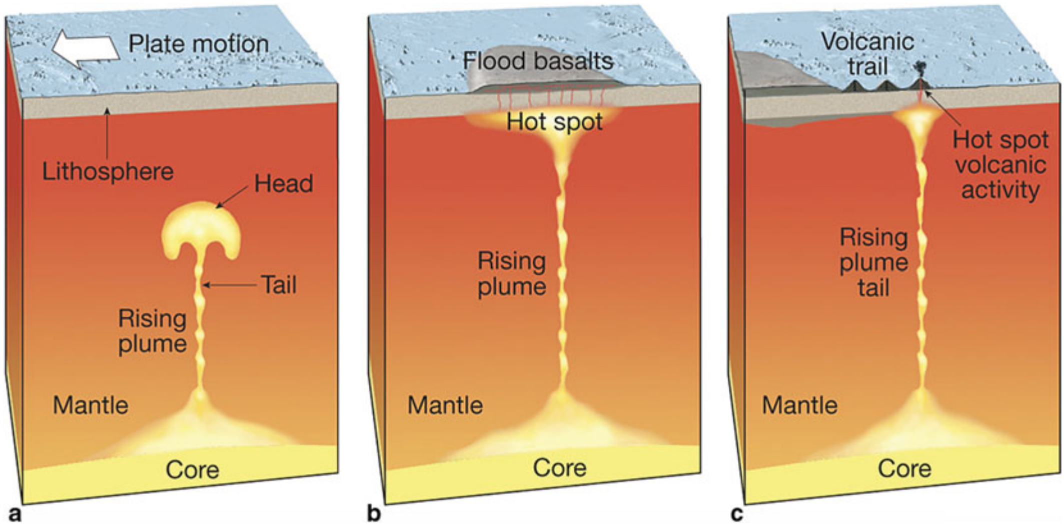

Fig. 10.4 Stages of mantle plume development: **a** a plume rises form the mantle-core boundary; **b** the head of the plume generates massive volcanism at the surface; **c** the plume forms a hot spot that generates a chain of volcanoes parallel to plate motion (Jaffe and Taylor 2018; Image from Tasa Graphics)

of the upper mantle along the boundary couples with the upwelling of the plume. The general processes occurring at hot spots are introduced in Sect. 10.5, whereas their regional features are described in Chap. 13.

In summary, divergent and convergent plate boundaries, as well as hot spots, directly or indirectly explain the distribution of most of the volcanoes on our planet through different processes, which are discussed below. This distribution closely matches that of the Earth's seismicity, as seismic activity also focuses along plate boundaries. This seismicity generally results from regional tectonic processes related to the deformation of the plate boundary (as along transform faults, slabs or in collisional zones), although in some cases shallow seismicity may be also associated with magmatic activity (as along some divergent plate boundaries).

There are three main modes of obtaining partial melting of a peridotite rock in the upper mantle: pressure decrease, temperature increase and addition of volatiles (as H_2O and CO_2). Through one or more of these conditions a solid peridotite just below the dry melting curve (solidus) may experience partial melting (Fig. 10.5; e.g., Schmincke 2004). These conditions controlling the generation of magma and the associated geothermal gradients may significantly differ at divergent and convergent plate boundaries and hot spots (Fig. 10.5). Under ordinary intraplate conditions, and away from hot spots, partial melting is not expected to occur at any lithospheric and asthenospheric level: as a result, the geotherm here is below the solidus of the rocks at any depth. Along divergent plate boundaries, as for example in an oceanic ridge, lithospheric thinning induces decompression within the asthenosphere, which becomes shallower and experiences partial melting. This creation of melt in conditions of rapid decrease of pressure and at a constant temperature is known as **decompression melting**. The generated magma rises and heats the overlying lithosphere, so that here the geotherm overcomes the solidus of the rocks. Along convergent plate boundaries, the fluids released from the partially melting slab shift the solidus of the rocks towards lower temperatures. In particular, in the lower lithosphere the rocks solidus shifts below the geotherm, promoting partial melting and the generation of magma above the subducting lithosphere. Below hot spots, the rise of hot mantle plumes increases the geothermal gradient,

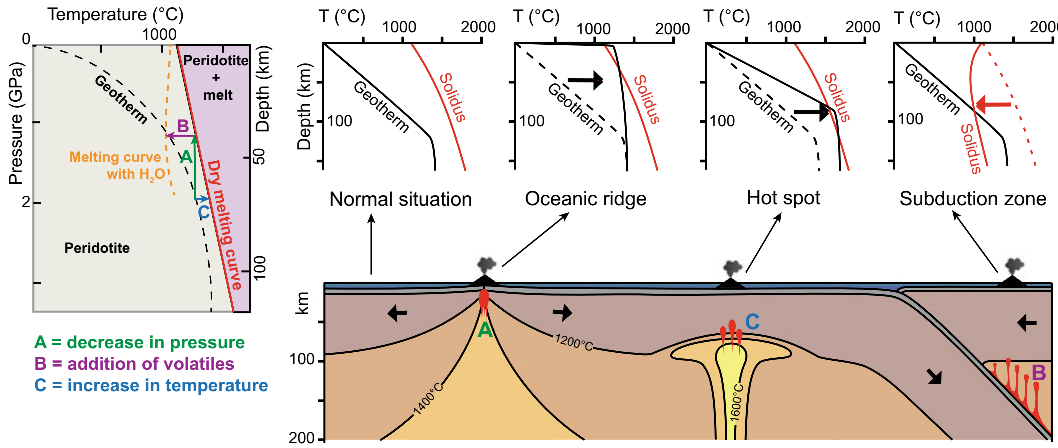

Fig. 10.5 Left: modes of partial melting of a peridotite rock through pressure decrease (A), addition of volatiles (as H_2O and CO_2; B) and temperature increase (C). Through one or more of these conditions, a solid peridotite just below the dry melting curve (solidus) may experience partial melting (modified after Schmincke 2004). Right: these three conditions predominate at divergent boundaries (as oceanic ridges), convergent boundaries (as subduction zones) and hot spots, respectively (see A, B and C in bottom panel). In each of these domains the predominant condition results in the geotherm partially overcoming the rock solidus, producing melt (see top diagrams; modified after: https://commons.wikimedia.org/w/index.php?curid=10433862)

which overcomes the solidus of the rocks in the lower lithosphere: this also results in partial melting and magma generation.

The distribution of the relative volumes erupted along divergent and convergent plate boundaries, as well as hot spots, shows that approximately two thirds of the magma are erupted along divergent plate boundaries, approximately one quarter is erupted along volcanic arcs, and more than one tenth along hot spots Fig. 10.6 (Fisher and Schmincke 1984). Clearly, most of the magma is erupted along oceanic ridges, where new crust is being continuously created over tens of thousands of kilometres. The creation of new crust does not only occur at the surface, through volcanism, but also at depth. Indeed, the amount of erupted magma is usually one tenth of the amount of magma permanently intruded in the crust. Similar ratios between erupted and intruded magma are also inferred for convergent plate boundaries and hot spots. This indicates that volcanic activity is the final and limited manifestation of much larger processes ultimately responsible for magma accumulation at depth, whose release at the surface is largely independent of the specific tectonic setting.

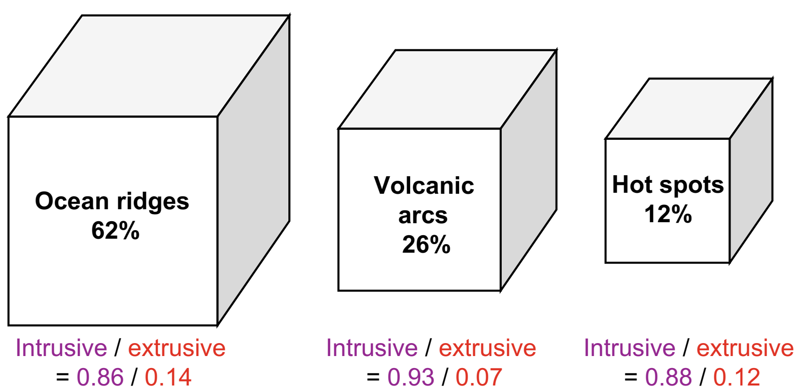

Fig. 10.6 Relative volumes of magma, and their inferred intrusive to eruptive ratios, erupted at ocean ridges, volcanic arcs and hot spots (modified after Fisher and Schmincke 1984)

The following sections introduce the general tectono-magmatic processes representative of the relationships between the regional tectonic context and volcanic activity occurring along divergent and convergent plate boundaries, as well as at hot spots.

10.3 Magmatic Processes Along Divergent Plate Boundaries

A divergent plate boundary identifies a **rift zone**, or rift, which is the lithospheric volume experiencing extension at the contact between the two diverging plates. Note that the term rift is more broadly used for any zone experiencing extension, also in intraplate settings, as for example the Rhine Graben (western Europe), the Baikal Rift (central Asia) and the Rio Grande Rift (southern USA).

Rift zones may be perpendicular or at some angle to the direction of plate divergence. **Orthogonal rifting** occurs when the direction of divergence is orthogonal to the trend of the rift axis, and thus to the trend of the major normal faults identifying the rift zone. However, this condition is rarely verified, as approximately 80% of the plate boundaries are oblique with regard to plate motion, forming an angle between 10° and 80° and highlighting a predominant **oblique rifting**. In particular, the direction perpendicular to the rift trend and the direction of plate divergence define the angle δ of obliquity of the rift, with orthogonal rifts having $\delta \sim 0°$ and oblique rifts $\delta > 0°$ (Fig. 10.7; Withjack and Jamison 1986; Tron and Brun 1991; Brune et al. 2018; Sani et al. 2019). In an oblique rift three structural directions are usefully defined: (a) the rift trend, (b) the direction of relative motion between the two plates (or direction of divergence), and (c) the trend of the greatest principal horizontal strain axis ε_1 of the finite strain ellipsoid. For instantaneous strain, the latter coincides with the minimum principal stress σ_3, or the local extension direction. In orthogonal rifts, ε_1 is perpendicular to the normal faults and parallel to the divergence direction. In oblique rifts, for low amounts of extension, ε_1 is approximately bisector of the angle between the regional direction of divergence and the normal to the rift axis; only for large stretching factor β (with $\beta \geq 5$) ε_1 tends to become parallel to the direction of divergence (Withjack and Jamison 1986; Tikoff and Teyssier 1994; Fournier and Petit 2007). Typically, fault systems generated during oblique rifting show an en-echelon arrangement, with fault segments oblique to the rift axis and to the direction of divergence. As oblique rifts are characterized by a shear component along the rift, faults striking orthogonal to the local extension direction display dominant dip-slip kinematics, whereas those oblique have a

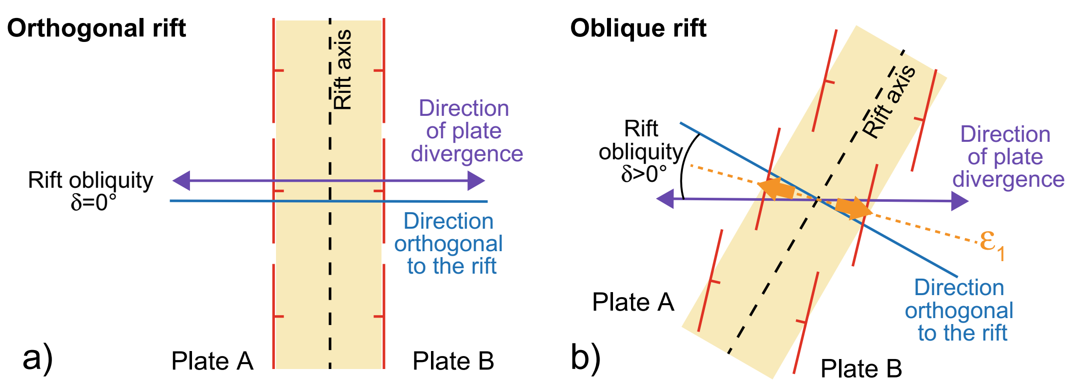

Fig. 10.7 Geometrical relationships (in map view) between the rift axis (and its perpendicular, in blue), the direction of relative motion between two plates (in purple) and the trend of the greatest horizontal principal strain ε_1 (or local extension direction, in orange) during orthogonal (**a**) and oblique rifting (**b**); yellow areas schematically indicate rift valleys (modified after Sani et al. 2019)

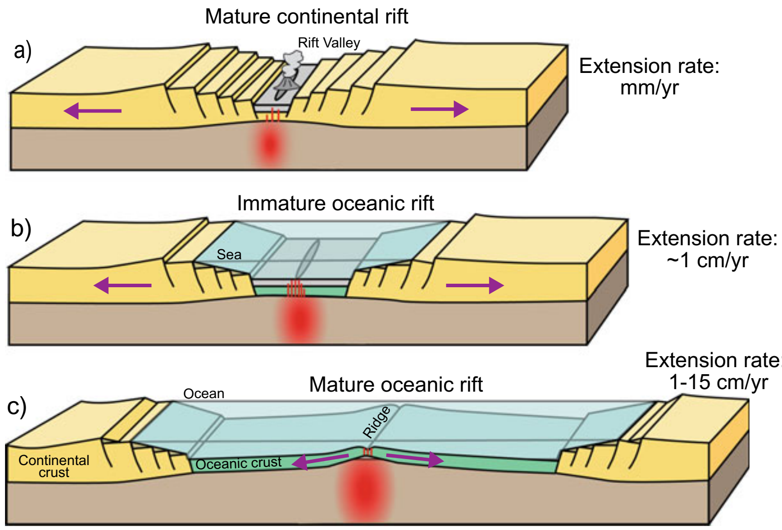

Fig. 10.8 Main stages of rifting, from mature continental rifting (**a**) to immature oceanic rifting (**b**) and oceanic spreading developing an oceanic ridge (**c**)

strike-slip component of motion that depends upon the degree of obliquity (e.g., Sani et al. 2019, and references therein).

