CHAPTER 9

Magma Ascent and Emplacement: Field Relations of Intrusions

Fundamental Questions Considered in This Chapter

1. How does magma rise from its site of generation in the deep crust and upper mantle and move to shallower depths in the solid lithosphere: That is, how does magma *ascend?*
2. How is room created in solid rock for magma to intrude: That is, how is magma *emplaced?*

INTRODUCTION

It has been recognized for at least a century that the fundamental driving force causing magma to rise is its buoyancy—the difference between the density of the magma and its surrounding country rock. Magma generated by partial melting of solid source rock in the deep crust or uppermost mantle is less dense than the surrounding rock and is, therefore, gravitationally unstable and capable of rising. But whether a mass of magma can actually ascend buoyantly depends on the relative magnitude of the resistive force that is dictated by the magma rheology, basically its viscosity. Large volumes of silicic magma are required to provide the necessary buoyant force to overcome the viscous resistive force at the margin of the magma body where it contacts solid country rock. Smaller volumes of less viscous basaltic magma can ascend with greater facility and more speed, even in large surface area dikes that have been created by tectonic forces or by the pressure of the magma itself.

Ascending magma constitutes a classic interaction of the two fundamental but opposing energy sources—thermal and gravitational—within the Earth. Melting at the magma source might proceed without bound until all of the source rock is melted and the causative mechanism (e.g., decompression, volatile influx) stops. But in a gravitational field, the partially melted rock or a segregated partial melt usually becomes sufficiently buoyant to rise out of the source before complete melting occurs. Gravity-driven ascent is commonly arrested as the magma loses heat to the country rocks, becomes more viscous, and stops flowing. Or magma may ascend all of the way to the surface of the Earth and exit as a volcanic extrusion.

During ascent and intrusion of magma, nature exercises its principle of parsimony, taking the path of least work that consumes the least energy.

Magma ascent and final emplacement cannot always be separated and distinguished; rather, they are a dynamic continuum. Thus, a vertical dike through which magma ascended from a deep source also constitutes an intrusion, unless somehow the magma drained out or the crack closed together to eliminate the filling magma.

The challenge for the petrologist is to try to understand the physical and thermal dynamics of ascent and emplacement of a magma body from the field relations, fabric, and composition of a partially exposed and long-dead cold intrusion.

Magma ascent constitutes a significant advective transfer of heat from deep levels of the lithosphere to shallower and, therefore, plays a major role in the thermal evolution of the cooling Earth.

☼9.1 MOVEMENT OF MAGMA IN THE EARTH

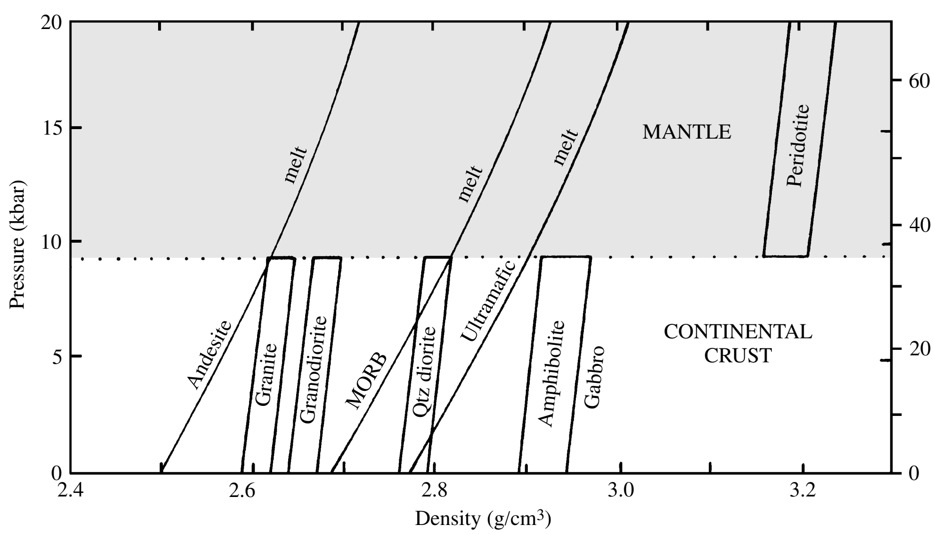
9.1.1 Neutral Buoyancy and the Crustal Density Filter

Mafic to ultramafic magmas in the *uppermost* mantle are less dense than the mantle peridotite in which they might have been generated (Figure 8.17). They are, therefore, positively buoyant and can potentially rise. However, above the Moho, crustal rocks are dominantly feldspathic rather than olivine-rich, and consequently their density is much less than the underlying mantle. Whether mantle-derived magmas are positively, negatively, or neutrally buoyant in the crust depends entirely on their *T,* their *P,* and especially their composition and the mineralogical and modal composition of the crustal rock (Figures 8.15 and 9.1). Subtle differences of a few tenths of a gram per cubic centimeter in the density of the magma or the crustal rock are sufficient to change the buoyancy from positive to negative.

The density of the oceanic basaltic crust (about 7 km thick) increases with depth because of progressively decreasing vesicularity, closure of pore spaces, and expulsion of water. Mantle-derived, olivine-rich basalt (picrite) magma and crust are of similar density at a depth of about 1–3 km which is, therefore, a **horizon of neutral buoyancy** at which magma may stagnate and accumulate (Ryan, 1994). Lateral enlargement of magma chambers at oceanic spreading junctures may occur at this horizon. Evolved, less dense magma can rise via dikes to the seafloor where it extrudes.

Continental crust is far more heterogeneous than oceanic crust, making generalizations of magma ascent less certain. Beneath a variably thick veneer of sedimentary rock whose density ranges from 2.2 to 2.7 g/cm^3, crustal igneous and metamorphic rocks have densities ranging from about 2.6 to 2.9 g/cm^3 and averaging aproximately 2.7 g/cm^3, corresponding to granodiorite to diorite bulk compositions (Figure 9.1). In many places, deeper continental crust has densities of approximately 2.9 g/cm^3, corresponding to more mafic compositions, such as amphibolite. Andesite and less dense silicic magmas are probably positively buoyant in crust of any composition and can rise all the way to the surface. Volatile undersaturated basaltic magmas have densities >2.7 g/cm^3 and would not be expected to rise buoyantly through less dense continental crust of granite to granodiorite composition. There is strong geologic evidence (discussed in later chapters) for appreciable **underplating** of the lower continental crust, as well as intrusion into it, by basaltic magmas. Feldspathic crustal rocks can, therefore, serve as an effective **density filter** blocking ascent of denser mafic, mantle-derived magmas. These buoyantly blocked basaltic magmas are believed to be responsible for partially melting the already hot lower crust as they cool and transfer heat. During cooling, they also are likely to fractionate olivine, so their residual melts, which can be less dense (Figure 8.16), may be buoyant enough to rise. Fractionating magmas that contain dissolved, and especially exsolved, volatiles or magmas assimilating silicic crustal material may also have low enough density that they can rise higher into the overlying crust. Repeated stagnation and solidification of mafic magmas within the crust may densify it sufficiently that progressively denser, more primitive basalt magmas may subsequently ascend farther.

Yet, paradoxically, rather primitive, unevolved basalt magmas that have densities greater than that of continental rock are commonly extruded onto the surface as lavas in continental areas. Some magmas have even intruded through porous alluvium (Figure 8.14) whose density can be <2.0 g/cm^3. How can this happen? One possibility is volumetric expansion upon melting in the source, which drives the magma upward



9.1 Density relations between some rock compositions and volatile-free melts in the continental crust and uppermost mantle. Range of densities for rock types is indicated in rectangular boxes. Note the smaller compressibilities of rocks compared to melts as *P* (depth) increases. MORB, mid-ocean ridge basalt. Amphibolite is a metamorphic rock composed of hornblende and plagioclase that is formed by recrystallization of mafic igneous rocks (basalt, gabbro) under hydrous conditions. (Redrawn from Herzberg et al., 1983.)

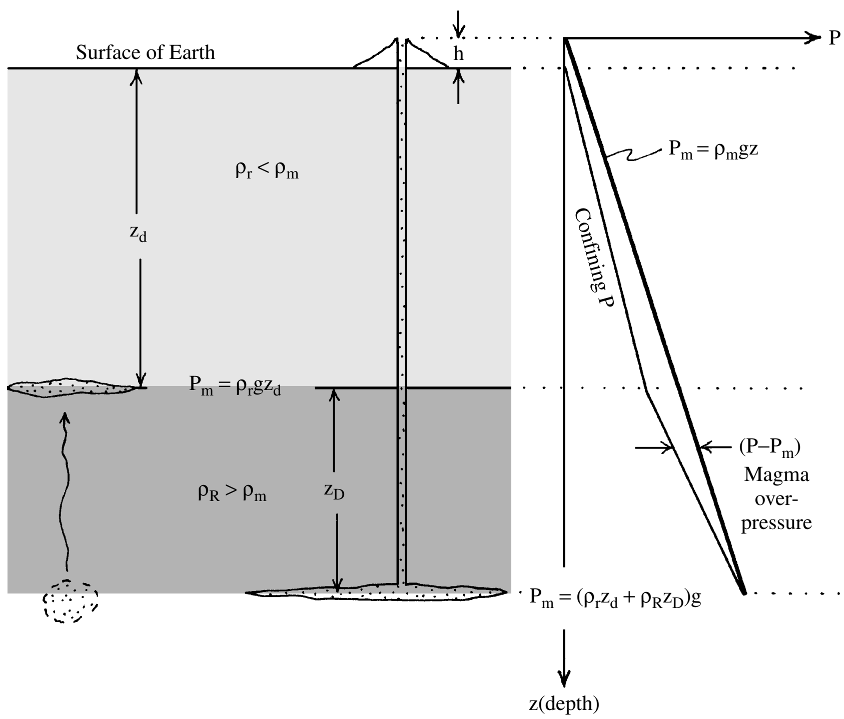
toward a lower P regime. However, it is unlikely that this effect is capable of driving the magma very far from its source. Other driving forces are exsolution and expansion of bubbles in volatile-saturated magmas, which can greatly reduce magma density, enhance buoyancy, and cause volcanic eruption. But magmas must have another driving mechanism, independent of volatile exsolution, that can commonly propel them upward through less dense rock. This driving force is magma overpressure.

9.1.2 Magma Overpressure

Consider a lens-shaped body of magma buoyantly blocked at a density contrast in the lithosphere after having ascended from deeper in the lithosphere (left side of Figure 9.2). The pressure in the magma, P_m, is equal to the weight of the rock column overlying the lens, $\rho_r g z_d$, which is the lithostatic (confining) pressure, P, at that depth (Section 1.2). In a deeper body of magma (center of Figure 9.2) at a depth $z_d + z_D$, the magma pressure, $P_m = \rho_r g z_d + \rho_r g z_D$. Now suppose a conduit is accessible from the deeper magma lens all the way to the surface. Ignoring viscous drag of the magma in the conduit and assuming the density of the magma, ρ_m is constant regardless of depth, how far can the magma rise? The P-z (pressure-depth) diagram on the right of Figure 9.2 indicates that the magma will rise to a height h above the surface, satisfying the equality between lithostatic and magma pressure, $P = P_m = \rho_m g(h + z_d + z_D)$. In other words, the magma has an excess hydraulic head, h, due to the load of the *overall* denser rock column on the less dense body of magma at the base of the column. Provided conduit continuity between subterranean magma body and volcanic vent is maintained, less dense magma is pushed out of the ground by the weight of the overall more dense rock column. Theoretically, the deeper the magma column, the greater is the height to which a volcano can grow above the surface; this may have a bearing on the maximum summit heights of Hawaiian and other volcanoes.

In the P-z diagram on the right of Figure 9.2 note that the magma pressure, P_m, *at any depth* exceeds the confining (lithostatic) pressure, P. The magnitude of this **magma overpressure,** or hydraulic head, at any depth z is $(P - P_m)$. Magma overpressure can be sufficient to cause hydraulic fracturing in brittle rock (Figure 8.2c) or, if not, at least can drive magma into and through existing cracks, overcoming viscous resistance. Or overpressure (in effect, buoyancy) can slowly force magma through ductile overlying rock.

Thousands of Cenozoic basaltic extrusions worldwide in extensional tectonic regimes contain xenoliths of dense mantle peridotite. The basaltic magmas must have ascended rapidly (Problem 8.12) through essentially continuous passageways from a mantle source in order to have lifted the xenoliths. Additionally, copious volumes of magma extruded from fissures in continental and oceanic flood basalt plateaus testify to the



9.2 Origin of magma overpressure. See text for explanation.

tapping of huge subterranean reservoirs and efficient transport to the surface.

9.1.3 Mechanisms of Magma Ascent

Most magmas move upward through solid rock in basically two ways: as diapirs and as dikes. **Diapirs** are bodies of buoyant magma that push slowly through surrounding ductile, highly viscous country rock in the lower crust or mantle. Diapir originates from the Greek verb *diaperien,* "to pierce." The existence and nature of magmatic diapirs are inferred from examination of field relations of intrusive magma bodies, model studies of viscous fluids, and theory. Magma can also rise rapidly through subvertical cracks in brittly fractured rock as **dikes.** Their existence is a matter of simple observation. Movement of magma through fractures has been tracked seismically. Table 9.1 compares these two end-member processes that operate at vastly different time scales and depend on contrasting magma and host rock rheologies. Ascent of a particular mass of magma can involve either mechanism at different depths. Shallow crustal processes including stoping and "drilling" of gas charged magma are considered further in the section on magma emplacement (Section 9.4).

9.2 SHEET INTRUSIONS (DIKES)

9.2.1 Description and Terminology

Sheet intrusions, as the name implies, are tabular bodies having very small **aspect ratios** of thickness/length, generally $10^{-2}-10^{-4}$. A **dike** is a sheet intrusion that cuts discordantly across planar structures, such as bedding, in its host rock. Dike also refers to a sheet intrusion hosted within massive, isotropic rock, such as granite. In contrast, a **sill** is a concordant sheet intrusion that parallels planar structures in its host rocks (Figure 9.3). Some geologists define a sheet intrusion, regardless of country rock concordancy, as a sill if horizontal, or nearly so, and as a dike if vertical, or nearly so.

Compositionally diverse dikes and sills are commonly associated together beneath volcanoes and near margins of larger intrusions (Figures 9.4 and 9.5) where they inflate the volume of rock into which they are intruded.

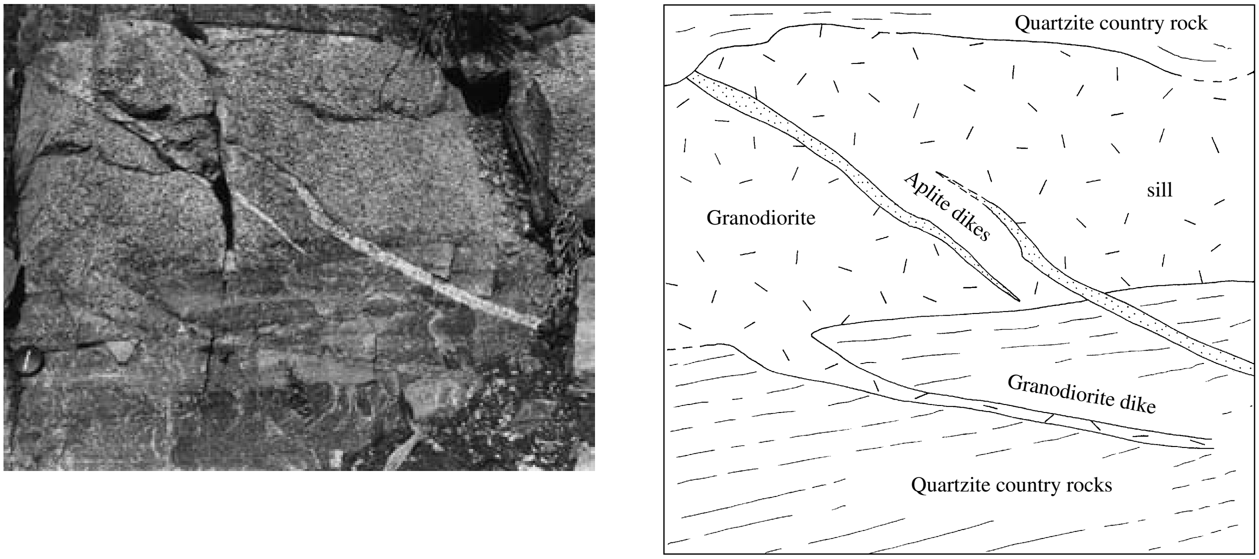
Dike swarms consist of several to hundreds of dikes emplaced more or less contemporaneously during a single intrusive episode. Dikes in swarms may be irregular in orientation, more or less parallel, or **radial,** arrayed in map-view-like spokes of a wheel from a central point (Figure 9.6a). Huge radial swarms (Ernst and Buchan, 1997) are believed to form above mantle plumes associated with continental extension and ocean opening (Figure 9.7). Individual dikes in such swarms can be >2000 km long and tens to rarely 100 meters in width. However, most common dikes are less than 10 m wide; 1- to 2-m-wide dikes are typical.

