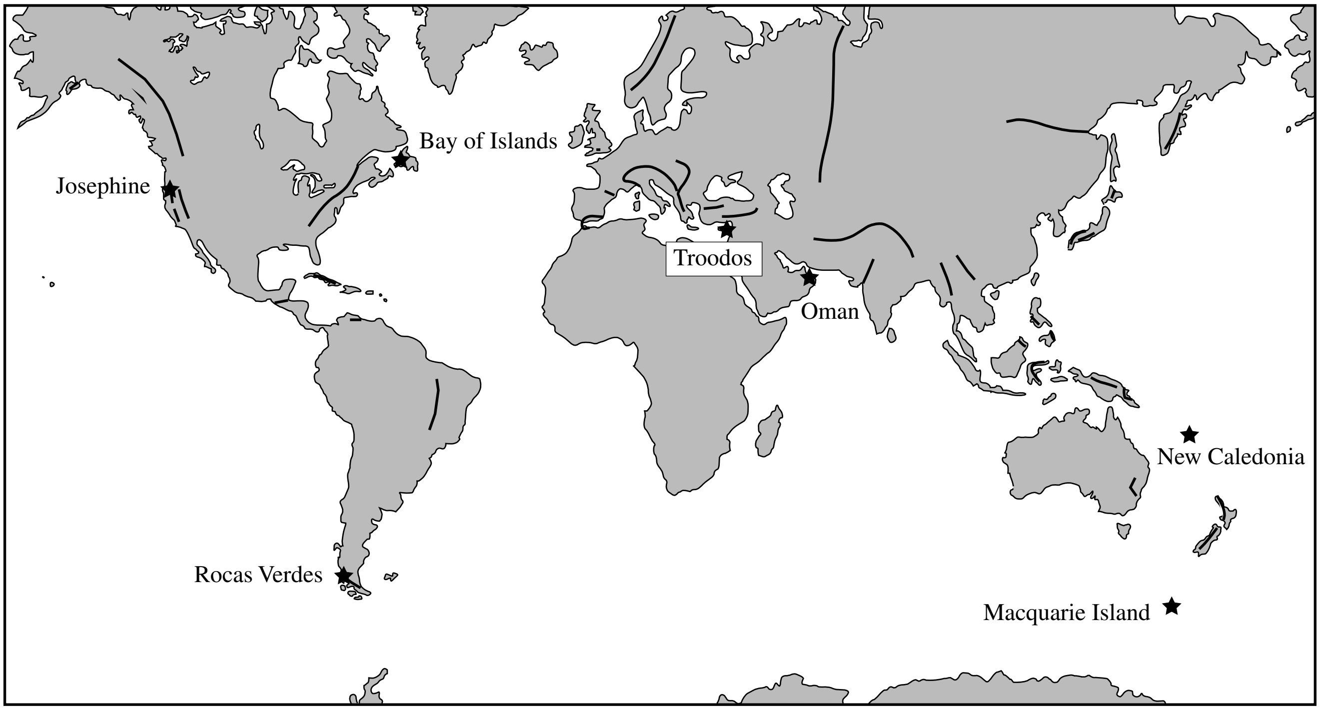
Chapter 6

Oceanic Magmatism

6.1 Introduction

Because it is covered by kilometers of water, ocean crust was long inaccessible to direct observation by geologists. Today, however, our knowledge of the ocean floor comes from two sources, the study of fragments of the ocean floor that have been thrust onto the land, called **ophiolites**, and from ship-based geophysical and geological studies that burgeoned during the Second World War and were followed by the Deep Sea Drilling Program (DSDP), which began in 1968. These investigations provided the foundation that underpins our understanding of oceanic magmatism. This chapter first discusses the structure and stratigraphy of ophiolites and to what extent they provide models that help understand the ocean crust. A description of advances achieved by recent research of the ocean floor based on geophysical studies and ocean drilling follows. Finally, this chapter describes the magmatic suites that compose ocean islands and oceanic plateau.



Map 6.1 Map showing select ophiolite belts around the world. Ophiolites occur along the trends indicated by bold lines. Stars show particularly well-known occurrences. Data from Irwin and Coleman (1974).

6.2 The Petrology and Structure of the Ocean Crust

6.2.1 Ophiolites as a Model of the Ocean Crust

Geologists have long recognized that an association of peridotite (in many places hydrated to serpentinite), gabbro, basalt, and deep-water chert are exposed in many places around the world (Map 6.1). In some localities, these rocks form a complete stratigraphic section, but in many places one or more of these rock types exist within fault-bounded tectonic slices. As early as the 1820s this association was called an **ophiolite**, but before the advent of plate tectonics, the significance of these rocks was cryptic. Geologists attending the September 1972 Penrose Conference defined the stratigraphy of a typical ophiolite, shown in Figure 6.1 (Anonymous, 1972). Implicit in the definition is the assumption that ophiolites are fragments of oceanic crust thrust onto the continents, and thus the stratigraphy described at the Penrose Conference represents an idealized cross-section of the oceanic crust.

The uppermost layer in an ophiolite is composed of deep-water sediments, mostly pelagic mud, although chert may be common in some places. The thickness of

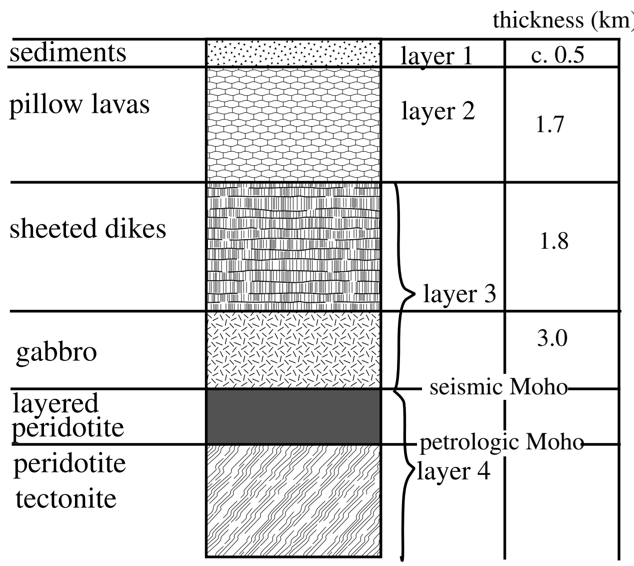


Figure 6.1 Petrologic and seismic profile for an ideal ophiolite (Anonymous, 1972).

this layer depends on the age of the crust. On juvenile oceanic crust there are no sediments; the thickness of the sediment layer generally increases with age. A kilometer or so of pillow basalts, which represent lavas that were erupted directly onto the ocean floor, underlie the sediments. The pillow basalts grade into sheeted dikes, a horizon that may be over a kilometer thick. Sheeted dikes are dikes that consistently chilled on one side only. They are interpreted to have been emplaced into a spreading center, with each new dike intruded into the core of a preceding dike. Below the sheeted dikes lie several kilometers of gabbro. The top of the gabbro is directionless, but toward the bottom it may be layered or foliated. This layer is interpreted to have crystallized from an intrusive body of basaltic magma. Below this lies layered peridotite, which is much denser than the overlying gabbro. The contact between peridotite and gabbro is the location of a distinct change in seismic velocity marking the Moho. However, because these peridotites are interpreted to have formed as cumulates from the basaltic magma, they are unrelated to underlying mantle and actually represent an ultramafic portion of the crust. Below the cumulate peridotites is a highly deformed peridotite, which is interpreted as mantle depleted by partial melting during basalt genesis. Petrologically this is true mantle, even though it is impossible to distinguish it seismically from the overlying cumulate peridotite.