In an ideal progression consistent with the Wilson cycle, rifting and thus plate divergence first develops on continental domains (**continental rifting**), determining the split of a larger plate into two or more smaller plates, as observed in the East African Rift System, which generates the larger Nubia (African) and Somali plates, as well as additional microplates. The persistence of rifting promotes the progressive thinning of the continental lithosphere, with augmented decompression, partial melting and magma generation (Fig. 10.8). As a consequence of crustal thinning and the progressive rise of more primitive mantle melts, the erupted magma becomes finally characterized by an increasing mafic component, evolving into transitional crust. Then mafic magmatism produces progressively newer crust and lithosphere within the rift, eventually replacing the previous lithosphere that shifts to the sides following plate divergence. This results in an **oceanic rifting** stage characterized by seafloor spreading, which may be initially immature (as observed in the Red Sea) and then mature (as in the Mid-Atlantic Ridge). This evolutionary continuum describes an ideal full spectrum of development of a divergent plate boundary, as not all continental rifts create oceanic lithosphere and evolve into seafloor spreading. The main stages of this ideal progression are described below.

The initiation of a divergent plate boundary may derive from two main processes. In fact, the extensional stresses within the lithosphere required for plate divergence may be initially imparted by either relative plate motion or doming of the lithosphere by a mantle plume. On these premises, two models have been proposed to explain the origin of rifting: passive and active rifting (Fig. 10.9; Bott 1995; Corti et al. 2003; Ziegler and Cloetingh 2004; van der Pluijm and Marshak 2004, and references therein). **Passive rifts** are initiated along plate boundaries and are tectonically driven: here regional extension generated by plate boundary forces and/or convective drag at the base of the lithosphere is responsible for thinning. Thinning reduces the lithostatic load on the asthenosphere rocks, enhancing decompression and partial melting herein. The asthenospheric melts replace the previously unstretched lithosphere, promoting magma emplacement and rise at the base and within the extended crust. Passive rifts thus undergo an earlier phase of tectonic extension and subsequent magmatism. In other cases, the initiation of continental rifting may be aided by a mantle plume, which provides the geodynamic forces to drive extension within the lithosphere

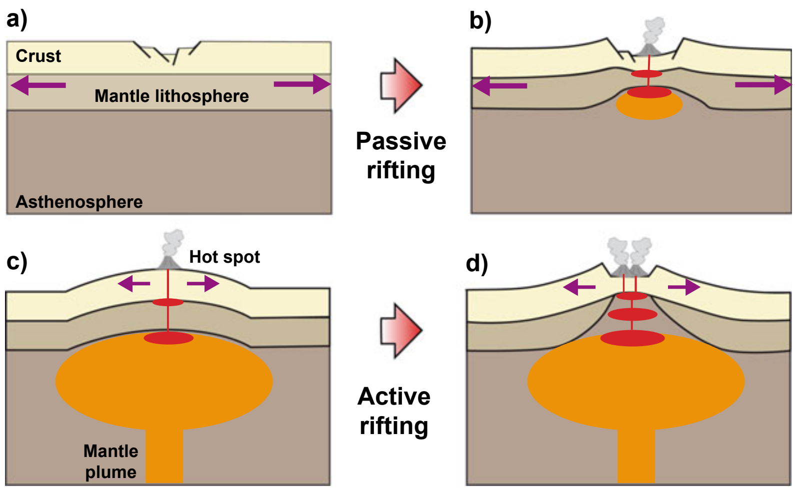

Fig. 10.9 Schematic diagrams illustrating the passive (**a**, **b**) and active (**c**, **d**) hypotheses for the initiation of rifts. In passive rifting extension is driven by a tensional regional stress originating from remote plate boundary forces: generation of a hot, low-density region in the asthenosphere (orange area) results from extension, followed by magmatic underplating (red areas) and volcanism. In active rifting the ascent and emplacement of a hot, low-density body in the sublithospheric mantle (orange area) is a primary cause for extension: this accounts for pre-rift uplift, magmatic underplating (red areas) and volcanism (not to scale; modified after Corti et al. 2003)

and the magma to accommodate the extension. These **active rifts** are thus initiated and driven in the mantle by the activity of plumes, emplacing hotter material below the lithosphere, inducing uplift (doming) and failure. Doming, much wider than the area to be affected by extension, and widespread volcanism predating extension of several million years, are distinctive features of active rifts. In particular, the volcanism fed by the mantle plume initiates the active rift erupting high volumes of basaltic lavas, or flood basalts (see below), as for example observed in Afar. This phase of doming and widespread volcanism anticipates the phase of lithospheric extension associated with more limited volcanism. Therefore, the relative timing of three main components (uplift/subsidence, extension, volcanism) determines whether rifting can be attributed to passive or active processes, although the application of this procedure to real settings is often difficult (Ruppel 1995).

Considering in detail the development of a divergent plate boundary, stretching associated with rifting thins the lithosphere, decreasing the lithostatic pressure in the asthenosphere below. The asthenosphere is so hot (>1280 °C) that, when such decompression occurs, partial melting begins, promoting magma production. The proportion between the amount of extension and percentage of created melt is not linear, with an effective melt production being already available for moderate extension, the latter quantified through the stretching factor β (White and McKenzie, 1989). The produced magma is less dense than the overlying lithosphere, and thus rises through it. A part of this rising magma may get trapped and solidify at the base of the crust, a process called **magmatic underplating**, which thickens the crust by adding mass to its base. In the case of emplacement of a hot mantle plume beneath a rifting lithosphere, as with active rifts, the higher temperature of the plume induces a higher amount of partial melting in the asthenosphere, and thus a higher production of magma. This condition may ultimately explain the generation of **flood basalts**, which are very voluminous (≥ 1000 km^3) amounts of erupted basaltic lavas. Flood basalts likely form where a mantle plume underlies a rift, for the peridotite in the plume is significantly hotter than the peridotite comprising normal asthenosphere, and thus undergoes a greater degree of partial melting during decompression (see Sect. 10.5; e.g., Rooney 2020a, and references therein).

The mafic magma that does not solidify at the base of the crust rises further, intruding into the crust and eventually erupting at the surface. Mafic magma is so hot (>1000 °C) that when it intrudes the continental crust it also conducts enough heat into the adjacent crust to cause partial melting. This melting takes place because the rock comprising the continental crust contains minerals with relatively low melting temperatures (<900 °C), generating lighter felsic magmas. As a consequence, the source and composition of erupted magmas also evolves. Magmas erupted during the initial stages of rifting are typically tholeiitic basalts formed through decompression melting of the mantle. As the rift evolves, magma differentiation towards felsic compositions may occur in the continental crust, forming the entire petrologic compositional spectrum of magmas. In some cases, the crustal emplacement of large volumes of relatively hot basaltic magma from a mantle source and the generation of crustal granitic magmas by partial melting may result in the eruption of a **bimodal volcanic suite**, characterized by the association of rhyolitic and basaltic products. Bimodal volcanism is for example observed in the mature portion of the continental Main Ethiopian Rift (see Chap. 11; e.g., Rooney 2020b, and references therein).

With the increased generation and rise of melts, magmatism plays a progressively more important role and the continental plate boundary system begins to switch from one that is dominated by tectonic processes to one where magmatism accommodates divergence. In fact, although regional tectonic structures (as normal faults) dominate in the early stage of continental rift growth, once magmatism is initiated these structures become less important in accommodating plate motion, as exemplified by the East African Rift System. Here the decrease in the activity of regional normal faults is accompanied by the increase in magmatic activity through repeated dike emplacement and dike-induced normal faults, which progressively accommodate plate motion. Following the increased crustal thinning and the encouraged injection of deeper feeder dikes, the intruded and erupted magmas start to show dominant mafic composition (see Sect. 10.7; Ebinger and Hayward 1996; Ebinger et al. 2010; Acocella 2014; LaFemina 2015).

At this stage, continental break-up may occur, especially if the progressively dominating magmatic activity within the rift is supported by the flux of an underlying mantle plume. **Continental break-up** consists of the fragmentation of the plate, which splits into smaller plates whose original contact is being replaced by the creation of a new and mainly mafic crust promoted by the magmatic activity of the rift and the plume. During continental break-up the crust evolves from continental to transitional to proto-oceanic, as exemplified in the Afar region (e.g., Ebinger and Hayward 1996).

Increased rates and amount of extension and increasing magmatic activity develop an oceanic divergent plate boundary with oceanic lithosphere, whose crust is mainly made of basalt and gabbro, and the upper mantle is made of ultramafic rock, as peridotite. The oceanic lithosphere is thinner than the continental lithosphere, although its thickness is a function of its age. As oceanic lithosphere moves away from the ridge axis, it cools and thickens to between 50 and 140 km, following the simplified relation:

$$z \approx 2\sqrt{\alpha t} \qquad (10.1)$$

where z is the thickness of the lithosphere, α is its thermal diffusivity, and t its age: the latter can be approximated by the ratio between the distance from the ridge axis and the relative rate of plate motion (e.g., LaFemina 2015, and references therein). In addition to becoming thicker with distance from the ridge axis, the lithospheric mantle becomes denser. After ~15 Ma, oceanic lithosphere becomes denser than the asthenospheric mantle and can therefore subduct into the mantle, a crucial condition for convergent plate boundaries (see Sect. 10.4). Oceanic divergent plate boundaries are defined by the relative rate of plate motion between the diverging plates and are subdivided into ultrafast (full spreading rate of >12 cm/year), fast (<12 to >8 cm/year), intermediate (<8 to >5 cm/year), slow (<5 to >2 cm/year), and ultraslow (<2 cm/year). These differences in spreading rate give rise to distinctive morphologies, structures and magmatic activities, as discussed in Chap. 11.

As anticipated in Chap. 2, the architecture of the rift zone characterizing the divergent plate boundary may be associated with a symmetric pure shear or an asymmetric simple shear mode of extension, or a combination thereof (McKenzie 1978; Wernicke 1985). Nevertheless, magmatically productive divergent plate boundaries are usually associated with a symmetric pure shear type of extension, rather than an asymmetric simple shear type. This follows from the fact that focused lithospheric extension, as determined by pure shear conditions, promotes melt generation much more than delocalized extension associated with simple shear (see Sect. 2.7.1; Latin and White 1990). Asymmetric rifting may be more common in the early continental stages, where the rift is mostly characterized by half-graben structures. Conversely in the advanced oceanic stages rifts develop a more symmetric architecture, mainly consisting of graben-like structures.

The rise of magma along divergent plate boundaries commonly occurs through the emplacement of regional dikes perpendicular to the local direction of extension, or minimum principal stress σ_3. In the case of rifts experiencing orthogonal extension, the dikes strike parallel to the rift axis, with an overall collinear configuration; in the case of rifts experiencing oblique extension, the dikes strike at some angle with the rift zone, with an overall en-echelon configuration (Fig. 10.10).

With the possible exception of portions of fast oceanic ridges (as the East Pacific Rise), dikes, and more in general magma, are not uniformly distributed along the entire rift length. Indeed, the rise and storage of magma along both continental and oceanic divergent boundaries focuses in elongated crustal portions, called **magmatic systems** or magmatic segments (Gudmundsson 1995; Ebinger and Casey 2001). These are usually located along the rift axis, and their length depends upon the extension rate of that portion of rift: the higher the extension rate, the longer the magmatic segment, with typical lengths in the order of 10^1 km. Each magmatic system consists of a dominant polygenic volcano and several monogenic vents aligned parallel (in case of orthogonal extension) or slightly oblique (in case of oblique extension) to the direction of the rift (Fig. 10.10). Volcanic activity is accompanied by diffuse fracturing, mostly consisting of extension (tensile) fractures and normal faults striking parallel to the alignment of the volcanoes. Magmatic segments indicate that magmatic activity and extension are not uniformly distributed along the plate boundary. Rather, they show that the diverging lithosphere is split along

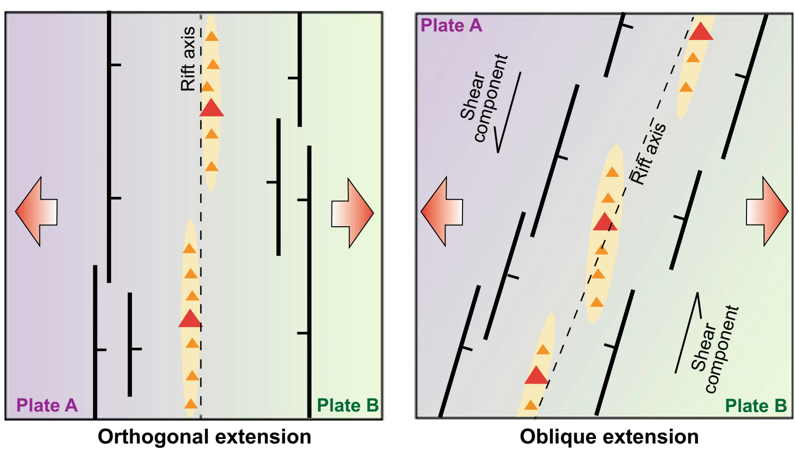

Fig. 10.10 Schematic structure (map view) of a portion of a divergent plate boundary experiencing orthogonal (left) and oblique (right) extension: magmatic systems are represented by yellow ellipses, polygenic volcanoes by red triangles and monogenic volcanoes by orange triangles. See text for details

$\sim 10^1$ km long wedges of intruded magma, which are periodically activated, determining a temporally and spatially incremental mode of opening of the plates. In the fast oceanic ridge of the East Pacific Rise the magmatic contribution is more stable in time and continuous in space: as a result, the magmatic systems are longer, on the order of 10^2 km, and more frequently active (Acocella 2014, and references therein).

In some cases, significant volcanic activity may be found also off-rift, or along the sides of the rift marking the plate boundary. This is for example the case of the extinct polygenic trachytic volcanoes, up to 1500 m high, lying on the flanks or the Main Ethiopian Rift. The unusual location of these volcanoes is interpreted to result from the deflection of the magma propagation paths during the early stages of rifting, under the unloading induced by the rift depression; before the establishment of a central (in-rift) propagation path, the unloading determines the lateral propagation of dikes, feeding the off-rift volcanoes (Fig. 7.3; Maccaferri et al. 2014).