On a global perspective, most dikes are of basaltic composition and manifest ascent of a vast volume of mantle-derived magma through fractured lithosphere throughout Earth history (see Figure 19.12).

Feeder dikes supply magma to connected sills or other intrusions and overlying volcanoes. The huge radial swarms just mentioned likely fed copious extrusions of lava, perhaps forming vast thick flood

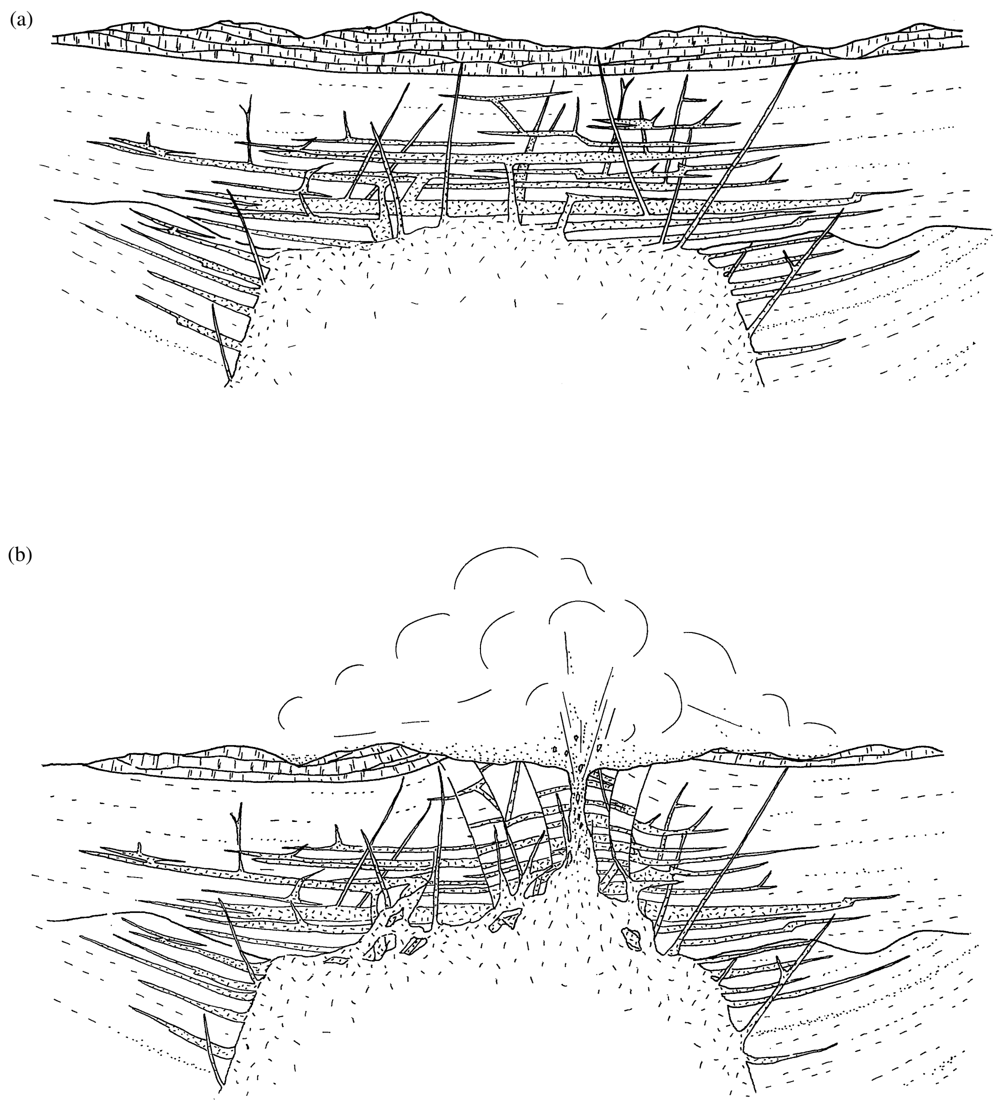
Table 9.1 Comparison of Sheet Intrusion (Diking) and Diapirism in the Continental Crust

Aspect	Sheet Intrusion	Diapirism
Most common magma composition	Basalt	Granitic
Rheologic behavior of country rock	Brittle (elastic)	Ductile or viscoplastic
Viscosity contrast between country rock and magma	Many orders of magnitude	A few orders of magnitude
Ascent velocity	0.1–1 m/s	0.1–50 m/y
Time for magma ascent	Hours to days	10^4–10^5 y
Factors controlling ascent velocity	Magma viscosity and density contrast with country rock; dike thickness	Country rock ductile strength and thickness of boundary layer around diapir
Effect of state of stress on path of magma transport	Sheet perpendicular to least principal stress, σ_3	Probably slight
Country rock deformation	Nil	Substantial penetrative ductile, chiefly in boundary layer
Nonmagmatic example	Hydrothermal quartz vein	Salt dome

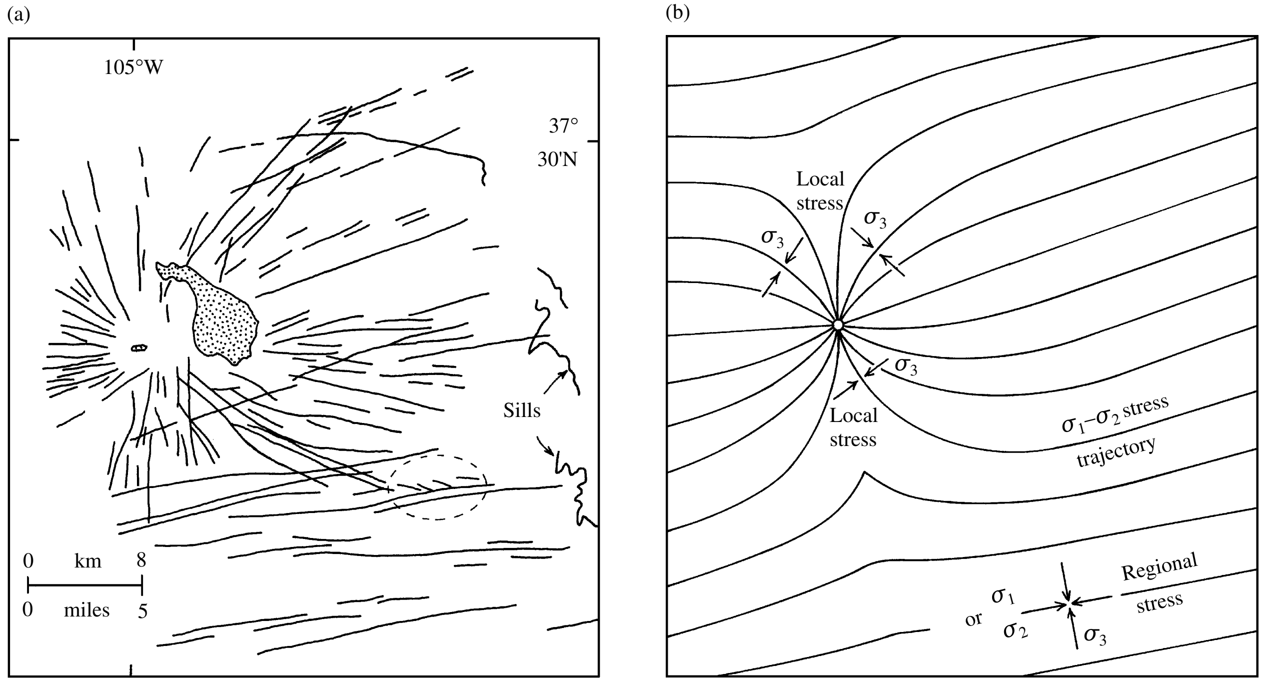


9.3 Sill and dikes. A **sill** of granodiorite intruded concordantly with layering in quartzite host rock and a smaller offshoot **dike** penetrating discordantly across layering. Thin subparallel dikes of leucocratic granite aplite cut across more mafic granodiorite and layered country rock. Note sharp contacts. Camera lens cap in lower left for scale. (a) Photograph. (b) Annotated sketch.

9.4 Sill-dike swarm in subhorizontally layered country rocks surrounding granodiorite pluton, Alta, Utah. Most of the leucocratic aplite-pegmatite granitic rock forms **sills** parallel to layering in darker-colored recrystallized shale, but smaller **dikes** cut discordantly across layers. The original stratigraphic section of shale is approximately doubled in thickness by the inflating sills. Pocket knife 8 cm long for scale in center of photograph.



9.5 Schematic cross sections illustrating inferred relations among main intrusion, overlying dike-sill swarm, and layered volcanic rocks in the Miocene-Pliocene Tatoosh complex in Mount Rainier National Park, Washington. (a) In an early stage in the rise and emplacement of the intrusion, magma is lodged in an overlying dike-sill complex, advectively heating the roof of older volcaniclastic rocks. (b) In a later stage, the main mass of magma has continued to rise by **stoping** into its dike-sill complex and by **doming** the sills, their older host rock, and overlying volcanic rock layers. Note stoped blocks of roof rock in the main intrusion. Magma has broken through to the surface in an explosive eruption. A still later stage can be envisaged in which the still ascending magma intrudes its own volcanic cover. (Redrawn from Fiske et al., 1963.)



9.6 Radial and **parallel dike swarms.** (a) Subvertical dikes were emplaced at 28 to 20 Ma around **central intrusions** of the Spanish Peaks (stippled) in south central Colorado. Flow markers (aligned tabular phenocrysts, elongate vesicles) in the dikes indicate the central intrusions as the source of the radially diking magma. Most dikes consist of segments a few meters to several kilometers long; many segments are *en echelon* but cannot be shown on this small-scale map except some unusually well-expressed ones enclosed by the dashed-line ellipse. An origin for *en echelon* dikes is shown in Figure 9.9. (Redrawn from Smith, 1987.) (b) Theoretical stress analysis. Central intrusion (open circle) is responsible for a *local stress* field that allows for radial diking. The central intrusion perturbed a *regional stress* field that controlled emplacement of the mostly older swarm of subparallel east-northeast-striking dikes mainly of more mafic magma. Trajectory lines are traces (intersections) in the horizontal plane of vertical surfaces parallel to σ_1 and σ_2. Because these surfaces are perpendicular to σ_3, they are potential avenues for magma intrusion. Note that most radial dikes are oriented nearly parallel to the regional $\sigma_1 = \sigma_2$ trajectory. (Redrawn from Odé, 1957.)

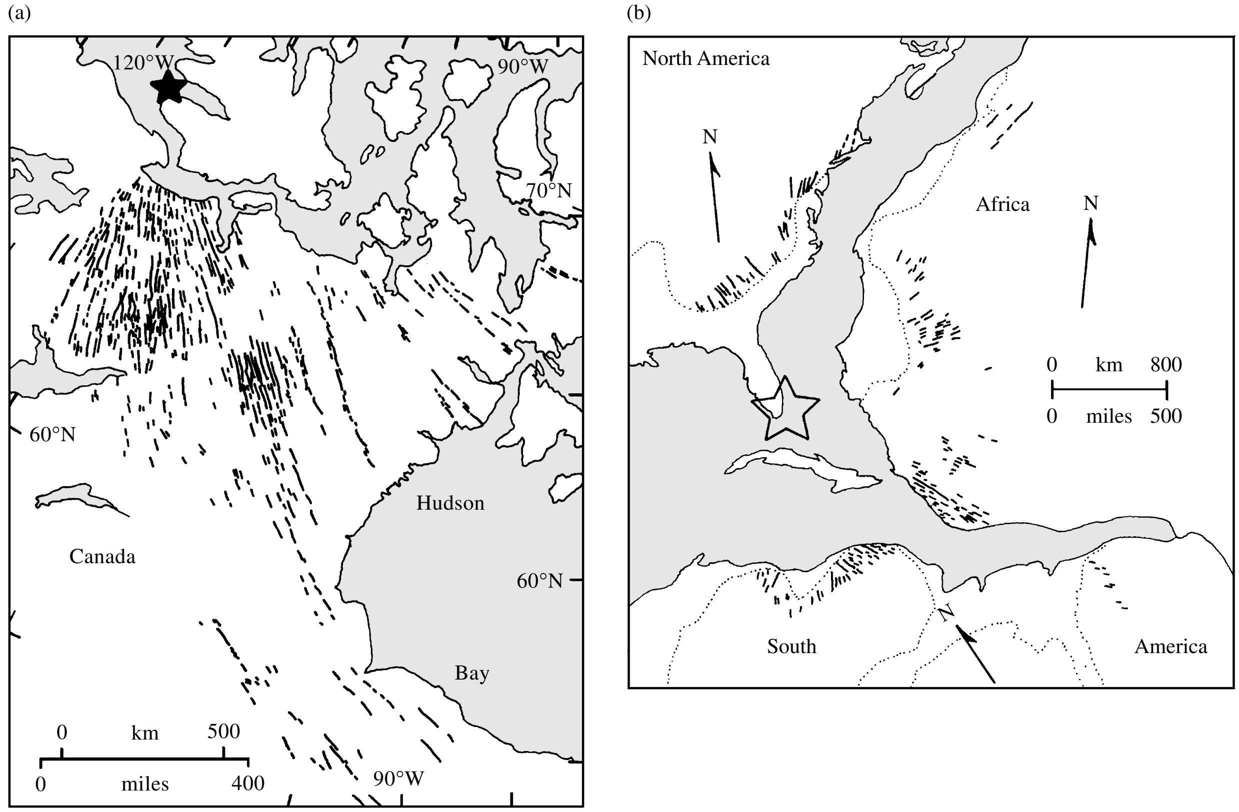
basalt plateaus. Basalt dikes are also the means for growth of continental and oceanic island volcanoes (such as Hawaii, Figure 9.8) and for growth of oceanic-ridge submarine volcano systems related to seafloor spreading and, therefore, for growth of the entire seafloor itself. **Sheeted dike complexes** formed at ocean ridges in the extending oceanic crust consist of subparallel dikes intruded into older dikes. They testify to long-term prolific diking in the dilating ridge.

Some dikes never reach the surface of the Earth where the magma can "see" the light of day; they are "blind" intrusions.

Basalt **sill swarms** underlie flood-basalt fields and thick plateaus and can have volumes comparable to those of dike swarms. A huge swarm of Jurassic diabase sills, together with an accompanying feeder dike swarm, formed during the breakup of Gondwanaland. Segments of these swarms are found in Antarctica, South Africa (see Figure 13.20), and Tasmania, where individual sills are as much as 300 m thick and the swarm segment crops out over 25% of the 65,000-km^2 surface area of the island (Walker, 1993). The famous Carboniferous Whin sill in the British Isles has an areal extent of >5,000 km^2 and an average thickness of 40 m. Sills accompany the dike swarm on continental margins in the mid-Atlantic (Figure 9.7b). Some sills preserve evidence of episodic replenishment during growth. Other sheet intrusions include cone sheets and ring dikes described later.

