Chapter 10

Igneous rock associations

10.1 IGNEOUS ROCK ASSOCIATIONS

The purpose of this chapter is to relate igneous rock associations to a petrotectonic framework, incorporating information presented in Chapters 7–9. **Petrotectonic associations** are suites of rocks that form in response to similar geological conditions. These associations most commonly develop at divergent plate boundaries, convergent plate boundaries and hotspots (Figure 10.1). Hotspots can occur at lithosphere plate boundaries (e.g., Iceland) or in intraplate settings (e.g., Hawaii). Fisher and Schmincke (1984) estimate the percent of magma generated at modern divergent, convergent and hotspot regions as 62, 26 and 12%, respectively.

While plate tectonic activity plays a critical role in the development of petrotectonic associations, it is not the sole determining factor. For example, the earliest onset of modern plate tectonics continues to be debated, with some researchers (Kusky et al., 2001; Parman et al., 2001) favoring Archean (>2.5 Ga) onset and others (Hamilton, 1998, 2003; Stern, 2005, 2008; Ernst, 2007) proposing Proterozoic initiation of deep subduction ~1 billion years ago (1 Ga). If the latter is true, then over 75% of Earth's magmatic history occurred under conditions that pre-date the onset of modern plate tectonic activity. In addition to questions regarding magmatism at Precambrian plate tectonic boundaries, Phanerozoic intraplate magmatism may or may not be influenced by lithospheric plate boundaries (Hawkesworth et al., 1993; Dalziel et al., 2000). So while the plate tectonic paradigm is very useful, it does not address all igneous rock assemblages produced throughout Earth's tumultuous history. In the following sections we will address major igneous petrotectonic associations, beginning with divergent plate boundaries.

10.2 DIVERGENT PLATE BOUNDARIES

Decompression of the asthenosphere in response to lithospheric extension results in partial melting of mantle peridotite at divergent margins. Basic (basaltic) melts rise and

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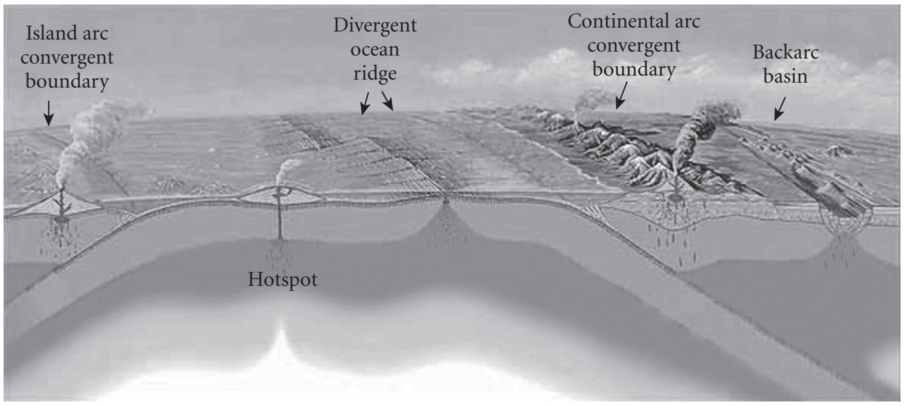

Figure 10.1 Major tectonic environments where igneous rocks occur. (Courtesy of the US Geological Survey and US National Park Service.)

solidify to produce oceanic crust, while refractory residues cool below a critical temperature to form the thickening mantle layer of ocean lithosphere. Ocean lithosphere is created primarily at spreading ridges such as the Mid-Atlantic Ridge, East Pacific Rise and Indian–Antarctic ridge systems. A small percentage of ocean lithosphere is generated in backarc basin spreading ridges (e.g., Marianas Trough) and ocean hotspots (e.g., Hawaii). In all cases, anatexis of ultramafic mantle is the primary magmatic source of ocean lithosphere.

Ocean lithosphere contains four distinct layers as indicated in Figure 10.2a. Layer 1 contains well-stratified marine pelagic sediments and sedimentary rocks that accumulate on the ocean floor. Layer 2 can be subdivided into two basaltic rock layers. An upper layer contains pillow basalts that develop when basic lavas flow onto the ocean floor, rapidly cool in the aqueous environment and solidify in spheroidal masses (Chapter 9). Beneath the pillow basalt pile, basic magma injects into extensional fractures producing steeply inclined diabase dikes as the magma cools and contracts. Repeated horizontal extension and magma intrusions generate thousands of dikes arranged parallel to one another in a sheeted dike complex (Chapter 8). Beneath the sheeted dike layer, basic magma cools slowly, allowing phaneritic crystals to nucleate and grow as layer 3. Layer 3 contains massive (isotropic) gabbro in the upper section, layered (cumulate) gabbro in a middle section, and increasing amounts of layered (cumulate) peridotite towards the bottom of the section, marking the base of ocean crust. The Mohorovičić discontinuity (Moho) occurs at the contact between cumulate rocks in layer 3 and non-cumulate, metamorphosed rocks in layer 4, marking the rock boundary between the ocean crust and mantle. Layer 4 is composed of depleted mantle peridotite refractory residue (e.g., harzburgite, dunite). Layer 4 mantle peridotite is marked by high temperature, solid state strain fabric (metamorphosed) and represents the lowest layer of the oceanic lithosphere.

Layers 3 and 4 are generally unexposed on ocean floors because they are overlain by layers 1 and 2. In rare locations, these deep layers are exposed on the ocean floor in ultra slow (<1 cm/yr), or magma-starved, spreading ridges and transform zones where brittle faulting and uplift processes bring them to the surface. Slices of ocean lithosphere are also preserved in alpine orogenic belts as ophiolite sequences. Let us now consider petrotectonic assemblages that form at ocean ridge spreading centers.

10.2.1 Mid-ocean ridge basalts

At ocean spreading centers (Figure 10.2b–c), partial melting of lherzolite (peridotite) generates voluminous, geochemically distinct, **mid-ocean ridge basalts** (MORB) and gabbros containing minerals such as plagioclase, augite, hypersthene, pigeonite, diopside and olivine. MORB are the most abundant volcanic rocks on Earth. Typical major and

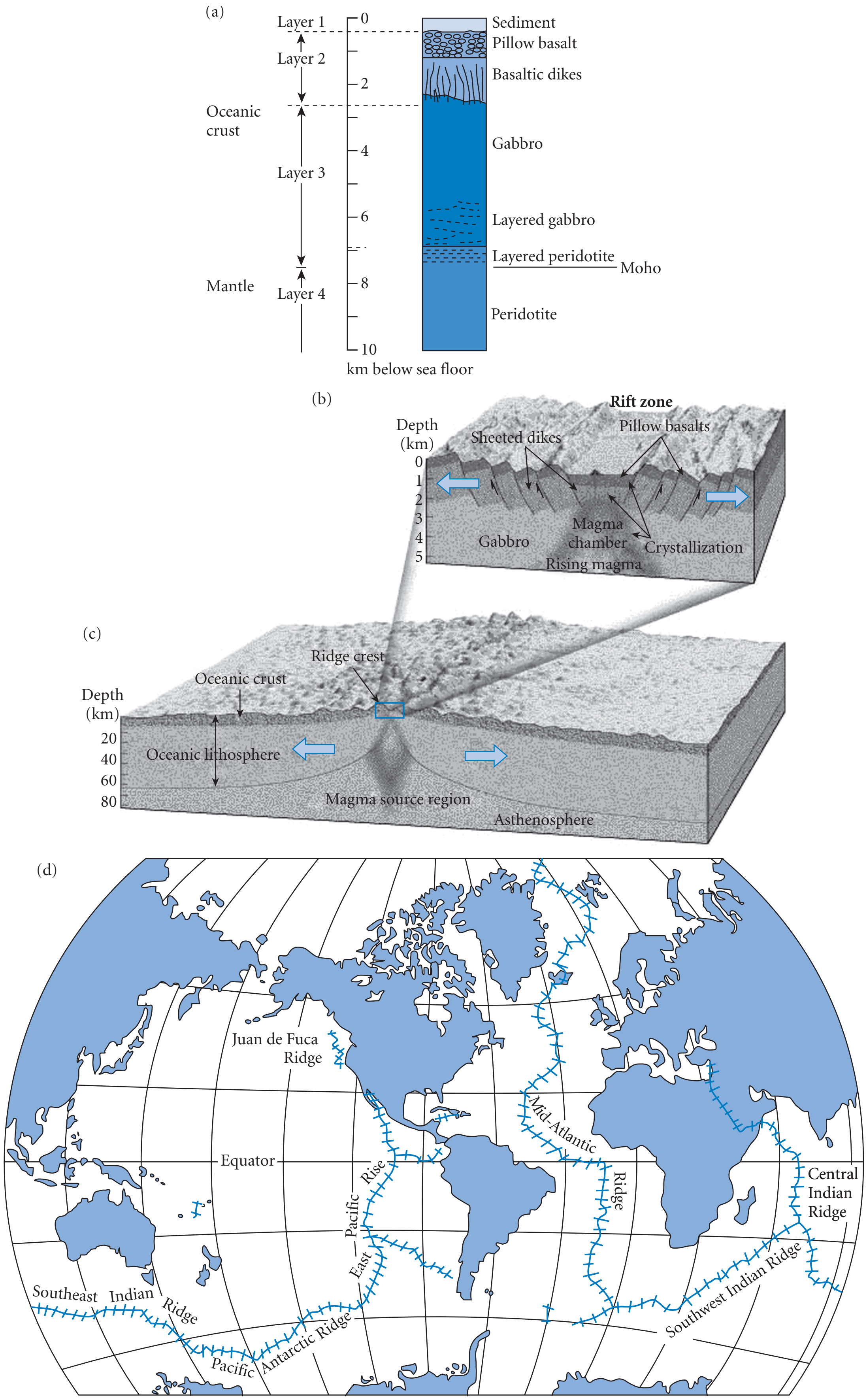
(a)
Layer 1
Layer 2
Oceanic crust
Layer 3
Mantle
Layer 4
0
2
4
6
8
10
km below sea floor
Sediment
Pillow basalt
Basaltic dikes
Gabbro
Layered gabbro
Layered peridotite
Moho
Peridotite
(b)
Depth (km)
0
1
2
3
4
5
Rift zone
Sheeted dikes
Pillow basalts
Gabbro
Magma chamber
Crystallization
Rising magma
(c)
Ridge crest
Oceanic crust
Depth (km)
20
40
60
80
Oceanic lithosphere
Magma source region
Asthenosphere
(d)
Juan de Fuca Ridge
Equator
Mid-Atlantic Ridge
East Pacific Rise
Central Indian Ridge
Southeast Indian Ridge
Pacific Antarctic Ridge
Southwest Indian Ridge

Table 10.1 Trace element abundances for N-MORB and E-MORB in parts per million (ppm). (After Best, 2003; data from Sun and McDonough, 1989.)

	LIL					HFS		LREE					HREE				
	Cs	Rb	Ba	Th	U	Nb	Ta	La	Ce	Pr	Nd	Sm	Zr	Eu	Gd	Yb	Lu
N-MORB	0.007	0.56	6.3	0.12	0.47	2.33	0.132	2.5	7.5	1.32	7.3	2.63	74	1.02	3.68	3.05	0.455
E-MORB	0.063	5.04	57	0.6	0.18	8.3	0.47	6.3	15	2.05	9	2.6	73	0.91	2.97	2.37	0.354

E-MORB, enriched mid-ocean ridge basalt; HFS, high field strength; HREE, heavy rare Earth elements; LIL, large ion lithopile; LREE, light rare Earth elements; N-MORB, normal mid-ocean ridge basalt.

minor element concentrations are indicated in Table 10.1. MORB are low SiO_2 (45–52%), low potassium (<1% K_2O) tholeiites with high MgO (~7–10%), Al_2O_3 (15–16%) and compatible element concentrations (Ni and Cr ~100–500 ppm). MORB develop from partial melting of a depleted mantle source, as indicated by low $^{87}Sr/^{86}Sr$ ratios (0.702–0.704), low volatile and incompatible element concentrations, and high compatible element concentrations (Cann, 1971). "Depleted source" refers to mantle lherzolite that has undergone previous melt cycles that largely removed mobile incompatible elements (Chapter 7).

Mid-ocean ridge basalts can be subdivided into **normal MORB** (N-MORB) and **enriched MORB** (E-MORB) based upon minor and trace element abundances (Figure 10.3; Table 10.1). N-MORB are strongly depleted in highly incompatible elements such as large ion lithophile (LIL) elements (such as Cs, Rb and Ba), high field strength (HFS) elements (such as Nb and Ta) and light rare Earth elements (LREE, such as La, Ce, Pr, Nd and Sm). These geochemical characteristics imply that

Figure 10.2 (a) Idealized stratigraphy of ocean lithosphere and ophiolites. Note the petrological Moho between layers 3 and 4, separating the base of the crust from the upper mantle. (Courtesy of the Ocean Drilling Project.) (b) Block diagram of ocean ridge divergent margins. (c) Ocean ridges are primary sites for the generation of ocean lithosphere. (d) The global distribution of divergent margins.

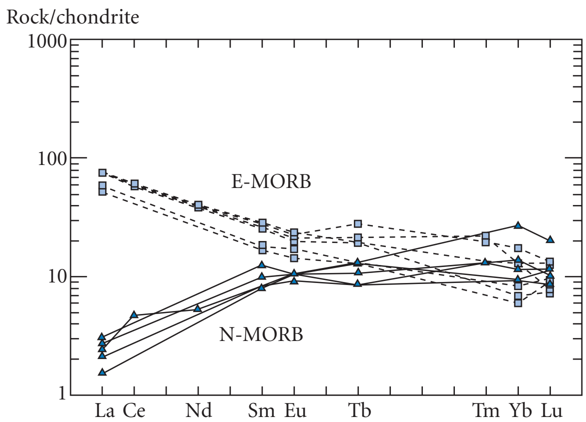

Figure 10.3 Chondrite-normalized rare Earth element patterns for enriched and normal mid-ocean ridge basalt (E-MORB, squares; N-MORB, triangles) samples collected from the Mid-Atlantic Ridge. (From Schilling et al., 1983; with permission of the *American Journal of Science*.)

N-MORB magma represent 20–30% partial melting of a well-mixed, depleted mantle source (Frey and Haskins, 1964; Gast, 1968).

Although the major element and heavy rare Earth elements (HREE, ranging from Eu to Lu) concentrations are comparable, E-MORB have higher incompatible element (LREE, HFS, LIL) concentrations relative to N-MORB. Specifically, E-MORB are defined by having chondrite normalized La/Sm ratios of >1. Lanthanum may occur in concentrations of 1–5 ppm in N-MORB but up to 100 ppm in E-MORB.

How can we account for chemical variations between N-MORB and E-MORB?

Several different hypotheses have been proposed. First, E-MORB may represent smaller degrees (~10–15%) of partial melting of residual mantle rock so that the incompatible elements are more highly concentrated in E-MORB magmas. Second, E-MORB could be tapping a deep mantle source that has not been previously melted. Third, E-MORB could represent magma enriched from magma mixing, assimilation or partial melts derived from subducted ocean lithosphere. For example, Eiler et al. (2000), based on a study of 28 basalt samples from the Atlantic, Pacific and Indian ridges, propose that E-MORB include a component of partially melted oceanic lithosphere that has been recycled into the upper mantle from ancient subduction zones.