6.2.2 Refinements of the Ophiolite Model

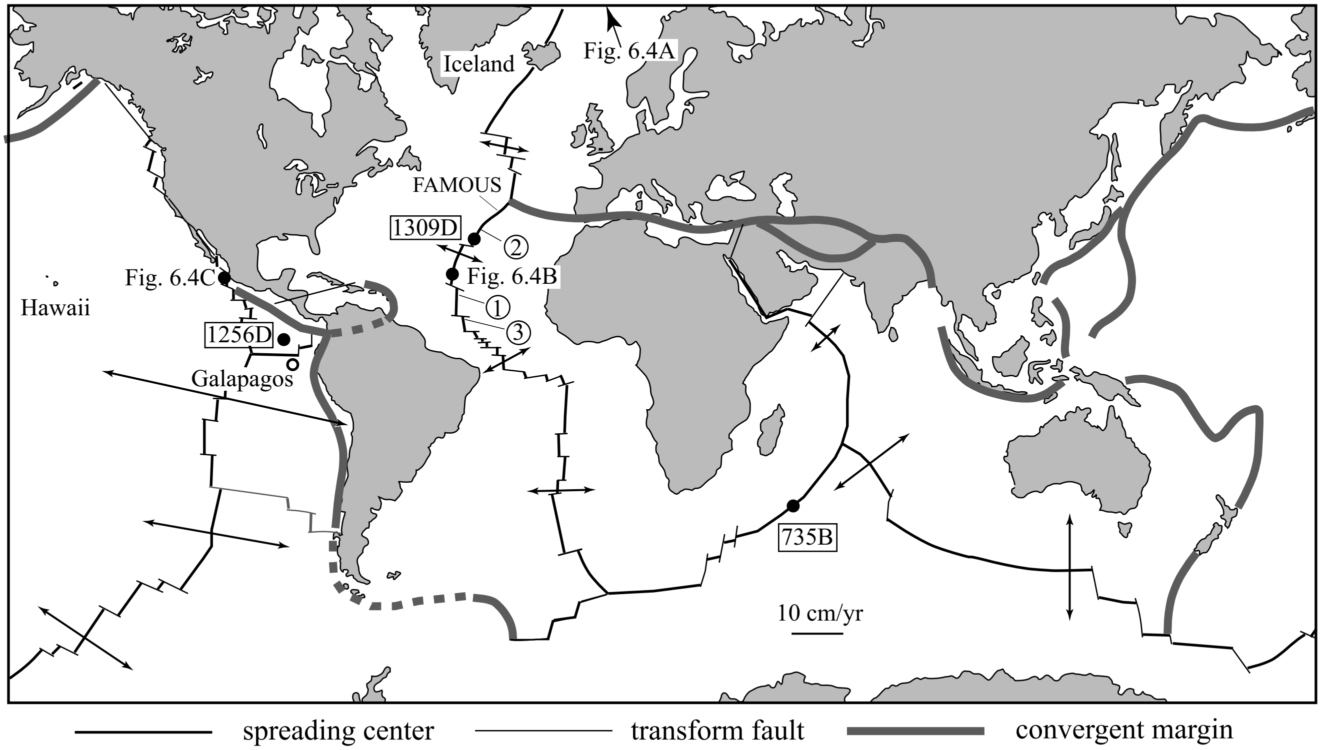
Nearly as soon as the Penrose ophiolite model was proposed, geologists began to debate whether the model describes a true picture of the ocean crust (Miyashiro, 1975; Moores, 1982). It quickly became evident that ophiolites form in diverse tectonic environments, and not all reflect ocean-floor stratigraphy produced at mid-ocean spreading centers. Some ophiolites, such as the Troodos ophiolite in Cyprus, contain basalts more closely related compositionally to arc basalts than to mid-ocean ridge basalts (Miyashiro, 1973) and evidently formed above newly initiated subduction zones. These are called **suprasubduction-zone** ophiolites (Pearce, Lippard, and Roberts, 1984). Observations suggest ophiolites form in a wide range of tectonic environments and thus resist a simplified, "one-size-fits-all" model. In addition to forming above subduction zones, ophiolites form by back-arc spreading as did the Rocas Verdes ophiolite in Chile (Stern and de Wit, 2003), at the contact between a back-arc and an arc as did the Bay of Islands ophiolite in Canada (Kurth-Velz, Sassen, and Galer, 2004), or in an oceanic spreading center as did the Macquarie Island ophiolite in the south Pacific (Varne, Brown, and Faloon, 2000) and the Oman ophiolite on the Arabian Peninsula (Boudier and Nicolas, 2011). Map 6.1 shows the global distribution of these and other major ophiolites.

A second problem with the ophiolite model arose in the 1990s and 2000s when seismic surveys and deep-ocean drilling showed the stratigraphy of the oceanic crust is far more complex than the ophiolite model suggested. Geophysical studies revealed significant differences in spreading rates among oceanic ridges (Map 6.2) and that ridges with different spreading rates have different morphology (Figure 6.2), which translates into differences in crustal cross-section.

Fast-spreading centers. The East Pacific Rise (EPR) is an example of a fast-spreading center (half-rate 6–7 cm/yr). Fast-spreading centers are characterized by a 2.5 to 3.0 kilometer-wide zone of magma extrusion, which forms a smooth topographic high of around 200 meters (Figure 6.2A). Flat lava plains made of ponded lava lakes and small volcanic hills composed of sediment-free pillow lavas occur along the ridge axis. There is either no axial valley or only one that is poorly developed.

Seismic studies of the EPR appear to image large sub-axial magma chambers beneath fast-spreading centers. These magma chambers appear to be periodically replenished from below with fresh batches of mantle magma. Between additions of magma, fractional crystallization takes place. Lavas erupt when the magma pressure exceeds the lithostatic pressure and the strength of the chamber roof, probably coincident with addition of magma into the chamber. The crustal cross-section beneath a fast-spreading ridge is similar to that described by the ophiolite model.

Slow- and ultra-slow-spreading centers. The Mid-Atlantic Ridge (MAR) is a typical slow-spreading center (half-rate 1–2 cm/yr), and the Gakkel Ridge under the Arctic Ocean is an ultra-slow-spreading ridge (half-rate 0.1 cm/year) (Figure 6.2B, C). Unlike fast-spreading centers, slow- and ultra-slow-spreading centers tend to have a well-defined axial valley. The slow-spreading center is characterized by a twenty-five to thirty-kilometer-wide axial valley bounded by mountains. Within this broad valley is a second, well-defined inner valley, three to nine kilometers wide, where volcanic activity is concentrated (Figure 6.2B, C). Small volcanic hills occur within this inner valley, showing that volcanic activity is neither spatially nor temporally continuous.



Map 6.2 Tectonic map of the ocean basins showing mid-ocean ridges, convergent margins, transform faults, and areas discussed in the text. The length of the spreading rate vector arrows is proportional to the spreading rate. Numbers refer to cross-sections shown in Figure 6.3. Numbers in boxes refer to IODP drill holes shown in Figure 6.4. Modified from Brown and Mussett (1981) with additional data from Dick and colleagues (2000), Teagle and colleagues (2006), and Blackman and colleagues (2011).