In synthesis, in a divergent plate boundary tectonics and volcanism focus in magmatic systems along the rift axis (Fig. 10.10): tectonic structures consists of extension (tension) fractures and normal faults, whereas volcanism is represented by polygenic and monogenic volcanoes. While monogenic volcanoes are fed by dikes perpendicular to the opening direction, the feeding system of polygenic edifices is less straightforward. In many cases, there is evidence that the same regional dikes feeding monogenic volcanoes also feed the polygenic volcanoes. This may explain the elongation of the polygenic volcanoes parallel to the rift axis, as for example Boseti, Erta Ale (Ethiopia) and Hekla volcanoes (Iceland; Nakamura 1977). However, while frequent, this condition is not the rule and the different orientation of the volcanic edifices in rift zones, for example parallel to the opening direction, may be explained by other processes, as discussed in Sect. 5.5. The composition and type of volcanism associated with divergent plate boundaries is largely dependent on the amount of extension. Lower extension rates (and amounts of extension) are usually found on continental divergent boundaries. These are associated with a thicker continental granitoid crust (thicker than 20–30 km), which affects the composition of the erupted magma through storage, assimilation and differentiation towards felsic compositions: volcanism may be explosive, and ignimbritic eruptions are not uncommon. The associated volcanic edifices include polygenic, dominantly felsic stratovolcanoes and calderas. Conversely, higher extension rates and amounts of extension are typical of oceanic divergent boundaries, where the thinner and mafic crust does not affect the original composition of the basaltic magmas, determining a primitive and predominantly effusive volcanism. In this case, volcanic activity is more frequent and commonly associated with fissure eruptions, although mafic calderas and central volcanoes may be present.

10.4 Magmatic Processes Along Convergent Plate Boundaries

Magmatic activity along convergent plate boundaries is related to the development of subduction zones, with volcanic arcs located above slabs. Subduction results from the density difference between the two converging plates, where the denser oceanic lithosphere underthrusts the lighter lithosphere and sinks into the mantle: this may occur between an oceanic and a continental plate, as well as between two oceanic plates, with the denser one subducting. Once the oceanic lithosphere penetrates the upper mantle, the body forces driven by the density contrast between the oceanic plate and the mantle (slab pull forces) take over, promoting subduction.

Seismic tomography and electrical resistivity studies across several subduction zones indicate serpentinization of the slab. Serpentinization is the result of hydrothermal circulation and low-temperature metamorphism, and produces a suite of hydrated mineral phases. As the slab sinks into the mantle, the increasing temperature and pressure drive metamorphic reactions that release these fluids (mainly H_2O and CO_2) from the crust into the overlying mantle wedge. This addition of fluids to the asthenosphere (hydration) has the

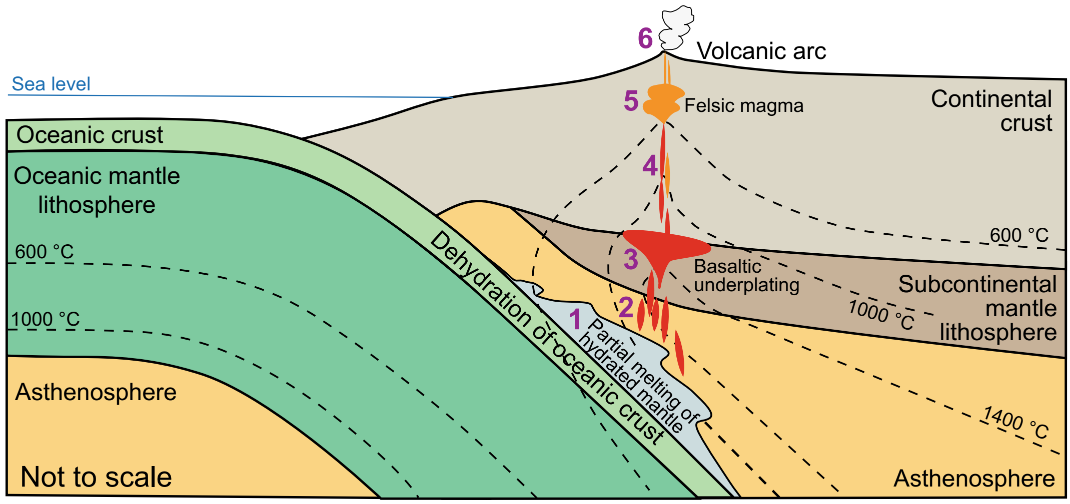

Fig. 10.11 Processes accompanying the generation and rise of magma in subducting convergent plate boundaries (section view), from volatile flux melting above the slab (1) to magma generation (2), magma underplating (3), magma rise through the crust (4), magma differentiation (5) and magma eruption (6)

effect of lowering its solidus temperature, developing a melting front above the slab and generating mafic magma. This process, termed **volatile flux melting**, typically occurs between depths of 80 and 150 km (Fig. 10.11; Turner et al. 2000; Cagnioncle et al. 2007; LaFemina 2015, and references therein). Once these melts are formed in the mantle wedge, they rise for buoyancy through the wedge and to the overriding plate. Here the magmas may accumulate at the Moho (underplating), heating the base of the crust and promoting partial melting and the production of more evolved felsic and lighter magmas, which will tend to rise. At the same time, the mafic magma may also continue to rise through dikes in the lower crust. While rising and emplacing in the crust, this magma may undergo a number of processes that lead to its differentiation towards more evolved felsic (andesitic to rhyolitic) compositions. These processes include assimilation and melting of crust, fractional crystallization, and magma mixing with more evolved magmas. In particular, the mixing of basaltic and rhyolitic magmas is the main process responsible for the formation of andesitic magma, with andesite being the most common rock type along volcanic arcs. Eventually, the magma rising in the crust reaches the surface and erupts along the arc. The location of the volcanic arc results from the location of the thermal anomaly within the wedge above the slab, which in turn is constrained by the depth of release of the volatiles from the slab, itself a function of the slab dip. Therefore, steeper slabs develop volcanic arcs nearer to the trench and shallower slabs develop volcanic arcs farther from the trench. Very shallow dipping (i.e., flat) slabs may not reach the temperature and pressure conditions to release volatiles, thus inhibiting magmatism: this is currently the case of the Peruvian portion of the Andes and the western portion of Honshu, Japan.

Volcanic arcs may consist of island or continental arcs. **Island arcs** form where the denser oceanic plate subducts beneath the lighter oceanic one (as for example the Marianas), or where the volcanic arc grows on a sliver of continental crust that has separated from a continent (as for example Japan). **Continental arcs** grow where an oceanic plate subducts beneath continental lithosphere, as the Andes or the Cascades Arc. Volcanism at island arcs formed on oceanic plates tends to produce mostly mafic and intermediate igneous rocks, whereas volcanism at continental arcs may also produce intermediate and felsic igneous rocks, including granitic batholiths. In fact, while partial melting of mantle peridotite yields mafic magma, partial melting of mafic or intermediate continental crust yields intermediate to felsic magma (e.g., LaFemina 2015, and references therein). In continental arcs, the hot mafic magma rising from the mantle, transferring heat into the continental crust and causing partial melting, may also generate large volumes of felsic magmas.

The convergence between two plates may be orthogonal or, more commonly, oblique. The amount of obliquity is proportional to the angle of obliquity ϕ between the direction of the plate convergence motion V_{con} (which derives from the sum of the velocities of the subducting and overriding plates) and the direction perpendicular to the contact between the two plates, or to the trench (Fig. 10.12).

In addition to the angle ϕ, the degree of convergence between the two plates can be expressed through the ratio between the trench parallel (V_p) and trench orthogonal (V_n) convergence velocities, both derived from the plate convergence V_{con}. All these kinematic parameters are independent of any global selected reference frame, as they are described relatively to the overriding plate (e.g., Heuret et al. 2007, and references therein). Based on the relative values of V_p and V_n, the kinematics of convergent plate boundaries is defined by **orthogonal** (where V_p is negligible with regard to V_n) or **oblique** (V_p not negligible with regard to V_n) **convergence**. Even though plate convergence V_{con} is defined and accommodated at the contact between the two plates, a minor part of this motion is commonly transferred to the front of the overriding plate, including the region of the volcanic arc. Indeed, the overall

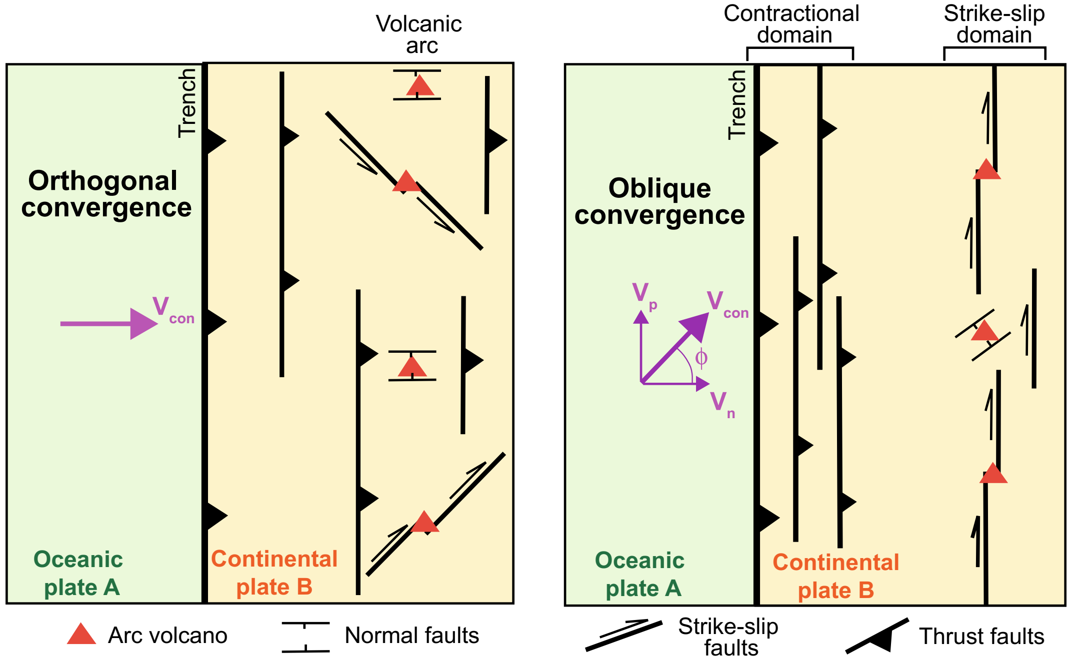

Fig. 10.12 Orthogonal and oblique convergence along portions of convergent plate boundaries experiencing subduction (map view). In orthogonal convergence, the motion resulting from both plates defines the convergence V_{con}, which is orthogonal to the contact between the plates, or trench. In oblique convergence the motion resulting from both plates V_{con} makes an angle ϕ (angle of obliquity) with regard to the direction orthogonal to the trench, defining a trench parallel V_p and trench orthogonal V_n component of convergence. Orthogonal and oblique convergence result in distinct structural styles on the overriding plate: in particular, the oblique motion is partitioned between predominant contraction on the accretionary prism and predominant strike-slip motion along the magmatic arc (see text for details)

motion of the volcanic arc usually reflects that deriving from the convergence between the two plates. If the convergence is accommodated through a predominant orthogonal component, this is present also in the arc region, as for example observed in Northeast Honshu (Japan). This arc-perpendicular motion may be characterized by extension or contraction, as a function of the variation (decrease or increase, respectively) in the trench-normal convergence velocity (e.g., Acocella and Funiciello 2010). Conversely, in presence of an important oblique convergence component, the volcanic arc may experience a predominant trench parallel motion, as for example observed at Sumatra (Indonesia). This behaviour is interpreted to result from a process of strain partitioning (Fig. 10.12). **Strain partitioning** here refers to the separation of the two components of the overall oblique motion of convergence across the portion of the overriding plate between the trench and volcanic arc. As a result, the trench portion of the overriding plate accommodates predominant arc-perpendicular motion, mainly developing thrust faults, whereas the arc portion mainly accommodates the arc-parallel motion, developing strike-slip systems consistent with the left- or right-lateral component of convergence between the plates. Depending on the obliquity angle and the convergence velocity V_{con}, the strike-slip faults along the arc may also display a contractional or extensional component, creating transpression or transtension. All these thrust and strike-slip fault systems are parallel to the trench (McCaffrey 1992, 1996). The spatial coincidence in the location of the volcanic arc and the zone of strike-slip faulting likely results from magma-induced thermal weakening, which focuses the deformation in the upper plate. A possible consequence of oblique convergence and strain partitioning is the creation of a microplate, or **sliver** in the forearc, with arc-parallel motion. The forearc side of the approaching sliver experiences contraction, whereas the forearc side of the departing sliver undergoes extension (Fig. 10.13; DeMets 1992; e.g., Acocella et al. 2018). More in general, available structural data show that, depending on the convergence values and their arc-normal and arc-parallel components, any type of motion may be found along volcanic arcs, including extension, contraction, oblique (transtension or transpression) and strike-slip motions. Therefore, in addition to the overall plate motion, a crucial parameter to consider for understanding volcanic processes along arcs is the overall arc kinematics (Acocella and Funiciello 2010).