While MORB is the dominant volcanic rock type at divergent margins, other rock types occur in varying proportions. Ocean ridges also produce high aluminum basalts, where the Al_2O_3 concentrations are >16%. Other rocks such as andesite, icelandite, ferrobasalt, trachyte, hawaiite, mugearite, trachybasalt, trachyandesite, dacite and rhyolite can occur as minor components at ocean ridges as well as in "leaky" transforms, continental rifts and ocean islands. The andesitic to rhyolitic volcanic rocks at ocean ridges have higher TiO_2 (>1.3%) concentrations compared to more common convergent margin varieties and are always subordinate to basalt (Gill, 1981).

Divergent margins (see Figure 10.2d) generate the bulk of ocean floor rocks, which represent ~70% of Earth's area. As a result, the ocean ridge basalt and underlying gabbro and peridotite are widespread in our relatively young ocean basins, all of which are less than 200 million years old. What happens to old ocean lithosphere? At least for the past 1 billion years, it has been subducted and recycled at convergent margins, as discussed below.

10.3 CONVERGENT PLATE BOUNDARIES

While divergent plate boundaries are dominated by MORB, chemically diverse igneous assemblages erupt in the convergent margins widely distributed in the Pacific Ocean, eastern Indian Ocean and the Caribbean and Scotia Seas (Figure 10.4).

Convergent margin magmatism may occur for thousands of kilometers parallel to the trench, and up to 500 km perpendicular to the trench in the direction of subduction (Gill, 1981). Plutonic rocks at convergent margins include diorite, granodiorite, quartz diorite, granite, gabbro, tonalite and rocks referred to as trondhjemite. This plutonic suite of rocks occurs in batholiths above subduction zones and provides magma to overlying volcanic arcs. The spectrum of possible volcanic rock types varies widely from youthful island arc environments – dominated by arc basalts and basaltic andesites – to mature continental arc systems – comprised largely of andesites, with lesser amounts of basalt, dacites, rhyodacites and rhyolites.

As opposed to relatively simple decompression melting of the mantle at divergent margins, convergent margin magmatism is affected by more variables, each of which can diversify magma composition. These variables include:

- Composition (continental versus oceanic) and thickness of the overlying converging plate: thinner ocean lithosphere in the overlying plate generally produces metaluminous, mafic to intermediate rocks. Thicker continental lithosphere overlying the subduction zones commonly yields peraluminous, potassic, intermediate to silicic rocks.
- Composition of rock material experiencing anatexis: Earth materials experiencing partial melting may include overlying ultrabasic mantle wedge, basic to silicic forearc basement, subducted basic–ultrabasic ocean lithosphere, and marine sedimentary material. The relative proportion of each of these components affects the composition of plutonic and volcanic rocks generated in the arc system.
- Flux melting whereby volatile-rich minerals such as micas, amphiboles, serpentine, talc, carbonates, clays and brucite release H_2O, CO_2 or other volatile vapors that lower the melting temperature of mantle peridotite and eclogite (high pressure metagabbro) overlying the subduction zone.
- Diversification processes such as fractionation, assimilation and magma mixing (Chapter 8) as well as metamorphic reac-

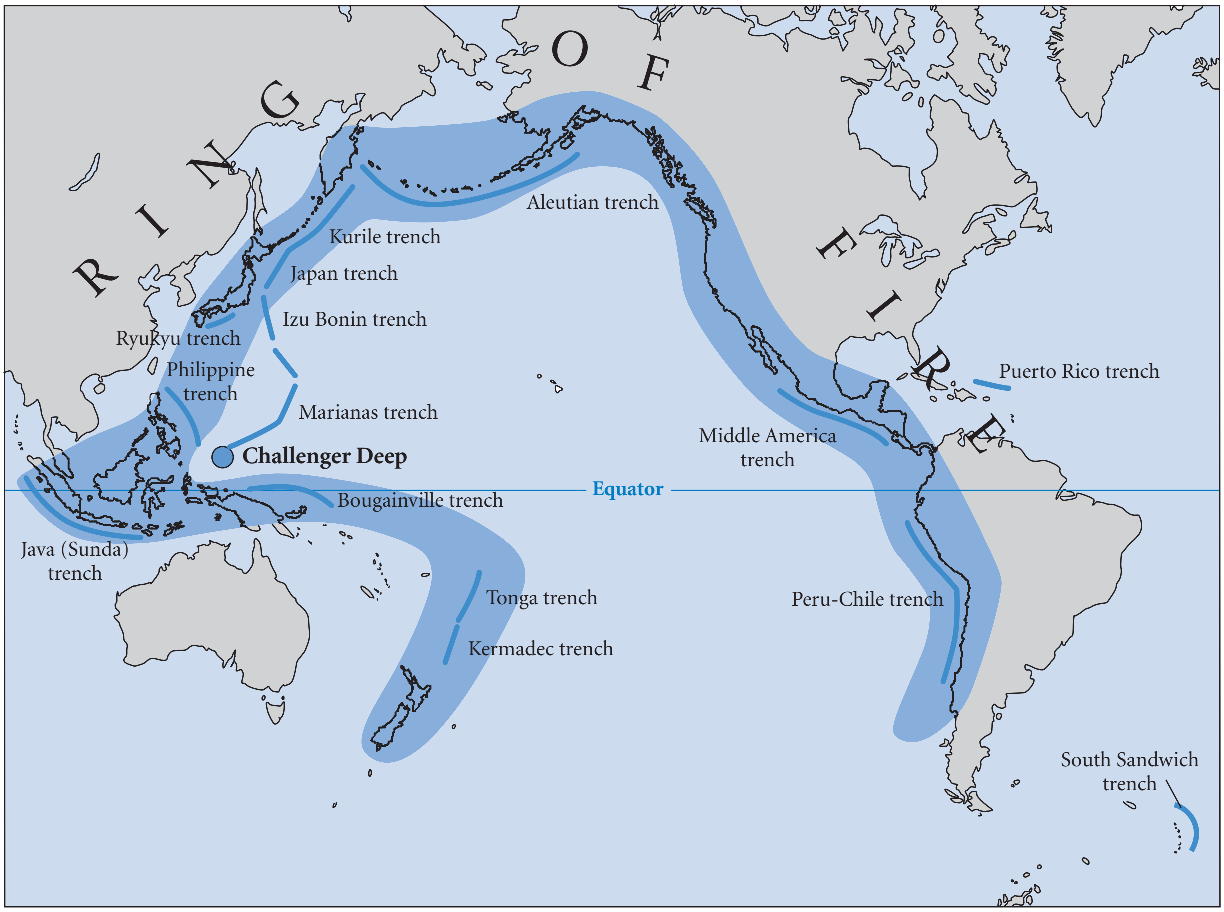

Figure 10.4 Earth's convergent margins. (Courtesy of the US Geological Survey.)

tions (Chapter 15) strongly alter magma composition generated in the overlying wedge of the arc system.

- Dip angle of the subduction zone wherein old, cold, dense lithosphere favors steep subduction and young, warm, buoyant lithosphere produces shallow subduction zones (Figure 10.5). Steeply inclined subduction zones allow for the melting of thick wedge-shaped mantle slabs in the overlying plate. Shallowly dipping subduction zones allow only thin wedge-shaped mantle slabs to intervene above the subducting plate, minimizing overlying mantle wedge input. The negative buoyancy of old, cold, dense ocean lithosphere is the key force driving deep lithosphere subduction in modern plate tectonics over the past 1 Ga. This negative buoyancy may not have been present in the hot, buoyant Archean ocean lithosphere such that deep subduction may not have been possible (Davies, 1992; Ernst, 2007; Stern, 2008). This may explain the origin of some unique Archean rock assemblages as well as the virtual absence of Archean blueschists discussed later in this text.

While magma composition is highly variable based on the factors described above, Phanerozoic convergent margins are dominated by the calc-alkaline suite of rocks whose chemistry is enriched in SiO_2, alkalis (Na_2O and K_2O), LIL, LREE and volatiles and is relatively depleted in FeO, MgO, HFS and HREE concentrations (Miyashiro, 1974; Hawkesworth et al., 1993; Pearce and Peate, 1995). The presence of hydrous minerals such as hornblende and biotite indicates that arc magmas contain >3% H_2O. Volatiles play an important role in subduction zone flux melting.

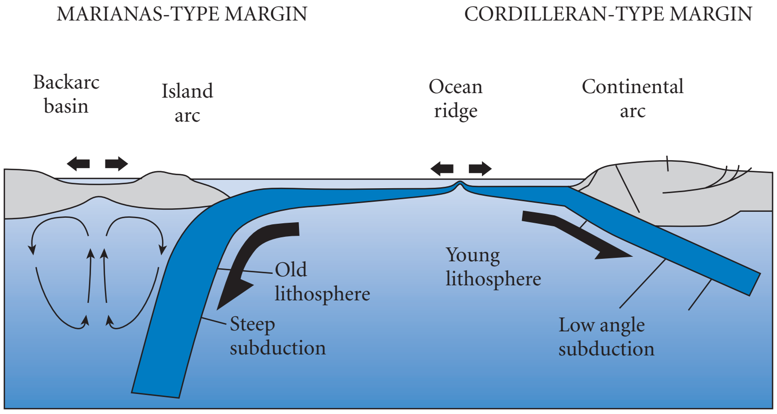

Figure 10.5 The steeply dipping Marianas-type island arc subduction model and the shallowly dipping Cordilleran continental arc subduction model. Note the thick, deep, mantle wedge overlying the Marianas-type margin and the thin mantle wedge in the Cordilleran model.

The calc-alkaline association of **basalt, andesite, dacite and rhyolite** (**BADR**) is the signature volcanic rock suite of convergent margins and constitutes one of the most voluminous rock assemblages on Earth, second only to MORB (Perfit et al., 1980; Grove and Kinzler, 1986). Harker diagram plots of major elements (see Figure 8.11) generally indicate a liquid line of descent from a common source, such that BADR rocks are derived from a common parent magma of basaltic composition. Andesite, named for South America's Andes Mountains, which overlie the Peru–Chile trench, is by far the most common calc-alkaline volcanic rock forming at convergent margins (see Figure 10.4). The more silicic (dacite, rhyolite) members of the BADR group represent more highly fractionated daughter products. We will discuss each of these below.

The major rock types in volcanic arc systems can be distinguished based upon major element concentrations such as SiO_2, K_2O and Al_2O_3 content. Basalts contain 45–52% SiO_2 and can be subdivided into a number of different varieties based upon major and minor element concentrations. Basalts common in convergent margins include aphanitic and aphanitic–porphyritic varieties of arc tholeiites (low K_2O) and calc-alkaline basalts (moderate K_2O). Plagioclase phenocrysts are common. The arc tholeiites differ from other tholeiitic basalts (MORB and ocean islands) in containing higher concentrations of Al_2O_3, typically in concentrations greater than 16 wt %. As a result, arc tholeiites are also referred to as high aluminum basalts. The calc-alkaline basalts differ from tholeiites in having higher alkali (notably K_2O) concentrations and not displaying iron enrichment typical of tholeiitic fractionation trends (Figure 10.6). As discussed in Chapters 7 and 9, magma viscosity and explosiveness are proportional to SiO_2 increases. As a result, the more siliceous volcanic rocks described below commonly produce pyroclastic tuff and breccia deposits in addition to aphanitic to aphanitic–porphyritic crystalline textures.

Andesites are volcanic rocks containing >52–63% SiO_2. Andesites can be subdivided based upon the range of SiO_2: basaltic andesites, common in youthful island arc systems, contain >52–57% SiO_2 while more silicic andesites, common in mature continental arc systems, contain >57–63% SiO_2 (see Figure 7.24). Andesites commonly occur as gray, porphyritic–aphanitic volcanic rocks with phenocrysts of plagioclase, hornblende, pyroxene or biotite. Plagioclase phenocrysts are most common and may display euhedral, zoned crystals. Hornblende phenocrysts are also common and may display reaction rims. Pyroxene (principally augite, hypersthene or pigeonite) and biotite phenocrysts less commonly occur. Quartz, potassium feldspar, olivine or feldspathoid phenocrysts are rare. Interestingly, the bulk composition of andesite (and its plutonic equivalent diorite) approximates that of terrestrial crust, suggesting that subduction zone processes have played a significant role in the development

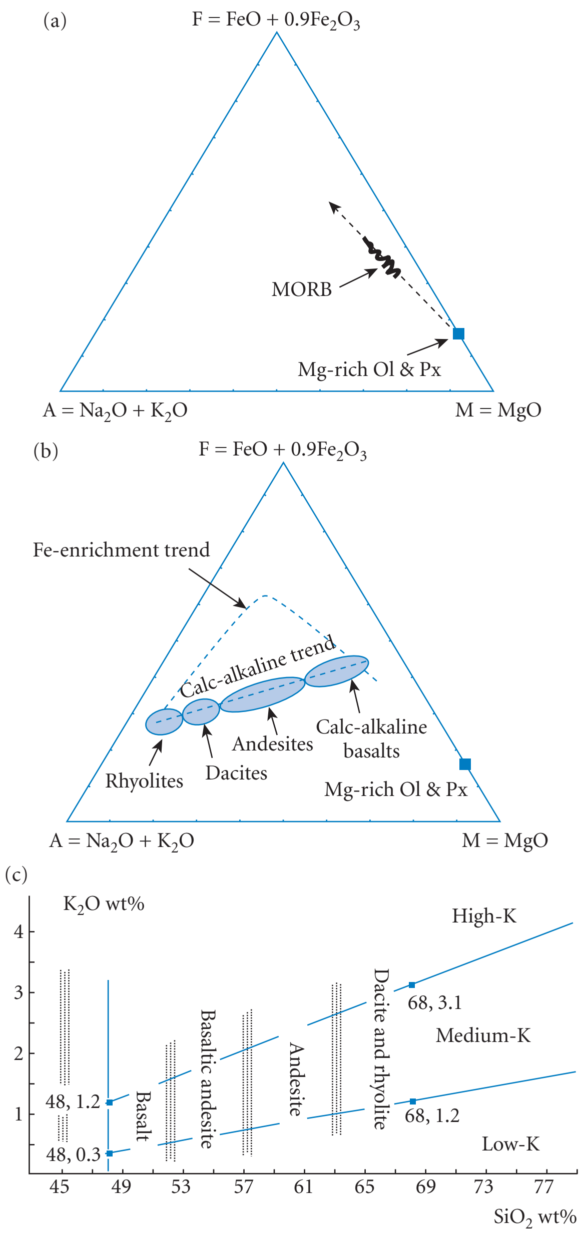

Figure 10.6 (a) Tholeiitic mid-ocean ridge basalts (MORB) display iron enrichment due to the early crystallization of magnesium-rich olivine and pyroxenes. (b) The calc-alkaline suite does not display significant iron enrichment but displays alkali enrichment with progressive crystallization. (c) Three volcanic rock suites are recognized on the basis of percent Si_2O and KO_2: low potassium assemblages consist of tholeiitic basalt; medium potassium assemblages contain calc-alkaline assemblages; high potassium suites consists of high potassium calc-alkaline rocks and trachyandesite rocks referred to as shoshonites (Gill, 1981; LeMaitre, 2002).

of continental crust (Hawkesworth and Kemp, 2006). The generation of voluminous andesite is favored by subduction angles greater than ~25°, anatexis of thick (greater than ~25 km), continental, hanging wall plates and partial melting of subducted slabs at depths of 70–200 km (Gill, 1981).

Dacites (Chapter 7) are quartz–phyric volcanic rocks, intermediate between andesite and rhyolite (Gill, 1981). While most dacites contain 63–68% SiO_2, the total alkali to silica (TAS) dacite classification extends to 77% SiO_2 (see Figure 7.24). Dacites are enriched in plagioclase and are the volcanic equivalent of granodiorites, in which alkali feldspars are subordinate to plagioclase. When present, phenocrysts are commonly subhedral to euhedral, zoned and generally consist of oligoclase to labradorite plagioclase or sanidine. Minor minerals commonly include biotite, hornblende, augite, hypersthene and enstatite.