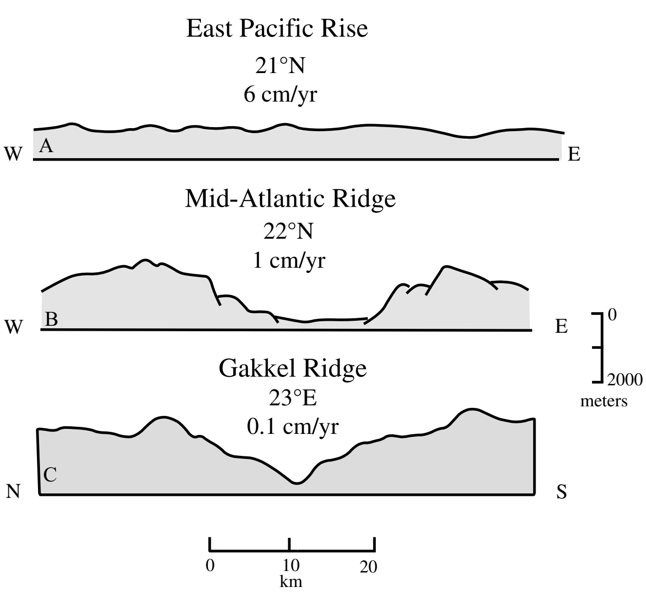


Figure 6.2 Morphology of (A) fast (East Pacific Rise), (B) slow (Mid-Atlantic Ridge), and (C) ultra-slow (Gakkel Ridge, Arctic Ocean) spreading centers. Data from Basaltic Volcanism Study Project (1981) and Cochran (2008).

A sub-crustal magma chamber beneath slow-spreading centers must either be about the width of the inner valley floor (three kilometers), or each volcanic hill may have a small (<1.5 kilometer-wide) magma chamber beneath it. The second hypothesis is more consistent with geophysical data, which does not observe attenuated S-waves as would be expected if a large magma chamber existed.

The crust beneath the slow-spreading centers is more poorly layered and more heterogeneous than the ophiolite model predicts. This is shown in surveys of fractures zones off of the Mid-Atlantic Ridge, where cross-sections of the oceanic crust are exposed (Figure 6.3). Because magma is not constantly supplied, extension at slow- and ultra-slow-spreading is partially or completely accommodated by faulting. These extensional faults produce crustal sections that eliminate some of the units in the ophiolite stratigraphy. In some places the basalt flows from the spreading centers are in fault contact with gabbro; in others the basalt is in contact with serpentinized peridotite. In many places in slow- and ultra-slow-spreading ridges this

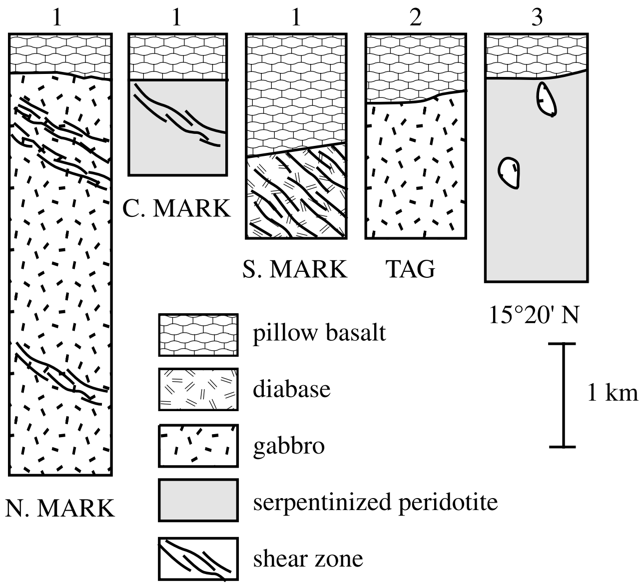


Figure 6.3 Cross-sections of oceanic crust beneath the Mid-Atlantic ridge. Numbers refer to locations in Map 6.2. After Karson (1998).

extension has stripped the crust from the mantle, exposing serpentinized peridotite directly on the sea floor.

Four decades of ocean drilling have revealed a great petrologic variability to the ocean crust (see Box 6.1). Deep drill cores in fast-spreading crust, such as Integrated Ocean Drilling Program (IODP) hole 1256D, show relations similar to what the ophiolite model predicts (Teagle et al., 2011) (Figure 6.4). However, drill holes into gabbroic crust exposed in slow-spreading ridges (IODP holes 375B and 1309D) show relations that are much more complex (Dick et al., 2000; Blackman et al., 2011) (Figure 6.4). Core from hole 375B from the Southwest Indian ridge contains mainly gabbro with minor amounts of oxide gabbro. The section is cut by several large, ductile shear zones. In contrast, core from hole 1309D from the Mid-Atlantic ridge contains a complex series of gabbro and oxide gabbro interlayered with screens of peridotite. Magmatic differentiation produced the oxide gabbro retrieved from holes 375B and 1309D. As explained in Chapter 2, crystallization of olivine (and any other Fe-Mg silicate) removes Mg from a melt preferentially to iron (Figure 2.11). Eventually this saturates the melt in Fe-Ti

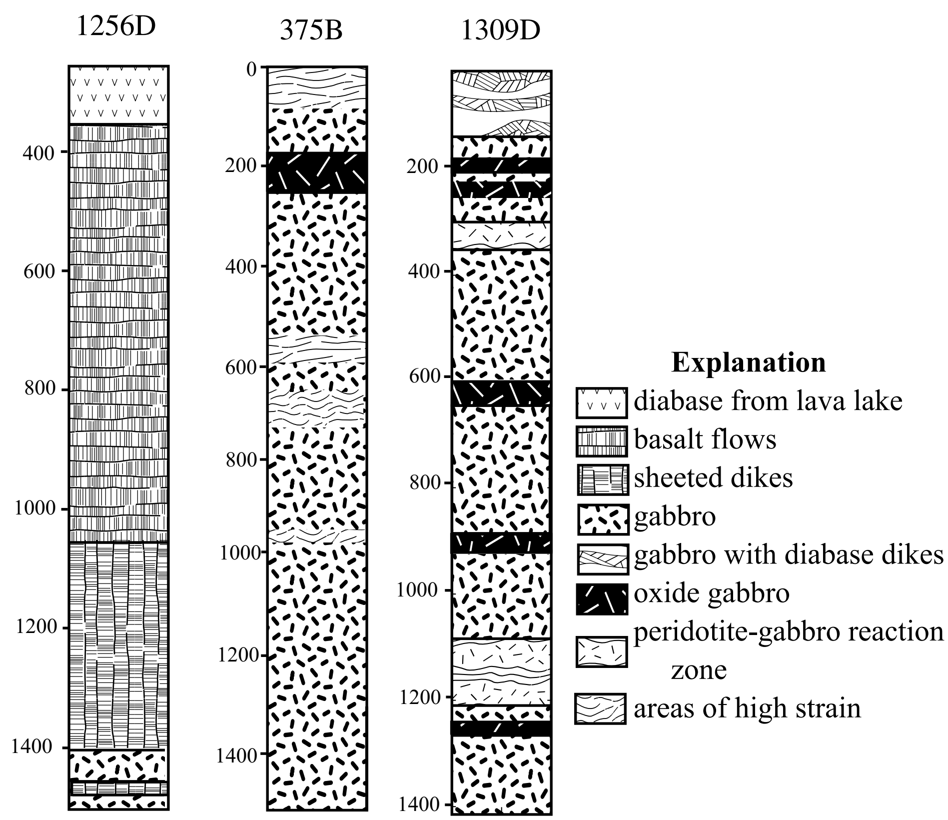


Figure 6.4 Cross-sections of the ocean crust as obtained in several IODP drill holes. Data from Dick and colleagues (2000), Teagle and colleagues (2006), and Blackman and colleagues l. (2011).