9.2.2 Some Thermomechanical Concepts Pertaining to Emplacement of Sheet Intrusions

Magma can invade existing fractures in shallow crustal rock if the normal stress perpendicular to the fracture is less than the pressure exerted by the magma and if that pressure is sufficient to overcome the resistance to viscous flow. However, most dikes, even in the mantle and ductile lower crust, are probably created as the magma itself rapidly stresses and fractures the rock (Shaw, 1980) and fills the propagating crack as it advances (Plate VI). This is the hydraulic fracture mechanism shown in Figure 8.2c. The work to fracture rock is less than that required to push magma through a crack, especially for viscous magma. The stress at the tip of the opening crack is sufficient to continue the fracturing process. Evidence for magma-generated



9.7 Radial dike swarms of basalt and diabase related to mantle-plume-induced continental breakup and ocean opening. (a) Basalt dikes of the gigantic 1.27-Ga radial Mackenzie swarm northwest of Hudson Bay, Canada, associated with opening of a middle Proterozoic ocean. Flow markers in the dikes show that the direction of magma transport was subvertical within 500 km of the postulated mantle-plume source (star) but subhorizontal at distances up to 2000 km from the source. Dikes near the source fed a flood basalt province and the large differentiated Muskox intrusion (see Figure 12.17). (Redrawn from Earth and Planetary Science Letters, v. 96, A. N. LeCheminant and L. M. Heaman, Mackenzie igneous events, Canada: Middle Proterozoic hotspot magmatism associated with ocean opening, pp. 38–48, 1989 with permission of Elsevier Science NL, Sara Burgerharstratt 25, 1055 KV Amsterdam, The Netherlands.) (b) Radial swarm of Triassic-Jurassic dikes along the margins of the North American, African, and South American continents. (Redrawn from May, 1971; see also Puffer and Ragland, 1992.) Continents have been restored to an early Mesozoic, predrift configuration. The swarm indicates a possible mantle-plume source at the southern tip of the Florida peninsula (star) (Ernst and Buchan, 1997). Dotted lines show the inland contact of post-Jurassic sedimentary deposits, which likely conceal many additional dikes.

fractures can be found in the time-space relations of dikes, such as a radial swarm centered around a contemporaneous central intrusion (Figure 9.6) and in the absence of similarly oriented joints in the country rocks far from the dikes.

The magma flow velocity through a dike is (Rubin, 1995)

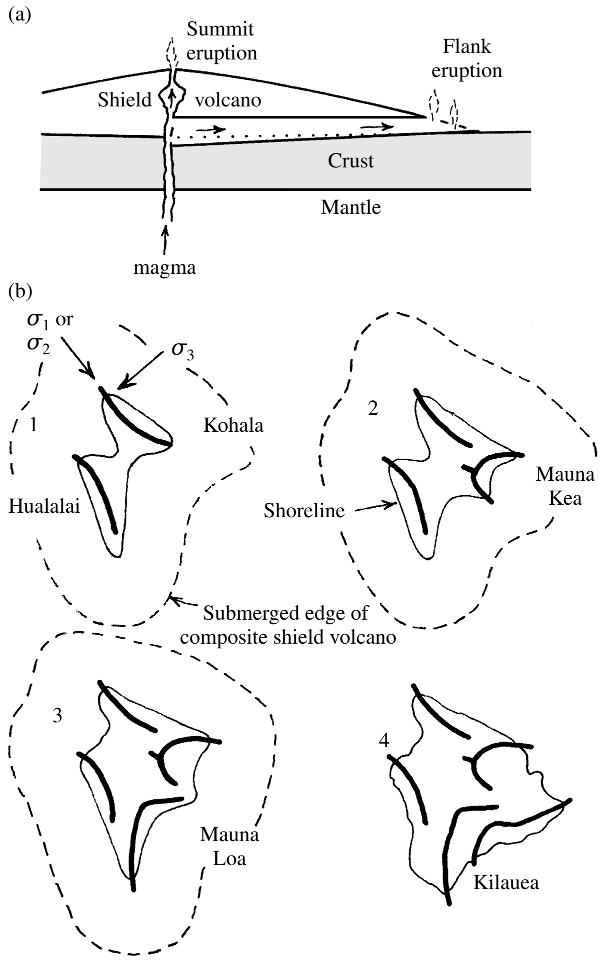
$$9.1 \quad v = \left(\frac{w^2}{3\eta}\right)\left(\frac{dP_m}{dz}\right) = \frac{w^2 g \Delta\rho}{3\eta}$$

where w is the width of the dike, g is the acceleration of gravity, and η is the magma viscosity. The vertical magma pressure gradient, dP_m/dz, that drives ascent equals $\Delta\rho g$ where $\Delta\rho$ is the density contrast between magma and host rock. A pressure gradient of only 0.1 bar/km can maintain a flow velocity of almost 1 m/s (about as fast as a person can walk) in a 5-m-thick basalt dike (Spera, 1980). Doubling dike thickness allows magma to flow four times as fast because there is proportionately less viscous resistance along dike walls for movement of a larger volume of magma.

The distance magma can be transported through a dike from its source depends critically on the competition between two rates:

1. The rate of magma cooling, and accompanying viscosity increase, by conductive (Figure 8.18) and advective heat transfer into the wall rocks
2. The rate of magma flow

Near source, the magma still contains most of its thermal energy and flows readily; but "downstream" farther from the source an increasing amount of heat has been lost to the wall rocks, increasing viscosity and arresting flow. One way to model the distance of magma transport, d, is by multiplying the characteristic conductive cooling time by the magma flow velocity (equation 6.7 by 9.1)



9.8 Evolving magma injection paths during growth of the five basalt shield volcanoes that the exposed part of the island of Hawaii comprises. (a) Schematic cross section of a Hawaiian basalt shield volcano built of porous lava flows and minor intercalated elastic deposits. Basalt magma rises in a subvertical central conduit through the uppermost mantle and lower crust. Some magma spreads laterally along subvertical extensional fissures in a **rift zone** and feeds a flank eruption. (b) Schematic growth of the five coalesced shield volcanoes from oldest to youngest (1 through 4) by rift-zone controlled eruptions (thick lines). After the composite Kohala and Hualalai shield had formed, the orientation of subsequent extensional rifts was governed by local rather than regional stresses. The flanks of the shield are gravitationally unstable and pull away from the center, forming arcuate extensional fractures, like gigantic landslides. (From Shaw, HR. The fracture mechanisms of magma transport from the mantle to the surface. In: Hargraves RB, ed. Physics of magmatic processes. Princeton, NJ: Princeton University Press, 1980:201–264. Copyright © 1980 by Princeton University Press. Reprinted by permission of Princeton University Press.)

$$9.2 \quad d = \left(\frac{w^2}{\kappa}\right)\left(\frac{w^2 g\Delta\rho}{3\eta}\right) = \frac{w^4 g\Delta\rho}{\kappa 3\eta}$$

Therefore, the distance of magma transport is sensitive to dike width, w (to the fourth power), and viscosity, η. Therefore, in terms of heat loss and solidification, doubling the thickness of a dike is equivalent to moving it 16 times closer to the source of the magma, all other factors being the same (Delaney and Pollard, 1982). Because magma viscosities range over many orders of magnitude, this factor also strongly influences the effectiveness of dike transport of magma, as higher viscosities reduce transport distance. But increasing viscosity can be compensated for by widening the dike. Wada (1995) found that the thickness of 44 dikes in Japan and Peru correlates with apparent viscosity (calculated from magma composition), but not exactly as predicted by equation 9.2. Mafic magmas whose viscosities were 10–100 Pa s formed dikes 1 m thick, whereas felsic magmas whose calculated apparent viscosities were 10^6–10^7 Pa s formed dikes 100 m thick. Crystal-rich, water-poor granitic magmas, which have viscosities several orders of magnitude greater than 10^6–10^7 Pa s, probably move distances measured in kilometers only if the conduit is several kilometers wide. Whether such magma intrusions can be called dikes at all is debatable, as this is the probable order of magnitude of diapir diameters (discussed later). Silicic dikes less than a meter or so in thickness are uncommon, except in the immediate vicinity of larger granite plutons. Granitic counterparts of the huge basalt dike swarms (Figure 9.7) do not exist.

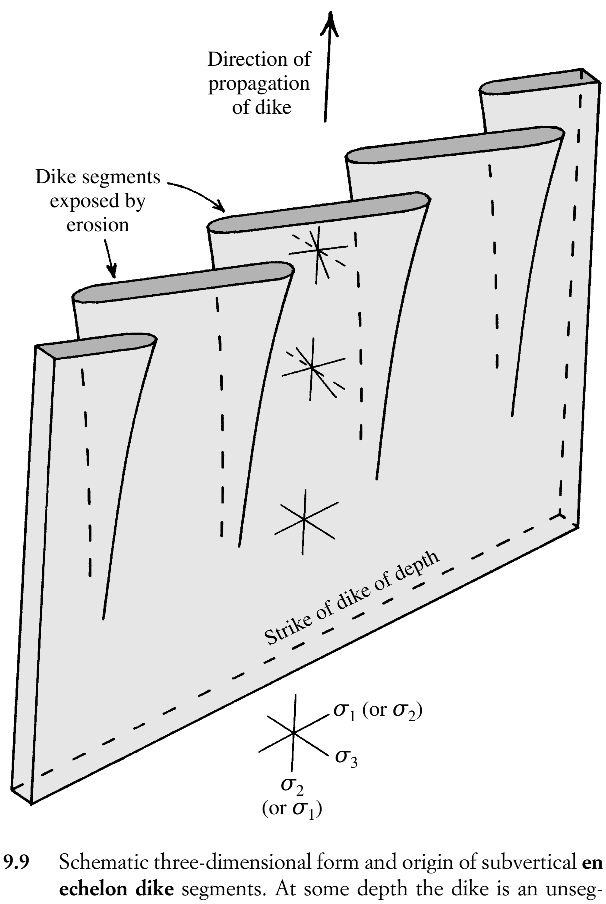
Aplite dikes (Figures 7.48 and 9.3) are virtually ubiquitous within, or are closely associated with, more mafic granitic plutons and deserve special mention. Compositional and textural observations indicate that these fine-grained phaneritic and leucocratic dikes originate from minimum T (Figures 5.24–5.27) residual melts sucked into self-generated extensional fractures in the cooling and contracting, mostly crystalline host magma body. Although the dike walls are planar and the contact apparently sharply defined between the aplite and the coarser-grained, less evolved host, close examination reveals interlocking unfractured crystals across the textural-compositional contact (Hibbard and Watters, 1985). Hence, the dike host rock behaved in a brittle manner (Section 8.2.2) so it could fracture yet was actually a crystal-rich mush with a small percentage of interstitial residual melt. In terms of equation 9.2, the typical aplite dike thickness of a few centimeters is possible because the residual, virtually crystal-free melts are enriched in water, so their viscosities are not excessive, but also because the melts probably migrated through the dike no more than several meters.

9.2.3 Geometry and Orientation of Sheet Intrusions

The orientation of self-induced, sheetlike magmatic pathways through the crust is governed by the state of stress. According to nature's least work principle, hydraulic tensile fracturing by overpressured magma creates extensional fractures parallel to σ_1 and σ_2 (Figure 8.2c), which open in the direction of the least compressive principal stress, σ_3. These magma-filled dilatant cracks, therefore, serve as paleostress indicators. Magma transported upward from deep sources, such as mantle-derived basalt magma intruded into near-vertical dikes in the crust, implies that σ_3 is horizontal, the typical stress orientation above mantle plumes and in other regimes of tectonic extension. Swarms of subparallel, subvertical dikes exposed over large areas indicate a uniform regional state of extension in the crust at the time of intrusion.

Dikes commonly occur in segments that may be arrayed *en echelon,* one explanation for which is a shift in the orientation of σ_3 with depth (Figure 9.9).

Sheet intrusions are common in shallow crustal, subvolcanic environments, where they surround and overlie a more massive **central intrusion** (Figure 9.6a). These more or less upright, bottle-shaped or cylindrical central intrusions can perturb a uniform *regional*



9.9 Schematic three-dimensional form and origin of subvertical **en echelon dike** segments. At some depth the dike is an unsegmented sheet intrusion whose orientation is controlled by the nonhydrostatic state of stress indicated by the thin-line orthogonal principal stresses below the dike. At progressively shallower depths the orientation of the least principal horizontal stress, σ_3, rotates progressively counterclockwise, as shown in the upper part of the dikes. Consequently, the least-work dike configuration there is an *en echelon* system of dike segments. Note that the other horizontal principal stress must also rotate. (Redrawn from Delaney and Pollard, 1981.)

state of stress, which is dictated by tectonic environment, because of their buoyant magma pressure, compounded by thermal expansion of the wall rocks. The resulting superposed *local* state of stress is spatially variable in orientation around the intrusion. Sheet intrusions created in this perturbed stress regime by magma supplied from the central intrusion include radial dikes, cone sheets, and ring dikes.

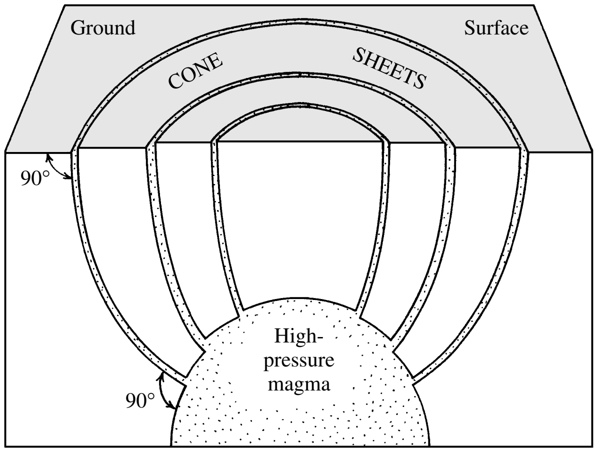
Formation of Radial Dikes and Cone Sheets. The local state of stress laterally *around* a central magma intrusion differs from the state *above* it. Around an intrusion swelling by magma overpressure, wall rock is compressed so that the maximum horizontal compressive stress, σ_1, is oriented perpendicular to the wall rock-magma contact. The least compressive stress, σ_3, which can be tensile for high magma pressures, is oriented horizontally and tangentially to the contact in the stretched wall rock. Trajectories of planes parallel to σ_1 and σ_2 and perpendicular to σ_3 are accordingly arrayed vertically and radially around the subvertical central intrusion (Figure 9.6b) so that magma-filled extensional fractures form a **radial dike swarm** centered at the central source intrusion. Farther from the influence of the central intrusion, stress trajectories assume a regional orientation and, if this far-field state of stress is uniform, dikes become subparallel.

Above the central intrusion, trajectories of curved surfaces representing planes parallel to σ_1 and σ_2 define the orientation of potential concentric conical extensional fractures perpendicular to σ_3. Magma driven along these conical fractures from the apex of the central intrusion forms one or more commonly concentrically nested **cone sheets** (Figure 9.10). Because of their inward dip, the magma intruded into a cone sheet elevates the segment of roof rock inside the cone. Considerable buoyancy-related magma pressure is required to accomplish this work against gravity.

The orientation of radial dikes and cone sheets is governed by two principles enunciated by Anderson (1951):

1. Both the surface of the Earth and the country rock–magma contact are solid-fluid interfaces that cannot support any shear stress; they are, therefore, principal planes, and local principal stresses must be normal to them (Figure 9.10).
2. Farther away from these two interfaces, principal stresses must bend to conform to the regional state of stress.

These geometrical constraints can be seen in Figure 9.10 and especially Figure 9.6b. The geometry of cone sheets differs from that of radial dikes because the magma pressure pushes upward against the free surface of the Earth above a central intrusion but sideways against confined rock around the intrusion.



9.10 Idealized geometry of **cone sheets** above a shallow, forcefully intruded **central magma intrusion.** Inward-dipping cone sheets develop as high-pressure magma in the central intrusion invades conical extensional fractures that follow $\sigma_1 = \sigma_2$ trajectories above the apex of the intrusion. This geometry applies to situations in which the depth to the top of the intrusion is comparable to its width; in such cases, some of the magma commonly extrudes and the intrusive complex is referred to as *subvolcanic.* Many intrusions are emplaced farther beneath the surface and for this reason and other factors do not have cone sheets.

The fact that many central intrusions never create radial dikes or cone sheets must reflect something of the particular nature of the roof rock and magma overpressure in intrusions.

Effects of Topography on Dike Configuration. Regional states of stress can also be perturbed in topographically high land masses, such as large volcanoes. So-called gravitational stresses in the huge shield volcanoes that form the island of Hawaii create a reorientation of intravolcano stresses relative to regional intraplate stresses. Extensional fissures (called *rifts* by Hawaiian geologists) through which magma rises thus become reoriented during island growth (Figure 9.8).