Trachyandesites (also known as latites and **shoshonites**) are generally composed of ~66–69% SiO_2, although the lower TAS limit begins at 57% SiO_2 (see Figure 7.24). Trachyandesites commonly contain phenocrysts of andesine to oligoclase plagioclase feldspar amidst a groundmass of orthoclase and augite.

Rhyolites (>69% SiO_2) and **rhyodacites** (~68–73% SiO_2) are associated with explosive silicic eruptions producing fragmental, glassy and aphanitic to aphantic–porphyritic textures. Rhyodacite is a rock term, not recognized by the IUGS system, for intermediate volcanic rocks that bridge the dacite/rhyolite boundary. These rocks can occur as glasses (obsidian or pumice), pyroclastic tuffs and breccias, or as aphanitic to aphanitic–porphyritic crystalline rocks. Common phenocrysts include alkali feldspar or quartz, with minor concentrations of hornblende and biotite.

In addition to variations in SiO_2, arc rocks display significant variation in K_2O concentrations, ranging from low (tholeiitic), medium (calc-alkaline) and high K_2O (calc-alkaline to shoshonite) rock suites (Gill, 1981). The progression from tholeiite to calc-alkaline to shoshonite (trachyandesite) reflects increasing K_2O and K_2O/Na_2O and decreasing iron enrichment (Jakes and White, 1972; Miyashiro, 1974). K_2O content in convergent margin volcanic suites broadly correlates with the thickness of the overlying slab in convergent margin systems (Figure 10.6c). Low

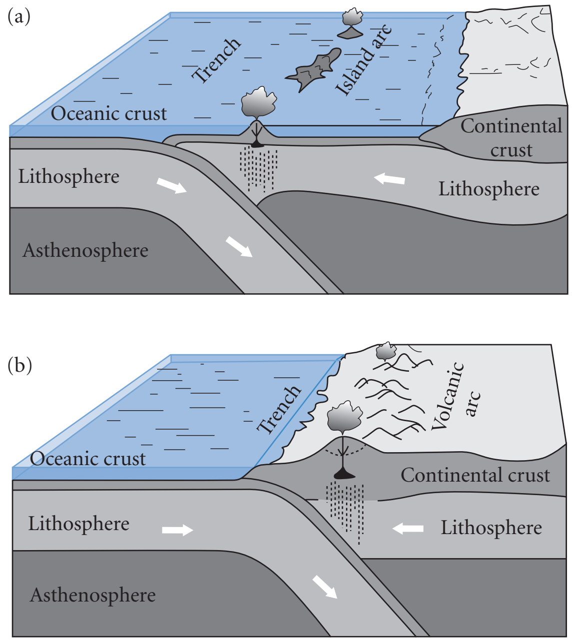

Figure 10.7 (a) Ocean–ocean convergence producing island arc volcanoes and backarc basins. (b) Ocean–continent convergence producing continental volcanic arcs. (Courtesy of the US Geological Survey.)

potassium tholeiites dominate with overlying slab thicknesses ranging from ~0 to 20 km; medium- to high potassium calc-alkaline andesites are associated with overlying slab thicknesses of ~20–40 km; high potassium shoshonites commonly develop where the overlying slab is >40 km thick (Gill, 1981).

Three major types of convergent margins occur: (1) ocean–ocean convergence generating youthful island arc volcanic complexes (Figure 10.7a), (2) ocean–continent convergence generating mature continental arc complexes (Figure 10.7b), and (3) continent–continent convergent margins marked by the cessation of subduction and consequent continental collision. Many of the same rock types can be found in all three environments; however, continental systems contain a greater proportion of silicic calc-alkaline rocks enriched in quartz and potassium feldspar; in contrast, island arcs contain a greater proportion of mafic to intermediate rocks as described below.

10.3.1 Island arcs

Ocean lithosphere is subducted beneath an overlying plate composed of oceanic lithosphere (Figure 10.7a) producing island arc chains in the eastern Indian Ocean, the Caribbean and Scotia Seas and the western Pacific Ocean, notably the Marianas Islands. Island arcs develop on the overlying ocean lithosphere plate, above the subduction zone. Island arc volcanoes are underlain by intermediate to mafic plutonic suites dominated by diorite, quartz diorite, granodiorite, tonalite and even gabbro. **Diorites** contain <5% quartz and **quartz diorites** contain 5–20% quartz (see Figure 7.20); both of these rocks are enriched in plagioclase and hornblende, with lesser amounts of pyroxene and biotite. Hornblende (and to a lesser degree pyroxene and biotite) imparts dark (mafic) colors, while the plagioclase tends to result in lighter (felsic) hues; together, these mineral suites tend to occur in approximately equal concentrations, producing a speckled light and dark coloration. Diorite, quartz diorite and granodiorite batholiths intrude beneath youthful volcanic arcs. **Granodiorites**, which represent the plutonic equivalent of dacites and rhyodacites, contain >20% quartz and more plagioclase than potassium feldspar. Island arc granodiorites are generally metaluminous, containing hornblende, biotite and minor amounts of muscovite. Island arc plutons can also consist of tonalites and trondhjemites, which are plutonic rocks enriched in plagioclase feldspar and quartz. **Tonalites**, first described from Monte Adamello near Tonale in the eastern Alps, contain calcium plagioclase and quartz with minor amounts of potassium feldspar, biotite and hornblende. **Trondhjemites**, also known as plagiogranites, are granodioritic rocks in which sodium plagioclase represents half to two-thirds of the total feldspar component.

In addition to the voluminous calc-alkaline rock suite dominated by andesites and basaltic andesites discussed earlier, young island arc systems also produce low potassium arc tholeiite basalts as well as relatively rare rocks named boninites and adakites. Low potassium arc tholeiites occur on the oceanward side of the volcanic arc, nearest the trench. Tholeiitic magmas commonly form at subduction zones where the overlying plate is relatively thin. Major element concentrations of the tholeiitic island arc basalts are very similar to MORB, as indicated by their relatively low K_2O concentrations and iron enrichment, suggesting a similar depleted mantle source – most likely by flux melting of the ocean lithosphere wedge overlying the subducted slab as well as the subducted slab itself. Island arc tholeiite basalts can be distinguished from MORB by greater concentrations of potassium and other LIL elements (such as Ba, Rb, Sr, Cs, Rb and U) and lower concentrations of HFS elements (such as Th, Hf, Ta, Ti, Zr, Nb and Y) (Perfit et al., 1980; Hawkesworth et al., 1993; Pearce and Peate, 1995). Tholeiitic island arc magmas commonly produce basalts, basaltic andesites and andesites in the volcanic arc and diorite, tonalite (plagiogranite) or lesser granodiorite plutons in the underlying magmatic arc.

Boninites, named for the Bonin Islands in the western Pacific Ocean, are high magnesium (MgO/MgO + total FeO > 0.7) intermediate volcanic rocks that contain a SiO_2-saturated (52–68% SiO_2) groundmass. These rare rocks contain phenocrysts of orthopyroxene, and notably lack plagioclase phenocrysts (Bloomer and Hawkins, 1987). Boninites are enriched in chromium (300–900 ppm), nickel (100–450 ppm), volatile elements and LREE as well as zirconium, barium and strontium. Boninites are depleted in HREE and HFS elements. These unusual rocks occur proximal to the trench and bear the geochemical signature of primitive mantle-derived magmas produced early in the subduction cycle (Hawkins et al., 1984; Bloomer and Hawkins, 1987; Pearce and Peate, 1995). Thus, boninites are a product of subduction-related melting in the forearc of youthful island arc systems. Van der Laan et al. (1989, in Wyman, 1999) suggest that boninites are produced by high temperature, low pressure remelting of previously subducted ocean lithosphere. Interestingly, boninites can be associated with rare ultrabasic komatiites that we will discuss later in this chapter.

Adakites are silica-saturated (>56% SiO_2) rocks with high Sr/Y and La/Yb ratios (LREE enriched relative to HREE) and low HFS (such as Nb and Ta) concentrations. Adakites, named for Adak Island of the Aleutian Island chain, have been thought to be derived by slab melting of eclogite and/or garnet amphibolite from the descending ocean lithosphere (Kay, 1978; Defant and Drummond, 1990; Stern and Killian, 1996; Reay and Parkinson, 1997). While it was initially believed that adakites only form where young (<25 Ma), thin, hot ocean lithosphere is subducted beneath island arc lithosphere, adakites are now known to form at continent–continent collision sites as a result of shallow slab subduction of continental lithosphere (Chung et al., 2003). Shallow slab subduction and lithosphere recycling at subduction zones may play a significant role in the development of adakites as well as their plutonic equivalents trondhjemites and tonalites. Research continues to determine possible relationships of adakite formation with Archean **tonalite, trondhjemite and granodiorite (TTG) associations** and the evolution of continental crust (Drummond and Defant, 1990; Castillo, 2006; Gomez-Tuena et al., 2007) (Box 10.1).

Insofar as the overlying arc lithosphere is relatively thin in immature island arc systems, young volcanic arcs are dominated by basalts and basaltic andesites with rare boninites and adakites. Prolonged subduction in island arc systems generates increasingly thicker arc lithosphere. As island arc lithosphere thickens, andesites and dacites predominate as the Si_2O and K_2O contents of all rocks increase with the development of continental-type arc lithosphere (Miyashiro, 1974). In nearly all convergent margins, the calc-alkaline association is generated by fractional crystallization of basaltic magma derived by partial melting of overlying mantle peridotite – fluxed by fluids released from the dehydrated subducted oceanic lithosphere slab. The continued removal of crystals from melt leads to continuous variation in the residual liquid (liquid line of descent) generating basalt, andesite, dacite and rhyolite. In addition to fractionation, open-system diversification processes (Chapter 8) such as assimilation and magma mixing alter magma chemistry (Grove and Kinzler, 1986) and alter the chemical composition of magmas generated in the mantle wedge overlying the subducted slab.

In addition to magmatism within the island arc complex, igneous activity can also occur behind the island arc in backarc basins.

Box 10.1 Tonalite, trondhjemite and granodiorite (TTG) association

Plutons containing TTG are found in subduction zone environments ranging from the Archean to the Recent. However, Archean (>2.5 Ga) subduction zone plutonic rocks consist dominantly of TTG. Trondhjemites were named by V. M. Goldschmidt in 1916 for holocrystalline, leucocratic Norwegian rocks enriched in sodium plagioclase and quartz and depleted in biotite and potassium feldspar (Barker, 1979). Trondhjemites are similar to tonalites but contain greater concentrations of sodium plagioclase (oligoclase to albite) and more variable potassium feldspar concentrations (Figure B10.1a). Tonalites and trondhjemites are also known as plagiogranite. **Charnockites**, an orthopyroxene-bearing suite of rocks of generally granitic composition, also occur with the TTG association.

TTG associations occur in Archean rocks such as the Pilbara Craton of Australia, and the Beartooth and Big Horn Mountains of Wyoming. In contrast, Proterozoic and younger convergent margin granitic rocks consist predominantly of granite and granodiorite, with TTG associations representing a very small component in ocean–ocean convergence. What conditions changed at ~2.5 Ga? Shallow-dipping subduction zones "pinch out" the overlying mantle wedge such that the wedge component plays a relatively minor role in magma genesis. Archean subduction involved higher geothermal gradients, shallow subduction and melting of downgoing ocean lithosphere, with minimal input from the overlying lithosphere wedge. Shallow subduction of oceanic lithosphere at unusually low angles (Figure B10.1b) has been proposed as a model for the growth of Archean continental crust through the generation of TTG plutonic rocks and their volcanic equivalents – adakites (Martin, 1986; Smithies et al., 2003).

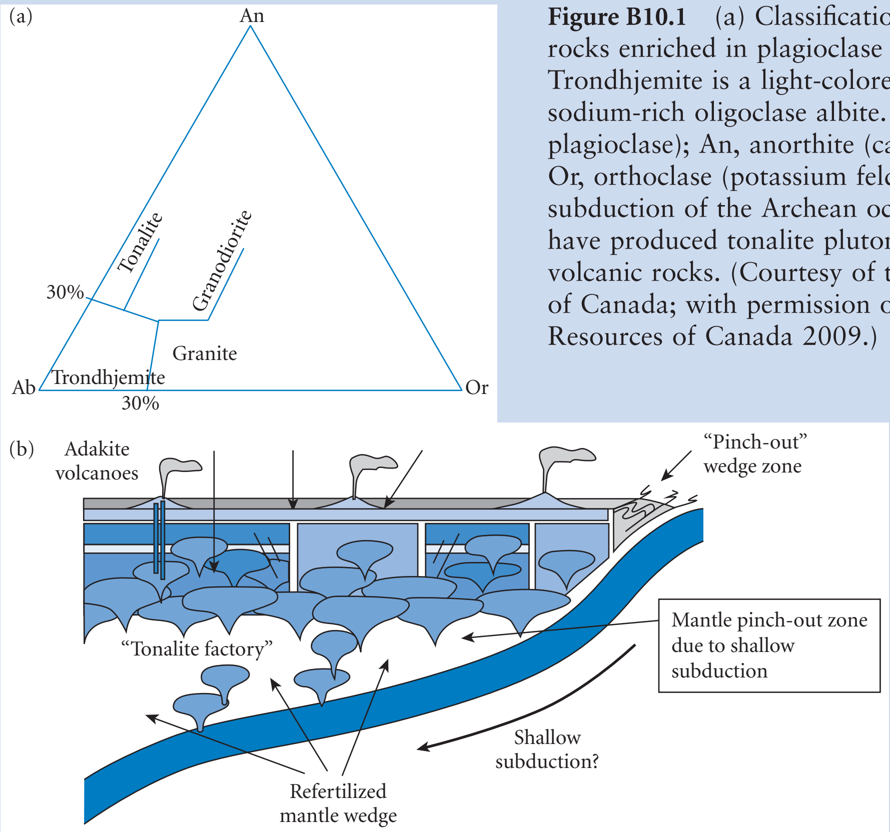

Figure B10.1 (a) Classification of some granitoid rocks enriched in plagioclase and >20% quartz. Trondhjemite is a light-colored tonalite containing sodium-rich oligoclase albite. Ab, albite (sodium plagioclase); An, anorthite (calcium plagioclase); Or, orthoclase (potassium feldspar). (b) Shallow subduction of the Archean ocean lithosphere may have produced tonalite plutons and adakite volcanic rocks. (Courtesy of the Geological Survey of Canada; with permission of the Natural Resources of Canada 2009.)

Backarc basins

Although compressional forces dominate island arc settings, lithospheric extension can occur in the overlying plate, behind the arc, resulting in the development of **backarc basins** (Figure 10.8). How does backarc extension occur? "Trench pull" forces move the volcanic arc towards the subduction zone resulting in the seaward movement of the trench and volcanic arc (Chase, 1978). Extension is manifested as normal faults and backarc spreading. The western and northern Pacific Ocean (Figure 10.8a) provide excellent examples of backarc basins, including the Sea of Japan, the Bering Sea, the Lau Basin–Havre Trough, Manus Basin and the Marianas Trough.