BOX 6.1 | OCEAN DRILLING

Before ocean drilling programs commenced, the only information scientists had about the composition of the ocean floor was obtained by dredging. Beginning in 1968, the United States' National Science Foundation (NSF) began a program to obtain samples of the ocean floor by drilling. The initial drilling project, entirely funded by the NSF, was called the Deep Sea Drilling Project (DSDP). In 1975, France, Germany, Japan, the Soviet Union, and the United Kingdom jointly funded the drilling program. DSDP ran from 1968 until 1983 using the research vessel *Glomar Challenger*. The *Glomar Challenger* was retired in 1983 and the drilling program resumed in 1985 as the Ocean Drilling Program (ODP) then, in 2003, as the Integrated Ocean Drilling Program (IODP), which uses the research vessel *Joides Resolution* and is ongoing today. Currently the drilling program is supported by twenty-six countries, including the United States, the European Union, the United Kingdom, Japan, China, India, Australia, and New Zealand.

One of the major scientific themes of the IODP is to study the petrology of the ocean crust to understand the geochemical and geodynamic processes involved in the solid Earth system. The scientific value of ocean drilling became apparent within the first years of drilling. The first cores substantiated the young age of the ocean crust and the dynamics of sea floor spreading, geologic observations and processes that now underpin discussions of plate tectonics. Ocean drilling verified that the primary transfer of energy and material from the deep Earth to the surface occurs via sea floor spreading and the creation of oceanic crust at mid-ocean ridges, as well as by upwelling magmas that form ocean islands, ocean plateau, and island arcs. Further drilling documented that sea floor spreading involves not only magmatic addition to the crust but locally, may include tectonic denudation as well. As a result of sea floor tectonics, a considerable area of the ocean floor is underlain by serpentinized mantle peridotite. Ocean drilling has enabled descriptions of the kinds of reactions involved in the alteration of the sea floor, including serpentinization; these reactions have proven critical to modeling the geochemistry of ocean water. Recent findings suggest that MORBs interact extensively with the mantle through which they move, producing hybrid troctolites whose existence was previously unexpected. In total, these observations help petrologists understand the evolution of mantle-derived basaltic magmas and the formation of oceanic crust.

Box 6.1 Rainbow over the drilling rig of the *Joides Resolution* 30°N on the Mid-Atlantic Ridge.

oxides (magnetite and ilmenite), producing the oxide gabbro. The presence of multiple oxide gabbro horizons in holes 375B and 1309D means the holes penetrated several discrete igneous bodies, each of which had differentiated to oxide gabbro. A significant amount of troctolite and olivine-rich troctolite is also present and these are inferred to have formed by reaction between the peridotitic mantle and the basalt emplaced into the spreading center (Blackman et al., 2011).

In summary, the data obtained from recent drilling of the sea floor shows that, although some or all of the components of an ophiolite may be present at any given locale, oceanic crust is immensely more complex than the ophiolite model implies. Fast-spreading centers produce crustal stratigraphy closest to the ophiolite model, but in slow-spreading areas faults truncate the stratigraphy and the gabbroic sequence consists of multiple injections of magma that interacted with the peridotite host rocks and differentiated in place.

6.3 Petrography and Geochemistry of Oceanic Magmatism

Oceanic magmatism occurs in two distinct environments: at mid-ocean ridges and at off-ridge locations where ocean islands and oceanic plateau are formed. The basalts erupted in each of these two environments are chemically and petrographically distinct. Volumetrically the most important environment is along mid-ocean ridges where new oceanic crust continually forms as tectonic plates diverge (Map 6.2). The basalt erupted here is olivine and quartz-normative tholeiite; these basalts are referred to as **mid-ocean ridge basalts** (**MORB**). A significant volume of basaltic magma is also erupted from vents not located on ridges; this **off-ridge magmatism** occurs on ocean islands and oceanic plateau. The rocks erupted off-ridge are referred to as **ocean island basalts** (OIB), and they include both tholeiitic and alkali basalts.

6.3.1 Mid-ocean Ridge Basalt

The fine-grained groundmass of MORB reflects rapid cooling of magma extruded into a cold submarine environment. Phenocryst assemblages in glassy basalts suggest the first minerals to crystallize are olivine + spinel. As the magma differentiates, plagioclase joins the crystallizing assemblage. Finally a groundmass consisting of

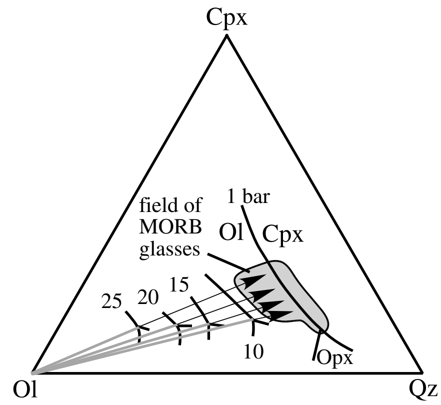


Figure 6.5 Pseudoternary projection from plagioclase onto the olivine-clinopyroxene-quartz plane showing the composition range of MORB glasses. Also shown are the eutectics for aluminous peridotite at 1 bar and 10, 15, 20, and 25 kilobars. See text for discussion. Modified after Elthon (1989).

plagioclase + Ca-rich clinopyroxene (augite) + olivine forms. Olivine compositions are typically Fo_{65}–$_{91}$. The spinel is Mg and Cr rich, and is frequently found as inclusions in olivine. Plagioclase is typically An_{88}–$_{40}$ (Grove and Bryan, 1983) and is commonly more Ca-rich in basalts erupted on the Mid-Atlantic ridge than those erupted on the East Pacific rise. The presence of more sodic plagioclase on the East Pacific rise suggests magmas differentiate to a greater extent at fast-spreading centers where large magma chambers may be present (Hekinian, 1982).

Trace element characteristics of MORB suggest the type of mantle source rock from which partial melts are extracted is spinel or plagioclase lherzolite, rather than the high-pressure phase, garnet lherzolite. This mineralogy is consistent with geophysical studies of P- and S-wave attenuation, which suggest that the melting begins at depths of sixty to eighty kilometers, and the melt segregates at about twenty kilometers to rise and feed the magma chambers. At minimum, 20 percent partial melting is required to produce the most MgO-rich MORB compositions (Wilson, 1989).

The origin of MORB is summarized in Figure 6.5, which shows the system olivine-clinopyroxene-quartz-plagioclase as projected from anorthite to the olivine-clinopyroxene-quartz plane (see also Figure 5.5). Heavy lines on this pseudoternary projection are the locations of the olivine-clinopyroxene, olivine-orthopyroxene, and orthopyroxene-clinopyroxene cotectics in the presence of an aluminous phase at one bar, and ten, fifteen, twenty, and twenty-five kilobars. At one bar the aluminous phase is plagioclase, at ten, fifteen, and twenty kilobars it is spinel, and at twenty-five kilobars it is garnet. This diagram clearly illustrates that increasing pressure stabilizes clinopyroxene with respect to olivine and orthopyroxene. This occurs because at increasing pressure clinopyroxene is progressively enriched in Na and Al.