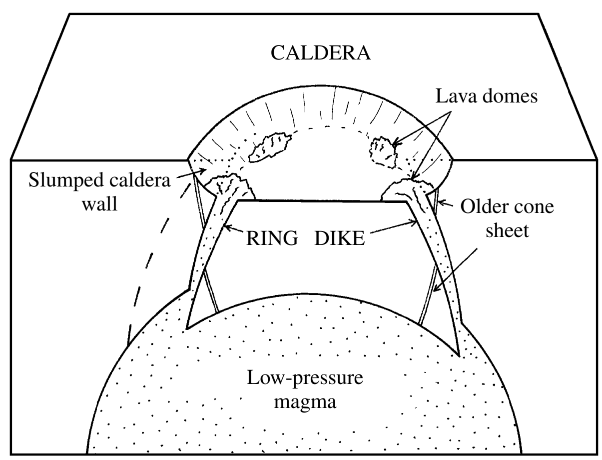
Ring Dikes. During the history of a magma intrusion the pressure in the chamber is likely to vary. After an initial overpressured state sufficient to cause magma ascent and possibly produce radial dikes and cone sheets, pressure may decrease. In evolving magma chambers this may be the result of release of exsolved volatiles or simply contraction during cooling and crystallization. Partial intrusion into nearby country rock or extrusion of magma can create potential voids in the partially evacuated chamber. With loss of supporting magma pressure, the roof over the partially evacuated chamber can subside, creating a topographic depression known as a **caldera.** (This is sometimes called a **cauldron** and the foundering process **cauldron subsidence.**) The **ring fault** (in some places a ring-fault zone) bounding the subsiding roof slab more or less follows the outline of the chamber. Magma can well up between the subsiding roof block and the undisturbed country rock, forming a **ring dike** (Figure 9.11). This is an arcuate, subvertical sheet intrusion that, in some cases, may form a complete 360° circular structure. In order for the denser roof rock to subside intact into the less dense, low-pressure magma, its margin must be vertical or dip outward. (Inward dips have been noted in some ring dikes, but such dips require special conditions for roof subsidence to compensate for diminished downward diameter of space into which the roof block can subside.)

More deeply eroded calderas expose increasing proportions of plutonic rock, including the ring dike, relative to extruded volcanic rock (see Figure 13.36).

9.2.4 Basalt Diking in Extensional Regimes

As a broad generalization, extruded basalt magmas are widespread in extensional tectonic regimes, whereas andesitic, or at least intermediate-composition calc-alkaline, extrusions predominate in compressional regimes. In the Basin and Range province of western North America, middle Tertiary andesitic volcanism was gradually supplanted by late Cenozoic basaltic activity as the tectonic regime changed from compressional to extensional. Magma extrusions are discussed in the next chapter, but the way basalt magmas ascend to the surface is relevant here.

Extensional and compressional tectonic regimes have fundamentally different states of stress. In extensional regimes, the lesser two principal stresses, σ_2 and σ_3, are horizontal and the greatest, σ_1, is vertical. In compressional regimes, the greater two principal



9.11 Hypothetical **ring dike** and **caldera** above a shallow magma chamber. Postcollapse caldera fill consists of landslide debris that is produced by slumping of the unstable caldera wall and epiclastic (sedimentary) deposits shed off the eroding caldera wall. Magma rising in the ring dike locally extrudes to form lava domes or flows; their vents mark the position of the usually concealed underlying ring dike and the ring fault it followed.

stresses, σ_1 and σ_2, are horizontal and the least, σ_3, is vertical. These contrasting states have different consequences for magma ascent and evolution in the lithosphere, as shown in Figure 9.12. In this figure, the vertical principal stress, σ_v either σ_1 (in extensional regimes) or σ_3 (compressional), equals $\rho_r gz = P$ where the density of rock, ρ_r, is assumed to be constant with depth.

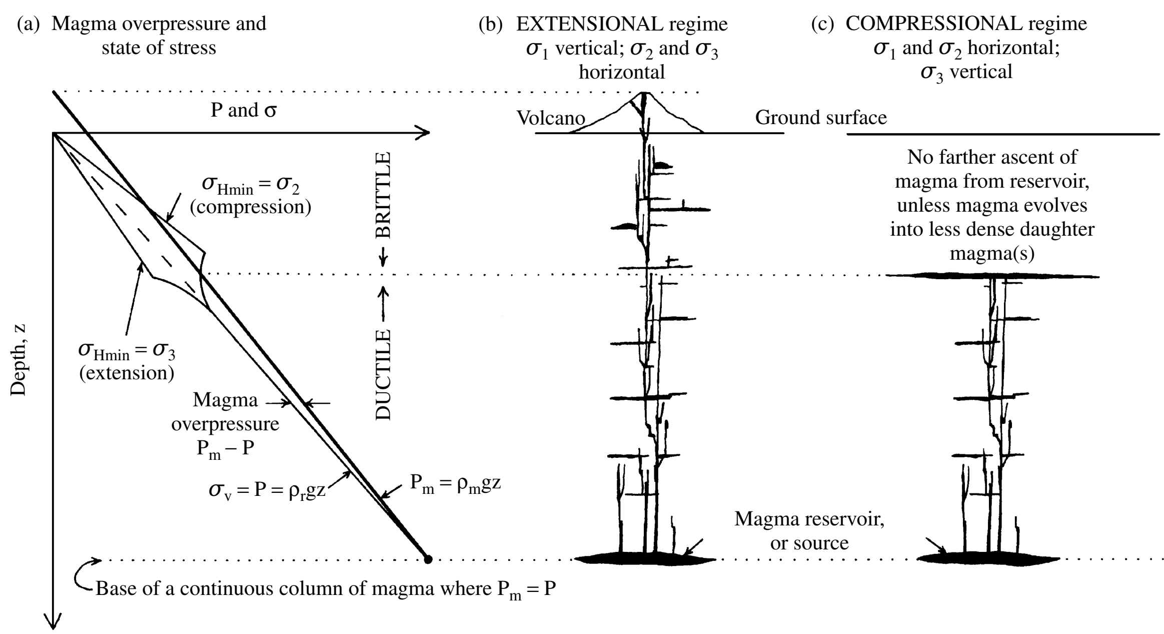
In a compressional regime, where σ_3 is vertical, the *minimum* horizontal principal stress, σ_{Hmin}, is σ_2. Its magnitude is fixed by the brittle and ductile rock strengths, as in Figure 8.9. In the lower part of the brittle crust in Figure 9.12a, the straight heavy line representing the magma pressure (compare Figure 9.2), *Pm,* is less than σ_2. Therefore, magma cannot invade this part of the crust by vertical diking and ascent is prevented. Magma can, however, spread horizontally in sills, lifting or inflating the overlying crust against σ_3 (Figure 9.12c). Sills of mantle-derived basalt magma can evolve into less dense differentiates and these evolved magmas can then ascend. Although vertical diking would be precluded in this depth interval for the relative values of σ_{Hmin} and P_m in the diagram, variations in rock strength, geothermal gradient, and magma density might locally permit vertical diking in compressional settings, as actually observed in some instances.

In contrast, in an extensional regime, with σ_1 vertical, the *minimum* horizontal principal stress, σ_{Hmin}, is σ_3. Its magnitude, fixed by the brittle and ductile rock strengths, is less than σ_v and much less than the magma pressure. Therefore, magma can vertically dike through this part of the crust (Figure 9.12b). The magma overpressure is sufficient to create extensional fractures, independently of any additional contributing motive force, such as might be provided by expanding exsolved volatiles, and regardless of how much low-density material, such as alluvium, might exist at the surface. These deductions agree with geologic observations.

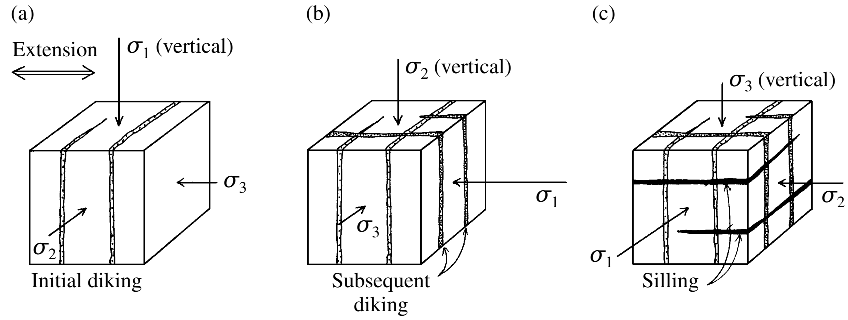
In the ductile regime in Figure 9.12a, sustainable stress differences, that is, rock strengths, are small and the state of stress is essentially hydrostatic; thus, the two horizontal principal stresses equal the vertical, σ_v and $\sigma_1 = \sigma_2 = \sigma_3 = \sigma_v = P = \rho_r gz$. Fracturing can occur accompanied by vertical magma ascent.

Horizontal sills of basalt are commonly intermingled with essentially contemporaneous vertical dikes that must have served as magma feeders. Such coexisting dikes and sills would seem to defy the concept that state of stress dictates the orientation of sheet intrusions, as just discussed. However, prolonged vertical diking and magma inflation in vertical dikes can produce an interchange or switching of local principal stresses (Parsons et al., 1992), which allows emplacement of magma in horizontal sills (Figure 9.13). Alternatively, vertical dikes passing through layers of weak rock, such as shale sandwiched between stronger sandstone, can balloon into them, forming sills.

State of Stress and Petrotectonic Association. In extensional stress regimes of continental rifts, basalt is one of the most common rock types, and in oceanic



9.12 Schematic relations between static magma pressure and state of stress in the lithosphere that govern magma ascent and stagnation. See also Marrett and Emerman (1992). It is assumed for simplicity that there is no reduction in magma pressure due to viscous loss that occurs during upward flow in a dynamic system and that the tensile strength of rock is nil.



9.13 Intrusion of horizontal sills in an extensional tectonic setting after vertical diking. (a) Initial intrusions are vertical dikes perpendicular to horizontal least compressive principal stress, σ_3. Wedging of magma reinforced by thermal expansion of heated wall rocks increases the stress perpendicular to dikes so that σ_3 becomes σ_1 in (b). Relative magnitude of other two principal stresses remains the same: Vertical σ_1 becomes σ_2 and horizontal σ_2 becomes σ_3. In this new state of stress, additional magma is emplaced in vertical dikes perpendicular to initial ones. After this subsequent diking, magnitudes of principal stresses are again interchanged to yield a third state of stress in (c) where σ_3 is now vertical, allowing horizontal sills to develop as more magma is introduced. The sills lift the overlying crust against gravity. (Redrawn from McCarthy and Thompson, 1988.)

rifts basalt is virtually the only rock type. On the other hand, along continental margins subject to compressional states of stress at convergent plate junctures basalt is commonly subordinate to andesite and more silicic rock types, especially where the crust is thicker. Although reasons for this contrast will be considered further in later chapters, it can be noted here that in compressional regimes (Figure 9.12c) basalt magma that is stalled in the lower to middle crust has ample opportunity to diversify into more evolved, more silicic magmas, including andesite. Basalt magma can crystallize and yield more evolved residual melts, it can partially melt surrounding country rock as a result of the transferred heat and create silicic magmas from that rock, and it can assimilate this rock or mix with its partial melts. These diversification processes are less common, but by no means absent, in extensional settings, where the basalt magma tends to be erupted rather than stagnating in the crust.

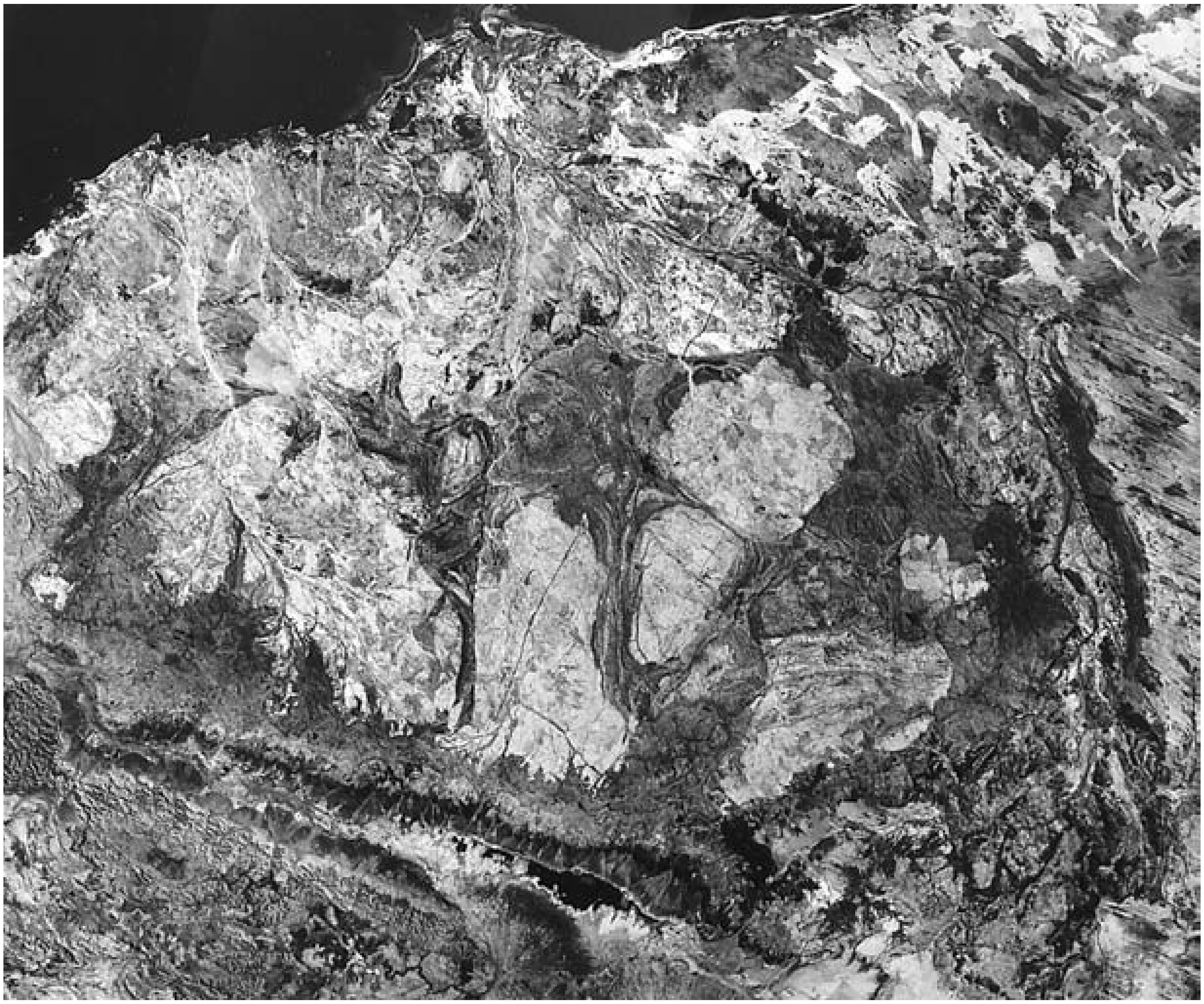
Hence, states of stress related to plate tectonic motion have a profound impact on the associated types of magmatic rocks. This is one facet of the concept of **petrotectonic associations.**

9.3 DIAPIRS

Diapirs and plumes are terms used for bodies of buoyantly rising material. Mantle plumes are long-lived columns of ascending less-dense mantle *rock.* Diapir is used for columns of rock salt (also called *salt domes*) in sedimentary basins and for bodies of magma rising in the lithosphere. Diapirs of felsic magma ascending in the continental crust are emphasized here. Diapirs have been the subject of numerous laboratory model studies (e.g., Ramberg, 1981) and theoretical and numerical analyses. Field relations of some granitic intrusions are believed to be compatible with emplacement as diapirs (Figure 9.14).

Any layer of less dense material overlain by denser material is gravitationally unstable. The upper boundary of the less-dense unstable layer develops sinusoidal bulges, known as *Rayleigh-Taylor instabilities,* which grow until the density inversion is stabilized in some way (Figure 9.15). As the bulges extend upward, smaller ones may die whereas larger ones grow and separate from the "mother" layer, forming diapirs that continue buoying toward the surface. The rise of thunderclouds from heated near-surface air on hot summer days is an example of this buoyant instability. The wavelength of the bulges in the low-density layer depends upon the thickness of the layer and the viscosity and density contrasts with the overlying denser layer. Thus, diapir spacing provides insight into these parameters (Lister and Kerr, 1989).