Backarc basins (Figure 10.8b) erupt a diverse suite of volcanic rocks including

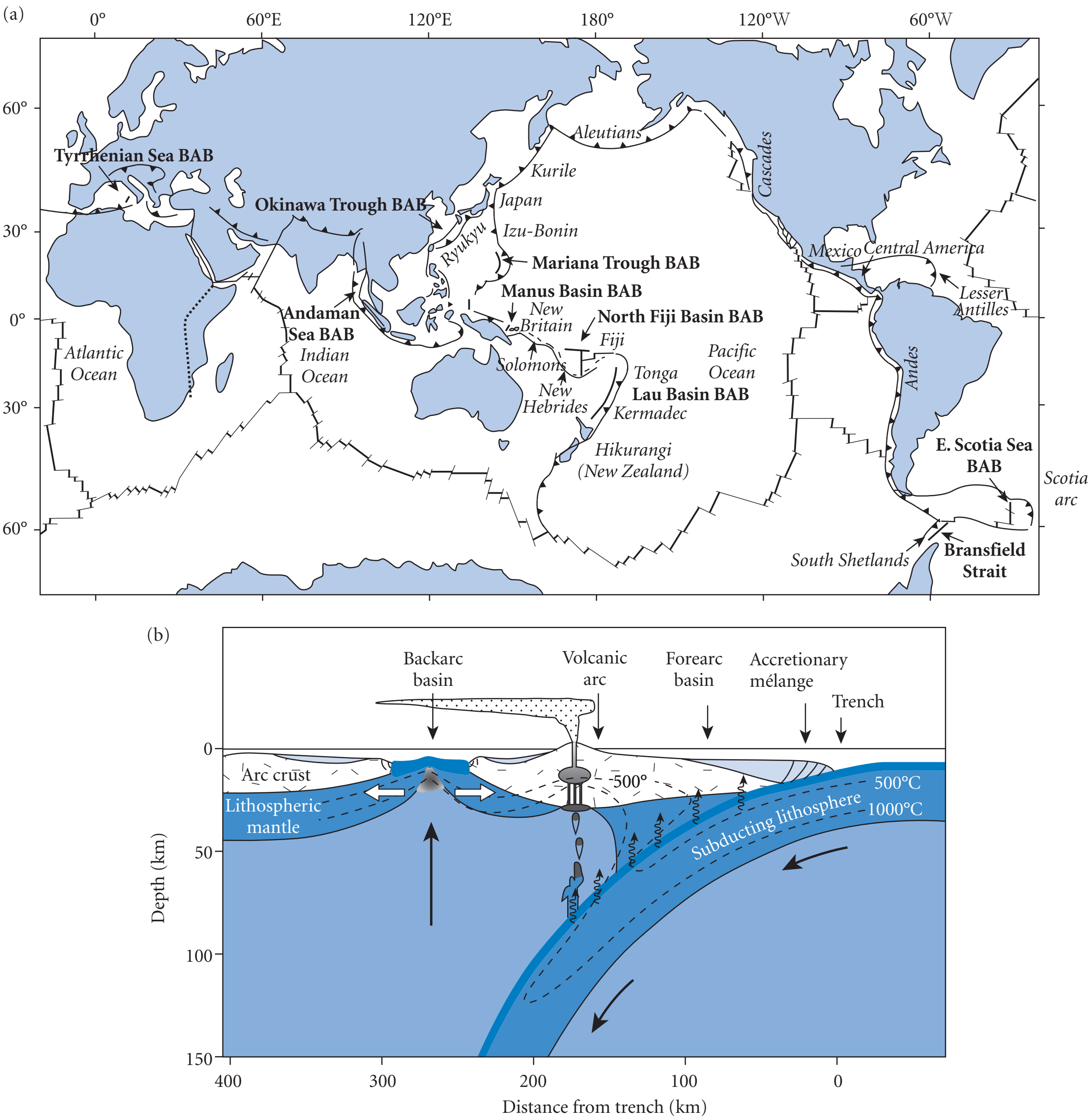

Figure 10.8 (a) Modern backarc basins (BAB) are concentrated in the western Pacific Ocean. (b) Backarc basins form by extension within the arc crust. (Courtesy of Wikipedia.)

basalt, basaltic andesite, andesite and dacite; however, tholeiitic and alkalic basalts commonly dominate. As suggested in our earlier discussion of island arc tholeiitic basalts, the relative proportions of andesitic versus basaltic magmas is related to the nature and thickness of the lithospheric wedge above the subduction zone (Gill, 1981). In some island arc settings (e.g., Kermadec, Marianas, Scotia and Vanuatu arcs), basalts dominate over andesites in the backarc and the volcanic arc regions. Fryer et al. (1981), on the basis of trace element chemistry, identified a distinctive group of rocks known as **backarc basin basalts** (BAB). BAB are tholeiitic, with geochemical similarities with both MORB and arc tholeiite trends. Relative to MORB, BAB display greater enrichment of H_2O, alkali elements and LIL elements. BAB are slightly depleted in titanium, yttrium and niobium and display flat rare Earth element patterns 5–20 times that of chondrites (Sinton et al., 2003). BAB may show relative enrichment in volatile elements, thorium and LREE, which suggests the involvement of subduction-related fluids in magma genesis (Pearce and Peate, 1995).

Why do backarc basins produce a wide array of rock types that range from near MORB to calc-alkaline compositions? Extension in the backarc (Figure 10.8b) results in partial melting of mantle peridotite, producing MORB-like magmas. However, these magmas interact to varying degrees with calc-alkaline sources. Calc-alkaline magma sources include the hydrated mantle wedge situated above the downgoing slab, recycled subducted lithospheric slab, and subducted marine sediment. Thus, BAB are produced by a combination of partial melting of lherzolite upper mantle wedge that has been fluxed by volatiles released by the subducted ocean lithosphere, as well as decompression melting of mantle peridotite at backarc spreading ridges (Fretzdorff et al., 2002). As indicated in the discussion above, distinction between different types of basalts generated in different tectonic environments is largely dependent upon geochemistry (Box 10.2).

10.3.2 Continental margin arcs

Mature convergent margins, involving the subduction of ocean lithosphere beneath thick continental lithosphere, occur along the eastern Pacific region extending from the Cascades southward to the Andes Mountains (Figure 10.8a). Ascending hydrous melts from the subducted ocean slab chemically react with the overlying wedge composed of mantle and thick continental lithosphere. These magmas produce continental arc plutons that are more silicic than island arc plutons as a result of the thick overlying continental lithosphere through which subduction zone fluids must penetrate. Extensive assimilation and magma mixing within the overlying continental slab result in K_2O and SiO_2 enrichment within plutons.

Box 10.2 Geochemical approaches to petrotectonic associations

Basalt, basalt, basalt! As indicated in the preceding discussion, basalt can be produced in a number of different tectonic environments. Petrotectonic studies utilize a number of different approaches to identifying sites of basalt genesis. One of these approaches utilizes geochemical indicators discussed in Chapter 7. Pearce and Cann (1973) combined geochemical variations of minor and trace elements to infer tectonic origin. The Pearce and Cann (1973) classification (Figure B10.2a) is widely used in the tectonic analysis of basalts and is particularly useful in accretionary terranes (Chapter 1) where the original source of basalt is ambiguous. The petrotectonic environments defined by these discrimination diagrams include: within plate basalts, island arc tholeiites, calc-alkaline basalts, mid-ocean ridge basalts, ocean island tholeiites and ocean island alkaline basalts. Shervais (1982) also developed discrimination diagrams using minor element concentrations such as vanadium and titanium (Figure B10.2b). These geochemical techniques designed to determine petrotectonic origin are extremely useful if used in conjunction with field studies – and if metamorphism has not chemically altered the rock.

Box 10.2 *Continued*

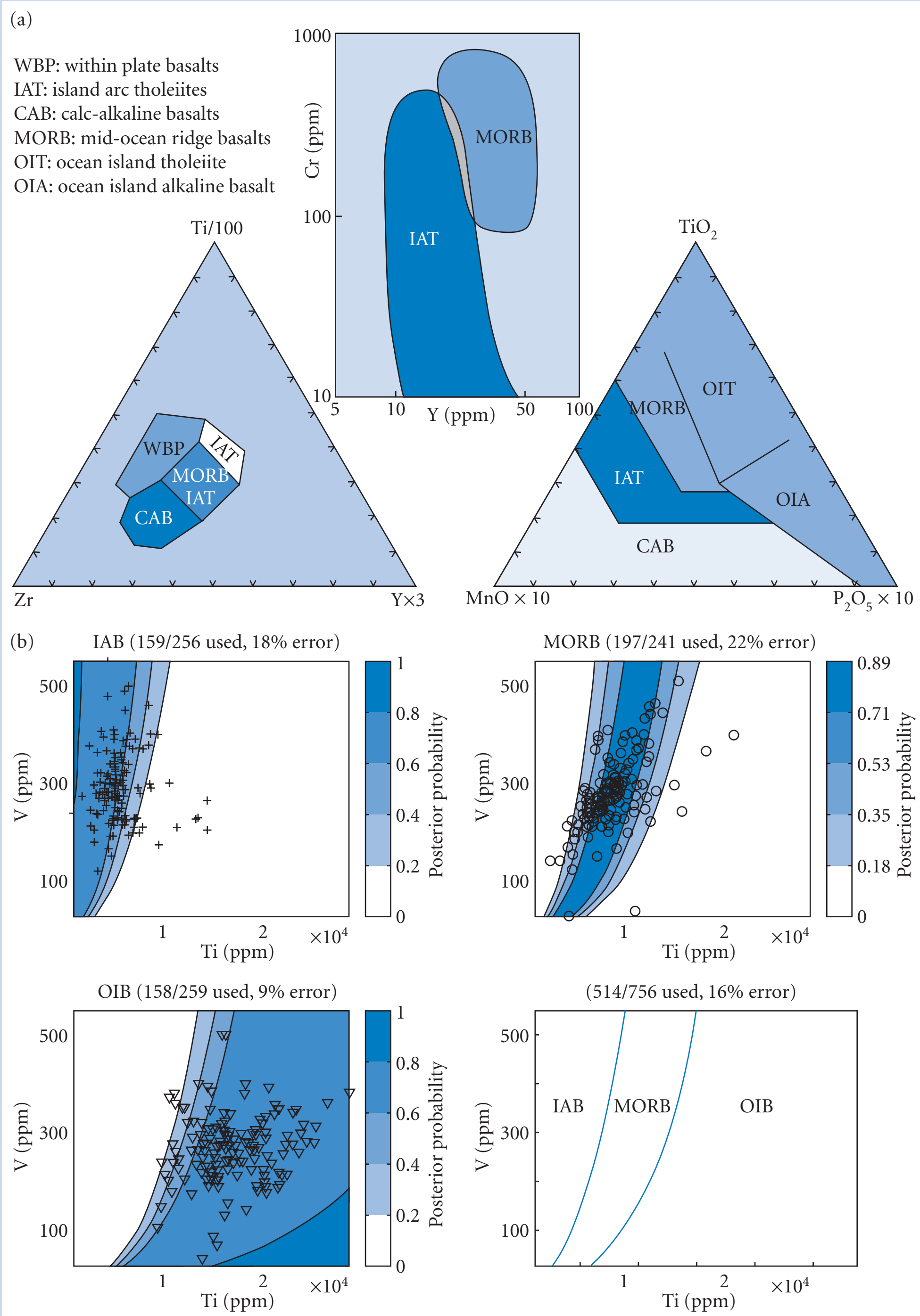

Figure B10.2 (a) Discrimination diagrams for basalt using minor and trace element concentrations such as Zr, Ti, Y, Cr, MnO and P_2O_5. (After Pearce and Cann, 1973; with permission of Elsevier Publishing.) (b) Discrimination diagrams using vanadium and titanium concentrations as discriminating elements. IAB, island arc basalts; OIB, ocean island basalts; MORB, mid-ocean ridge basalts. (After Shervais, 1982; with permission of Elsevier Publishers.)

Ocean–continent convergent margins produce voluminous granodiorite, diorite, granite and tonalite plutons. Large granodiorite plutons commonly dominate ocean–continent convergent plate boundaries. For example, the Sierra Nevada consists of 25–30 km thick upper crustal rocks containing hundreds to thousands of individual granodioritic, dioritic and tonalitic plutons. The Sierra Nevada composite batholith developed due to eastward-dipping subduction beneath western North America. Arc magmatism continued from 220 to 80 Ma (Fliedner et al., 2000; Ducea, 2001).

Magma from these intermediate–silicic plutons erupts onto Earth's surface producing composite volcanoes. Together with voluminous andesites, rocks such as dacites, rhyodacites, rhyolites and latites display aphanitic–porphyritic to pyroclastic textures in composite volcanic settings such as Mt St Helens and Crater Lake in the Cascade Range (USA).

In late stages of ocean–continent subduction, highly alkalic shoshonites can erupt over thick continental lithosphere. **Shoshonites** are dark-colored, potassium-rich trachyandesites, commonly containing olivine and augite phenocrysts with a groundmass of labradorite plagioclase, alkali feldspar, olivine, augite and leucite. Shoshonites occur in thickened lithosphere farthest from the trench region, in continent–continent collisions and in some backarc basins.

10.3.3 Continental collision zones

In long-lived convergent margins, the termination of ocean lithosphere subduction marks the transition from mature arc–continent convergence to a final collision of two continental blocks. The Alpine–Himalayan orogenic system is the modern classic model of continent–continent collision following tens of millions of years of ocean lithosphere subduction. Ancient examples include the complete subduction of ocean basins 250 million years ago, resulting in continent–continent collision creating the supercontinent Pangea.

In continent–continent collisions, the lower continental lithosphere does not subduct to great depths but essentially breaks off and underplates the overlying continental lithosphere plate producing a doubly thick lithosphere. Melting at the base of this lithospheric stack produces Al_2O_3-, K_2O- and SiO_2-rich igneous rocks such as rhyolites, rhyodacites and shoshonites and plutonic rocks of increasingly granitic composition. These magmas are the result of relatively flat subduction (<25°), thick continental lithosphere (>25 km), higher degrees of partial melting of continental lithosphere and/or arc basement, and the diminished role of ocean lithosphere subduction. Alkaline basalts also occur in continent–continent collisions as a result of upwelling mantle melts.

Rhyolites and rhyodacites are characterized by high viscosity, which retards lava flow, resulting in thick accumulations of limited aerial extent. Common minerals include quartz, potassium feldspar, biotite, plagioclase, anorthoclase and magnetite. Shoshonitic magmas are generated farthest from the trench, wherein melts assimilate K_2O and Na_2O as they rise through thick slabs of overlying continental lithosphere. Volcanic eruptions of rhyolite to shoshonite lavas can erupt explosively, generating voluminous pyroclastic tuffs and breccias, or produce lava flows that solidify to produce glassy and/or aphanitic–porphyritic textures.

The plutonic equivalent of rhyodacite and rhyolite are granitic rocks or granitoids (Box 10.3). The term "granitic" or "granitoid" is loosely used for silica-oversaturated plutonic rocks that contain essential potassium feldspars and quartz. Granitoids include the IUGS fields of quartzolite, quartz-rich granitoid, alkali feldspar granite, granite, quartz granitoid, granodiorite and tonalite (see Figure 7.20). Granitoid rocks that form at mature convergent margins tend to be peraluminous to metaluminous, containing hornblende, biotite and/or muscovite. Although granites of variable composition occur, S-type and I-type granites tend to predominate. I-type magmas form by partial melting of basic to intermediate igneous rocks in or above the subduction zone at ocean–ocean or ocean–continent convergent margins. Peraluminous, potassium-rich, S-type granites and granodiorites are particularly common at continent–continent collisions. The peraluminous sedimentary component is derived from phyllosilicate minerals in graywackes and mudstones of the continental crust and accretionary wedge. These sedimentary materials melt to

Box 10.3 Granite classification

Strictly speaking, the term "granite" is restricted to plutonic rocks containing 20–60% quartz and 35–90% alkali to plagioclase feldspars (see Figure 7.20). Thus, the two essential mineral groups in granite are quartz and feldspars. Other minor minerals include hornblende, biotite and muscovite. Accessory minerals include magnetite, rutile, tourmaline, sphene, apatite, molybdenite, gold, silver and cassiterite. Granite plutons are genetically associated with Precambrian cratons and convergent margins (Pitcher, 1982). Phanerozoic granitic plutonic belts are found along continent–ocean subduction zones or at continent–continent suture zones. Within orogenic settings, granites may be emplaced synchronous (syn-kinematic) with convergence, as late-stage collisional plutons or as post-kinematic intrusions.