To understand how to read Figure 6.5, consider a dry mantle diapir that begins melting at twenty-five kilobars. The first melt has a composition that lies on the olivine-orthopyroxene-diopside "eutectic." (The quotation marks recognize that in a pseudoternary projection, the eutectic does not lie in the plane of the diagram.) If this high-pressure melt moves to shallower crustal levels (i.e., to lower pressure), for example to a pressure of ten kilobars, then the original melt will no longer lie on the "eutectic." Rather it will lie in the field of primary olivine crystallization. As olivine crystallizes, it drives the residual melt composition directly away from the olivine apex as the light grey arrow originating at olivine and passing through the "eutectic" composition shows. All melts derived by fractional crystallization of olivine from the twenty-five-kilobar "eutectic" lie along the dark portion of the arrow. Partial melting at a range of mantle depths and fractional crystallization of olivine during magma ascent will produce the observed compositional range of MORB glasses (gray field in Figure 6.5).

Figure 6.5 implies that the range in compositions of MORB results from a number of processes. First, magmas form by partial melting at various depths, producing parent magmas with a limited but varying composition. These magmas start to crystallize during ascent into the ridge center. When the magmas pond at low pressures (i.e., less than ten kilobars), magmas originating from different depths mix. These composite melts move to shallow magma chambers where their compositions are further changed by low-pressure differentiation. Finally, observations from IODP hole 1309D suggest the magmas interact with their mantle host rocks at relatively shallow crustal levels and that this process further affects the compositions of MORB (Blackman et al., 2011).

These multiple processes result in a small but significant range in the major element composition of erupted MORB. The composition of basalt liquids erupted on the sea floor can be determined from the composition of basalt glass. Compositions of basalt glasses from the Narrowgate region of the FAMOUS valley, Mid-Atlantic Ridge (Figure 6.6) show that although SiO_2 contents are relatively constant, MgO varies between 7 to 9 percent. As the observed phenocrysts – olivine, plagioclase, and augite – crystallize, MgO in the remaining melt decreases. Thus, plotting the basalt glass compositions as a function of MgO indicates the behavior of the other major elements as a function of magmatic differentiation. The decrease in Al_2O_3 and CaO with decreasing MgO is consistent with the crystallization of plagioclase and olivine. Crystallization of these phases alone would not result in the observed decrease in CaO; another calcium-rich phase, augite, is required to account for the observed compositional range of basalt glasses. Both FeO^{tot} and TiO_2 behave incompatibly in MORB and increase with decreasing MgO, indicating that neither ilmenite nor Ti-magnetite were crystallizing during the limited fractionation of these rocks. This geochemical dataset indicates that minor differentiation occurs during emplacement of MORB onto the sea floor. More extensive differentiation by crystal fractionation would not be expected, given the rapid cooling of the magma in contact with seawater.

There are two places where hot spot activity has produced enough tholeiitic lava to produce islands that rise above sea level – Iceland and the Galapagos Islands. Iceland is located on the Mid-Atlantic Ridge, and the Galapagos Islands sit slightly off axis of the Galapagos spreading center. Both localities represent places where basalts formed by upwelling and decompression of mantle at a spreading center and are also affected by complex interaction with a mantle plume or hot spot (Harpp et al., 2002; O'Connor et al., 2007). Whereas basalt generated at mid-ocean ridges is commonly considered "normal," or N-MORB, basalts at locations on or near hot spots exhibit higher concentrations of incompatible elements and may be referred to as "enriched." or E-MORB. In addition to enrichments in incompatible elements, both Iceland and the Galapagos contain volcanoes that have erupted lavas with a wide range of compositions, from basalt to rhyolite (Carmichael, 1964; McBirney and Williams, 1969). Lavas from both centers show similar differentiation trends (Figure 6.7). The magma undergoes extensive differentiation when a large change in FeO*/(FeO*+MgO) ratio coincides with a minimal change in silica content. This trend results from the crystallization of olivine and pyroxenes without participation of Fe-Ti oxides. Such differentiation

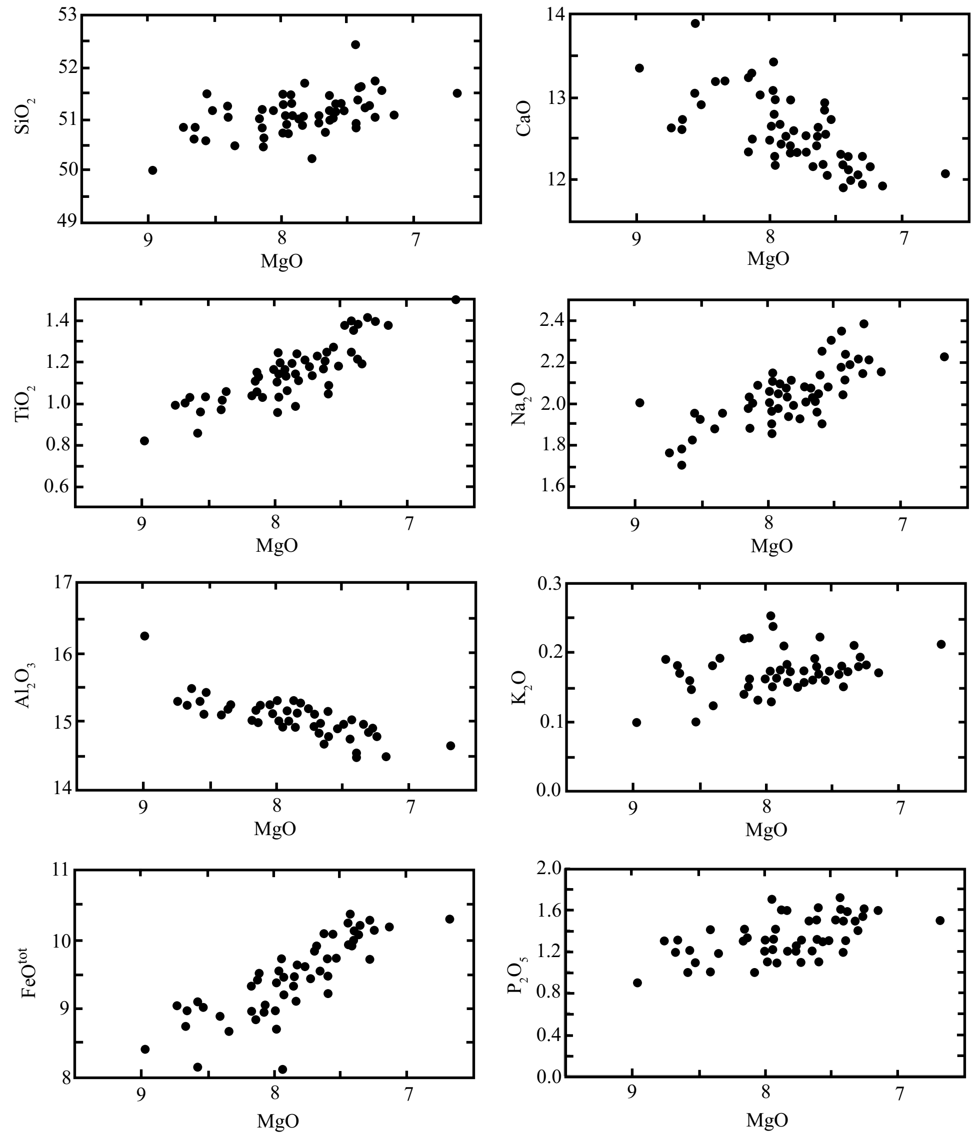


Figure 6.6 Compositions of basalt glasses from the Narrowgate region of the FAMOUS valley, Mid-Atlantic Ridge. Data from Stakes, Shervais, and Hopson (1984).

enriches the magma in iron and leads to the formation of ferrobasalt, a basalt with more than 13 percent FeO* and less than 6 percent MgO (McBirney and Williams, 1969). As such, the rhyolites formed by differentiation of these basalts are ferroan, metaluminous, and calc-alkalic (in Iceland) to alkali-calcic and alkalic (in the Galapagos) (Figure 6.7).