The velocity of ascent and diapir longevity are complex functions of many parameters, not the least of which are diapir shape and size. A magma diapir rising through ductile crust must have a relatively large ratio of volume to surface area so that the buoyant body force—a function of its volume—is maximized, whereas the resistive drag force—a function of its surface area—is minimized. Therefore, the "ideal" diapir shape is a perfect sphere, which also happens to be the most thermally retentive for conductive heat loss. But a sphere is only approximated in nature, because, among other possible factors, the drag during ascent maintains a tail in the wake of the rising sphere. The relative magnitudes of the driving and resistive forces determine the ascent velocity of the diapir, as for solid particles in a viscous material (Stokes's law, equation 8.8). Thus, doubling the mean diapir diameter can increase ascent velocity by a factor of 4 if other factors remain constant. Resistive drag also depends on the rheology and thickness of a boundary layer of thermally perturbed country rock adjacent to the ascending diapir (Marsh, 1982). More heat transferred from the magma into the



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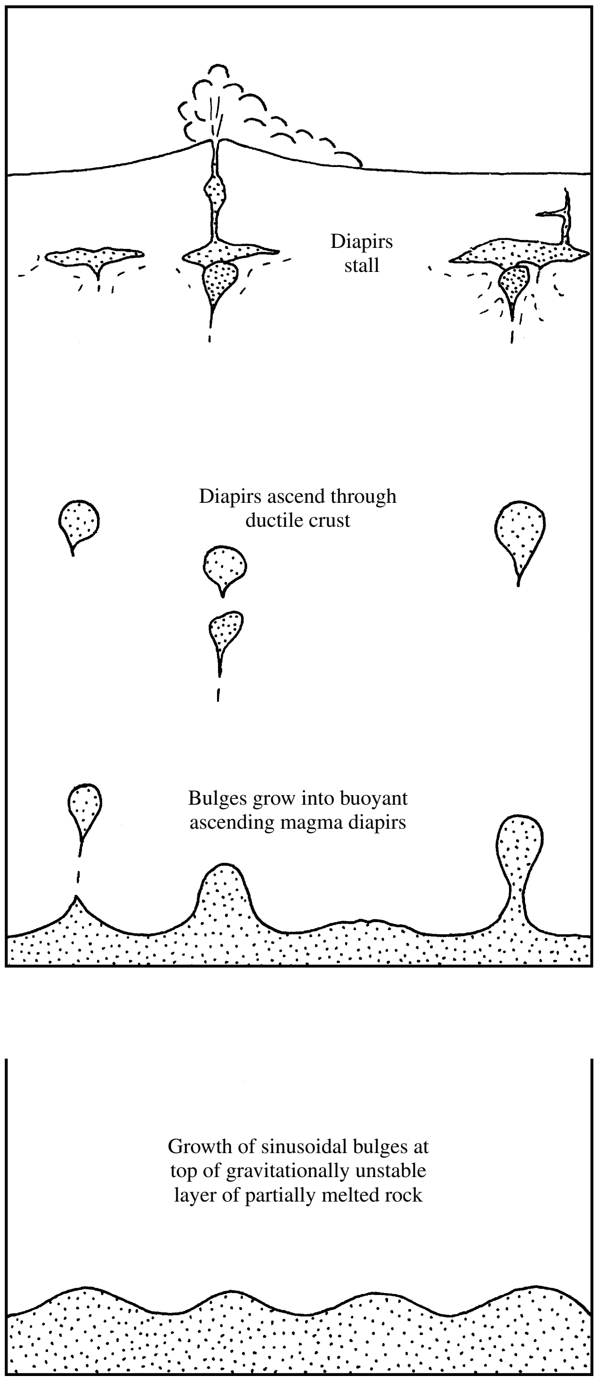
9.14 Elliptical plutons (diapirs?) in the Archean Pilbara craton, Western Australia. Light colored 3.4 and 3.0 Ga granitic rock is surrounded by darker colored 3.5 to 3.0 Ga metamorphic greenstone belts. Overlying these rocks in marked angular discordance is a gently dipping cover sequence deposited 2.7 to 2.4 Ga visible on the east (right). Indian Ocean is to the north. Area is about 400 km wide. Image furnished through the courtesy of Clive A. Boulter, University of Southampton, UK, and provided by the Australian Centre for Remote Sensing (ACRES), AUSLIG, Canberra and SPOT Imagine Services, Sydney and digitally enhanced and produced by Satellite Remote Sensing Services, Department of Land Administration, Perth, Western Australia Copy Licence 629/2000.

boundary layer can reduce its ductile strength and possibly induce partially melting, allowing the diapir to slip through the country rock with greater ease. But the amount of available thermal energy in the diapir is finite. Small-volume diapirs and those with larger surface area would be expected to stall sooner than larger subspherical ones. Even large diapirs may stall at a crustal level where the ductile strength increases exponentially (Figures 8.8 and 8.9). In the overlying stronger brittle layer, other mechanisms of movement and emplacement of large bulbous masses of magma must come into play, as discussed later.

Because of the transferred heat into country rocks, an ascending diapir leaves in its wake higher-T, softened country rock. Subsequent diapirs can rise significantly faster and farther in this thermally perturbed, preconditioned column of rock. This may explain why major centers of magmatism commonly have lifetimes of several million years. As one ascending mass of magma is intruded and stalls, another follows in its wake.

Some geologists doubt that diapirs of highly viscous felsic magma are thermally and physically capable of ascending very far in the continental crust.



9.15 Highly schematic diagram (not to scale) showing the growth, ascent, and stalling of buoyant magma **diapirs.** Beginning at bottom of diagram, a layer of partially melted rock (magma) in source region in upper mantle or lower crust of lesser density than the overlying rock develops sinusoidal Rayleigh-Taylor instabilities. In next higher frame of diagram, these bulges grow and separate from the source layer, forming inverted "teardrop"-shaped diapirs of magma that ascend through denser ductile country rock, as do hot-air balloons rising into the atmosphere. Eventually (top of diagram), magma diapirs stall at a density barrier or where they encounter stronger brittle rock. Subsequent diapirs may follow in the wake of earlier ones. Some magma may erupt.

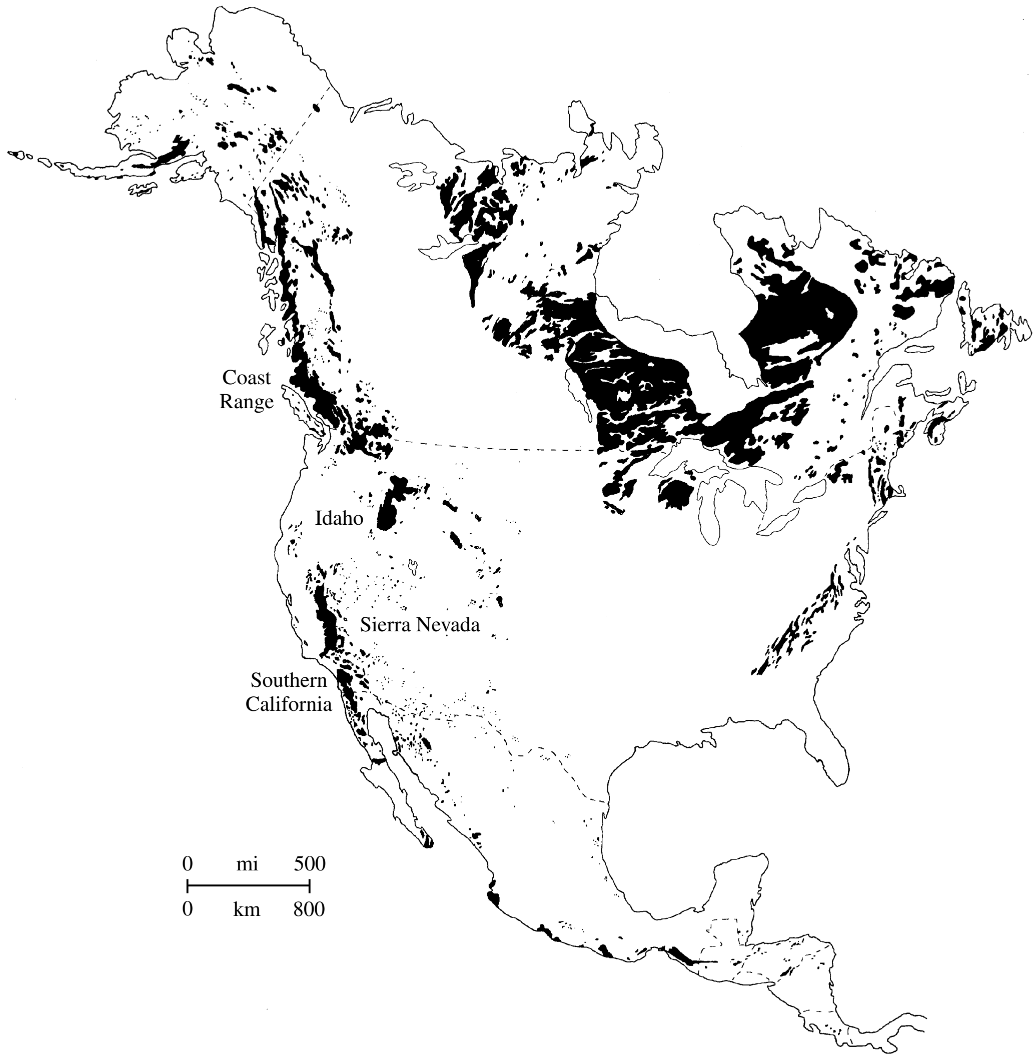
Doubts stem, at least in part, from the theoretical models, which depend strongly on parameters chosen and simplified boundary conditions assumed. Further doubt stems from the lack of unequivocal field evidence for vertical movement. Except in unusual instances, such as where a major sector of the crust has been tilted substantially after magma emplacement, the geologist can see only a subhorizontal section of limited vertical extent through an intrusion and its wall rock. Complete exposures from the top of an alleged diapir to its tail are not seen. In these vertically limited exposures (2 km of relief is exceptional), the only evidence for diapirs lies in highly strained country rock immediately surrounding the magma intrusion, separating the generally less deformed rock within it from more distant country rock. In overly simplified terms, the question facing the geologist is whether the ductile deformational fabric in the country rock (Cruden, 1990) is the result of passage of an ascending diapir or of ballooning of a body of magma that ascended in some other manner, such as by unexposed dikes. However, the paucity of felsic dikes at all levels of the crust but widespread occurrence of more equidimensional intrusions having small aspect ratios strengthens the case for felsic diapirs.

❊9.4 MAGMA EMPLACEMENT IN THE CRUST: PROVIDING THE SPACE

Once magma generated in deep sources has ascended to shallower depths, final emplacement occurs at a particular position within the lithosphere. The concern in this section is how space is provided for nonsheet intrusions, which are referred to as **plutons.**

On a lithospheric scale, mantle-derived magma can be emplaced into the crust by thickening it, displacing the Moho downward to replace the volume of melted mantle source rock and/or lifting the surface of the Earth. Room for magma generated in the lower crust and emplaced in the upper crust involves an exchange in position of material, rock for magma. Deep magma moves up and displaces shallow crustal rock. At an observed level of exposure, creation of the space that is now occupied by hundreds or tens of thousands of cubic kilometers of magmatic rock is a nontrivial **"room problem."** For example, the belts of batholiths that characterize orogenic subduction zones, such as the Cordillera of western North America (Figure 9.16), involve intrusion of a vast amount of granitic magma into the presently exposed crustal level, an estimated 10^6 km^3 in the case of the Mesozoic Sierra Nevada batholith of California (Paterson and Fowler, 1993). Intrusion into orogenic zones that are typically under a compressional state of stress only compounds the paradox.

However, the room problem diminishes if the entire intrusion is considered from a three-dimensional per-

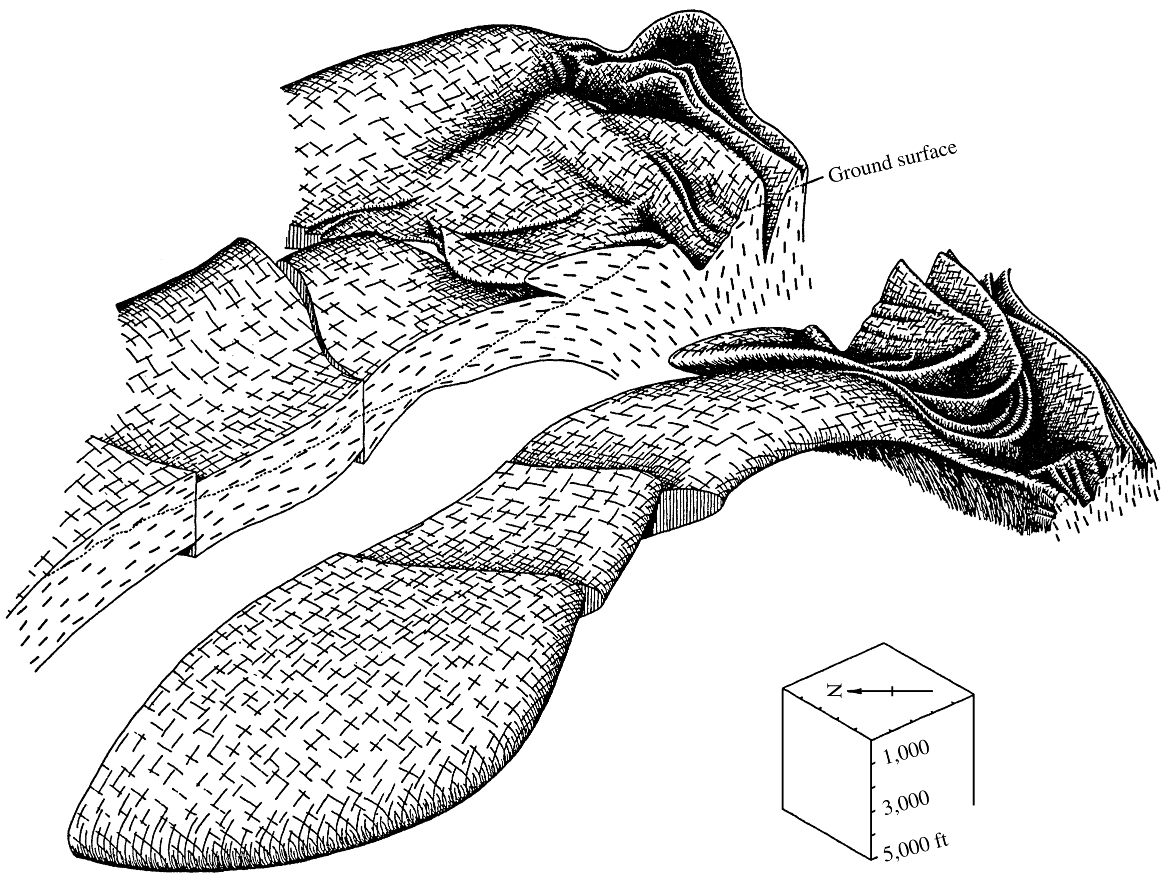


9.16 Distribution of exposed granitic rocks in North America. Granitic rocks in Precambrian Canadian shield are generalized and include some metamorphic rocks. Rocks in the Appalachian orogen along the U.S. East Coast are mostly Paleozoic. Labeled **batholiths** along the west coast in the Cordilleran orogen are mostly Mesozoic. (Redrawn from the Tectonic Map of North America, U.S. Geological Survey.)

spective. The view that plutons extend indefinitely downward with constant horizontal dimensions into the deeper crust, lacking a floor, is unquestionably flawed. Large areal extent at a particular level of erosion does not imply large vertical dimension (e.g., Figure 9.17). Perceived difficulties in accounting for the space occupied by a pluton at a particular level of exposure may be reduced if all three dimensions of it and the surrounding country rocks can be examined.