The arc-perpendicular amount of convergence may be used to define two ideal and end-member behaviours of subducting plate boundaries (e.g., van der Pluijm and Marshak 2004). In a **coupled**

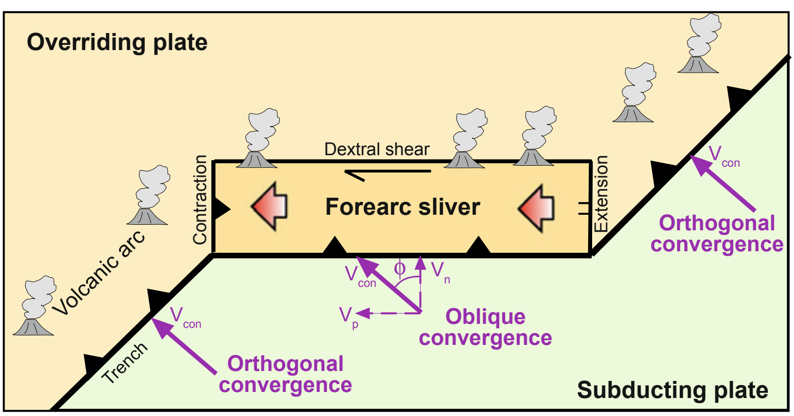

Fig. 10.13 Tectonic features associated with a portion of convergent plate boundary with strong obliquity (map view): in this case the forearc may be decoupled with regard to the subducting plate and a forearc sliver (in orange), characterized by overall arc-parallel motion, may develop (modified after DeMets 1992). A similar situation may apply to the southern Kuril Arc

convergent margin, the downgoing plate pushes tightly against the overriding plate, so the plate boundary overall is under compression (Fig. 10.14a). As a consequence, significant stresses develop across the contact, triggering large magnitude earthquakes; even the back-arc region may experience contraction. Because of the compression, the rise of magma is slower and its storage is not shallow. Therefore, magma has time to fractionate and/or cause partial melting of adjacent continental crust before intruding at shallower depth or erupting. Since partial melting of continental crust produces intermediate to felsic magma, and since fractionation removes mafic minerals from a melt, intermediate to felsic igneous rocks predominate at coupled convergent margins. The difficulty in the rise and shallow emplacement of magma may also result in lower eruptive rates. In an **uncoupled convergent margin**, the slab does not push tightly against the overriding plate, so compression across the margin is limited (Fig. 10.14a). As a consequence, the stress coupling across the plate boundary is smaller and earthquakes have smaller magnitude; the back-arc region may even experience extension. Because of the lower compression within the volcanic arc, mantle-derived magmas may rise more easily to the surface and intrude at shallower levels before any significant fractionation or crustal contamination; thus, mafic rocks are more common. Also, the enhanced conditions for the rise and shallow storage of magma may result in higher eruptive rates (Acocella and Funiciello 2010; Chaussard and Amelung 2012). While Northeast Japan and the Taupo Volcanic Zone of New Zealand may resemble coupled and uncoupled convergent margins, respectively, these two types of margins should be better seen as simplified end-members, as there is no textbook example of arc showing a purely coupled or uncoupled behaviour.

A further source of complexity in the structure of the arc, not necessarily depending from plate coupling, derives from the presence of extension

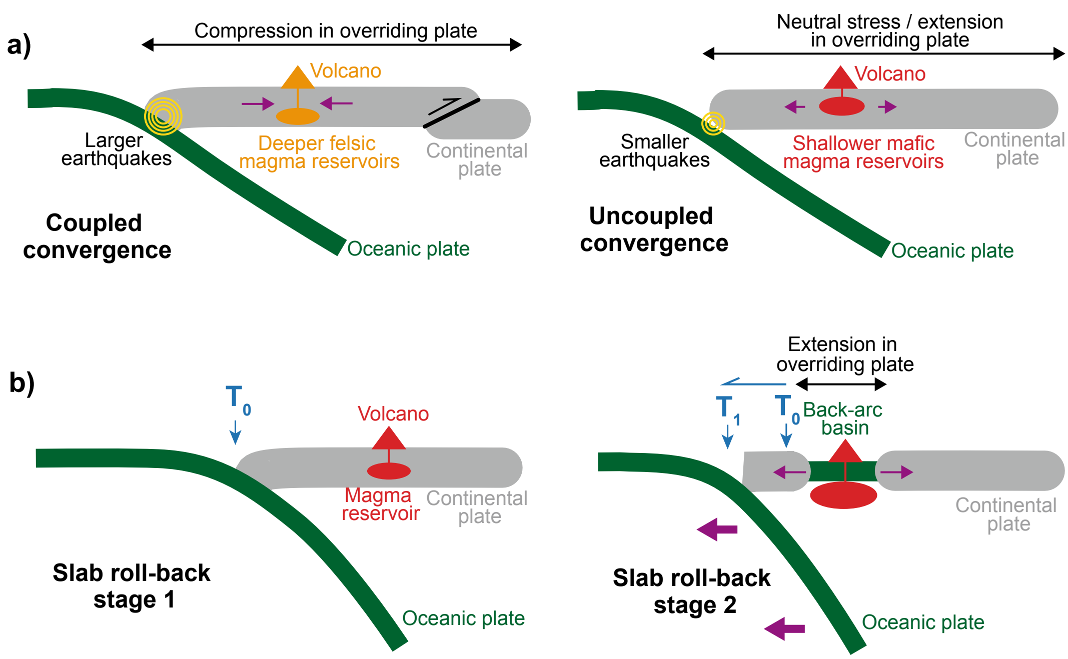

Fig. 10.14 **a** Main features of a coupled (left) and uncoupled (right) convergent plate boundary, illustrating the seismicity along the plate boundary, the state of stress within the overriding plate and depth and composition of magma reservoirs (section view). **b** Two progressive stages of slab roll-back, with the trench moving backward from position T_0 to T_1, determining the opening of a back-arc basin with oceanic crust and widespread mafic volcanism

in the arc associated with a wider area of back-arc opening, which develops a back-arc basin. **Back-arc opening** results from the differential kinematics of the trench, the subducting plate and the overriding plate. This condition is often related to the **roll-back** of the negatively buoyant subducting lithosphere, which occurs when the slab migrates oceanward, inducing the oceanward migration of the trench (Fig. 10.14b; e.g., Uyeda and Kanamori 1979; Sdrolias and Muller 2006). During the development of a back-arc basin, extension of the overriding plate enhances the decompression melting of the asthenosphere and, in turn, magma generation. Slab roll-back is thus responsible for an extensional back-arc basin with significantly thinned crust, which may also experience oceanization, as observed for example in the Tyrrhenian Sea (Italy), the Japan Sea and the Havre Trough-Lau Basin in the southwest Pacific.

In general, extension along a volcanic arc is the ultimate requisite for significant volcanic activity, so that the higher the amount of extension, the higher is the arc output rate. The extension along the arc may be regional, as along extensional arcs, or local, as in arcs where more limited extension may be related to the activity of strike-slip structures in releasing bends, pull-apart structures, dilational jogs or horsetail structures along principal displacement zones. Contractional conditions may also accompany volcanism, as testified by volcanic arcs undergoing predominant shortening. Different mechanisms may be invoked to explain the emplacement and rise of magma in arcs experiencing shortening. For example, contraction promotes the emplacement and accumulation of magma, as through sills feeding stacked systems or tabular intrusions (as laccoliths, lopoliths) at different crustal levels. Also, stacked magma may locally increase the vertical component of the principal stress, which may pass from the least component σ_3 to the intermediate component σ_2 and finally to the maximum component σ_1, promoting strike-slip and extensional structures, respectively. The latter may also enhance local arc-parallel extension. In addition, transient variations in the stress field induced by major regional earthquakes (or mega-earthquakes, with magnitude $M > 8.5$) may induce temporary extension in arcs otherwise experiencing contraction or strike-slip motions, also encouraging volcanic activity (see Chap. 12). Moreover, the activity of strike-slip or transtensive fault systems related to variations in the amount and direction of shortening in contractional arcs may locally decrease the minimum principal stress σ_3, increasing the deviatoric stress and further encouraging the rise of magma through dikes (Fig. 10.12; Tibaldi 2005; Galland et al. 2007; Acocella et al. 2008; Shabanian et al. 2012).

In some cases, the location and/or relevant activity of large volcanoes may be not directly related to the presence of a slab in a convergent setting, as inferred for example for the volcanism occurring along the northern Tonga Arc (southwest Pacific Ocean), Mount Etna and the cluster of large volcanoes within the Kamchatka depression (Russia), which includes the imposing Klyuchevskoy and Sheveluch volcanoes. In fact, the location of these volcanoes and, for the Etna and Kamchatka cases, also the high erupted volumes have been related to their peripheral position at the slab edge, a condition that promotes the suction of asthenospheric material from under the neighbouring plate to the side. Such a lateral flow is expected when descending slabs migrate backwards in the mantle (roll-back), leaving low-pressure regions to the front. In this case, the asthenospheric return flow from the back to the front position at the edge of the slab induces mantle upwelling, eventually driving decompression melting and promoting volcanism. In particular, the generation of magma at Mount Etna, in a clear off-arc location, is largely explained by this process (Fig. 10.15; Wendt et al. 1997; Gvirtzman and Nur 1999; Peyton et al. 2001; Faccenna et al. 2011).

Magma may be also produced through different processes in non-subducting collisional orogens. When subduction ceases and continental collision takes over, slab breakoff may promote magma generation. **Slab breakoff** consists of the tearing of the subducting plate due to the buoyancy-driven detachment of the denser subducted oceanic lithosphere from the lighter continental

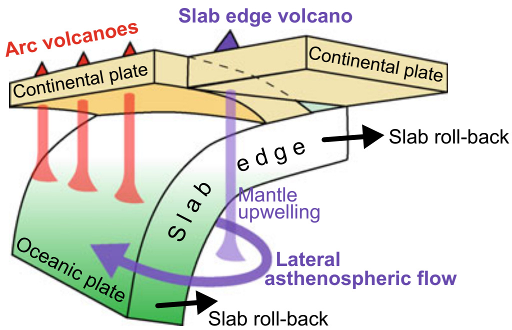

Fig. 10.15 Scheme of asthenospheric window (in purple) formed at the lateral edge of a slab experiencing roll-back, oblique view. The lateral asthenospheric flow induces local mantle upwelling at the slab edge; the asthenospheric material penetrates the plate boundary and reaches the crust in the forearc region, generating volcanism above the slab edge

lithosphere that follows it during continental collision (Fig. 10.16; e.g., Davies and von Blanckenburg 1995). Breakoff leads to heating of the overriding lithospheric mantle by the upwelling asthenosphere intruded within the tear and also by decompression due to the removal of the slab. This thermal perturbation induces the melting of the metasomatised overriding mantle lithosphere, producing basaltic magmatism that leads to granitic magmatism in the crust (bimodal magmatism). Slab breakoff has been invoked to explain the existence of anomalous volcanism within collisional and subduction domains, as in Eastern Anatolia, Central Italy and at Toba (Sumatra; Keskin 2003; Rosenbaum et al. 2008; Hall and Spakman 2015; Koulakov et al. 2016).

Melting and thus volcanism may also be caused by **lithosphere delamination**, when the deep keel of lithosphere that develops during thickening drops off (Fig. 10.17; e.g., van der Pluijm and Marshak 2004; Fossen 2010). Warm asthenosphere rises to take its place and heats the remaining lithosphere, possibly partially melting a broad portion of the mid crust. Lithospheric delamination has been invoked to explain the presence of the isolated but imposing Damavand volcano, 5600 m high, within the Alborz collisional orogen of Iran. At a different scale, piecemeal lithospheric delamination has been also suggested to explain part of the impressive

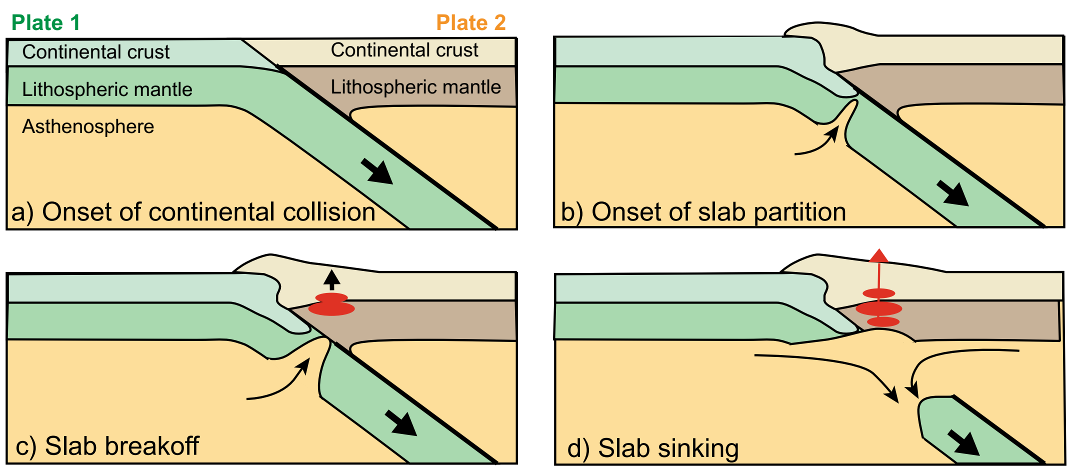

Fig. 10.16 Schematic slab breakoff process, section view. **a** Continental collision begins when all the oceanic lithosphere of plate 1 has already sunk and the continental lithosphere of both plates comes into contact. **b** The density difference between the heavier oceanic slab at depth and the lighter continental lithosphere at the surface determines the partitioning of the slab. **c** The slab undergoes breakoff; the impingement of hot asthenosphere at the base of metasomatised continental lithosphere leads to magmatism (in red) and the release of the load leads to uplift. **d** As the slab sinks away, a deep return flow develops above, maintaining a high temperature at the base of the lithosphere, promoting magmatism (modified after Davies and von Blanckenburg 1995)