Granites have been subdivided by a number of methods, one of which attempts to infer source rock origin. Chappell and White (1974) and others recognize four distinct types of granite (M, I, S, and A types) based upon the nature of the inferred parental source rock (Table B10.3). M-, I-, and S-type granites are orogenic granites associated with subduction, whereas A-type granites are anorogenic in origin.

- **M-type granites** (Pitcher, 1982) are derived from mantle-derived parental magmas, as indicated in the low Sr^{87}/Sr^{86} ratios (<0.704). M-type granites are associated with calc-alkaline tonalites, quartz diorites and gabbroic rocks. In addition to quartz and feldspars, hornblende, clinopyroxene, biotite and magnetite are among the major minerals. M-type granites develop in island arc settings. Copper and gold mineralizations are associated with M-type granites.
- **I-type granites** (Chappell and White, 1974) are generated by the melting of an igneous protolith from either the downgoing oceanic lithosphere or the overlying mantle wedge. I-type granites are enriched in Na_2O and Ca_2O and contain lower Al_2O_3 concentrations. I-type granites have Sr^{87}/Sr^{86} ratios of less than 0.708, usually in the range 0.704–0.706, indicating magma derived from a mantle source. Because they are primarily derived from a mantle source, I-type granites may be enriched in mafic minerals such as hornblende, biotite, magnetite and sphene. Porphyry copper, tungsten and molybdenum deposits are associated with I-type granites. I-type granites are prevalent along the Mesozoic–Cenozoic Andes Mountains (Chappell and White, 1974; Beckinsale, 1979; Chappell and Stephens, 1988).
- **S-type granites** (Chappell and White, 1974) are produced by the melting of sedimentary crustal rocks in collision zones. S-type granites are depleted in Na_2O but enriched in Al_2O_3 (peraluminous). S-type granites have Sr^{87}/Sr^{86} ratios of >0.708, indicating that source rocks had experienced an earlier sedimentary cycle. S-type granites are also known as two-mica granites in that they commonly contain both muscovite and biotite, reflecting the peraluminous content of the sedimentary source rock rich in phyllosilicate minerals. Hornblende is conspicuous by its absence. Other minerals include monazite and aluminosilicate minerals such as garnet, sillimanite and cordierite. Tin deposits are associated with S-type granites (Chappell and White, 1974; Beckinsale, 1979).

Table B10.3 The major features of M-, I-, S- and A-type granitoids.

	M	I	S	A
SiO_2	54–73%	53–76%	65–79%	60–80%
Na_2O	Low, <3.2%	High, >3.2%	Low, <3.2%	>2.8%
K_2O/Na_2O	Very low	Low	High	High
Sr^{87}/Sr^{86}	<0.704	<0.706	>0.706	0.703–0.712

Continued

Box 10.3 *Continued*

- **A-type granites** (Loiselle and Wones, 1979) are anorogenic rocks produced by activities that do not involve the subduction and collision of lithospheric plates. A-type granites are enriched in alkaline elements with high K/Na and (K + Na)/Al ratios as well as high Fe/Mg, F, HFS elements (Zr, Nb, Ga, Y, Zn) and rare Earth element concentrations. A-type granites are depleted in Mg, Ca, Al, Cr, Ni and have lower water contents and high Ga/Al ratios (Collins et al., 1982; Whalen et al., 1987). Relative to I-, S- and M- type granites, A-type granites are more enriched in LIL elements and depleted in refractory elements (Creaser et al., 1991). A-type granites are peralkaline and commonly contain biotite, alkali pyroxenes, alkali amphiboles and magnetite (Collins et al., 1982). Associated with A-type granites are alkali-rich, relatively anhydrous rocks that can include alkali granite, syenite, alkali syenite and quartz syenite.

produce two-mica granites containing biotite and muscovite.

Convergent margins contain a diverse range of rock types. In addition to intermediate and silicic igneous rocks prominently discussed thus far in this chapter, basic and ultrabasic assemblages also occur at convergent margins due to magmatic processes and/or tectonic displacement. In some cases, these involve metamorphic processes that we will address in Chapters 15–18. Let us first consider tectonically emplaced Alpine orogenic complexes and then we will briefly discuss basic–ultrabasic zoned intrusions.

10.3.4 Alpine orogenic complexes

Alpine orogenic complexes are fault-bounded, deformed rock sequences that mark the site of present or former convergent margins. Unlike the intermediate to silicic igneous rocks that develop in situ (in place) as a result of subduction-induced magmatism, alpine orogenic complexes have been transported far from their site of origin by thrust faulting and shearing. Because such tectonism can be intense, these complexes are commonly dismembered into fault blocks and jumbled together in a haphazard fashion such that their original layering may be disrupted. Alpine orogenic bodies contain disrupted pelagic sediment layers, basalt, cumulate basic and ultrabasic layers as well as tectonized mantle slices of ocean lithosphere and calc-alkaline intrusive and volcanic assemblages. Alpine orogenic bodies are commonly associated with tectonic mélanges.

A tectonic **mélange** (from the French word for mixture) is an intensely sheared, heterogeneous rock assemblage embedded within a highly deformed mud matrix. Mélanges form at subduction zones where rocks and tectonic blocks are sliced from the downgoing oceanic lithosphere and often mixed with rocks formed in forearc settings. The diverse suite of rocks may include: (1) deformed and altered mid-ocean ridge, ocean island and ocean plateau basalts, (2) limestone, chert and other marine sedimentary rocks, and (3) slices of eclogite, peridotite and blueschist from subducted oceanic or forearc lithosphere. Eclogites and blueschists are high pressure metamorphic rocks characteristic of subduction zones (Chapter 18).

Ophiolites constitute one type of Alpine deposit in which the oceanic or backarc basin lithosphere or volcanic arc basement rocks are preserved in orogenic belts. The term ophiolite was first proposed by Steinmann (1905) for serpentinized rocks in the Alps. Over the next several decades, Steinmann recognized a suite of rocks that, thereafter, became known as the "Steinmann trinity". These three rock types consist of pelagic chert, serpentinite (hydrothermally altered peridotite) and spilites (altered pillow basalts). As the term is currently used, ophiolites are thought to represent coherent slices of oceanic lithosphere, volcanic arc basement or backarc basin lithosphere "obducted", or thrust, onto the edge of continents above subduction zones.

While it is intuitively obvious that ophiolites may originate by sea floor spreading at ocean ridges, petrologists recognize that many ophiolites represent oceanic fragments produced in either forearc or backarc settings. These ophiolites are referred to as **suprasubduction zone (SSZ) ophiolites**. SSZ ophiolites

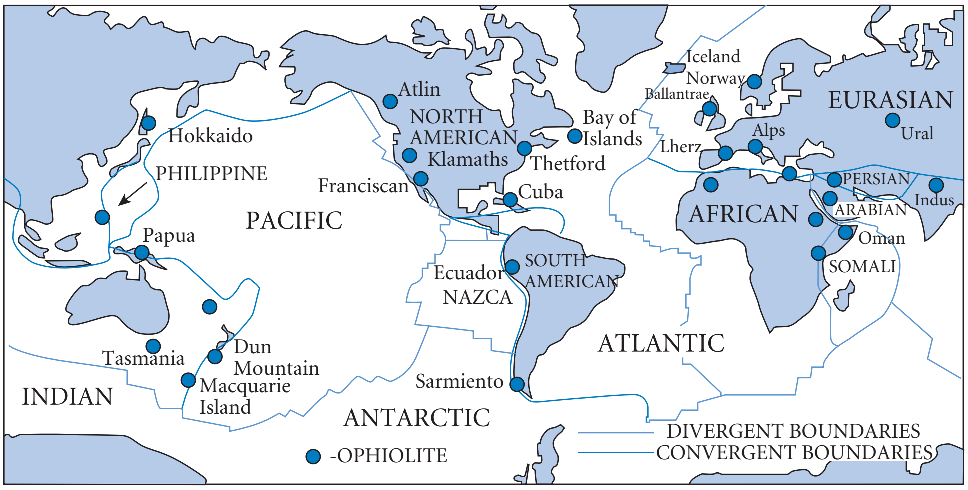

Figure 10.9 Ophiolite locations throughout the world. (Courtesy of William Church.)

develop due to extensional tectonics that result in backarc spreading or forearc spreading producing oceanic lithosphere. Researchers use immobile elements as petrogenetic indicators, such as chromium, to determine ophiolite sites of origin (Dick and Bullen, 1984). Both the origin of ophiolites and their means of emplacement on the edges of continents remain areas of intense research.

Complete ophiolite sequences display a stratigraphic sequence similar to that of ocean lithosphere (see Figure 10.2a). The stratigraphy of an idealized ophiolite sequence was defined by the first Penrose Conference in 1972 as follows:

- **Layer 1:** pelagic, marine sedimentary rock such as ribbon chert, thin shale beds and limestone derived from the lithification of siliceous ooze, clay and calcareous ooze sediments, respectively.
- **Layer 2A:** basic volcanic complex, which may contain pillow basalt.
- **Layer 2B:** basic sheeted dike complex.
- **Layer 3:** cumulate gabbroic complex with basal cumulate peridotites and pyroxenites.
- **Layer 4:** tectonized ultrabasic complex consisting of variably metamorphosed harzburgite and dunite. Podiform chromite deposits occur with dunite bodies. The tectonized ultrabasic complex overlies a metamorphic basal sole thrust.

While the idealized stratigraphic layering of ophiolites mimics ocean lithosphere layering, the complete four-layer stratigraphic sequence is rarely preserved. Tectonically disrupted ophiolites, missing one or more layers, are referred to as partial, or dismembered, ophiolites. Ophiolites occur throughout the world (Figure 10.9) and mark the former location of ocean lithosphere subduction. Excellent examples of ophiolites include the following localities: Oman, Troodos (Cyprus), Coast Range (California), Newfoundland and Morocco. Ophiolites are important in providing the following:

1. Valuable ore deposits containing Cu, Ag, Au, Zn, Ni, Co, Cr and other metals.
2. Evidence documenting oceanic lithosphere subduction dating from the Precambrian to the present. Well-documented ophiolites less than 1 Ga occur throughout the world. Archean examples, dating as far back as 3.8 Ga, are highly controversial and may (Furnes et al., 2007) or may not (Hamilton, 2007; Nutman and Friend, 2007) represent true ophiolites.

In addition to their occurrence in Alpine orogenic complexes and ophiolites, basic and ultrabasic magmas intrude convergent margin assemblages to form the concentrically zoned or layered plutons discussed below.

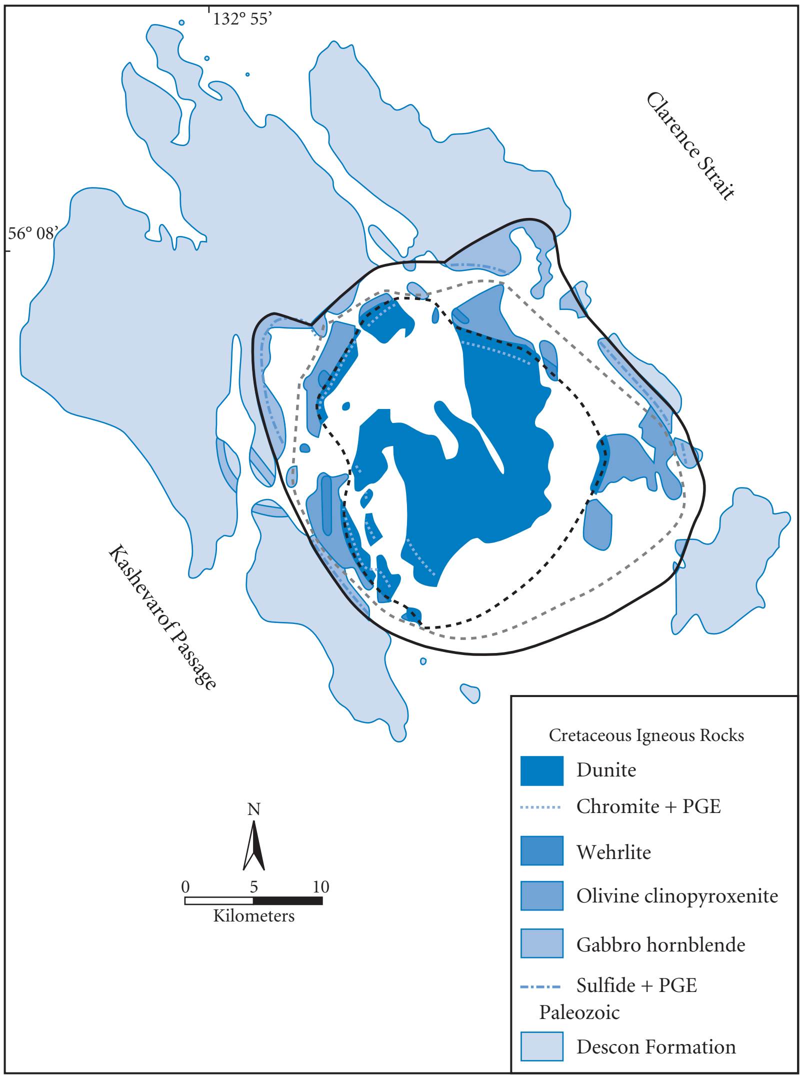

Figure 10.10 Zoned intrusion from the Blashke Islands Complex, southeast Alaska. PGE, platinum group elements. (After Kennedy and Walton, 1946.)

10.3.5 Alaska-type (zoned) intrusions

Alaska-type intrusions consist of concentrically layered (zoned) plutons formed in convergent margin settings. Alaska-type intrusions are commonly several kilometers in diameter, exhibit a dunite core and pyroxenite shell, and are surrounded by massive gabbro. Late granitic zones may also occur around the perimeter of the intrusive structure. In contrast to tectonically emplaced alpine suites, Alaska-type plutons form in situ by intrusion of magma into the surrounding country rock. Originally recognized in Alaska and in the Ural Mountains, these plutonic bodies have since been identified in many other localities throughout the world. Buddington and Chapin (1929) noted the concentric layers in the Blashke Island Complex of Alaska (Figure 10.10).

Irvine (1959) conducted extensive work on the Cretaceous age Duke Island Complex (Alaska). The Duke Island Complex consists of a 3 km diameter plutonic body at Judd Harbor and a 5 km diameter intrusion at Hall Cove. These two intrusions, which are likely to be continuous at depth, contain a dunite core surrounded by successive rings of peridotite, olivine pyroxenite, hornblende–magnetite pyroxenite and gabbro – all of which are cut by late granitic rocks. Irvine (1959) noted layers that exhibited graded beds within ultrabasic rocks at Duke Island, in which crystals are coarse grained at the base and fine upward within a layer. Duke Island remains an active exploration site with economic deposits of copper, nickel, platinum and palladium.

Alaska-type ultrabasic–basic plutons commonly occur as post-orogenic intrusions in volcanic arc or accretionary mélange terrains. A number of different processes have been suggested for their formation; these include fractionation of ultrabasic or basic parent magma from the upper mantle, magma mixing at convergent plate boundaries, or magmas from deep mantle plumes (Taylor, 1967; Tistl et al., 1994; Sha, 1995; Ishiwatari and Ichiyama, 2004). Alaska-type intrusions are economically important as sources of metals, particularly platinum group elements (PGE).