6.3.2 Off-ridge Magmatism

Off-ridge magmatism falls into three broad categories:

1. **Seamounts**: these submarine volcanic structures either never grow enough to breach sea level or if they do, they are eroded and subside. In the tropics, they may be capped with coral reef deposits, but as the volcano subsides only a guyot (a flat-topped seamount that lies just below sea level) remains. Seamounts are most abundant in the Pacific, where they number between twenty-two thousand and fifty-five thousand (Batiza, 1982). Many seamounts appear along fracture zones, which may provide conduits for the magma. Others form linear chains that show a progressive age relationship, and that suggest a genetic relationship with ocean island volcanoes.
2. **Oceanic island volcanoes**: these immense volcanoes rise up to ten thousand meters above the ocean floor and have dimensions greater than the largest

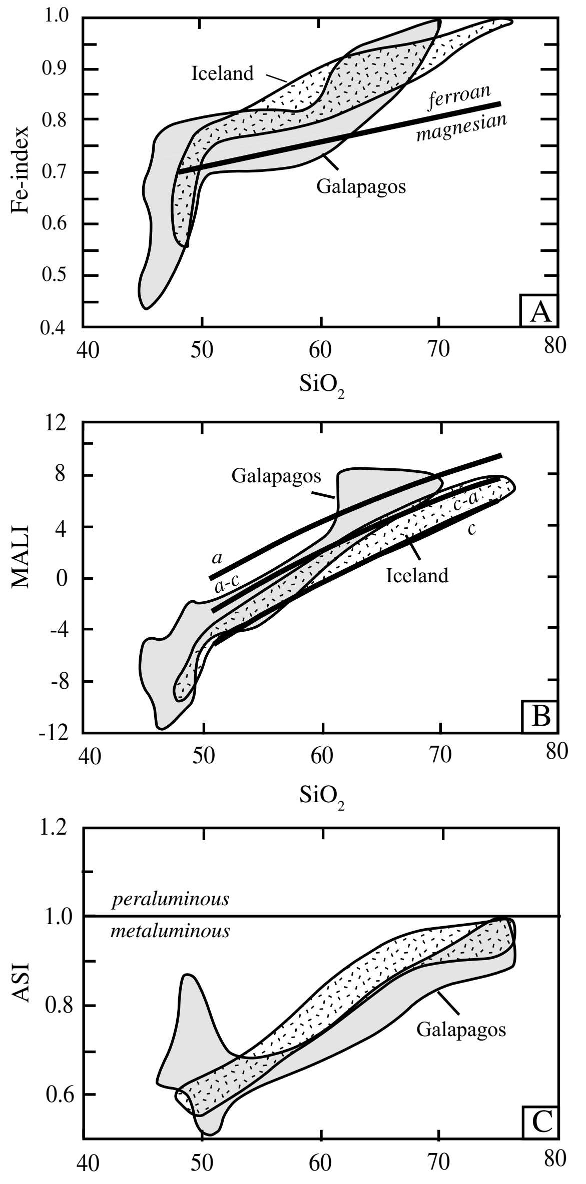
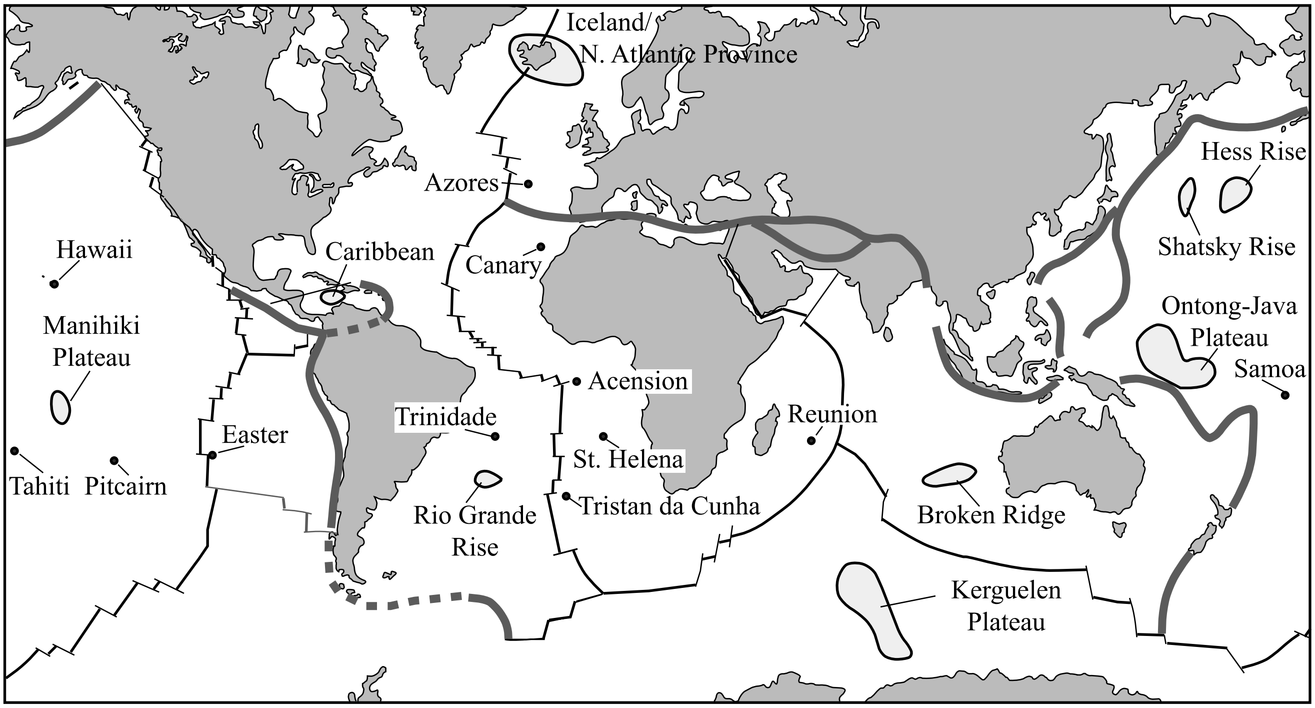


Figure 6.7 Geochemical trends of volcanoes on the Galapagos Islands and Iceland, both of which straddle oceanic ridges. Data from McBirney and Williams (1969) and Carmichael (1964).

mountains on the continents. Usually an ocean island volcano has several centers, suggesting the focus of magmatic activity migrates with time. Ocean island volcanoes may be single islands, or in fast-spreading oceans like the Pacific, they may form linear chains. A good example is the Hawaii-Emperor chain, a chain of seamounts and subaerial oceanic islands that stretches from Hawaii nearly to Kamchatka. In this chain, the oldest volcanoes lie at the northwest end, and the active, but still submarine, volcano of Loihi is situated at the east end. Locations of some ocean islands are shown in Map 6.3.