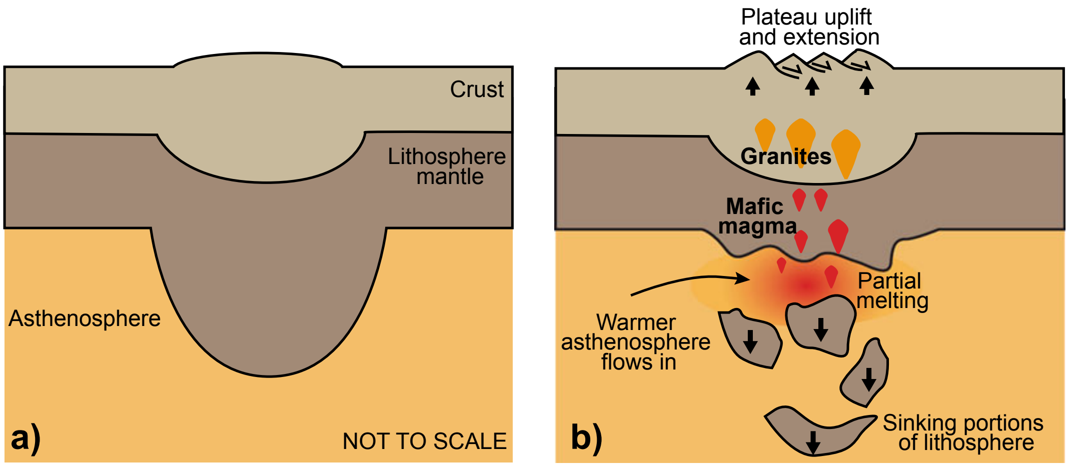

Fig. 10.17 Lithospheric delamination, section view. **a** Thickening of lithosphere forms a keel-shaped mass of cool lithosphere to protrude down into the asthenosphere. **b** The keel drops off and is replaced by warm asthenosphere, causing partial melting and formation of post-orogenic plutons. As a consequence, the crust may uplift (modified after van der Pluijm and Marshak 2004)

volcanism of the Central Andes, in a subduction setting (Kay and Kay 1993; Beck and Zandt 2002; Schurr et al. 2006; Liotard et al. 2008; Shabanian et al. 2012). Lithosphere delamination may also explain the anomalous uplift of portions of both subducting and collisional convergent boundaries, as the Central Andean and the Tibetan plateaus, respectively. In fact, the hot asthenosphere at the base of the remaining lithosphere would cause the lithosphere to heat up. Thus, to maintain isostatic equilibrium, the surface of the lithosphere would have to rise. In other cases, very thick orogens collapse under their own weight, spreading laterally (**orogenic collapse**). Because the extension associated with the collapse thins the upper crust, it causes decompression, which may promote partial melting of the deep crust and the underlying asthenosphere, producing magmas. Magma may also form and intrude the upper crust after the collision has ceased: in this case, the magma is **post-orogenic**.

Volcanic arcs mainly consist of andesitic stratovolcanoes, with subordinate but large rhyolitic caldera systems. The latter constitute the largest volcanic systems on Earth and are often associated with specific tectonic conditions promoting deep partial melting and inducing the pervasive rise of asthenospheric material, as explained in the following chapters. Arc volcanoes do not show frequent unrest, volcanic activity and open conduits, as magma is usually stored at higher depth with regard to divergent plate boundaries and occasionally reaches the surface to erupt (e.g., Chaussard and Amelung 2012). As a result of a moderately to highly viscous magma experiencing infrequent eruptions, explosive volcanism predominates along volcanic arcs. The volcanic edifices often show evidence of the regional convergence conditions. In fact, many volcanic edifices are elongated parallel to the direction of the maximum principal stress σ_1. In addition, several arc volcanoes show a local alignment parallel to the σ_1 direction. These features, as for example observed along the arcs of the Andes and Japan, result from the preferred development of feeder dikes parallel to the regional σ_1 direction, determining the elongation and alignment of the volcanic edifices (Nakamura 1977; Nakamura et al. 1977).

10.5 Magmatic Processes at Hot Spots

Hot spots are clusters of volcanoes whose origin cannot be easily reconciled with the activity of plate boundaries, in terms of location and distribution, or composition and flux of erupted magma. While most hot spots are located in intraplate settings, many are also located along plate boundaries, most notably divergent boundaries, affecting their geometry and the composition and volume of the erupted products (Fig. 10.18).

Hot spots are commonly considered as the surface expression of mantle plumes. **Mantle plumes** may originate at different depths within the mantle and consist of the localized rise of material hotter and less dense than the surrounding mantle rocks. Mantle plumes play a key role in transferring heat from the mantle to the lithosphere, also locally influencing the geometry of the plate boundaries and thus plate tectonics. Hot spot volcanism is extremely long-lived and may last for a hundred millions of years, or more. This is indicated, for example, by the age progression of volcanic chains formed during the hot spot's lifetime, such as the more than 100 Ma old Hawaiian-Emperor Chain in the Pacific Ocean.

The total number of hot spots is estimated to range between 45 and 70. Hot spots have irregular but non-random distribution over the Earth's surface. They primarily lie in oceans, often near oceanic ridges and are rare near subduction zones (Courtillot et al. 2003; Zhao 2004). This irregular distribution may result from several factors. First, as oceans cover most of the Earth's surface, most hot spots are expected to lie in oceans. Second, mantle plumes can weaken the continental

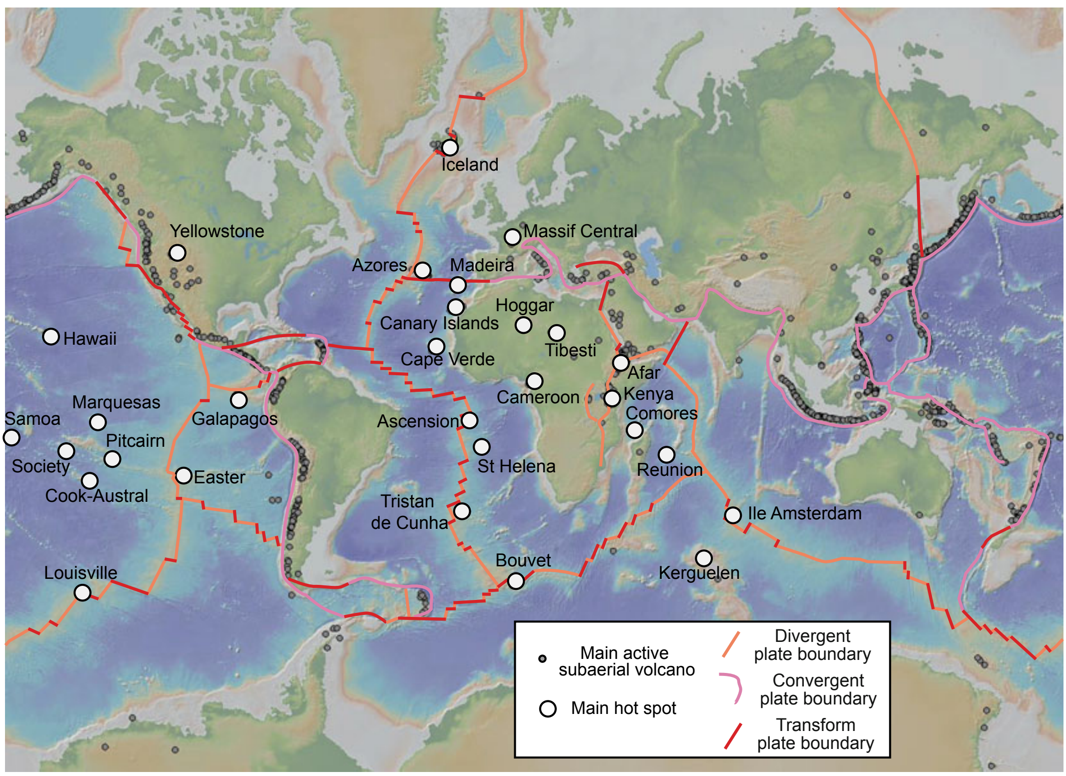

Fig. 10.18 Schematic distribution of the major hot spots in the frame of plate tectonics, with divergent (orange lines), convergent (purple lines) and transform (red lines) plate boundaries and major Holocene volcanoes. Base DEM provided by https://www.geomapapp.org

lithosphere and enhance its break-up, promoting the activity of divergent plate boundaries ultimately evolving in oceanic domains. Therefore, although many hot spots may have originated on continents, they currently appear on oceans. Third, the difference in thickness (100–150 km) between the oceanic and continental lithosphere may control partial melting. While weak plumes may not reach the conditions for melt generation beneath the thicker continental lithosphere, they may do so beneath a thinner oceanic lithosphere, at lower pressure.

The best-defined hot spots appear relatively stationary over time, as their respective motions (≤ 1 cm/year) are lower than the plate velocities. This feature, together with the long-lived hot spot volcanism, has been used to exploit hot spots as a near-fixed reference frame to determine absolute plate motions (Courtillot et al. 2003). Typically, the direction of plate motion is given by the direction of migration of the aligned hot spot volcanoes; in some cases, as in the Hawaiian-Emperor Chain, the alignment of volcanoes shows major changes in direction, indicating important variations in plate motion. The average rate of plate motion is given by the distance of a certain volcano from the active site of the hot spot with regard to the oldest age of activity of the volcano.

Plume-like slow seismic velocity anomalies are tomographically detected under major hot spot regions, as Hawaii, Iceland, South Pacific, and Africa. The slow anomalies under South Pacific and East Africa extend more than 1000 km, representing two **superplumes** in opposite locations of the Earth's mantle. The slow anomalies under hot spots normally do not show a straight pillar shape, but exhibit winding images, suggesting that even plumes can be deflected by mantle flow (Fig. 10.19; McNutt and Judge 1990; Romanowicz and Gung 2002; Zhao 2004; O'Connor et al. 2019).

Data acquired in the last decades have improved the understanding of the depth of

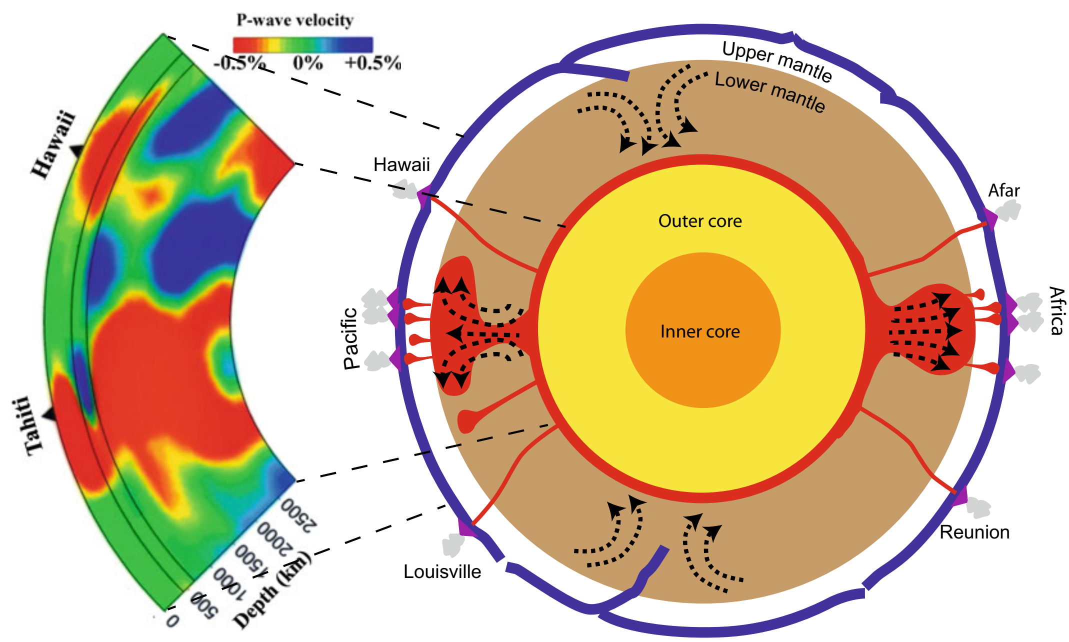

Fig. 10.19 Mantle plumes feeding hot spots. Right: schematic cross section of the Earth, outlining the sources of the different types of plumes/hot spots and the two antipodal superplumes upwelling below the South Pacific and East Africa (Courtillot et al. 2003). Left: north–south cross-section of P-wave velocity image along a profile passing through Hawaii and the South Pacific superswell; red and blue colours denote slow and fast velocities, respectively (Courtesy of Dapeng Zhao; Zhao 2004)

generation and shape of mantle plumes. The mantle plumes are thought to be generated by Rayleigh–Taylor instabilities from **thermal boundary layers**, or zones with higher temperature and viscosity gradient within the mantle. The most important of these layers, named *D″*, lies at the core-mantle boundary. Seven primary hot spots, namely Hawaii, Easter, Louisville, Iceland, Afar, Reunion, and Tristan, were initially inferred to have such a deep origin: all these hot spots, except Iceland, are close to or at the margins of the two superplumes beneath East Africa and South Pacific. The remaining, non-primary hot spots may form in the transition zone at the bottom of the upper mantle or at shallower levels, and then be linked to lithospheric break-up (Courtillot et al. 2003). Also, fluid dynamics studies have suggested narrower cylindrical conduits of radius of 50–150 km feeding large spherical heads, requiring important viscosity variations in the thermal boundary layer, with temperature differences between plumes and the surrounding mantle of a few hundred degrees. These large heads would reflect the dynamical requirements for a plume to ascend into a higher-viscosity environment, whereas the narrow trailing conduits would form because the low-viscosity fluid within the conduit rises faster than the plume head itself (Jellinek and Manga 2004; Farnetani and Hofmann 2011). More recent tomographic data support the model of multiple levels for the generation of mantle plumes, and suggest that at least nine plumes originate near the core-mantle boundary, whereas eight or more originate at the base of the upper mantle. This implies at least two convecting regions, possibly linked in a complex way. To be tomographically resolved, the plumes must be several hundreds of kilometres wide, conveying a substantial fraction of the internal heat. The imaged plumes show thick and subvertical conduits beneath many prominent hot spots, extending from the core–mantle boundary to about 1000 km below Earth's surface. These conduits are thus much broader than classical thermal plume tails, suggesting that they can carry more heat away from the core than it was previously thought (Montelli et al. 2004; Zhao 2004; French and Romanowicz 2015; Hand 2015).