Thus far in this chapter we have focused upon petrotectonic assemblages from divergent and convergent plate boundaries. While divergent and convergent margins produce the bulk of Earth's magmatism, igneous rocks also develop within lithospheric plates without any direct link to plate boundary processes, as discussed below.

10.4 INTRAPLATE MAGMATISM

Intraplate magmatism refers to magma generation and igneous rock suites generated within lithospheric plates, rather than at plate boundaries. Intraplate magmatism may be initiated by hotspot activity, continental rifts or overthickened continental lithosphere. Intraplate magmatism produces a wide range of igneous rock types including:

- Tholeiitic to alkalic basalt and related gabbros of hotspots and LIP.
- Siliceous anorogenic granite and rhyolite.
- Silica-undersaturated rocks.
- Basic–ultrabasic suites including komatiites and kimberlites.
- Carbonatites.

Large igneous provinces (**LIP**), encompassing volumes >10^6 km^3 (Mahoney and Coffin, 1997), are the greatest manifestation of intraplate magmatism on Earth (Figure 10.11). Most LIP are basaltic in composition although silicic examples, known as **SLIP** (**silicic large igneous provinces**), such as Yellowstone, also occur. The most widespread Phanerozoic intraplate magmatic features consist of massive tholeiitic flood basalts. These massive volcanic landforms occur as both **oceanic flood basalts** and **continental flood basalts**.

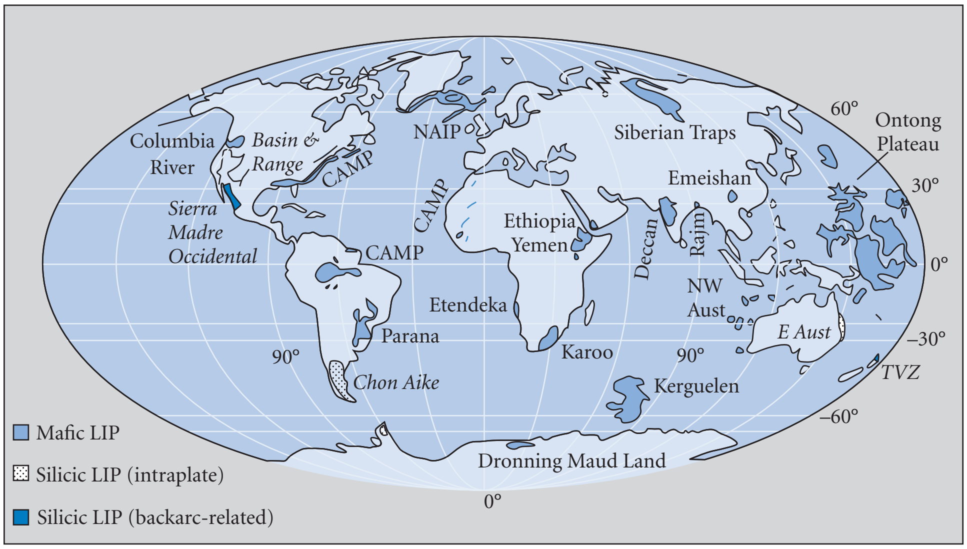

Figure 10.11 Earth's large igneous provinces (LIP). CAMP, Central Atlantic Magmatic Province; NAIP, North Atlantic Igneous Province; TVP, Taupo Volcanic Zone. (After Coffin and Eldholm, 1994; courtesy of M. F. Coffin.)

In the following section, we will first consider oceanic intraplate magmatism and then later discuss continental intraplate assemblages.

10.4.1 Oceanic intraplate magmatism

Ocean islands and ocean plateaus form above mantle hotspots that erupt anomalously high volumes of tholeiitic and alkalic basaltic lava onto the ocean floor. **Ocean islands** are volcanic landforms that rise upward above sea level. **Seamounts** are volcanically produced peaks below sea level. **Oceanic plateaus** are broad, flat-topped areas that result from massive outpourings of lava flowing laterally from source vents. Oceanic plateaus cover large areas of the ocean floor, ranging up to 10 million km^2.

Ocean island basalts

Ocean island basalts (**OIB**) are a geochemically distinct suite of rocks distinctly different from MORB (Figure 10.12). In contrast to MORB, OIB are more alkalic and are less depleted – and may in fact be somewhat enriched with respect to incompatible elements such as potassium, rubidium, uranium, thorium and LREE (Hofmann and White, 1982; Hofmann, 1997). The different geochemical signatures have been interpreted to represent different mantle source areas. While MORB were considered to represent partial melts of an upper, depleted (previously melted) mantle, OIB were considered to perhaps represent partial melts from a deeper, undepleted mantle source. However, ocean island basalts display large variations in strontium, neodymium and lead and other isotopic ratios, suggesting the role of multiple sources and processes. Various hypotheses proposed for OIB chemistry include:

- Small degrees of melting of a primitive mantle source.
- Melting of a mantle source enriched in alkali elements.
- Incorporation of subducted oceanic crust in the source region.
- Entrainment of subducted sedimentary rocks in the source region (Hofmann and White, 1982; Hofmann, 1997; Kogiso et al., 1998; Sobolev et al., 2005).

The isotopic signatures of many OIB indicate that magmas were derived from non-primitive sources of variable mantle composition. For example, Rb/Sr and Nd/Sm ratios are lower than primitive mantle ratios while U/Pb, Th/Pb and U/Th ratios are higher than primitive mantle sources (Hofmann and White, 1982). In fact, most OIB display isotopic ratios indicative of an enriched mantle source, particularly their elevated incompatible element and NiO concentrations. Why is the mantle composition so variable? One explanation involves mantle enrichment due to the incorporation of a recycled oceanic lithosphere derived from ancient subduction zones (Hofmann and White, 1982; Hofmann, 1997; Turner et al., 2007). Hirschmann et al. (2003) and Kogiso et al. (2003) propose that undersaturated, nepheline normative OIB magmas may be derived by partial melting of a garnet pyroxenite, which itself is derived by the mixing of subducted MORB basalt and mantle peridotite. Thus, at least some OIB hotspots tap mantle melts enriched by ocean lithosphere subducted up to 2.5 Ga (Turner et al., 2007). Let us consider the best known OIB location – the Hawaiian Islands.

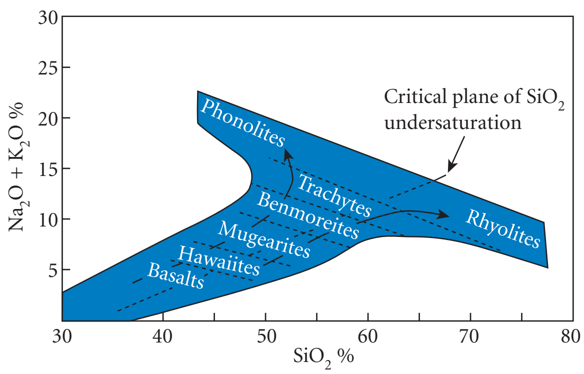

Figure 10.12 The fractionation sequence occurring at ocean islands produces a diverse suite of volcanic rocks with variable alkali and silica concentrations. (Courtesy of Stephen Nelson.)

Hawaii

From the ocean floor up, the Hawaiian Islands constitute the highest mountains on Earth with a relief of ~10 km high; Mt Everest in contrast has an elevation of just over 9 km.

The Hawaiian hotspot has been active for over 80 million years, generating a chain of seamounts and islands extending for a distance of 5600 km. The Hawaiian Islands are dominated by olivine tholeiites; tholeiitic basalts comprise ~99% of the exposed Hawaiian volcanic rocks with alkalic basalts contributing only a small fraction. Early eruptions of alkali basalts are followed by extensive tholeiitic basalts that generate massive shield volcanoes. Late-stage Hawaiian volcanism reverts to alkali basalts (hawaiites to benmoreites), perhaps indicating a lower degree of melting as the island moves away from the hotspot magma source and temperatures decline. The dominant tholeiitic iron-enrichment trend (Chapter 8) likely results from fractional crystallization of early formed, magnesium-rich olivine and pyroxene. As magmas become more enriched in iron, alkali-enriched hawaiite forms by fractionation of tholeiitic basalt. The addition of magnetite to the crystallizing assemblage causes the remaining magmas to become progressively less enriched in iron. Continued crystallization produces an iron-depletion and alkali-enrichment trend in which magmas evolve successively from hawaiite to mugearite and benmoreite (see Figure 7.24b). Depending upon the degree of silica saturation, the final minerals to crystallize may include feldspathoids in silica-undersaturated phonolite, potassium feldspar in silica-saturated trachyte or quartz in silica-oversaturated rhyolite (Figure 10.12).

Hawaiian basaltic magmas are thought to form by partial melting of a heterogeneous mantle plume source, perhaps composed of primitive garnet lherzolite, enriched by a subducted ocean lithosphere component. Using geochemical data such as Ni, Mg, Pb, O, Hf and Os isotopic ratios, researchers (Blichert-Toft et al., 1999; Sobolev et al., 2005) propose that the Hawaiian magma source is generated from a mantle enriched from the recycling of eclogite (subducted ocean lithosphere), which magmatically mixes with mantle peridotite. Using geochemical constraints such as high NiO concentrations, Sobolev et al. (2005) estimate that recycled (subducted) ocean crust accounts for up to 30% of the Hawaiian source material. These important studies suggest that the Hawaiian intraplate magmatism may be related to ancient subduction processes. While OIB volcanism at locations such as the Hawaiian Islands is impressive, humans have yet to witness the immense volcanism necessary to erupt gargantuan ocean plateaus.

Oceanic flood basalt plateaus

The Ontong–Java Plateau, located near the Solomon Islands in the western Pacific Ocean, is the largest oceanic flood basalt plateau on Earth. As summarized by Fitton and Godard (2004), the dates for the Ontong–Java eruptions are somewhat enigmatic. The Ontong–Java flood basalts erupted either in a single massive flood eruption (~122 Ma) or in a series of eruptions spread over 10 million years with the initial massive outpouring occurring ~122 Ma. The Ontong–Java Plateau encompasses a surface area of 2 million km^2 and a volume of 60 million km^3. On the basis of ten Ocean Drilling Project (ODP) rock core analyses, the Ontong–Java Plateau region is thought to consist largely of a relatively homogeneous low potassium tholeiite that erupted as massive sheet flows and pillow basalts, accompanied by minor volcaniclastic and vitric tuff deposits (Coffin and Eldholm, 1994; Fitton and Godard, 2004). On the basis of geochemical data (e.g., enrichment in incompatible elements Zr to Lu), Tejada et al. (2004) suggest that the Ontong–Java basalt magma was derived by 30% melting of a primitive, enriched, high magnesium (15–20 wt % MgO) mantle source.

The origin of oceanic intraplate magmatism continues to be an active area of research. At least in some cases, intraplate magmatism may be partially derived by deep convection cells involving ancient subducted ocean lithosphere. In other cases, intraplate magmatism may be driven by mantle plumes unrelated to plate boundary activities. As discussed in Chapter 1, hotspots tap magmas from different depths such as the lower crust, upper mantle or mantle–core boundary. Geochemical analyses of basalt and seismic tomography continue to provide insight into our attempts to understand these perplexing igneous processes. Our discussion of oceanic intraplate magmatism has centered on varieties of basalt, which dominate these settings. In sharp contrast, continental intraplate magmatism produces a wider range of igneous rock types, discussed in the following section.

10.4.2 Continental intraplate magmatism

Continental intraplate magmatism and volcanism produce:

- Continental flood basalts.
- Continental rift assemblages.
- Bimodal volcanism.
- Layered basic and ultrabasic intrusions.
- Ultrabasic suites that include komatiites and kimberlites.
- An unusual array of alkaline rocks and anorogenic granites.

Continental flood basalts

Examples of huge outpourings of continental flood basalts (CFB) include the Deccan traps of India, Karroo basalts of Africa, Siberian flood basalts of Russia and the Columbia River, Snake River plain and Keweenaw flood basalts of the United States. The three largest flood basalt events – the Permo-Triassic Siberian traps, the Triassic–Early Jurassic Central Atlantic Magmatic Province and the Cretaceous–Tertiary Deccan traps – correspond with the largest extinction events in Earth's history (Renne, 2002).

Although less common, silicic large igneous provinces (SLIP) also occur in association with continental break-up, intraplate magmatism and backarc basin magmatism. SLIP are silicic-dominated provinces containing rhyolite caldera complexes and ignimbrites. SLIP occur notably in the Whitsunday volcanic province of eastern Australia, the Chon Aike Province of South America and Yellowstone (Bryan et al., 2002). Below we briefly describe several well-known continental flood basalt provinces beginning with the CAMP.

Central Atlantic Magmatic Province (CAMP)

The CAMP formed during the Early Jurassic break-up of the Pangea supercontinent, which produced rift basins and flood basalts in North America, South America, Europe and Africa (see Figure 10.11). These once contiguous tholeiitic basalts are now widely dispersed across the Atlantic Ocean realm, encompassing a total area of more than 7 million km^2. The Ar^{40}/Ar^{39} ages indicate that the CAMP basalts erupted between 191 and 205 Ma, with a peak age of 200 Ma. CAMP rocks consist of tholeiitic to andesitic basalts, with rare alkaline and silicic rocks. CAMP tholeiites have low TiO_2 concentrations, negative mantle normalized nobium anomalies and moderate to strongly enriched rare Earth element patterns. These geochemical patterns indicate an anomalously hot mantle plume that resulted in the partial melting of the overlying lithosphere (Marzoli et al., 1999).

Siberian flood basalts

The **Siberian flood basalts** (see Figure 10.11) consist predominantly of tholeiitic basalt flows tens to a few hundreds of meters thick with minor trachyandesites, nephelinites, picrites, volcanic agglomerates and tuffs (Zolotukhin and Al'Mukhamedov, 1988; Fedorenko et al., 1996). The 251 Ma Siberian flood basalts were already recognized as one of the greatest known outpourings of lava when, in 2002, the western Siberian Basin flood basalt province was discovered which effectively doubled the aerial extent of the Siberian traps to approximately 3,900,000 km^2 (Reichow et al., 2002). It is analogous to burying half of the contiguous United States in lava. In the Maymecha-Kotuy region of Russia, Kamo et al. (2003) suggest that the entire 6.5 km thick basalt sequence erupted within ~1 million years, based upon U/Pb dates obtained from the base (251.7 ± 0.4 Ma) and top (251.1 ± 0.3 Ma) of the basalt sequence – truly mind boggling in scale.

Deccan traps

Over 1,000,000 km^3 of flood basalt erupted in southwestern India between 65 and 69 Ma (Courtillot et al., 1988). The Deccan traps (see Figure 10.11) encompass an area of 500,000 km^2 in western India. Individual lava flows generally vary from 10 to 50 m in thickness with total flow thicknesses varying from less than 100 m to more than 2 km (Ghose, 1976; Sano et al., 2001). The flood basalts and related dike swarms are interpreted to result from rifting as the Indian Plate migrated over a mantle plume (Muller et al., 1993). The Deccan traps are dominated by tholeiitic basalts with minor amounts of alkalic basalts. Geochemical studies suggest that the Deccan basalts originated by fractional crystallization of shallow magma chambers (~100 kPa, 1150–1170°C). The basaltic magma experienced variable degrees of contamination as it ascended and assimilated granitic crustal rocks (Mahoney et al., 2000; Sano et al., 2001).