3. **Oceanic plateau**: Oceanic plateau are topographic highs within ocean basins that have an area of several 100 km^2 and rise an average of a thousand meters above the ocean floor. Large, well-studied oceanic plateau include the Ontong-Java plateau in the western Pacific Ocean and the Kerguelen plateau in the southern Indian Ocean (Map 6.3). Oceanic plateau lie on thickened oceanic crust that may be between ten and thirty-five kilometers thick. Many form at or near mid-ocean spreading centers, and appear to have formed by immense, short-lived eruptions of tholeiitic basalt (Kerr, 2004). Because of their thickness, they are not easily subducted and instead fragments of oceanic plateau may accrete to continental margins. For example, the Ontong-Java plateau collided with the Solomon Islands (Neal et al., 1977; Petterson et al., 1997), and the Caribbean plateau collided with northwestern South America (Kerr et al., 1997). Oceanic plateau are similar to continental flood basalts (see Chapter 8) in that both involve large, rapid outpourings of basalt. Both are sometimes referred to as “large igneous provinces” (Coffin and Eldholm, 1992).

A plume or hot spot model appears to explain many of the intraplate volcanic features of the ocean floor, especially many oceanic island volcanoes and plateau. Plumes may originate from a thermal boundary layer at the core-mantle boundary, or at the base of the upper mantle at 670 kilometers. The rising plume of solid material undergoes decompression melting as it shallows. The composition of the basalt magma depends on the depth and extent of melting and the composition of the mantle diapir. A short-lived voluminous eruption of this basalt may form an oceanic plateau, whereas a plume that produces magma over a longer period of time will build an oceanic island volcano. As plate motion carries the overlying oceanic crust across the plume, the site of volcanic activity shifts to that part of the ocean floor that is directly above the plume. In this way, a chain of hot spot volcanoes develops across the ocean crust at the pace of plate motion.



Map 6.3 Map showing the location of oceanic islands (points) and oceanic plateau (shaded). Plateau locations are after Coffin and Eldholm (1992).

6.3.2.1 Hawaii: An Example of an Oceanic Island Volcano

The Hawaii-Emperor island chain is over two thousand kilometers long. The easternmost island, Hawaii, is the only volcanically active island, although to its east the submarine volcanic center of Loihi is developing (Figure 6.8). Five overlapping shield volcanoes build the island of Hawaii, of which only Kilauea and Mauna Loa are active. Mauna Loa, at 4,170 meters above sea level, and Mauna Kea, at 4,205 meters, have the highest relief above base level of any mountain on earth (>10,000 meters), which indicates the huge amounts of magma involved in their formation. The focus of volcanism on Hawaii moves six centimeters a year, which is essentially identical to the movement rate of the Pacific plate. A stationary hot spot appears to explain the spatial relationships of Hawaiian volcanism very well.

Structure. The Hawaiian islands are associated with a large-amplitude, free-air gravity anomaly, which can be explained in terms of a huge volcanic load that downwarps the oceanic lithosphere. The crust beneath Hawaii is fifteen to twenty kilometers thick, as opposed to five to six kilometers in the adjacent Pacific. The lithosphere is

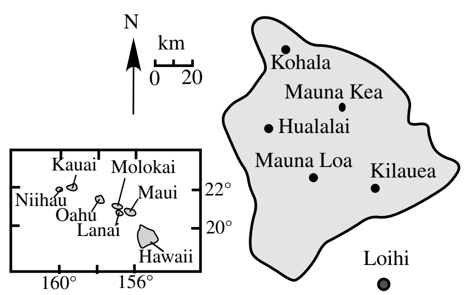


Figure 6.8 The island of Hawaii is a compound volcanic edifice formed by five overlapping shield volcanoes.

estimated to be relatively thin at less than ninety kilometers (Forsyth, 1977). The distribution of magma-related earthquake hypocenters has been used to construct the three-dimensional layout of the magma chamber and feeder conduit for the active volcanic center of Kilauea. It appears that melting occurs below sixty kilometers depth, and this magma is transported up through narrow conduits to shallow magma chambers (Wright, 1984). The main magma storage area is a zone, three kilometers in diameter, which extends from three to six kilometers depth. As magma fills the magma reservoirs

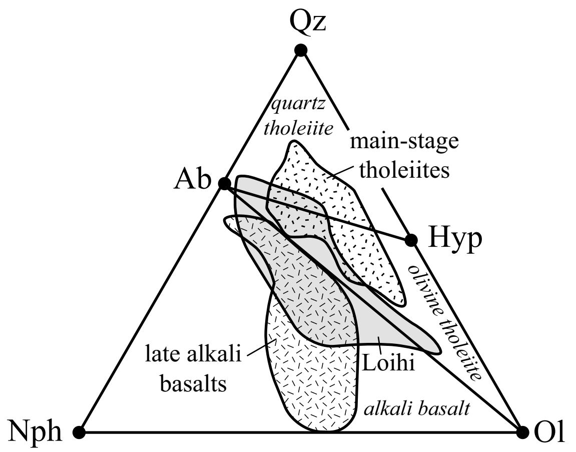


Figure 6.9 Comparison of the compositions of early Hawaiian basalts (as exemplified by Loihi), the main stage tholeiites, and the late-stage alkali basalts projected onto the normative Qz-Nph-Ol triangle. Data from MacDonald (1968), Fodor and colleagues (1992), Frey and Clague (1983), Hawkins and Melchior (1983), Muir and Tilley (1957), Ren and colleagues (2009), Wilkinson and Hensel (1988), and West and colleagues (1992).

the summit of the volcano inflates, as is measured by tilt meters. After a major eruption, it takes a period of months for the Kilauea magma chamber to refill (Ryan, Koyanagi, and Fiske, 1981; Dzurisin, Koyanagi, and English, 1984).

Evolution of an oceanic island volcano. Nearly all oceanic island volcanoes exhibit a complex pattern of igneous evolution, although the pattern is not the same for all ocean islands. At Hawaii, four characteristic stages can be identified. The first, the pre-shield stage, is seen at Loihi, the seamount south of Hawaii constructed of alkali basalt and tholeiite (Figure 6.9). The second, the shield-building stage, consists of large volumes of tholeiitic basalt. As volcanism wanes, the eruptions become more explosive and more alkaline in composition. The volume of magma produced during this post-shield stage is only about one percent of the total production of the volcano. A long period of dormancy and erosion initiates the post-erosional stage of volcanism. Again the eruptions are explosive, and the magmas produced are highly alkaline and silica poor. The rock types produced are alkali basalts and nepheline-bearing basalts. The amount of time between stages varies from volcano to volcano, and can be between one hundred thousand years to two million years. On the island of Hawaii, Mauna Loa and Kilauea are in the shield-building stage, while the three older volcanoes (Mauna Kea, Hualalai, and Kohaala) are between the post-shield stage and the post-erosional stage.

Petrography and geochemistry. Hawaii is a little unusual compared to other oceanic island volcanoes in the large volume of tholeiite and small volume of alkaline rock it has produced. Most other hot spots, such as the Azores, St. Helena, and Tristan da Cunha, have a larger proportion of alkaline rocks. Oceanic island tholeiites are similar to MORB, but may have orthopyroxene as well as olivine, spinel, clinopyroxene, and Fe-Ti oxides. In terms of major element chemistry, these tholeiites have higher K_2O and TiO_2 relative to MORB, but lower Al_2O_3 contents. Alkali basalts commonly contain ultramafic xenoliths, whereas ultramafic xenoliths are rare in tholeiites.

Magma sources for Hawaii volcanism. Neodymium, strontium, and lead isotopic data from Hawaiian volcanic centers strongly suggest that no single mantle source of magma can produce all the rock types found there. The interaction of an upper mantle source that also supplies the mid-ocean ridges with a lower-mantle plume source can account for much of the isotopic variability at oceanic island volcanic centers. In addition some ocean island lavas require the incorporation of a material that has isotopic characteristics like the crust. One possible contributor is the subduction of oceanic crust. The subcontinental lithosphere also may be recycled into the mantle and could become incorporated into oceanic island magmas. A combination of sources, along with different degrees of partial melting, seems to be involved, although the processes are complex and the details remain incompletely resolved.