The longevity and structure of the mantle plumes result from interactions between core cooling, dense low-viscosity material at the thermal boundary layer and plate tectonics. As anticipated, plumes may be affected by the flow regimes in the upper mantle. Weak plumes may be sufficiently deflected by large-scale mantle flow driven by subduction and plate tectonics, so that they may break up into diapirs (Jellinek and Manga 2004; Kerr and Meriaux 2004). Also, a mantle plume head can separate from its trailing conduit upon passing the interface between the upper and the lower mantle. The detached plume head may eventually trigger a first volcanic (usually flood basalt, see below) event at the surface. The remaining conduit forms a new plume head, which may cause a second flood basalt event at least 10 Ma after the first one. This interaction between the plume and the mantle discontinuities may explain distinct episodes of major eruptions, commonly separated by between 20 and 90 Ma within several of the world's flood basalt provinces (Bercovici and Mahoney 1994).

At shallower levels, the interaction between ascending plumes and the lithosphere includes, but it is not restricted to, the following aspects (Farnetani and Hofmann 2011).

(a) The control of the thickness, elasticity and age of the lithosphere on the spatial distribution of plume magmatism (Ebinger and Sleep 1998; Hieronymus and Bercovici 2001). As regards the lithosphere thickness, lateral flow from a plume impinged beneath a relatively thin lithosphere may be channelled along pre-existing rift-zones, explaining a wider and selective distribution of magmatism at the surface, as for example observed above the Afar plume in East Africa. As regards the elasticity of the lithosphere, the interaction of magma transport with lithospheric flexure may affect the spacing of volcanoes in hot spot chains through the generation of flexure-induced tensile stresses that determine where magma preferentially extrudes. In turn, plumes may also affect the structure and mechanical properties of the lithosphere, especially through magmatic underplating. As regards the age of the lithosphere, this is a main feature controlling the density of oceanic plates, in turn affecting the topography of hot spot tracks. In fact, the prolonged hot spot activity over a relatively fast moving plate develops a volcanic chain, or hot spot track, parallel to the plate motion direction. Topographically smooth and broad hot spot tracks, as the Galapagos, lie on young and lighter lithosphere, which favours magma intrusion. Conversely, rough and discontinuous hot spot tracks (as Hawaii) lie on older, thicker and denser lithosphere, which encourages volcanism (Orellana-Rovirosa and Richards 2017).

(b) A dynamic topography induced by the arrival of a plume head, commonly testified by a ~1000 km wide and ~1 km anomalously high topographic swell. Usually, the continental swells are much smaller than the oceanic ones. Hot spot swells have been used to estimate the flux of material of the mantle plumes (i.e., buoyancy flux) and this flux of material has been in turn used to estimate the amount of heat transported to the surface. The **buoyancy flux** of hot spots is a measure of the strength of the plume, which depends upon the buoyancy driven volume flux. The buoyancy flux B is defined as:

$$B = L H_e (\rho_{ma} - \rho_w) V_h \qquad (10.2)$$

where L is the width of the swell, H_e is the excess elevation averaged across the swell, ρ_{ma} is the mantle density, ρ_w is the water density (in the case of oceanic hot spots) and V_h is the plate velocity in the hot spot frame. The buoyancy flux, measured in Mg/s, ranges over a factor of 20, Hawaii being the largest; generally, the largest buoyancy fluxes are found in the Pacific region (e.g., Sleep 1990; King and Adam 2014). The topographic swell induced by the hot spot implies a rapid surface uplift, which may be often followed by flood volcanism by 1–2 Ma. Subsequently, there is a slower subsidence and, eventually (if the plume is below a continent), continental break-up, as in Afar (Farnetani and Richards 1994). The buoyancy flux allows quantifying hot spot activity, although its use becomes irrelevant over long periods, due to the own subsidence of the hot spot swell. In this case, to quantify hot spot activity it may be more relevant to determine the production rate of volcanic material, although this refers only to the surface manifestation of the hot spot (Vidal and Bonneville 2004).

(c) Hot spot-ridge interaction(s): there are strong feedbacks between the dynamics of slowly migrating ridges and deeply sourced plumes, also able to explain the repeated formation of large igneous provinces (see point d below). At least 21 hot spots are situated near a spreading ridge, with evident topographic, thermal and crustal thickness variations. The Oceanic Island Basalts (OIBs) produced by the hot spot are geochemically distinct from Mid-Ocean Ridge Basalts (MORBs). This suggests that the lower mantle, the inferred source of plumes, is compositionally different from the upper mantle, source of MORBs (Ito et al. 2003; Farnetani and Hofmann 2011; Whittaker et al. 2015).

(d) The onset of massive hot spot magmatism, often associated with a large igneous province, which results from the melting of the plume head, while the subsequent hot spot activity is associated with the long-lasting and narrower plume tail, as for example observed with the Yellowstone hot spot (see Sect. 13.8).

In particular, **Large Igneous Provinces**, or LIPs, are exceptional volcanic events responsible for a large total volume of dominantly mafic magma (up to millions of km^3) erupted over a brief period (usually < 1 Ma). Furthermore, the volume of magma erupted during each of the individual eruptions that make up a LIP (frequently 10^3–10^4 km^3) is also exceptional. Without LIP-forming igneous events, basaltic super-eruptions would not have occurred, nor indeed many of Earth's largest volcanic deposits: for example, the largest continental flood basalts province, the 252 Ma old Siberian Traps, might have totalled 4–5 million km^3 of lava. These large erupted volumes and the relatively frequent eruptions lead to the geologically rapid construction of extensive lava plateaus, up to several kilometres thick. Large igneous provinces are found on continental crust, in ocean basins and in transitional domains, and their rocks are distinguished from the products of other types of magmatism on the basis of petrologic, geochemical, geochronological, geophysical and volcanological data. LIPs come in two broad compositional varieties: basaltic LIPs, which form flood basalt provinces and are the most frequent, and silicic LIPs, which are few in number and form some of the major ignimbrite provinces (Fig. 10.20; Coffin and Eldholm 1994; Bryan et al. 2010; Self et al. 2015).

Owing to their surface exposure, continental flood basalts are the most intensively studied large igneous provinces. These are dominated by tholeiitic mafic lavas most likely originated in less than 100,000 years from prolonged eruptions, each lasting years to decades, with the largest flow fields fed by long fissures. A short-lived and isolated recent example of such fissure activity may be the 1783–84 Lakagigar eruption (Iceland). This, erupting 14 km^3 of basaltic magma in 8 months, provides the largest historical lava flow eruption. The voluminous lava flow-fields associated with large igneous provinces do not require rapid extrusion of mafic magma at rates much higher than the 1783–84 Lakagigar type of eruptions, nor do long fissures imply high eruption rates, as only single segments of the LIPs fissures were likely active at any one time. Taking the 1783–84 Lakagigar fissure eruption as a reference, the time required to emplace a large 1000–2000 km^3 lava flow at the range of its peak output rates would be approximately 10–20 years.

Large igneous provinces are not uniformly distributed over the Earth's history. Statistical studies suggest periodicities in LIP production at approximately 30–35 and 60 million years (Self et al. 2015). Such episodicity probably reflects variations in rates of mantle circulation. Large igneous provinces repeatedly formed at locations where oceanic ridges and plumes interact. In particular, slowly migrating ridge systems that have been stabilized by upwelling plumes have extracted large volumes of material from the same part of the upper mantle over periods up to hundreds of million years. This indicates a feedback between the dynamics of slowly migrating ridges and deeply sourced plumes (Whittaker et al. 2015).

The most common hypothesis to explain large igneous provinces is that these are generated when a rising mantle plume impacts the lithosphere, although alternative hypotheses also exist (Campbell 2005). As in the case of hot spot volcanoes, large igneous provinces are commonly attributed to decompression melting of hot and low-density mantle material ascending in mantle plumes. These plumes would initially transfer huge volumes (10^5–10^7 km^3) of mafic rock into localized regions of the crust over short intervals (10^5–10^6 years). This large-volume magmatism during LIP formation is commonly attributed to mantle plume heads reaching the crust, whereas the persistent (hot spot) magmatism is considered to result from mantle plume tails penetrating the lithosphere, which may be moving relative to the plume. Modelling of the impingement of a plume head below the lithosphere shows that the large head spreads out to form a thermal dome with the potential to form basaltic eruptions by decompression melting over an area that may be as much as 2000 km in diameter (Campbell 2005). The unusually large melt fractions formed in the mantle by this process cause the high extrusion rates that typify flood basalt provinces. As lithospheric plates move across a plume, the plume head disperses, leaving only the plume tail erupting basaltic magma at a much lower rate. These tails create chains of volcanic islands that terminate in active volcanoes marking the current position of the mantle plume, as at Hawaii or La Reunion.

The fact that not all large igneous provinces have obvious connections with mantle plumes, or even hot spot tracks, suggests that more than one

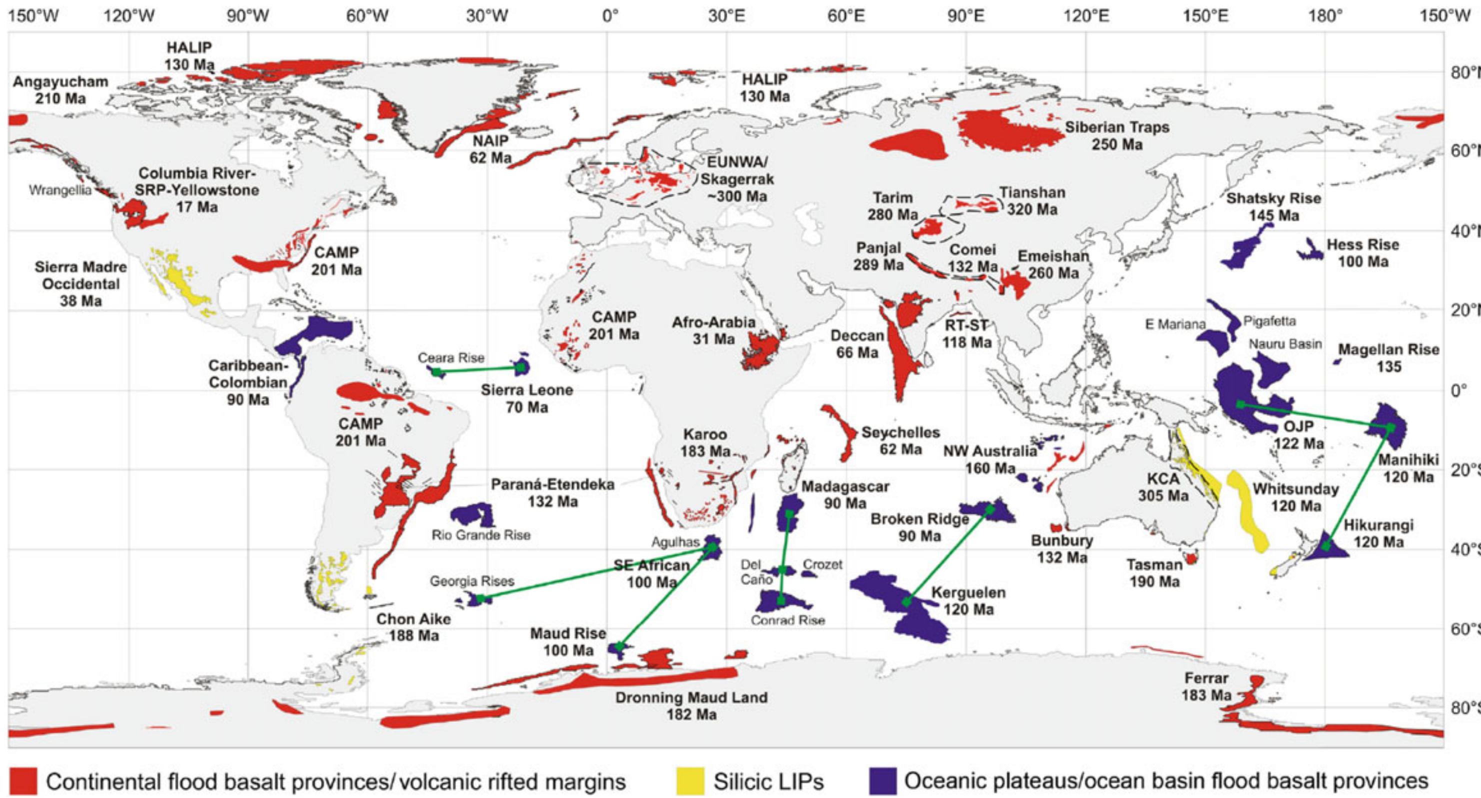

Fig. 10.20 Large igneous provinces (LIPs) of the world, including continental flood basalt provinces, oceanic plateaus, and silicic LIPs. Ages are those of main phase or pulse of volcanism. Inferred extent of some provinces shown by dashed lines; tie-lines join parts of same original province (Self et al. 2015)

source model is required for their formation (e.g., Self et al. 2015, and references therein).

10.6 Polygenic and Monogenic Volcanism

In any of the above-mentioned tectonic settings, magmatism may manifest at the Earth's surface through polygenic and/or monogenic volcanoes. Polygenic volcanoes consist of larger and long-lived volcanic edifices growing from repeated eruptions, whereas monogenic volcanoes consist of smaller volcanic edifices resulting from a single eruptive event (Fig. 10.21a, b).