Columbia River flood basalts

Although relatively small compared to the flood basalt provinces listed above, the Columbia River flood basalts (see Figure 10.11) are among the most studied CFB on Earth. The Columbia River flows consist largely of quartz tholeiites and basaltic andesite, with 47–56 wt % silica (Swanson and Wright, 1980; Reidell, 1983). Columbia River basalts crop out in the US states of Washington, Oregon and Idaho, encompassing an area of approximately 163,700 ± 5000 km^2 (Tolan et al., 1989). The total volume of erupted lava has been estimated to be approximately 175,000 ± 31,000 km^3 (Tolan et al., 1989). The Columbia River basalt group has been subdivided into five formations: the Imnaha Basalt, Grande Ronde Basalt, and coeval Picture Gorge, Wanapum Basalt and Saddle Mountain Basalt. The Grande Ronde Basalt, which erupted 15.5–17 Ma, comprises approximately 87% of the total volume of the Columbia River basalt (Swanson and Wright, 1981). Over 300 individual lava flows erupted from northwest trending fractures between 6 and 17 Ma, making this the youngest continental flood basalt province on Earth. Individual flows traveled as much as 550 km, erupting in north–central Idaho and flowing to the Pacific Ocean (Hooper, 1982). Unlike most other flood basalt provinces, the Columbia River basalts lack early picritic basalt eruptions and interbedded silicic lavas and have less than 5% phenocrysts (Durand and Sen, 2004). The low concentrations of phenocrysts are thought to be related to either rapid ascent of magma (McDougall, 1976) or to a high water content of ~4.4%, which effectively lowered the melting temperature and inhibited the early development of large crystals (Lange, 2002).

Various hypotheses have been proposed for the origin of the Columbia River basalt flows. One set of hypotheses propose that the basalts crystallized from primary magma (Swanson and Wright, 1980, 1981). The relatively low Sr^{87}/Sr^{86} (0.7043–0.7049) ratios indicate a mantle source (McDougall, 1976). However, the high SiO_2 concentrations, high total FeO (9.5–17.5 wt %) and low MgO concentrations (3–8 wt %) suggest that the parental magma was not primary (Hooper, 1982; Lange, 2002). A second set of hypotheses assert that the basalts are the product of diversification processes (McDougall, 1976; Reidel, 1983). Viable diversification models suggest that partial melting of pyroxenite (Reidell, 1983) or eclogite (Takahashi et al., 1998) parental rock was followed by the injection of separate magmatic pulses, subsequent magma mixing and assimilation of crustal rock. Trace element data suggest that the Columbia River basalts were not derived by fractionation of a single magmatic pulse. Thus, a unified model suggests that the Columbia River basalts were created by multiple pulses of heterogeneous mantle-derived magmas, contaminated by continental crust during magma ascent and magma mixing (Hooper, 1982; Riedel, 1983). The tectonic origin of the Columbia intraplate magmatism has been the subject of debate. Possible tectonic causes include: heating following subduction of the Juan de Fuca ridge, backarc spreading, the Yellowstone hotspot and continental rifting.

Continental rifts

Continental rifts produce a wide array of rocks that include alkalic basalt as well as alkaline and silicic rocks. Alkaline rocks include phonolite, trachyte and lamproite. Silicic rocks include rhyolite and rhyodacite, which occur in lava domes or as pyroclastic flow and ash fall deposits. Plutonic rocks vary from syenite and alkali granite to gabbroic rocks.

Continental rifting occurs in regions such as the East African rift basin, Lake Baikal (Russia), the Basin and Range and the Rio Grande rift system (USA). The East African rift system (Figure 10.13) erupts abundant alkali basalt as well as phonolite, trachyte, rhyolite and carbonatite lava. Ancient continental rifts include the Permian age (~250 Ma) Rhine Graben (Germany) and Triassic (~200 Ma) Oslo Graben (Norway) and rift basins of the Atlantic Ocean basin and the 1.1 Ga Keweenaw rift of the Lake Superior basin (USA). Continental rift zones can contain important hydrocarbon reservoirs because of

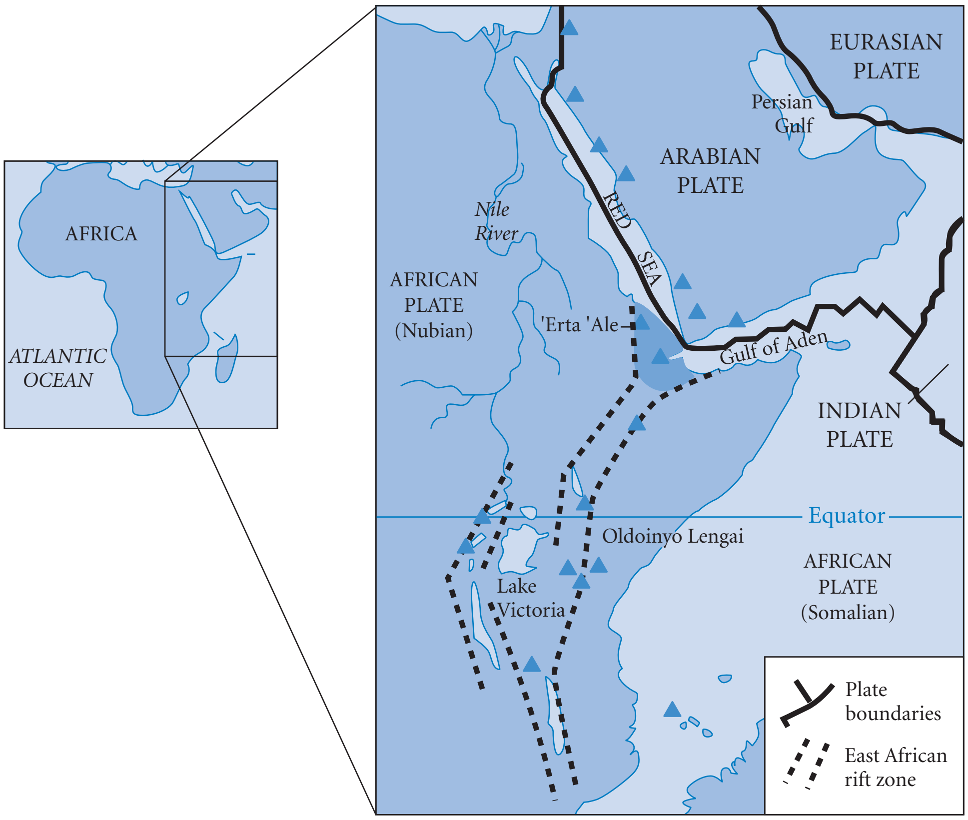

Figure 10.13 The East African rift system represents the third leg to the Gulf of Aden and Red Sea rift chain. (Courtesy of the US Geological Survey.)

the rapid deposition of organic-rich sediments. Volcanic flows and associated shallow intrusives can also provide valuable metallic ore deposits such as nickel and copper and platinum group elements.

What is the driving force behind the lithospheric extension that leads to the development of continental rifts? Various hypotheses have been proposed, which include (Figure 10.14):

- Upwelling of hot plumes generated by the return convective loop of downgoing oceanic lithosphere.
- Partial melting at great depths of overthickened continental lithosphere following supercontinent assembly.
- Subduction of ocean spreading ridges resulting in shallow sub-lithospheric melting producing backarc basin type extension within the continental lithosphere.

All of these forces have the potential to generate continental rifts.

Bimodal volcanism

The widespread occurrence of basalt and rhyolite without significant andesite is referred to as **bimodal volcanism** (Section 8.4). Bimodal volcanism occurs at continental rifts and hotspots underlying continental lithosphere. Partial melting of the mantle generates basaltic magma. The rising basaltic magma partially melts continental crust, resulting in the dual occurrence of basalt and rhyolite. A classic example occurs in Yellowstone National Park in Wyoming (USA). Yellowstone's magmatic source is related to a mantle hotspot that has been active for at least 17 million years. The Yellowstone hotspot may have provided the source material for the immense Columbia River flood basalts in Idaho, Oregon and Washington as well as the

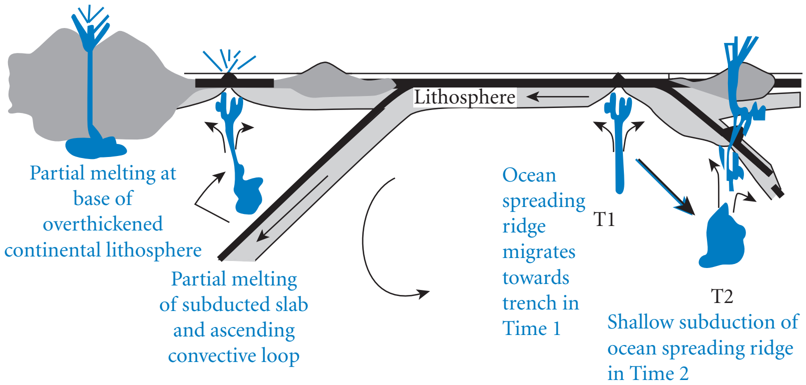

Figure 10.14 Possible tectonic causes for continental rifts.

northern parts of California and Nevada. Most of the magma producing the Columbia River flood basalts erupted 15–17 million years ago. Since that time, as North America has migrated in a southwest direction, the position of the active hotspot has migrated ~800 km in a northeasterly direction to its present location at Yellowstone.

Christiansen (2001) recognized three immense rhyolitic lava deposits at Yellowstone's silicic large igneous province: the 2.1 Ma Huckleberry Ridge Tuff, the 1.3 Ma Mesa Falls Tuff and the 640,000-year-old Lava Creek Tuff. Together, these three tuff deposits constitute the Yellowstone Group. The Huckleberry Ridge eruption dispersed 2450 km^3 rhyolite deposits over an area of 15,500 km^2 and produced a caldera over 75 km long. The Mesa Falls eruption produced tuff deposits largely within the Huckleberry Ridge Caldera. While the Mesa Falls eruptive deposits were restricted to the pre-existing caldera, a new 16 km caldera developed along the northwest end of the Huckleberry Ridge Caldera. The youngest Lava Creek cycle of eruptive activity began around 1.2 Ma and continued for approximately 600,000 years. The Lava Creek eruption produced a large caldera and scattered rhyolitic deposits over an area of 7500 km^2.

Thus the Yellowstone Caldera is a composite caldera generated by three separate rhyolitic eruptive events. In the intervening time between each of these rhyolitic eruptions, basaltic lava also erupted (Figure 10.15). The

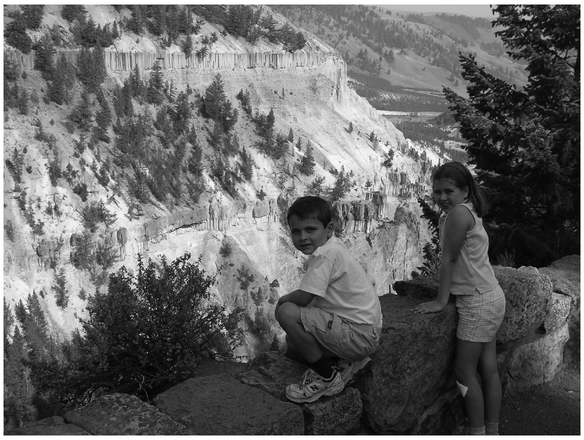

Figure 10.15 Dual columnar basalt flows are separated above and below by massive rhyolite ignimbrite deposits in the Yellowstone Caldera, USA. (Photo by Kevin Hefferan.)

basalt eruptions appear to be independent of the rhyolite eruptive cycles. The rhyolite and basalt eruptions represent two distinctly different magmatic sources. The rhyolitic magma is derived from the successive emplacement of granitic batholiths within the crust. The basaltic magma is generated by partial melting of the peridotite-rich upper mantle. The Yellowstone Caldera consists of two ring fracture zones within this composite caldera structure. Ring fractures are circular fracture sets generated by ground subsidence following the release of magma from a shallow pluton (Chapter 8).

Approximately 40 rhyolite eruptions have occurred in the past 640,000 years, since the last of the three cataclysmic Quaternary eruptions at Yellowstone. No lava has erupted in Yellowstone over the past 70,000 years. Two resurgent domes are currently being constructed within the Yellowstone Caldera and the ground surface is slowly being inflated, with uplift as much as 1 m since the 1920s. While the eruption of lava at Yellowstone is not anticipated in the next few thousand years, the area is presently experiencing uplift, perhaps the early warning signs of a new eruptive phase (Christiansen, 2001). Yellowstone has been the subject of a movie entitled *Supervolcano*, which is entirely appropriate: eruptions there were among the largest on Earth. The magma that erupted from Yellowstone 2.1 million years ago was approximately 6000 times greater than the volume released in the 1980 eruption of Mt St Helens. The smallest of Yellowstone's three Quaternary eruptive events released five times more debris than the massive 1815 Tambora (Indonesia) eruption. It is estimated that 25,000 km^3 of magma are contained within the 7 km deep Yellowstone batholith. Should a portion of that magma erupt from the Yellowstone Caldera, North America would experience a devastating eruption unlike any other witnessed in human history.

Layered basic–ultrabasic intrusions

Layered basic–ultrabasic intrusions are anorogenic bodies injected into stable continental cratons at moderate depths. Layered intrusions include shallow tabular sills and dikes as well as funnel-shaped lopoliths. These intrusions commonly contain layers of rocks such as norite, gabbro, anorthosite, pyroxenite, dunite, troctolite, harzburgite and lherzolite. Minor silicic rocks such as granite can also occur. Common major minerals include olivine, orthopyroxene (enstatite, bronzite, hypersthene), clinopyroxene (augite, ferroaugite, pigeonite) and plagioclase.

Layered intrusions develop by differentiation of eclogite–peridotite parent magmas resulting in mineral segregation within a pluton. In addition to closed-system differentiation processes, open-system diversification processes (Chapter 8) such as multiple injections, magma mixing or chemical diffusion can produce discrete layering in complex intrusive bodies. Layers generated by these processes occur on the scale of meters, centimeters or as microscopic cryptic lenses. Layers may occur as flat, planar structures or display features commonly associated with sedimentation such as cross-bedding, graded bedding, channeling (Figure 10.16) or slump structures. Cryptic (hidden) layering is revealed only by subtle changes in chemical composition.

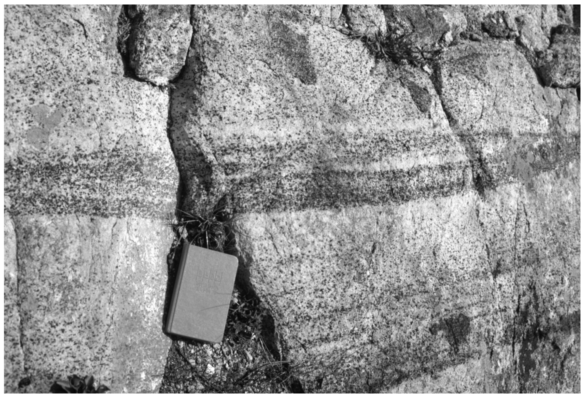

Figure 10.16 Close up of rhythmic layers within a channel structure in the Stillwater Complex. (Photo by Kevin Hefferan.)

As with Alaskan-type intrusions, layered basic–ultrabasc intrusions are highly valued for metal deposits, particularly platinum group elements (PGE) as well as chromium, nickel and cobalt. Metallic ore enrichment is likely due to a combination of factors that include original high concentrations of chromium, nickel, cobalt and PGE in magnesium-rich, refractory magmas as well as subsequent remobilization and concentration by halogen-rich (e.g., chlorine) fluids derived from the assimilation of crustal rock (Boudreau et al., 1997).