6.3.2.2 Ontong-Java: An Example of an Oceanic Plateau

The Ontong-Java plateau is the largest known oceanic plateau, with a volume of 44.4 million km^3 occupying an area of 1.9 million km^2 of Pacific sea floor (Kerr, 2004). It is composed of tholeiitic basalt erupted in two short episodes at 122 ± 3 and 90 ± 4 million years (Neal et al., 1997). The crust on the Ontong-Java plateau averages thirty-six kilometers thick, and the plateau sits two to three kilometers above the surrounding sea floor today.

Although all lavas are tholeiitic basalts with between six and eight percent MgO, Neal and colleagues (1997) identified two groups of lavas on the basis of incompatible

BOX 6.2 | ORE DEPOSITS IN OCEANIC CRUST

Most ophiolites lack economic mineral deposits, but two types of important ore deposits are locally found in ophiolites: chromite and copper-bearing, massive sulfide deposits.

Chrome: Concentrations of chromite are common in the peridotite tectonite portion of an ophiolite (see Figure 6.1). In some localities the chrome concentration is rich enough to be ore grade. Because the peridotite and chromite bodies within an ophiolite have been subjected to ductile deformation within the mantle, the ore bodies have a discontinuous, pod-like shape. Hence they are called **podiform chromite** deposits to distinguish them from chromite deposits associated with layered mafic intrusions, which are called **stratiform chromite** deposits (see Box 9.1). Although the stratiform deposits are much larger than podiform deposits, podiform deposits contain chromite much richer in chrome, and are therefore valuable despite their smaller size. The chromite deposits of Kazakhstan and Turkey, the third and fourth largest chrome producers in the world after South Africa and India, come from ophiolites.

Box 6.2 Copper ingot in the shape of an oxhide, from Cyprus, circa 1225–1150 BCE. Artifact held by the British Museum, London. Photo used with permission. © The Trustees of the British Museum / Art Resource, NY.

Copper: Associated with pillow basalts in some ophiolites are massive sulfide deposits consisting of pyrite, pyrrhotite, chalcopyrite, and sphalerite. How these deposits formed was a great mystery until the discovery of "black smokers" on the sea floor (Corliss et al., 1979). Geologists now recognize that the intrusion of gabbroic rocks at a ridge crest drives circulation of hydrothermal fluid through the overlying basalts. As these fluids react with the basaltic crust, they extract metals, mostly Fe, Cu, and Zn, as well as sulfur from the sulfides in the basalt. When these fluids are expelled into the sea, the sulfides precipitate as "black smoke," which cools and deposits as **volcanogenic massive sulfide (VMS)** deposits. The most famous ophiolite-hosted VMS deposits are associated with the Troodos ophiolite in Cyprus. Copper has been mined on Cyprus since the fourth millennium CE. Copper mining ceased in 1979, but there has recently been interest in restarting production.

BOX 6.3 | VOLCANIC HAZARDS FROM OCEANIC VOLCANISM

Despite the fact that oceanic volcanism is by far the most voluminous magmatism on Earth, it is associated with comparatively low volcanic hazard. This is partly because most eruptions take place under the ocean where eruptions are isolated from the atmosphere by a kilometer or more of seawater. Equally important, oceanic magmas are dominated by highly fluid basalt that generally lacks the explosive nature of arc magmatism. However, oceanic magmatism is not without its volcanic hazards, as any air traveler to Europe in the summer of 2011 could tell you. The eruption of the Eyjafjallajökull volcano in Iceland ejected a plume of ash that disrupted air travel to western Europe for weeks. This ash plume was small compared to the fissure eruption of Laki, also on Iceland, in 1783. Laki is the largest fissure eruption in history, and the volcanic gases emitted by the eruption killed many of the livestock on the island as well as approximately a quarter of the population because of famine and fluorine poisoning.

The islanders living on Hawaii are accustomed to the volcanic hazards of the island, which they attribute to activity of the goddess Pele. Volcanic eruptions and associated destruction of human structures are part of Hawaii's history. In the most recent of Pele's tricks, slow-moving lava flows from Kilauea overran the Kalapana Gardens Subdivision near Hilo. Much of the area was destroyed in 1986; the last house was overrun in 2012. One of the owners of the last houses to be engulfed in lava said they stayed until the last minute because "it is very easy to outrun a lava flow."

Box 6.3 Lava from Kilauea engulfing the Kalapana Gardens Subdivision in 1990, causing residents to abandon their community as it was consumed by the slow-moving flows. Photo from U.S.G.S. Hawaii Volcano Observatory, http://hvo.wr.usgs.gov/kilauea/history/1990Kalapana/.

element concentrations and isotopic compositions. They suggested the two types of basalt were likely the product of variable degrees of partial melting from multiple mantle sources. Based on the range of major element compositions recorded in the basalts, Neal and colleagues (1997) concluded that the parent magmas subsequently underwent crystal fractionation of olivine followed by clinopyroxene and plagioclase. Gabbroic enclaves in some of the flows evince removal of clinopyroxene and plagioclase during the later stages of differentiation. The apparent complexity of the processes that acted over geologically short periods of time is a reminder that much remains to understand about the genesis of oceanic plateau and how they differ from oceanic islands.

One potentially important aspect of the formation of oceanic plateau is their possible impact on Earth's environment. There is tantalizing evidence of temporal correlations between the formation of oceanic plateau and other large igneous provinces and oceanic anoxia, rapid global warming, and mass extinctions. Beyond the temporal coincidence, the causal links between these global impacts and the outpouring of large quantities of basaltic magma are not well established, representing a fruitful area for future research.

Summary

- Ophiolites are commonly used as a simplified model for ocean crust. The stratigraphy of an ophiolite from top to bottom is: pillow basalt, sheeted dikes, gabbro, cumulate peridotite, and deformed peridotite.
- Ophiolites may form in back-arc spreading centers, in suprasubduction zone settings, or in mid-ocean spreading centers.
- Oceanic crust is more complex than the ophiolite model suggests. Parts of the crust may be excised by faults, and the gabbroic horizon may involve multiple injections of melt that may interact with mantle host peridotite and evolve toward oxide gabbro.
- Ocean islands and oceanic plateau form above mantle plumes ("hot spots"). Basaltic magma forms by decompression melting within rising mantle diapirs. The composition of the basalt magma varies depending on the depth and extent of partial melting and the composition of the mantle involved.
- Tholeiites are found in MORB, ocean islands, and oceanic plateau.
- Alkali basalts are found on many ocean islands, and on some of these islands they have evolved to extremely alkaline magmas.

Questions and Problems

Problem 6.1. Compare and contrast fast mid-ocean spreading centers with slow- and ultra-slow- mid-ocean spreading centers. Include in your comparison:

- topography and dimensions of the ridge,
- stratigraphy and structure of the rocks produced,
- differences in plagioclase compositions, and
- magma volumes and differentiation histories.

Problem 6.2. How can hotspot tracks be used to determine plate motion? What assumptions are involved?

Problem 6.3. What types of basalt are found at mid-ocean ridges? At oceanic islands? What might account for any differences?

Problem 6.4. Plot data for MORB glasses from the FAMOUS area and basalts from Galapagos on Fe-index, MALI, and ASI diagrams. How do the two groups of basalts differ? How might these differences relate to the presence or absence of mantle plume activity?

Further Reading

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