9.4.1 Some Aspects of Granitic Plutons

Careful studies reveal that very few, if any, plutons are truly homogeneous in composition. This internal inhomogeneity is expressed in composite and zoned plutons. **Composite intrusions** have compositionally and/or texturally distinguishable parts reflecting emplacement of two or more contrasting magmas. In many instances, an appreciable time elapses between successive intrusions, as indicated, for example, by a chilled, finer-grained contact of the later intrusion against the earlier colder one. In other cases, subsequent magma may have been intruded before the first intrusion one cooled very far below its solidus *T,* so that their contact shows less evidence of a thermal contrast. **Zoned intrusions** have more or less concentrically arrayed parts of contrasting composition. In normally zoned plutons, more or less concentric parts are successively less mafic inward (Figures 9.18 and 9.19). Normal zoning might develop, for example, as a diorite magma diapir stalls at a particular level in the crust and, in its thermal wake, slower-moving, more silicic and viscous granodiorite and then granite magma diapirs are intruded, inflating the diorite envelope. These successive surges create gradational contacts due to in-



9.17 Three-dimensional projection of the Rattlesnake Mountain granitic pluton, California. The projection is split and separated to show internal structure of the pluton and position of ground surface. The wrinkled parts of the intrusion are screens of hornblende-quartz diorite embedded in more felsic rock. (From MacColl RS. Geochemical and structural studies in batholithic rocks of southern California. Part I. Structural geology of the Rattlesnake Mountain pluton. Geol. Soc. Am. Bull. 1964;75:805–822. Reproduced with permission of the publisher, The Geological Society of America, Boulder, Colorado, USA. Copyright © 1964 Geological Society of America.)

teractions between the incompletely crystallized magmas. Normal zoning might alternatively develop by assimilation of mafic wall rock or by sidewall crystallization processes in a homogeneous magma in which higher-T mafic minerals crystallized preferentially near the cool wall rock margin, allowing the felsic constituents in the magma to concentrate upward and in some manner inward. In some cases, a reverse zoning is evident in more mafic pluton interiors.

A **batholith** is a pluton or commonly groups of separately intruded plutons exposed over generally tens of thousands of kilometers (Figure 9.16). The composite Sierra Nevada batholith in California consists of hundreds of intrusions emplaced over about 130 million years (late Triassic to late Cretaceous) that range in composition from gabbro to granite (mostly granodiorite). Individual plutons crop out over areas ranging from <1 km^2 to $>10^3$ km^2; their average volume is about 30 km^3.

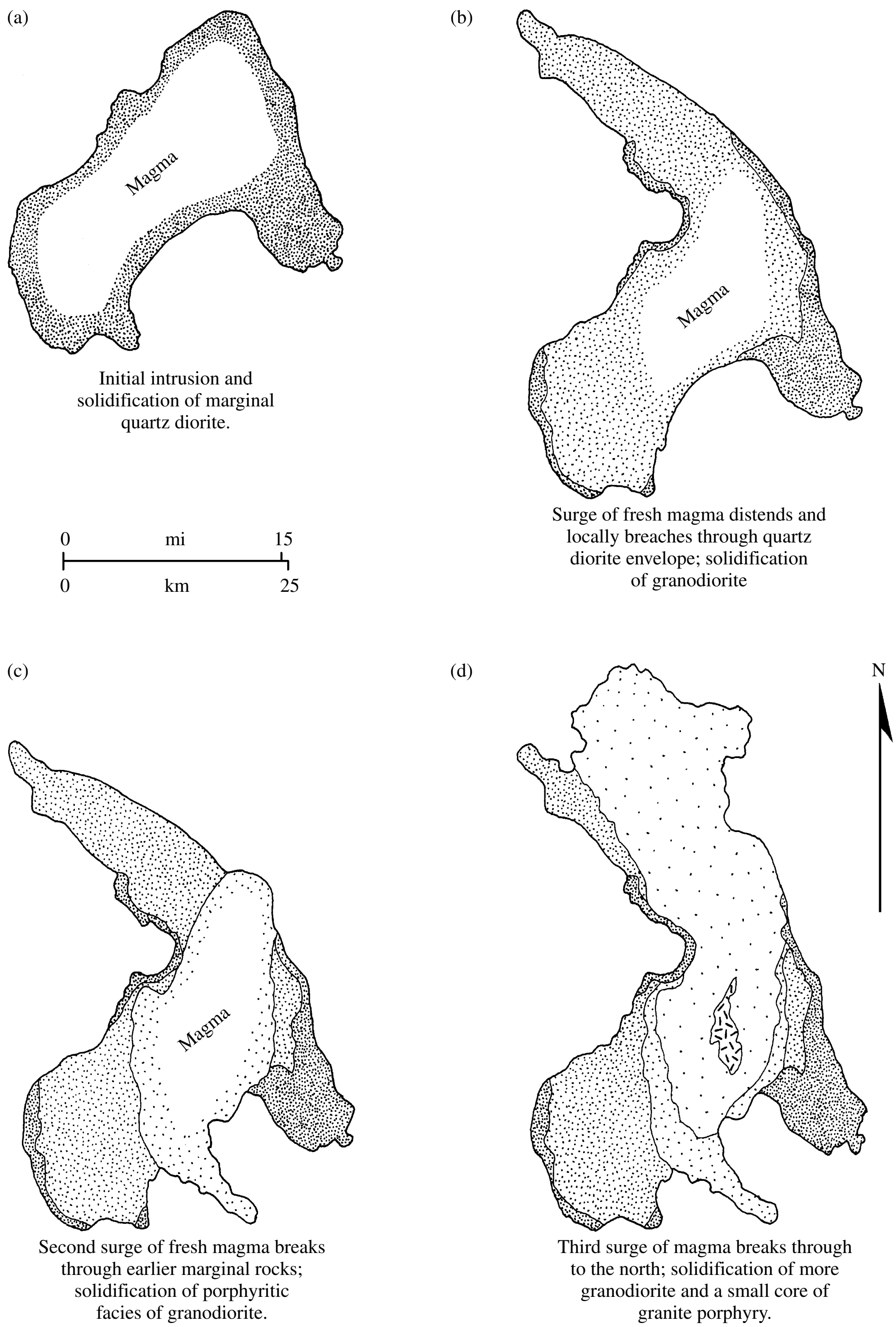
Plutons smaller than batholiths, commonly consisting of only a single intrusion, are called **stocks;** their outcrop area is generally <100 km^2. Features related to stocks and batholiths are shown in Figure 9.20.

9.4.2 Emplacement Processes and Factors

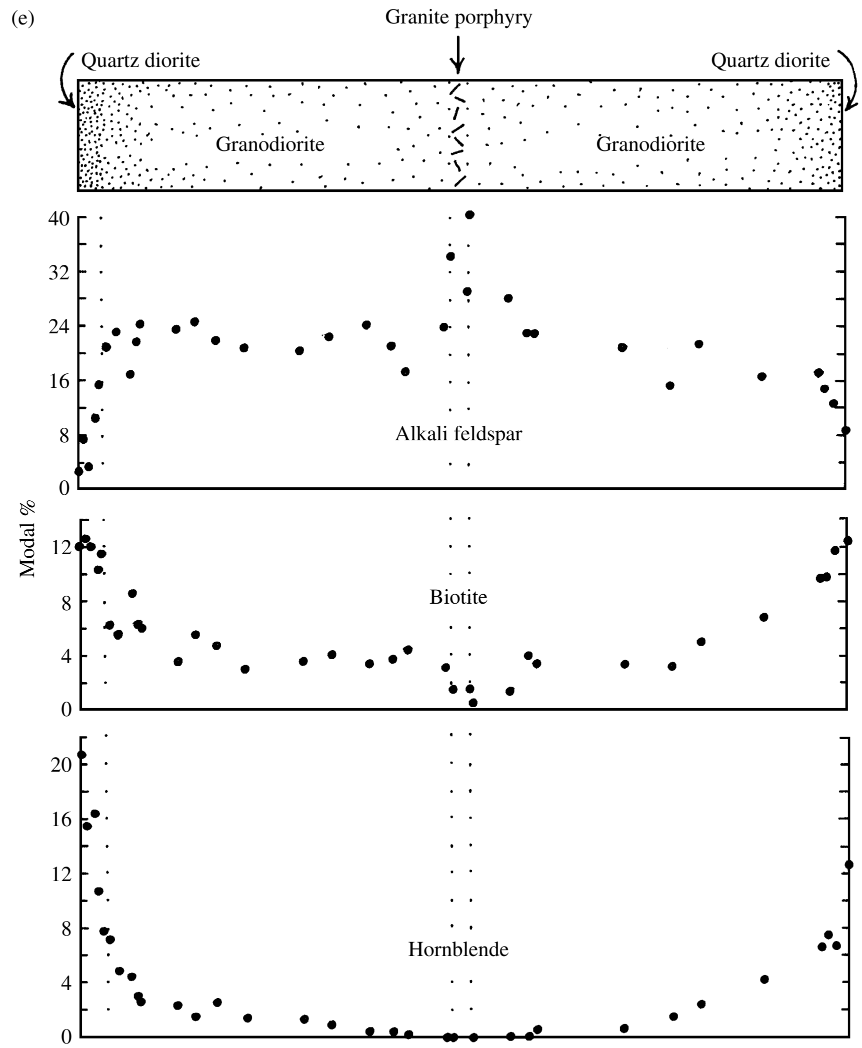
This section deals chiefly with emplacement of felsic plutons. Mafic magmas are mostly emplaced in sheet intrusions.

Several emplacement processes have been identified. Some space is created in country rocks through dewatering, removal of chemical constituents by migrating solutions, and growth of denser minerals during metamorphism of country rocks around the intrusion. However, these volume-reduction processes probably only contribute a small fraction of the total space occupied by an intrusion. Other more significant mechanisms (Figure 9.21) include the following:

1. Stoping: pieces of country rock that are physically incorporated into the magma, these xenoliths may be chemically assimilated ("digested") to varying degrees as well



9.18 Evolution of the Tuolumne Intrusive Series, a compositionally **zoned pluton** within the Sierra Nevada batholith, California. Mantle-derived basalt magmas contaminated by increasing amounts of partial melts of the lower continental crust were intruded into the shallower crust over a time span of several million years (Kistler et al., 1986). See also Table 13.8. (a–d) Schematic sequence of events during growth of pluton. (e) Compositional variations along a west-east line across the pluton. (Redrawn from Bateman and Chappell, 1979.)

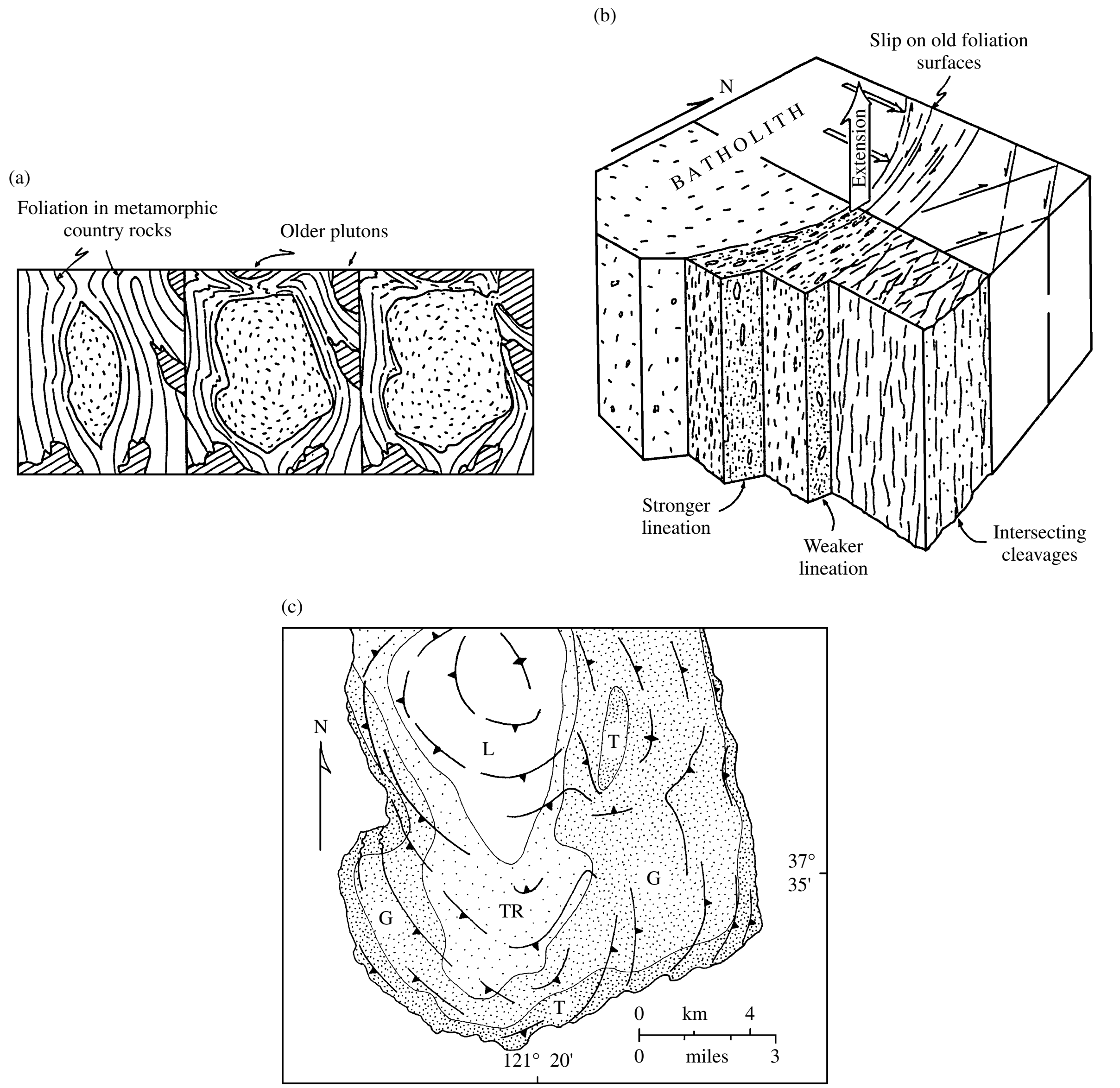


9.18 (*Continued*).

2. Brecciation and kindred phenomena involving expansion of a volatile fluid at generally low P
3. Doming: flexural and block-fault uplift of roof rocks
4. Ballooning: forcefully swelling magma that pushes aside ductile wall rocks and lifts roof rock
5. Magma invasion into tectonically favored "potential void" sites such as shear zones, fold hinges, and local extensional domains.

The emplacement history of a particular pluton has to be carefully evaluated on the basis of its field relations, fabric, and composition. A single pluton may be emplaced by more than one process, depending on the variable rheology of the country rocks and magma as its ascent slows and then stalls. Episodic emplacement and growth of a pluton might begin with brittle emplacement processes succeeded by more ductile ones as the country rocks are heated by the magma. Each new surge of magma can ascend higher, intruding former roof rocks (Figure 9.5), and in some cases intruding its own volcanic cover. Emplacement processes in the brittle upper crust differ from those in the weaker ductile lower crust.