Three of the largest layered intrusions on Earth are the Stillwater Complex in Montana, the Bushveld Complex in South Africa and the Skaergaard Intrusion in Greenland. Other significant layered intrusions include the Muskox Intrusion of the Northwest Territories (Canada), the Keweenaw and Duluth Intrusion of Minnesota (USA) and the Great Dike of Zimbabwe. The 1.1 Ga **Duluth Complex**, formed during the Keweenaw rift event, is a major undeveloped PGE source. Plans are currently underway to begin mining PGE in the Duluth Complex within the next few years.

Stillwater Complex

The 2.7 Ga **Stillwater Complex** is a large, layered basic–ultrabasic igneous intrusion in the Beartooth Mountains of southwestern Montana. The Stillwater Complex is exposed along a northwesterly strike for a distance of 48 km, with observable thicknesses up to 6 km. The Stillwater Complex, which formed when basic magma intruded metasedimentary rocks, is the finest exposed layered intrusion in North America and contains economic deposits of platinum group metals as well as chromium, copper and nickel sulfides (McCallum et al., 1980, 1999; Premo et al., 1990).

The Stillwater Complex consists of three main units, which include a lowermost basal zone, an ultramafic zone and an upper banded zone. The basal zone consists of norite, harzburgite and bronzite-rich orthopyroxenite layers. The ultramafic zone consists of dunite, harzburgite, bronzite-rich orthopyroxenite and chromite-rich peridotite layers. The basal and ultramafic zones contain copper, chromium and nickel sulfide ore deposits. The upper banded zone consists largely of repetitive layers of alternating norite, gabbro, anorthosite and troctolite and is enriched in copper, nickel and PGE ore deposits (McCallum et al., 1980, 1999; Todd et al., 1982). Chlorine-rich magmatic fluids played a key role in leaching background metal deposits within the intrusion and concentrating these metals in discrete enriched layers called reefs within the banded zone (Boudreau et al., 1986, 1997; Meurer et al., 1999).

Bushveld Complex

South Africa's 2.06 Ga **Bushveld Complex**, a massive laccolith or domal structure, is the world's largest layered igneous intrusion. Extending over 400 km in length, up to 8 km thick and underlying an area of 60,000 km^2, this complex contains a layered sequence of basic and ultrabasic rocks, capped locally by granite.

The Bushveld Complex consists of four main zones (Daly, 1928; Vermaak, 1976; Cawthorn, 1999). These include, from top to bottom, the following:

1 Upper zone consisting of gabbro and norite.
2 Main zone containing gabbro and anorthosite.
3 Critical zone consisting of anorthosite, norite and pyroxenite.
4 Basal zone consisting of orthopyroxenite, harzburgite, dunite and peridotite. A chromite horizon occurs at the top of the basal series.

The Bushveld Complex hosts the largest reserves of vanadium, chromium and platinum group metals in the world. PGE are concentrated within what is referred to as the Merensky Reef within the critical zone. Anorogenic granitic rocks capping the complex contain tin, fluorine and molybdenum. The Bushveld Complex layering formed through differentiation processes accompanied by a series of magmatic injections, resulting in a massive laccolith or domal structure. As in the Stillwater Intrusion described above, chlorine-rich magmatic fluids are thought to have played a role in concentrating PGE in the Bushveld Complex (Boudreau et al., 1986).

Skaergaard Intrusion

Whereas most layered ultrabasic–basic intrusions are Precambrian in age, Greenland's 55 Ma **Skaergaard Intrusion** is the youngest of the great PGE-enriched intrusions. The Skaergaard lopolith intrusion crops out along Greenland's eastern shores and offers exceptionally good exposures of layering formed by differentiation and convective current structures. The Skaergaard Intrusion, with a volume of 500 km^3, is heralded as the finest example on Earth of fractional crystallization, displaying layered sequences of euhedral to subhedral crystals as well as distinctive structures usually associated with sedimentary beds. These structures include cross-bedding, graded bedding and slump structures (Wager and Deer, 1939; Irvine, 1982; Irvine et al., 1998).

Zoned and layered ultrabasic–basic intrusive complexes provide rare but massive examples of magma diversification yielding segregated mineral zones and valuable metallic ore deposits.

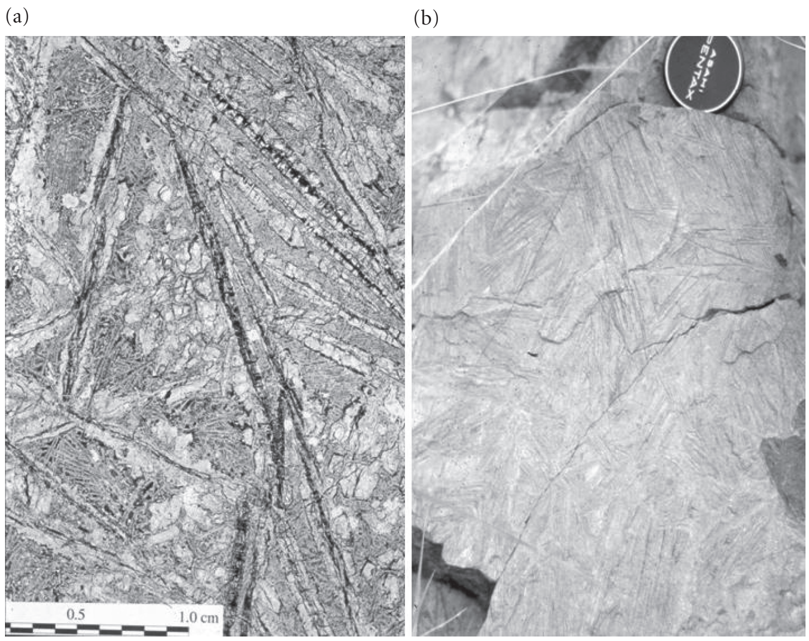

Figure 10.17 Microphotograph (a) and field photograph (b) of spinifex texture komattites. (Photos courtesy of Maarten de Wit.)

Other ultrabasic suites: intraplate volcanics and shallow intrusives

Komatiites

Komatiites are ultrabasic volcanic rocks found almost exclusively in Archean (>2.5 Ga) greenstone belts. Greenstone belts are metamorphosed assemblages of green-colored rocks that contain layers of ultrabasic and basic rocks overlain by silicic rocks and sediments (Chapter 18). Komatiites, named after the 3.5 Ga Komatii region of Barberton, South Africa, are high magnesium (>18% MgO), olivine-rich volcanic rocks, depleted in titanium and LREE. The high magnesium content and LREE depletion indicate a previously depleted mantle source (Sun and Nesbitt, 1978, in Walter, 1998). Komatiite flows, first recognized in the Barberton region of South Africa in 1969, commonly contain spinifex texture (Figure 10.17). **Spinifex texture** consists of needle-like, acicular olivine, pyroxene (augite and/or pigeonite) and chromite phenocrysts in a glassy groundmass (Viljoen and Viljoen, 1969; Arndt, 1994). Spinifex texture commonly occurs in the upper parts of komatiite flows or in the chilled margins of sills and dikes where rapid quenching produced skeletal, acicular crystals (Arndt and Nesbitt, 1982). In addition to spinifex texture, circular varioles, radiating spherulites and tree-like dendritic textures also occur. These textures are attributed to rapid undercooling (Chapter 8) or quenching of extremely hot lavas (Fowler et al., 2002).

Nearly all komatiites erupted during the Archean Eon when the early Earth was much hotter. Komatiites indicate elevated liquidus temperatures of 1575–1800°C (1 atmosphere pressure) in the Archean upper mantle (Green et al., 1975; Arndt, 1976; Wei et al., 1990; Herzberg, 1992, 1993, in de Wit, 1998). The virtual absence of Phanerozoic komatiites may be attributed to lower upper mantle temperatures which precludes the extensive mantle melting required to produce ultrabasic melts. The only known Phanerozoic (<544 Ma) komatiites occur on Gorgona Island, Colombia, where 88 Ma komatiites erupted as >1500°C ultrabasic lava flows. Gorgona Island, located 80 km west of Colombia in the Pacific Ocean, is composed largely of gabbro and peridotite (Echeverria, 1980; Aitken and Echeverria, 1984). Gorgona Island is also notable for the rare occurrence of ultrabasic pyroclastic tuffs which record explosive volcanism (Echeverria and Aitken, 1986).

Hypotheses for the origin of komatiites include:

- Melting in the hydrated mantle wedge above the subduction zones (Allegre, 1982; Grove et al., 1997; Parman et al., 2001).
- A deep mantle plume hotspot that led to large degrees of partial melting producing oceanic plateaus (Storey et al., 1991).
- Partial melting (10–30%) of a garnet peridotite at pressures of 8–10 GPa (Walter, 1998).

Komatiites, like layered gabbroic intrusions, are associated with valuable metallic ore deposits such as nickel, copper and platinum metals. For example, komatiite metallic ore deposits occur in the 2.7 Ga Yilgarn Craton of Western Australia, the 3.5 Ga South African Barberton region and the 2.7 Ga Canadian Shield. The nickel sulfide ore deposits are thought to have originated in ultrabasic lava tubes (Chapter 9) that concentrated high density metals in channel beds.

Kimberlites

Kimberlites are brecciated, magnesium-rich, ultrabasic rocks that rapidly rise to Earth's surface via cylindrical diatremes (Chapter 8) from deep within the mantle. Diatremes vary greatly in surface area, ranging from a few square meters to square kilometers. Most diatremes taper downward, resembling an inverted cone in cross-section view. Kimberlite pipes occur with other plutonic structures such as dikes and sills. Kimberlites, which originate at temperatures of 1200–1400°C and depths exceeding 150 km, rise explosively through thick continental lithosphere.

Volatile, enriched, very low viscosity mantle melts rocket upward towards Earth's surface at velocities of ~15–72 km/h (Sparks et al., 2006). The magma is propelled upward by either the degassing of CO_2-enriched magma or by phreatomagmatic processes (Chapter 9). Phreatomagmatic processes require a water source to interact with the kimberlite magma. The high volatile content serves two primary purposes in that (1) it lowers the melting temperature preventing crystallization, and (2) it provides the propellant "jet fuel" to accelerate kimberlite magma to Earth's surface. Sparks et al. (2006) suggest that kimberlite eruptions generate up to 10,000 m^3 of pyroclastic debris over hours to months, producing Plinian ash plumes up to 35 km high (Chapter 9). Strangely, no ultrabasic lavas have been documented with kimberlite deposits. This is probably due to their low preservation potential and the extremely high volatile (up to 20%) content of kimberlite magma, which can produce 70% vesiculation in the erupting lava (Sparks et al., 2006).

Kimberlite eruptions form maar craters (Chapter 9) that largely fill with brecciated, pyroclastic debris (Dawson, 1980; Mitchell, 1986; Sparks et al., 2006). Due to the association of high temperature, pressure, volatile content and velocity, kimberlites commonly exhibit extensive hydrothermal alteration and are intensely fractured (Dawson, 1980). Altered brecciated olivine and phlogopite phenocrysts occur within a fine groundmass of serpentine, calcite and olivine. Olivine constitutes the major mineral in the vast majority of kimberlites. However, in many samples olivine is completely replaced by serpentine, mica or clay minerals (Skinner, 1989). Kimberlites also contain the high pressure minerals pyrope garnet, jadeite pyroxene and diamond, which are stable at mantle depths >150 km.

Kimberlites were first discovered in the Kimberly region of South Africa where they are intimately associated with diamonds. Although best known from South Africa, kimberlites crop out in continental lithosphere throughout the world, commonly occurring with carbonatites and alkaline igneous rocks. Carbonatites – igneous rocks enriched in carbonate minerals such as calcite, dolomite or ankerite – are important CO_2 energy sources propelling kimberlites up from mantle depths. Kimberlites are also associated with reactivated shear zones and fracture zones (White et al., 1995; Vearncombe and Vearncombe, 2002). Kimberlites occur primarily in Early Proterozoic to Archean age cratons (2–4 Ga), although kimberlites as young as Tertiary age (~50 Ma) are known (Dawson, 1980).

Carbonatites, lamprophyre, lamproites and anorogenic granites

In addition to kimberlites, other rare and unusual rocks that occur in continental lithosphere include carbonatites, lamprophyres and lamproites. These SiO_2-undersaturated rocks typically occur in shallow (hypabyssal), volatile-rich dikes and may be associated with kimberlites.

Carbonatites are shallow intrusive to volcanic rocks that contain >20% CO_3 minerals such as natrolite, trona, sodic calcite, magnesite and ankerite as well as other minerals such as barite and fluorite. The origin of carbonatite was a contentious issue prior to the 1960 eruption of the Oldoinyo L'Engai Volcano in Tanzania. Oldoinyo L'Engai erupted unusually low viscosity pahoehoe carbonatite lava at temperatures of ~500°C. Carbonatites form in stocks, dikes and cylindrical structures primarily at continental rifts (Dawson, 1962).

Lamprophyres are magnesium-rich, volatile-rich, porphyritic rocks containing mafic phenocrysts such as biotite, phlogopite, amphibole, clinopyroxene and melilite. Lamprophyres are associated with kimberlites and continental rift zones, but also occur as dikes intruding granodiorite plutons at convergent margin settings.

Lamproites are potassium-rich, peralkaline rocks containing minerals such as leucite, sanidine, phlogopite, richterite, diopside and olivine. Lamproites are enriched in barium (>5000 ppm), lanthanum (>200 ppm) and zirconium (>500 ppm). In contrast to lamprophyres and carbonatites, lamproites are relatively poor in CO_2 (<0.5 wt %). Lamproites occur in areas of thickened lithosphere that have experienced earlier plate convergence or rifting episodes.

Anorogenic (A-type) granites are silicic plutonic rocks that are not associated with convergent margin tectonism. A-type granite environments include stable cratons, continental rifts, ocean islands and inactive, post-collisional continental margins. Anorogenic granite, alkali granite and syenite were particularly common 1.1–1.4 Ga following the assembly of the mid-Proterozoic Columbia Supercontinent. These A-type, granitic intrusions are widespread in North America, extending from Mexico to the Lake Superior region. Significant volumes of anorogenic granites occur in Precambrian cratons throughout the world. These mid-Proterozoic granitoid rocks are remarkably similar in age, composition and appearance, displaying rapakivi texture. **Rapakivi** texture refers to sodium plagioclase overgrowths on pre-existing orthoclase crystals.

A number of models have been proposed for the origin of A-type granites. One model proposes the overthickening of continental lithosphere such that the upper mantle and base of the crust partially melt generating silicic magma that subsequently rises and cools at shallower depths to form anorogenic granite. Other "residual source models" propose that A-type granites, such as Pikes Peak Batholith in Colorado (USA), develop from the partial melting of residual silicic granulite rocks (Chapter 18) that had previously generated I-type granites (Barker et al., 1975; Collins et al., 1982). Alternative models suggest that A-type granites are derived by melting quartz diorite, tonalite or granodiorite parent rocks (Anderson, 1983).

In Chapters 7–10 we have presented a logical approach to the description, classification and origin of igneous rocks and landforms. We have also demonstrated the tectonic relations in an understandable framework. Hopefully we have been somewhat successful in helping you understand igneous processes. The scope of this text requires us to limit our discussion of important topics. For more detailed discussions beyond the scope of this textbook, the reader is referred to excellent petrology textbooks by Winter (2009), Raymond (2007), McBirney (2007), Best (2003), Blatt and Tracy (1996), Philpotts (1990), Ragland (1989) and Hyndman (1985) among others.

In succeeding chapters, we will investigate how igneous rocks are altered in two ways: (1) by weathering and erosion at Earth's surface, and (2) through the effects of high temperatures, pressures and hot fluids via metamorphic reactions. In all of these reactions, water plays a critical role in altering and mobilizing elements within Earth's crust.