The local relationships between polygenic and monogenic volcanoes, including the conditions controlling the development of satellite monogenic volcanoes on the flanks of polygenic volcanoes, have been broadly discussed throughout Chap. 7. Here the focus is mainly on the regional context controlling the development of polygenic and/or monogenic volcanoes, including the monogenic volcanoes formed outside and independently of polygenic volcanoes. These volcanoes produce widespread **monogenic fields**, consisting of a set of vents, each of which has erupted once or only a few times, and whose number varies from tens to several hundreds. The composition of these monogenic fields is usually mafic (as for example the Auckland Volcanic Field, New Zealand, the Chaine des Puy, France, and Al-Madinah, Saudi Arabia), although felsic monogenic vents, often related to lava domes, are also found (as in the Central Andes; e.g., Le Corvec et al. 2013; de Silva and Lindsay 2015). While it is usually believed that monogenic volcanoes in a field erupt independently from each other, with only the nearest volcanoes being possibly related to a same eruptive fissure or event, there is also repeated evidence, as in the Auckland Volcanic Field, of contemporaneous but separated and structurally unrelated volcanoes. This suggests that spatial and temporal

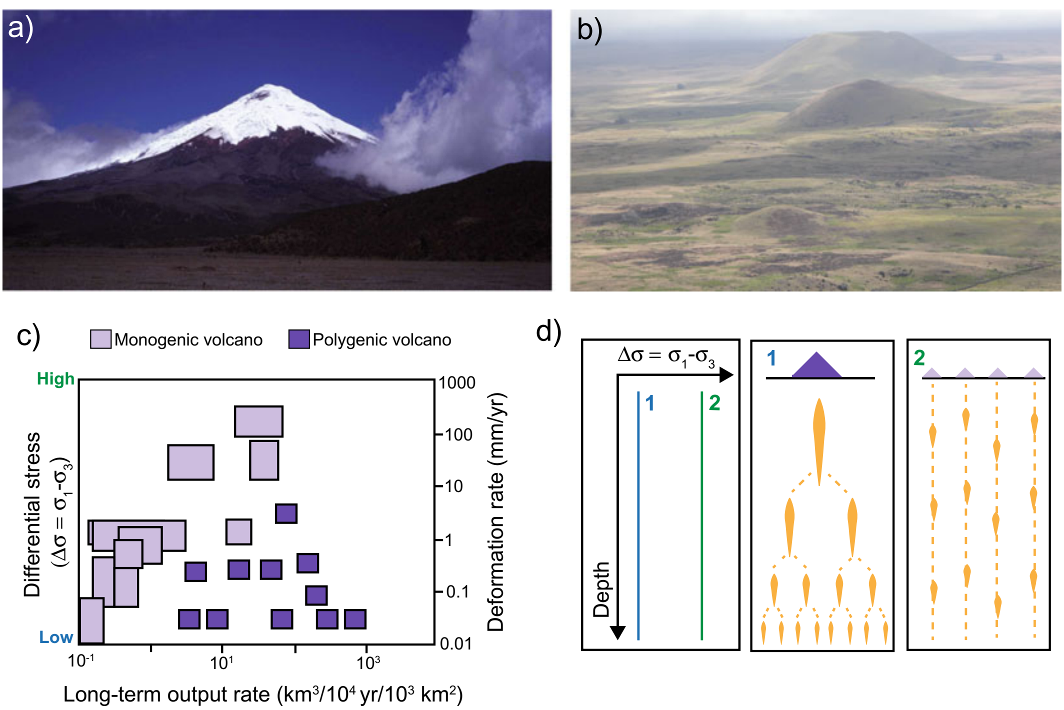

Fig. 10.21 Polygenic and monogenic volcanism. **a** View of the polygenic Cotopaxi volcano, Ecuador. **b** Aligned monogenic volcanoes at Rapa Nui (Easter Island, Chile). **c** Stress and magma flux conditions promoting polygenic or monogenic volcanism and **d** related theoretical model based on the crack interaction mechanism (Takada 1994)

heterogeneities can occur in monogenic fields, with important implications for statistically based hazard assessments (Cassidy and Locke 2010). The development of monogenic volcanic fields is mainly controlled by the regional tectonic setting: although found in any tectonic environment, most monogenic fields are in fact related to extensional regimes.

More in general, the regional tectonic setting controls the occurrence and the distribution, in terms of spacing and alignment, of polygenic or monogenic volcanoes. In fact, the possibility to develop polygenic or monogenic volcanoes has been inferred to depend upon the differential stress $\varDelta\sigma$, that is the difference between the maximum and minimum principal stresses σ_1 and σ_3, respectively (with $\varDelta\sigma = \sigma_1 - \sigma_3$). In particular, monogenic volcanoes focus in settings with higher differential stress, whereas polygenic volcanoes focus in settings with lower differential stress (Fig. 10.21c, d; Takada 1994). This follows from the fact that a higher differential stress is associated with a minimum principal stress σ_3 that is much smaller than the maximum principal stress σ_1, a condition that promotes the propagation of independent and non-coalescing dike paths, preferably feeding a monogenic volcanic field. Conversely, a lower differential stress results from a σ_3 approaching σ_1 in magnitude, a condition that promotes the coalescence of multiple dikes and magmatic sheets that tap magma over a wider area below a polygenic volcano. The possibility to develop polygenic or monogenic volcanoes depends, in addition to the tectonic setting, on magma availability (Takada 1994). In fact, a higher availability of magma, or magma flux, is more likely to result in the build up of a large and long-lived magma chamber that is usually required to feed a polygenic volcano. These two conditions suggest that monogenic volcanoes result from lower magma fluxes mainly in extensional settings and polygenic volcanoes result from higher magma fluxes mainly in neutral (neither compressive nor extensional) tectonic settings. This explains why monogenic volcanic fields are more frequent along divergent plate boundaries, although a same tectonic setting may show both polygenic and monogenic volcanoes, as observed on divergent (as the East African Rift System and the oceanic ridge of Iceland) and convergent plate boundaries (as the Trans Mexican Volcanic Belt and the Central Andes). Note that, while monogenic volcanism is commonly fed by dikes, there is also evidence, as for example at the Hopi Buttes Volcanic Field (Arizona, USA), of saucer-shaped sills feeding a significant amount of monogenic volcanoes. This indicates that shallow feeders in monogenic fields can form geometrically complex networks, particularly those intruding poorly consolidated sedimentary rocks (Muirhead et al. 2016).

The distribution of polygenic and monogenic volcanoes, as expressed through their spacing and alignment, also depends upon the tectonic context. The distance or **spacing** between nearby polygenic volcanoes may be related to several first-order factors. In general, the thinner the crust or the lithosphere, the more closely spaced are the volcanoes, as observed in the Main Ethiopian Rift—Afar area (Fig. 10.22a; Mohr and Wood 1976). This behaviour, supported by analogue models, is described by the development of Rayleigh–Taylor instabilities, where the wavelength associated with a dominant disturbance is proportional to the thickness of the upper layer, indicating that the thicker is the upper layer the more distant are the instabilities (see Sect. 3.2.1; Fig. 10.22b; e.g., Ramberg 1981; e.g., Turcotte and Schubert 1982). In a similar fashion to polygenic volcanoes, the distribution of the spacing of monogenic vents has been related to the thickness of the crust below various volcanic fields in different extensional tectonic settings. This highlights a relationship between the distribution of monogenic vents and the depth of origin of the magma (e.g., Mazzarini 2007; Mazzarini et al. 2010). These behaviours provide a convenient framework to explain volcano spacing as a consequence of variations in lithospheric or crustal thickness due to regional extension. As regards convergent plate boundaries, the spacing of the polygenic arc volcanoes does not show characteristic values and is more difficult to interpret, although probably related to random distributions

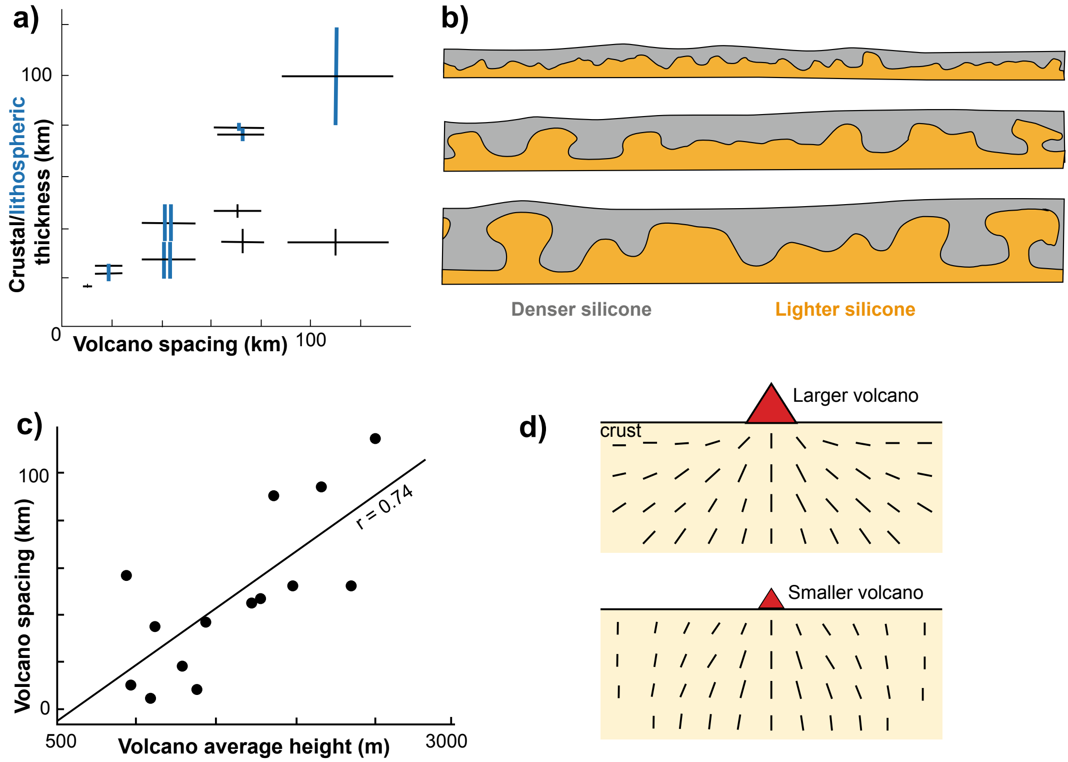

Fig. 10.22 Two models **a**, **b** and **c**, **d** used to explain polygenic volcano spacing. **a** The spacing of volcanoes along the Afar-Main Ethiopian Rift is proportional to the crustal (black lines) or lithospheric (blue lines) thickness (Mohr and Wood 1976). This behaviour is consistent with the three sections of analogue models in (**b**), characterized by two layers of silicone (viscosity of 10^4–10^5 Pa s), with the upper denser layer increasing its thickness from top to bottom (Ramberg 1981): the spacing between nearby intrusions of the lighter silicone increases with the thickness of the upper layer. Both behaviours are explained by the Rayleigh–Taylor theory, where the disturbance wavelength is proportional to the upper layer thickness. An alternative mechanism **c** relates volcano height to volcano spacing, considering the volcanoes of the Cascades Arc. This behaviour is explained in **d** by the fact that higher (or larger) volcanoes impose a higher load (load trajectories shown as small black lines in section view) on the crust, capable of attracting magma over a wider region (Muller et al. 2001)

resulting from repeated instability events associated with the rise of magma in different locations (de Bremond d'Ars et al. 1995). Other studies relate polygenic volcano spacing to the load of a volcanic edifice: larger volcanoes impose a higher load on the crust, inducing a stronger local stress field capable to attract the magma below in the form of dikes curving from progressively wider regions and focusing towards the volcano. This capability to attract magma over wider areas may explain why larger volcanoes are also more distantly spaced, as in the Cascades Volcanic Arc (Fig. 10.22c; Muller et al. 2001). The flexural response of the lithosphere due to the load of a volcanic edifice may also control the distance between nearby polygenic volcanoes forming in progression, as on plates moving above mantle plumes. In particular, compressive stresses within a crust flexured by the load of a volcano prevent new upwelling nearby, forcing a new volcano to develop at a minimum distance that is equal to the distance in which the radial stresses change from compressional to tensile, that is the inflection point of a flexured lithosphere. The distance to this inflection point is proportional to the thickness of the plate (e.g., Tenbrink 1991; Bianco et al. 2005, and references therein).

Alignments of active polygenic volcanoes are related to the tectonically-assisted availability of magma. For example, aligned polygenic volcanoes in divergent plate boundaries form as a consequence of localized decompression along the axis of rift zones (see Chap. 11). Similarly, aligned polygenic volcanoes in convergent plate boundaries result from the availability of magma above the zone of partial melting of the slab, with the location of the volcanic arc ultimately controlled by the slab dip (see Chap. 12). At a more detailed scale, the alignment of monogenic volcanoes along a given direction, as often witnessed by eruptive fissures up to several tens of kilometres long, commonly results from the shallow emplacement of a dike, whose direction is controlled by the tectonic and gravitational stresses and, partly, by pre-existing tectonic structures. In the case of tectonic stresses, dikes tend to orient perpendicularly to the direction of the least principal stress σ_3. In the case of gravitational stresses, the stresses due to topography at very shallow crustal levels may cause a dike to deflect from its path, in order to experience the local least buttressing conditions (see Sect. 7.2). Numerical modelling indicates that such near-surface effects are important to a depth of approximately 500 times the dike width. In the case of pre-existing tectonic structures, a dike may in theory deviate from its orientation in favour of a more energy-efficient path along a pre-existing discontinuity, as a fault plane. As discussed in Sect. 3.5, the orientation of the pre-existing structure relative to the regional principal stresses (σ_3 in particular) mainly determines whether a fracture is more or less likely to dilate in response to dike injection and redirect the ascending dike. However, even in the most favourable conditions, as with fractures perpendicular to σ_3, dikes may not need to reactivate pre-existing discontinuities. This is supported by the slight but systematic offset in the location of several tens of eruptive vents from important normal faults along the East African Rift System, suggesting that other processes have a stronger control on dike propagation (Maccaferri et al. 2015; Valentine and Connor 2015). Therefore, most monogenic fields display a statistically clustered distribution of their vents, with the alignment of volcanoes depending on the tectonic environment and other factors (i.e., topography, pre-existing structures, local stress changes due to older intrusions; Le Corvec et al. 2013). However, volcanoes in monogenic fields may also not show any evident major alignment, displaying an apparently random orientation, or minor and subtler orientations, as for example in the Auckland Volcanic Field. Here, although minor alignments may be still identified, most vents do not seem to follow evident preferred trends. In these cases, statistics can be applied to estimate eruptive recurrence rates, while geochemistry can characterize any variability in the magma source characteristics and magmatic processes. Vent clusters within volcanic fields may be explained by spatial magma supply variations, reflecting the scale of temperature, pressure, and compositional variations within the underlying mantle, as well as the development of shallow magmatic plumbing systems. In synthesis, the overall distribution of vent clusters within volcanic fields suggests that the propagation of the magma is the product of the interplay between deeper level influences (i.e., nature of the magma source) and shallower level influences of the crust (i.e., stress field, pre-existing crustal structures; Le Corvec et al. 2013; Valentine and Connor 2015, and references therein).