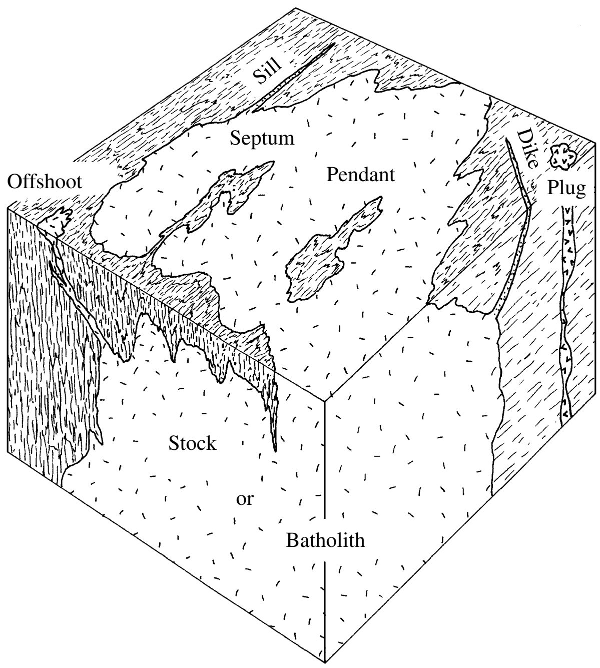
Stoping. The apt term *stoping* is derived from underground mining operations in which human-made caverns, called *stopes,* are created by removal of ore. As envisaged by R. A. Daly in the early 20th century (see Daly, 1968, Chap. 12) and by contemporary geologists, **stoping** is engulfment of pieces of brittle country rock by magma. In the usual case, fracture-bounded blocks of denser host rock, dislodged from the roof or wall, sink into the magma (Figures 9.5 and 9.21), allowing magma to move vertically upward or horizontally. Denser blocks of already fractured rocks can sink into the magma. In addition, new fractures in country rock can be created by the adjacent magma through thermal stressing of initially cool country rock (Marsh, 1982) and by hydraulic fracturing (Figure 8.2c). Magma can



9.19 Mostly **concordant** Bald Rock granitic pluton in the northern Sierra Nevada, California. Compton (1955) estimated that about one-fourth of its area at the exposed level was gained by stoping and assimilation and the remainder by **forceful emplacement.** Ductile metamorphic wall rocks were pushed aside and dragged upward during intrusion. (Redrawn after Compton, 1955.) (a) Three small-scale maps show hypothetical stages in progressive growth of pluton. (b) Accommodation of pluton by subhorizontal flattening combined with vertical stretching in the metamorphic wall rock that created a strong vertical stretching lineation and flattening foliation. (c) Generalized map of **zoned pluton** showing concordancy between internal foliation in pluton and its contact with the wall rock. Note, however, that magmatic foliation cuts across most internal compositional contacts in zoned pluton. L, leucotrondhjemite; TR, trondhjemite; G, granodiorite; T, tonalite. (Redrawn from Paterson and Vernon, 1995.)

insinuate into country rock surrounding the main body of magma, producing a network of dikes and sills (Figure 9.5). Magma-bounded chunks of the denser rock can fall into the main magma body. Exsolved magmatic fluids and heated groundwater in cracks and other pore spaces in country rock can expand and wedge blocks into the magma. Confined water in the upper few kilometers of the crust held at constant volume experiences an increase in pressure of 1.5 MPa for each degree Celsius increase. Stoping mechanisms are, thus, favored in shallower, cooler brittle crust.

Evidence for magma emplacement by stoping lies in the occurrence within the pluton of pieces of foreign country rock, or **xenoliths.** However, countless plutons neither have xenoliths nor offer evidence of alternate mechanisms of emplacement. For example, the strikingly discordant multiple intrusions at Mount Ascutney, Vermont (Figure 9.22), display no indication of emplacement by doming or ballooning. Thus it was

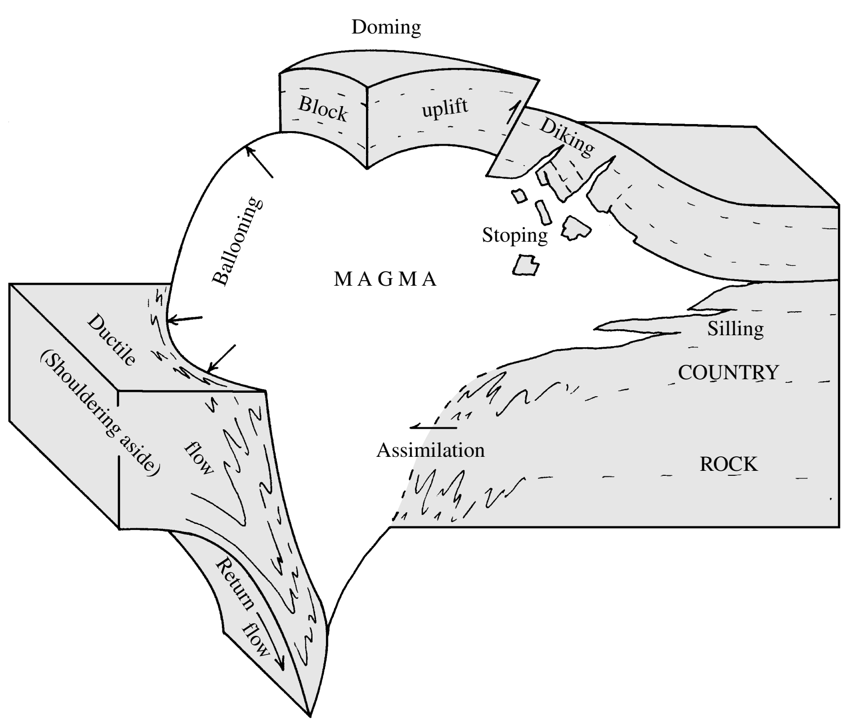


9.20 Nomenclature of some plutonic features. Offshoots, or **apophyses,** are any off-branching intrusions from the main pluton; they can be plugs or sheet intrusions such as dikes and sills. A **cupola** (not shown) is an upward-projecting part of the pluton into the roof rock. Erosional remnants of downward-projecting roof rocks completely surrounded by pluton are **roof pendants;** if they have peninsulalike connections to the main mass of roof rock, they are **septa,** or **screens.**

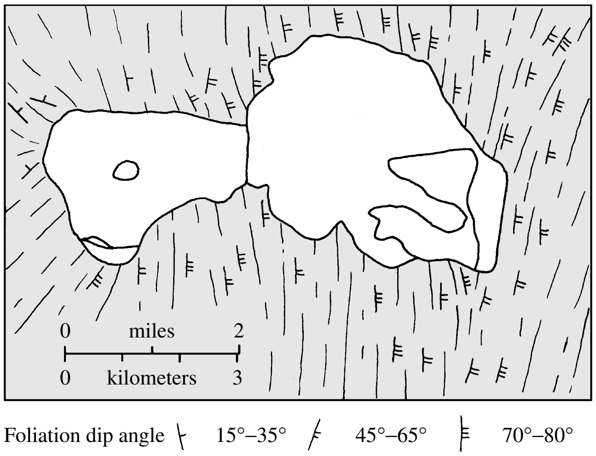
that Daly hypothesized the stoping mechanism: He believed the absence of xenoliths in these intrusions was due to their complete assimilation into the magma, leaving no trace of their former presence.

Stoping is an example of **passive emplacement** of magma. The intrusive rock tends to have an isotropic fabric that reflects dominance of crystallization over deformation, and the host rock is likewise little deformed.

Ring-Fracture Stoping. In contrast to the "piecemeal" stoping of outcrop-scale or smaller fragments of rock just described, much larger slabs of roof rock (measured in kilometers) may founder into a low-pressure magma chamber. An example is **ring-fracture stoping** (Daly, 1968), so called because the roof, initially supported by buoyant underlying magma, fails along a steeply dipping circular fault that follows the perimeter of the generally subcircular magma chamber (Figure 9.23). As the more or less intact slab sinks, less dense magma moves upward along the outward-dipping fault, invading the space between it and the wall rock, forming a **ring dike.** However, unlike in the situation shown in Figure 9.11, the ring fault and related dike do not reach all the way to the surface to create a caldera and magma extrusions. Rather, a fault-bounded roof slab detaches from higher roof rock, perhaps along a bedding plane or some other planar weakness. The resulting intrusion has the form of an inverted bowl in



9.21 Highly schematic perspective view of emplacement processes of a pluton in the crust. Note deeper, downward-directed, return flow and forcefully emplacing ballooning in more ductile rocks in contrast to shallower, brittle processes of stoping and fault-block uplift. (Redrawn from Paterson et al., 1991.)



9.22 Posttectonic plutons with highly **discordant,** sharp contacts against their foliated metamorphic host rock at Mount Ascutney, Vermont. Note the lack of deflection of the foliation in the country rock (shaded) adjacent to the six Cretaceous intrusions, as if a gigantic cookie-cutter had removed metamorphic rock and allowed magma to move into its place. Compare with the concordant plutons in Figures 9.19 and 9.28 emplaced in large part by forceful pushing aside of ductile wall rock. (Redrawn from Daly, 1968.)

three dimensions. Erosion just into the top of the inverted-bowl-like pluton gives the appearance of a discordant, "cookie-cutter" pluton. Only by deeper erosion are the foundered roof slab and surrounding ring dike revealed.

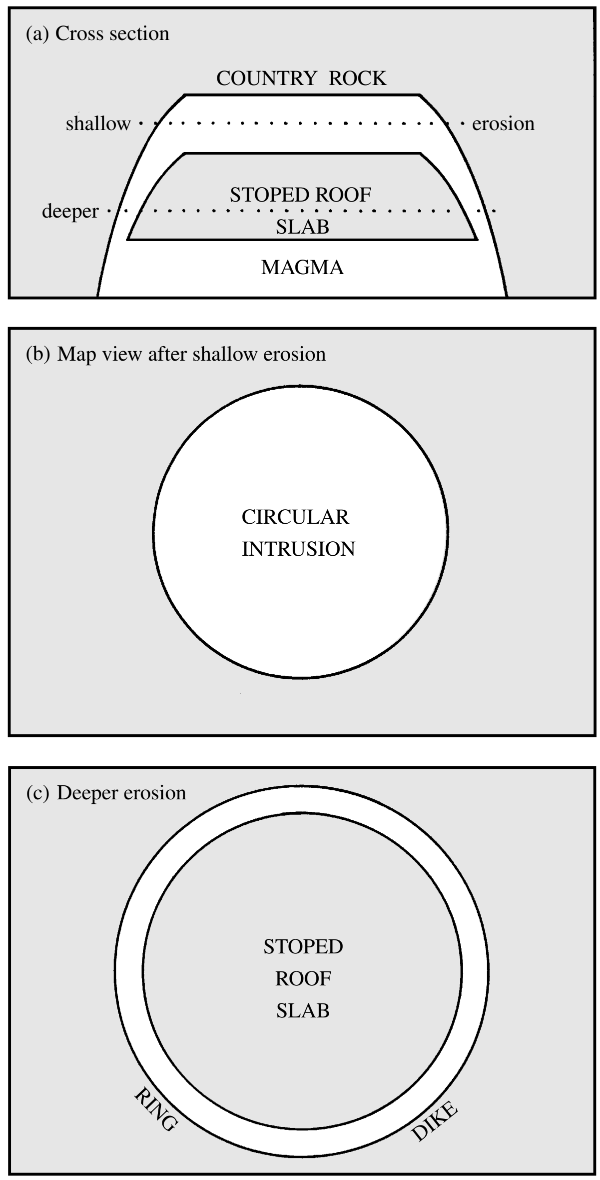
It is not unusual for multiple episodes of ring-fracture stoping to occur at nearly the same location, developing multiple ring dikes and circular intrusions (Figure 9.24).

Brecciation. Slender, subvertical columns or funnel-shaped bodies of intrusive magmatic rock, commonly emplaced into the brittle upper crust, are called **pipes,** or **plugs.** They have an elliptical or circular cross section whose diameter is generally <1 km. Some merge with dikes and sills laterally and/or at depth, and some have fed volcanic extrusions, in which case the term **volcanic neck** is used.

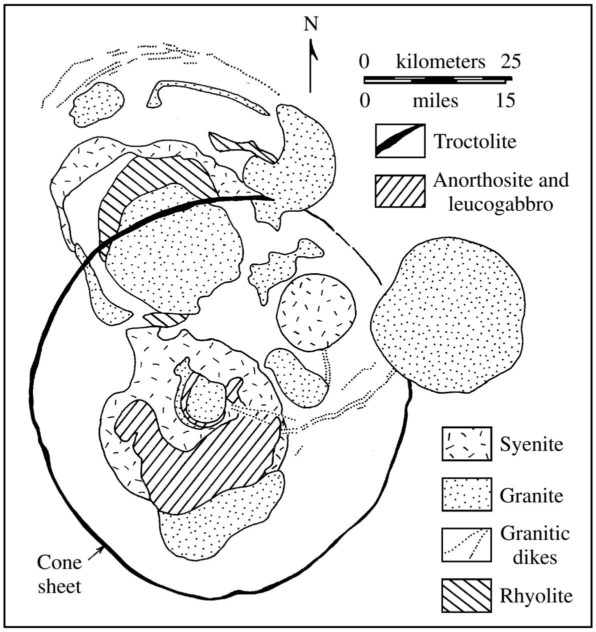
Delaney and Pollard (1981) compared three geometric and energetic constraints on the movement of magma through a dike versus a cylindrical pipe in a *brittle* host rock:

1. The conduit most likely to grow by extensional fracturing of the host rock
2. The conduit that accepts the greatest volume of magma for a given increase in magma pressure at the source
3. The conduit around which the least work is done on the wall rocks to accommodate a given volume of magma

For all three of these aspects, a dike is favored over a plug. So how do plugs develop? Significant observations of plugs include an absence of wall rock deformation, textures and mineral compositions indicating fluid over-saturation in the magma, and abundant xenoliths of wall rock in the plug; **breccia pipe** is a more apt label for a xenolith-rich intrusion. In some breccia pipes wall rock fragments from a higher stratigraphic level are concentrated along the margin, whereas fragments from greater depth prevail in the



9.23 Idealized **ring-fracture stoping.** (a) Vertical cross section of magmatic body showing foundered slab of roof rock. (b) Hypothetical geologic map of subhorizontal surface exposure after shallow depth of erosion reveals a "cookie-cutter" circular intrusion. (c) Hypothetical geologic map of subhorizontal surface exposure after deeper erosion reveals a **ring dike** surrounding the foundered slab of roof rock. Compare actual circular intrusions and ring dikes in Figure 9.24. (Redrawn from Hall, 1996.)

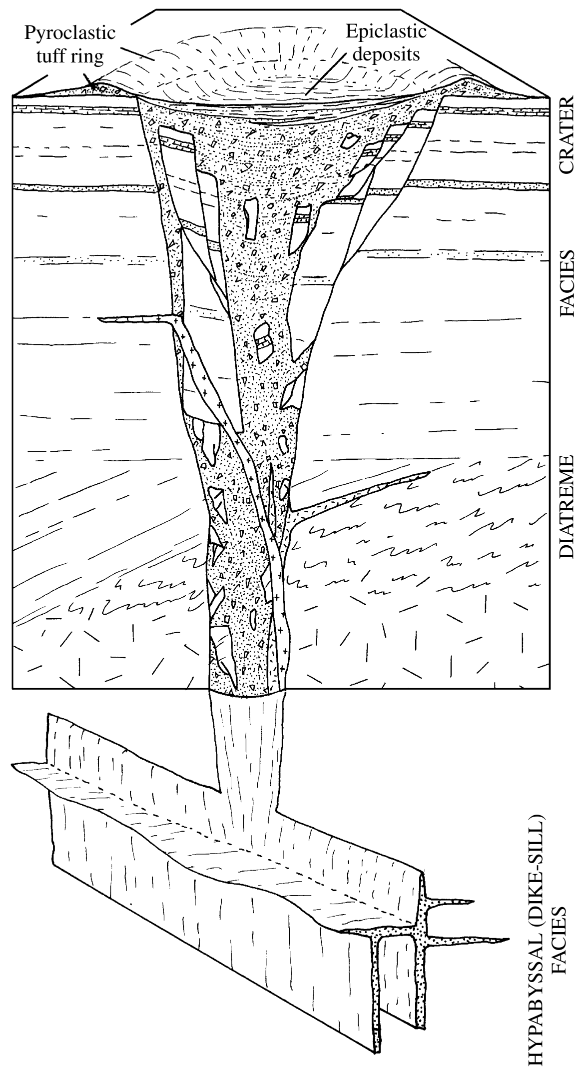


9.24 Ring complex and associated cone sheet in Niger, West Africa. Simplified geologic map of Silurian **ring complex** of circular intrusions and unusually large, nearly perfectly circular, inward-dipping Meugueur-Meugueur **cone sheet** of troctolite (olivine-plagioclase rock). Unpatterned area is chiefly older basement rocks. (Redrawn from Moreau et al., 1995.)