10.7 Magma Versus Regional Tectonics

Regional studies from the last decades, which are presented in Chaps. 11 and 12, have allowed gaining important insights on the role of magmatic processes with regard to regional tectonics in the development of plate boundaries, in some cases also changing significantly established perspectives. The general conclusions arising from these studies are anticipated and summarized below.

As regards divergent plate boundaries, as specified in Chap. 11, the established perspective that regional tectonics associated with plate motion controls the morphology, structure and development of rift zones has been repeatedly questioned by recent **rifting episodes** and **rifting events** (where a rifting episode consists of several rifting events clustering in time in the same rift portion) associated with the emplacement of tens of kilometres long dikes in continental, transitional and oceanic rifts. These rifting episodes and events include Krafla (Iceland, 1975–1984), Asal-Ghoubbet (Afar, Djibouti, 1978–1979), Dallol (Afar, Ethiopia, 2004), East Pacific Rise (2005–2009), Dabbahu (Afar, Ethiopia, 2005–2010), Lake Natron (Tanzania, 2007), Harrat Lunayyir (Saudi Arabia, 2009) and Bardarbunga (Iceland 2014; Jacques et al. 1996; Ruegg et al. 1979; Tarantola et al. 1979; Buck et al. 2006; Tolstoy et al. 2006; Wright et al. 2006; Rowland et al. 2007; Calais et al. 2008; Biggs et al. 2009; Pallister et al. 2010; Nobile et al. 2012; Sigmundsson et al. 2015). Each of these rifting episodes or events produced the rapid (lasting a few weeks) and significant (a few metres) opening of a portion of a magmatic system along the plate boundary and, with the possible exception of Lake Natron, was characterized by the lack of faulting due to regional tectonics. In fact, the observed generation or reactivation of normal faults at the surface, usually forming graben-like structures parallel to the rift axis, resulted from dike emplacement, and was thus magma-induced. The Lake Natron event, on continental crust, was the only observed "hybrid" rifting event possibly accompanied by significant regional seismicity promoting normal faulting in addition to dike emplacement. However, even in this case, the opening associated with the intruded dike was larger than the extension induced by regional faulting. Therefore, all these cases showed that dike-induced extension is the dominant mechanism responsible for splitting the plates along divergent plate boundaries. With the exception of the Harrat Lunayyir event, dike emplacement occurred along magmatic systems, emphasizing their central role in the activity of these boundaries. In addition, in several episodes (as at Krafla, Asal-Ghoubbet and Dabbahu) multiple rifting events insisted in the same area within a few years, highlighting repeated dike emplacement. As the emplaced dikes produced graben-like structures at the surface, repeated diking episodes may explain on the long-term most of the surface deformation and the rift morphology of divergent plate boundaries. This possibility is supported by the evidence that the relatively high frequency of diking and the associated deformation along divergent boundaries often hinders extension to be accommodated through regional tectonics, be it through creep, seismic or aseismic faulting (Fig. 10.23; Sigmundsson 2006; Ebinger et al. 2010; Acocella and Trippanera 2016). Possible exceptions to this behaviour are found in immature continental rifts (as in the above-

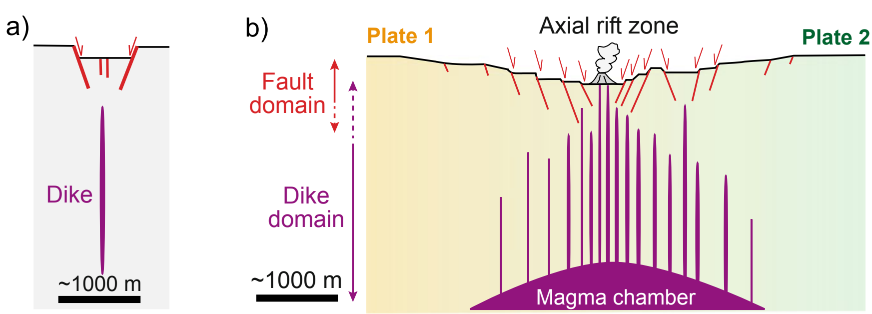

Fig. 10.23 Magmatic shaping of rift morphology and structure along divergent plate boundaries. **a** Example of surface deformation induced by a dike. **b** Upper crustal structure of the axial portion of a divergent plate boundary, as derived from field data in Iceland. Extension due to normal faults (propagating downward from the surface) becomes progressively less important at ~1 km depth, whereas at greater depths (>1 km) extension occurs almost exclusively through dike emplacement. Thus, the overall graben-like and depressed morphology of the plate boundary at the surface results from repeated dike emplacement at depth (modified after Acocella and Trippanera 2016)

mentioned Lake Natron event), where the regional tectonic component may be still relevant, and in the non-magmatic portions of oceanic rifts, as those experiencing detachment faulting (see Sect. 11.4). Therefore, it appears that while regional tectonic processes control the far-field plate motion and divergence, magmatic processes control the near-field evolution of mature plate boundaries and shape their structure and morphology.

There is evidence that magma plays an important role also along convergent plate boundaries, as specified in Chap. 12. Convergent plate boundaries characterized by extension along the arc, as the Taupo Volcanic Zone of New Zealand, have tectono-magmatic features similar to those of continental rifts, including proto-magmatic systems. This suggests an important, although probably not predominant, role of magmatic activity in controlling the evolution of the extensional arc and in shaping its structure (Rowland et al. 2010; Allan et al. 2012). These features are not observed along magmatic arcs experiencing weak extension, strike-slip motions, or contraction, where volcanic activity is restricted to the area of the volcanic edifice and magmatic systems are lacking. Despite this lack, at least two features highlight the non-negligible role of magmatism in strike-slip and contracting arcs (Fig. 10.24). The first is that, despite the lack of magmatic systems in strike-slip arcs, the magma intruded in the crust remains able to thermally weaken it, focusing along the volcanic arc the strike-slip deformation induced by strain partitioning. This shows how magma may control the structural architecture of the overriding plate. The second feature, as mentioned in Sect. 10.4, is related to the evidence that magma in contractional arcs finds its way to intrude from sills to stacked tabular intrusions, and even to rise to the surface, by locally increasing the principal vertical stress, which may switch from the minimum principal stress σ_3 to the maximum principal stress σ_1. This supports the idea that magmatic rates may locally overcome tectonic rates also in convergent plate boundaries, so that magma may intrude, accumulate and erupt largely independently of any regional tectonic context.

Nevertheless, despite the overall independence of magmatic processes from regional tectonics, magma may still benefit from favourable shorter-term tectonic variations occurring along volcanic arcs. This is exemplified by the volcanic activity (in terms of both unrest and eruptions) promoted by transient stress changes due to mega-earthquakes in various convergent settings. These changes are in fact able to temporarily affect the kinematics of a volcanic arc, even including shifts from overall inter-seismic contraction to co- and post-seismic extension (see Sect. 12.5; Walter and Amelung 2007; Pritchard et al. 2013; Takada and Fukushima 2013).

Independently of any specific tectonic setting, a further and more general question regards the role of pre-existing regional structures on the rise of magma, where a dike may intrude discontinuities deviating from its orientation perpendicular to the minimum principal stress in favour of a more energy-efficient path along a fracture or fault plane. The approach to this topic has shifted from an earlier perspective of fracture-controlled magmatism, where rising magma is captured by a pre-existing structure, to a modern vision of relative independence of dike ascent from pre-existing discontinuities. In fact, as discussed in Sect. 3.5 and mentioned in Sect. 10.6, only fractures oriented perpendicular to the least principal stress σ_3 seem likely to provide a low energy path to the surface and thus be intruded by magma, although this may still not be a sufficient condition (e.g., Rubin 1995; Maccaferri et al. 2015; Valentine and Connor 2015).

All these features indicate that magmatic activity, mostly in the form of dike propagation and subordinately as sill emplacement, is largely independent although not insensitive to the regional tectonic context and, under favourable conditions, it may even overcome non-magmatic regional tectonic processes. The details on this important role of magma are presented in the next two chapters.

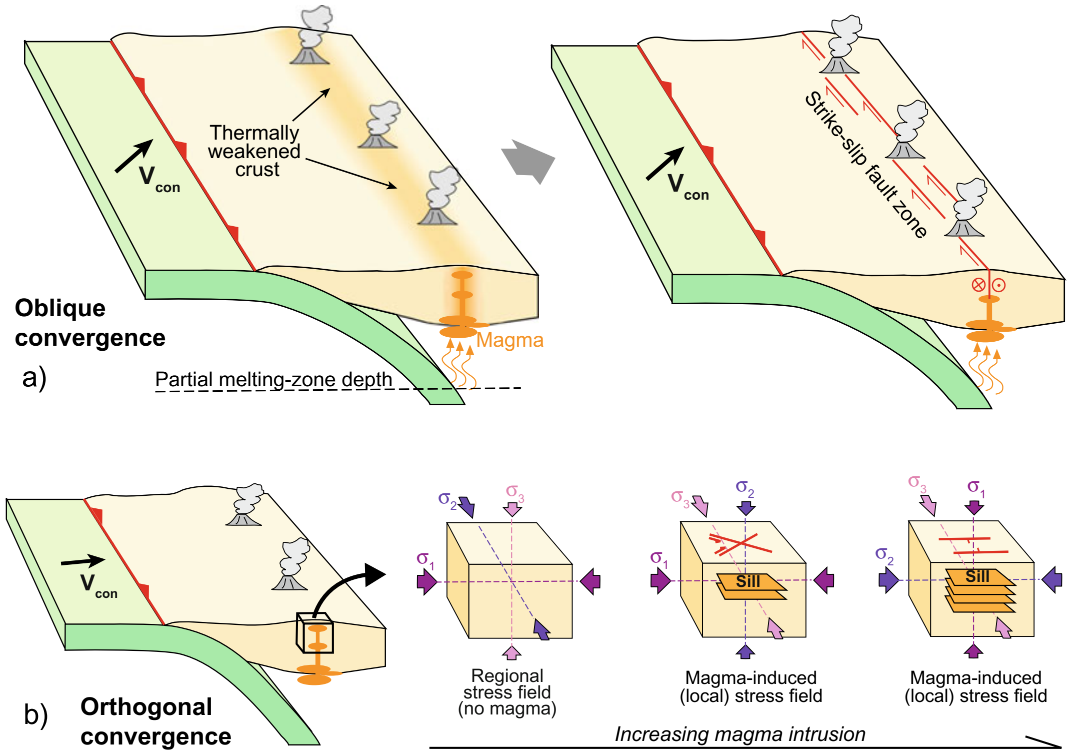

Fig. 10.24 Control of magma on the evolution of magmatic arcs in convergent plate boundaries. **a** Under oblique convergence, a crust thermally weakened by magma (dark yellow portion, left diagram) focuses the strain in the overriding plate experiencing strain partitioning, developing a strike-slip zone along the volcanic arc (right). **b** Under orthogonal convergence, the repeated emplacement of magma as stacked sills below the arc (in the zone highlighted by the cube, left diagram) increases the vertical component of the stress field, which passes from σ_3 to σ_1, promoting local arc-parallel extension and enhancing volcanism (right)

10.8 Summary

Plate tectonics theory provides the framework to understand the global distribution of volcanism, including the location and shape of volcanoes, the size and frequency of the eruptions and the composition of the erupted products. Most of the magma is being erupted along divergent plate boundaries, in continental, transitional and, especially, oceanic domains. In particular, the tectonically and magmatically active portions of divergent plate boundaries are the magmatic systems, where dikes actively split the plates. Convergent plate boundaries account for approximately a quarter of the erupted volumes. The structure of the volcanic arc overlying the slab may be characterized by various kinematics, mainly depending on the convergence motion between the two plates. In some cases, processes not requiring active subduction, as slab break-off and lithospheric delamination, may explain collisional magmatism. Hot spot volcanism explains a relatively lower amount of erupted volumes through the activity of mantle plumes. This volcanism is largely found in intraplate settings and along divergent plate boundaries, where the plumes interact with the rift segments. Despite the relatively low erupted volumes, volcanism from mantle plumes can be at times volumetrically impressive, as in the case of LIPs. The regional tectonic setting also controls, through different mechanisms, the occurrence and the distribution, in terms of spacing and alignment, of polygenic or monogenic volcanoes. Nevertheless, there is ample evidence of the predominant role of magmatism on regional tectonics in shaping the structure, morphology and evolution of divergent plate boundaries, whereas this role becomes less evident, although still affecting different processes at various scales, along convergent plate boundaries.

10.9 Main Symbols Used

B	Buoyancy flux
D''	Thermal boundary layer
H_e	Excess elevation averaged across the swell
L	Width of the swell
M	Magnitude
t	Age of lithosphere
V_{con}	Plate convergence motion
V_h	Plate velocity
V_n	Trench orthogonal convergence velocity
V_p	Trench parallel convergence velocity
z	Thickness of the lithosphere
α	Thermal diffusivity
β	Stretching factor
δ	Angle of obliquity of the rift
ε_1	Principal horizontal strain axis
ρ_{ma}	Mantle density
ρ_w	Water density
σ_1	Maximum principal stress
σ_2	Intermediate principal stress
σ_3	Minimum principal stress
ϕ	Angle of obliquity of convergence

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