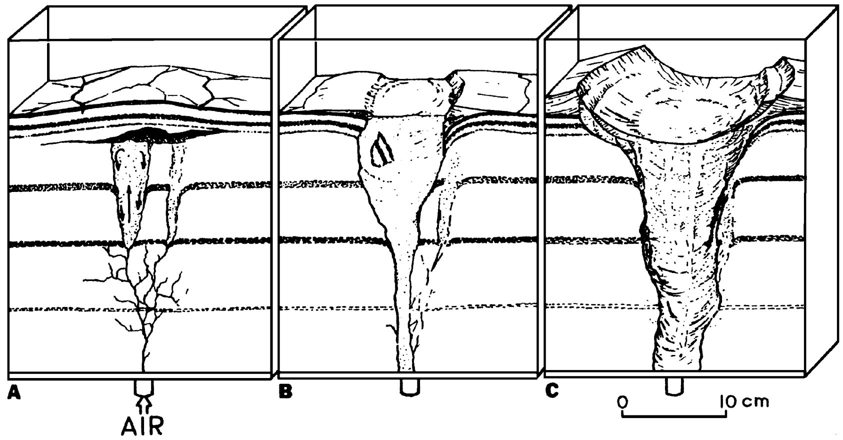
center (Williams and McBirney, 1979, p. 52). These observations suggest an upward-convecting column of fluid-oversaturated magma, at the top of which rock fragments were stoped from the roof and carried downward along the cooler margin of the column. The upward "drilling" process is accomplished by efficient, concerted hydraulic and thermoelastic-stress fracturing of roof rocks at the top of the magma column. Thus, a plug may begin as a magma-filled brittle crack, but as volatile saturation occurs the drilling action tends to create a more circular, thermally efficient conduit. Some breccia pipes may be "blind," having never erupted through the surface of the Earth.

Diatremes are funnel-shaped breccia pipes apparently emplaced at low temperature that contain a high concentration of mantle- and/or crust-derived fragments, even to the exclusion of a recognizable magmatic matrix. They are commonly overlain by a shallow, dish-shaped explosion crater called a **maar** that has an encircling low pyroclastic ring (Figure 9.25). Many diatremes originate from alkaline, mafic to ultramafic kimberlite magmas that are exceptionally enriched in H_2O and CO_2 so that they probably become oversaturated with CO_2 at subcrustal depths. Above a "root" zone of dikes and sills that contain unfragmented kimberlite the highly overpressured, fluid-charged magma creates its own fractures and ascends rapidly (10–30 m/s) and turbulently to the surface, carrying well-mixed dense mantle xenoliths and crustal fragments produced as the system drills upward through the lithosphere. The complexity of diatremes and their related crater and root zones makes generalizations difficult (for a description see Scott-Smith in Mitchell, 1996). Model experiments by Woolsey et al. (1975) have created many of the features of kimberlite diatremes (Figure 9.26).

Doming and Fault-Block Uplift. Clear evidence for overpressured magma that makes room for itself is seen



9.25 Idealized **diatreme** and overlying **maar.** The maar is a dishlike explosion crater surrounded by a ring of pyroclastic ejecta and is partly filled with lake sediments and alluvium (epiclastic deposits). Below this crater facies is the diatreme, here intruded by late dikes, which contains caved blocks of wall rock as much as 100 m long 1 km below their original stratigraphic level. Diagram not to scale; horizontal dimension (typically less than 1 km in diameter) exaggerated relative to vertical.



9.26 Diagrammatic summary of fluidization experiments bearing on the origin of a **diatreme.** (a) Introduced compressed air fractures cohesive clay layers and uplifts surface, forming ring and radial fractures in domed layers; fluidized convective cells develop in voids below uplifted layers, mimicking a blind diatreme. (b) Larger convective cell breaches surface; outcast ejecta accumulate in a ring around growing maar. Saucerlike stratification and wall-rock fragments are evident in the diatreme. (c) Continued bubbling circulation further enlarges maar and underlying diatreme; bedded ejecta sag into center of diatreme and faint cross-cutting channels manifest where ascending gas has flushed out finer particles. (From Woolsey et al., 1975.) (Reprinted from Physics and Chemistry of the Earth, Volume 9. Woolsey TS, McCallum ME, Schumm SA. Modeling of diatreme emplacement by fluidization, pp. 29–42. Copyright © 1975, with permission from Elsevier Science.)

in domed and fault-block-uplifted roof rock overlying shallow crustal intrusions. Undisputed doming has occurred over the classic laccoliths first described in 1877 by G. K. Gilbert in the Henry Mountains, Utah (Figure 9.27; see also Corry, 1988; Jackson and Pollard, 1988). A **laccolith** is a flat-floored intrusion with a domical upper surface essentially concordant with the layered rocks into which the magma was forcefully inserted. Laccolith intrusions apparently begin as tabular sills, generally with only a few kilometers of covering roof rock, but then inflate upward—by as much as 2 km—as more magma is injected. During the growth of some laccoliths, the arching roof rock breaks along one or more steeply dipping faults so that the roof rises in a trap-door fashion.

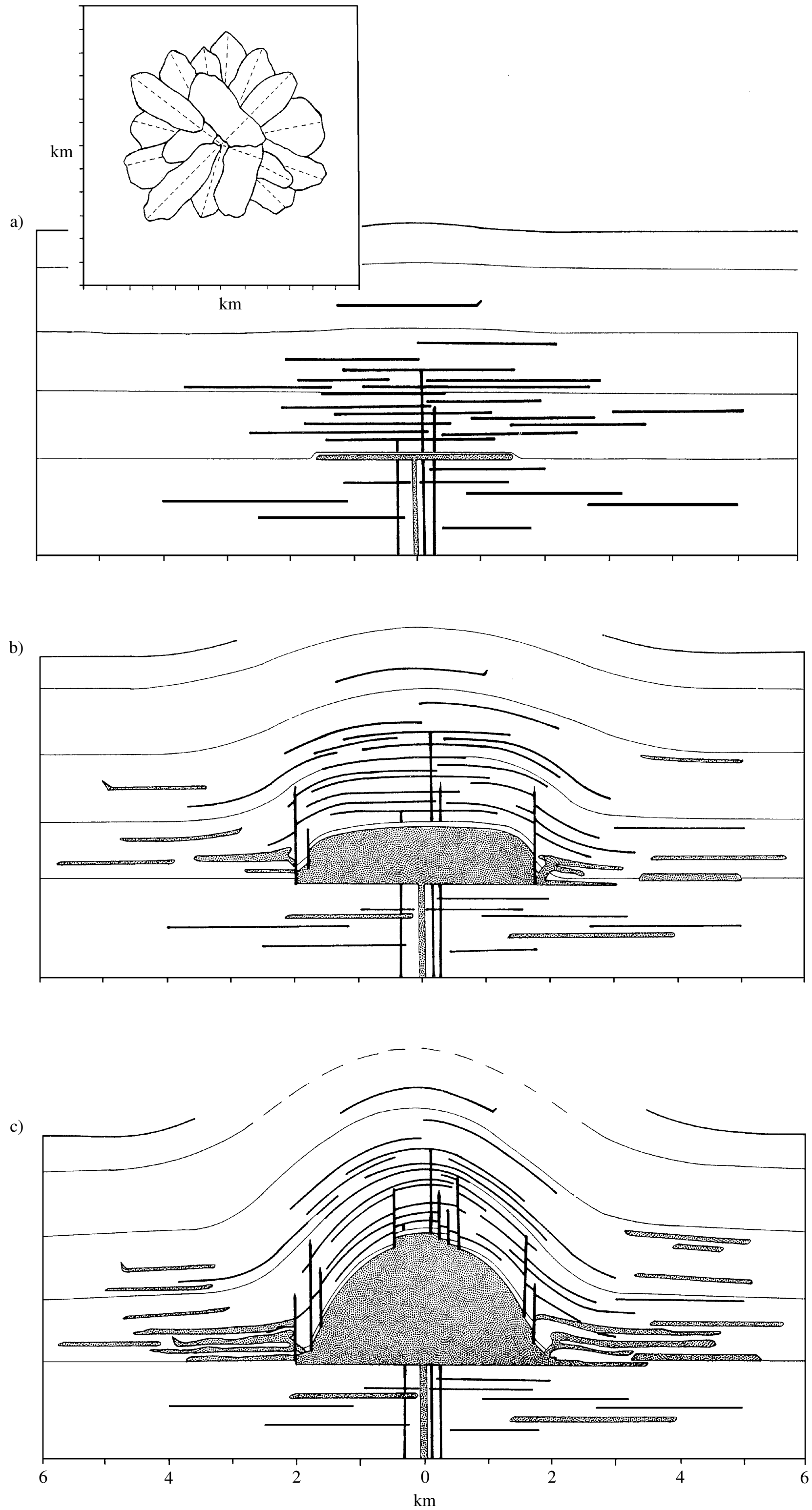
In the view of Hamilton and Myers (1967), the relatively lesser volume of granitic batholiths in the Paleozoic Appalachian Mountain belt than in the Mesozoic Cordilleran of western North America (Figure 9.16) reflects the laccolithic shape of plutons (e.g., Figure 9.17). The present 5- to 8-km depth of erosion in the Cordilleran is believed to expose more of their alleged laccolithic extent than does the deeper erosion in the older Appalachian region.

Ballooning. Many felsic plutons have the following characteristics, seen in Figures 9.18, 9.19, and 9.28:

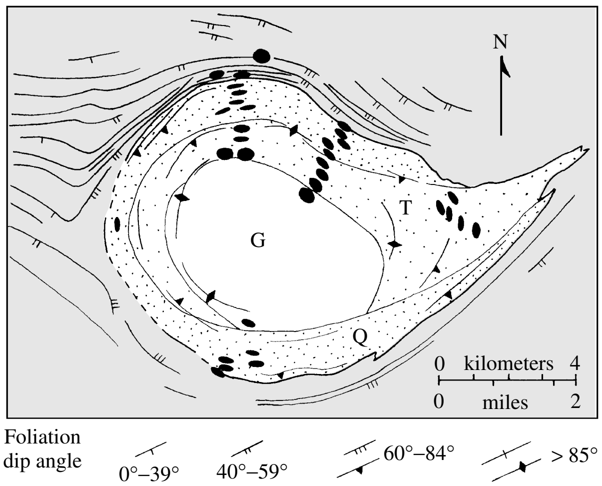
1. Roughly circular to elliptical shapes in map view
2. More or less concentric magmatic foliation within the pluton that is concordant with the contact and foliation in the wall rocks
3. Increasing strain toward the pluton-host rock contact, mainly evident in the wall rock but also locally seen in the intrusive rock; wall rocks flattened perpendicular to the contact and having steeply inclined stretching lineation parallel to it
4. Concentric compositional zonation within the pluton

A commonly invoked explanation of these characteristics is **ballooning,** a radially directed inflation of a magma chamber as additional magma is intruded. As an ascending magma diapir stalls, its “tail” continues to rise, inflating the surrounding cooler, more crystalline mass, producing a concentric foliation and a flattening strain within the pluton and the immediately adjacent, heated wall rocks. An important facet of the ballooning model is a return downward flow of the ductile wall rocks (Figure 9.21) as the ascending magma diapir stalls and inflates. This return flow can accommodate some of the lateral expansion and flattening of the wall rock that provide space for the pluton. However, Paterson and Vernon (1995) urge caution in several aspects of this model:

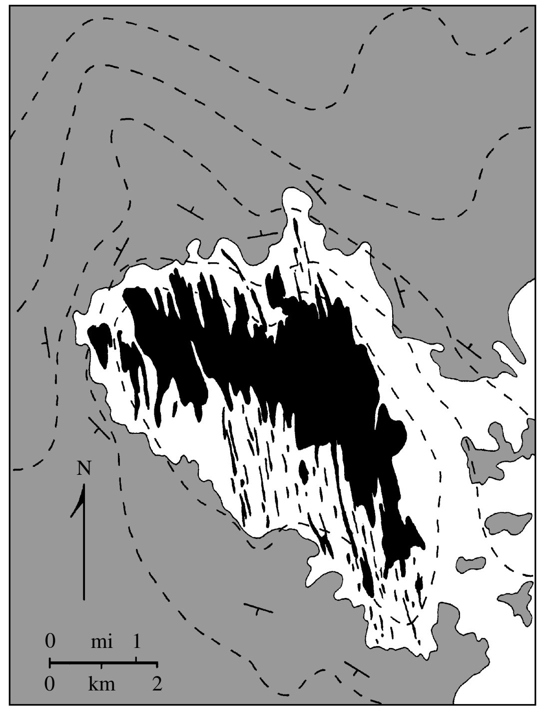
1. Concentric magmatic foliation patterns that cut across internal compositional contacts (e.g., Figures 9.19c and 9.28) suggest that magma flow may be less responsible than late-stage flattening strain of the largely crystallized magma.
2. Use of mafic inclusions as **strain markers** in the intrusive rock (Figure 9.28) can be misleading. Although commonly used to determine the amount



9.27 Growth of a **laccolith** by **doming** of overlying sedimentary strata in the Henry Mountains, Utah, according to Jackson and Pollard (1988). (a) Emplacement of sills and dikes (thick lines) and a thin protolaccolith (stippled). Inset is a plan view of the tongue-shaped, radial sills; tick marks at 1-km intervals. (b) Inflation of central laccolith induces bedding plane slipping and warping of overlying sills and 3.5-km thickness of Permian–early Tertiary strata. Additional diking and silling occur around the laccolith. (c) Increased inflation of the laccolith, now probably a composite intrusion formed of multiple pulses of magma, produces numerous faults (not shown) in overlying steepened, stretched strata. Sills below floor of main laccolith are conjectural. Horizontal and vertical scales are the same. (From Jackson MD, Pollard DD. The laccolith stock controversy: new results from the southern Henry Mountains, Utah. Geol. Soc. Am. Bull. 100:117–139. Reproduced with permission of the publisher, The Geological Society of America, Boulder, Colorado, USA. Copyright © 1988 Geological Society of America.)



9.28 Zoned concordant Ardara pluton, Donegal, Ireland, a result of **forceful emplacement.** Black ellipses show increasing flattening of mafic inclusions in the intrusion and mineral grains in the wall rock nearer the concordant contact. Dip of foliation in wall rocks (shaded) indicated by tick marks on strike lines and in pluton by triangles. G, granodiorite; Q, quartz monzodiorite; T, tonalite. (Redrawn from Pitcher, 1997.)



9.29 Forcible emplacement of granitic magma, Black Hills, South Dakota. Magma (black) wedged along steeply dipping foliation surfaces in country-rock schist (unshaded). Unconformably overlying, initially subhorizontal, sedimentary strata (shaded) were domed, as indicated by strike-dip symbols and by equal elevation structure contour lines (dashed) spaced 500 feet apart on the unconformity between the schist and sedimentary rock. Many faults are omitted. (Redrawn from Noble, 1952.)

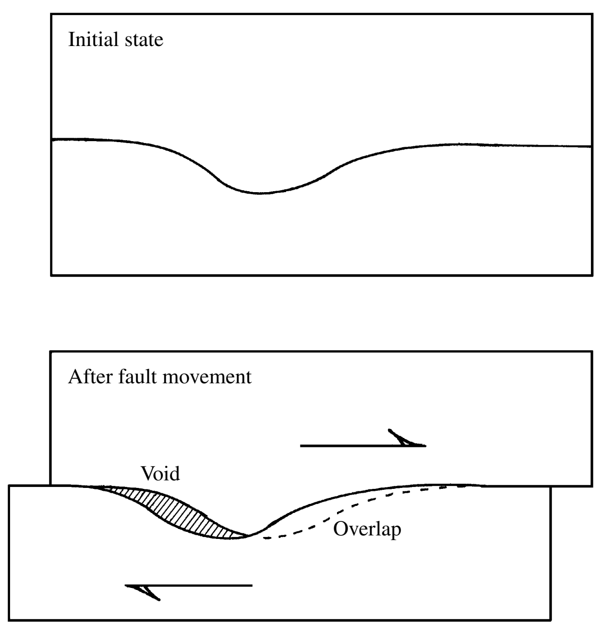
of strain during chamber inflation, the initial inclusion shape is rarely known with certainty and the final shape can be dependent on several factors.

3. Regional deformation may accompany pluton emplacement so that its fabric and that of the country rock may bear an imprint of far-field tectonic processes and these may well influence magma emplacement.

In a variant of the ballooning mechanism, magma forcefully injected along steeply dipping foliation surfaces in schist can accumulate in such volume as to dome unconformably overlying strata (Figure 9.29).

Ballooning is an example of **forceful emplacement** of magma. Both host and intrusive rock, but especially the former, display effects of deformation; their contact tends to be concordant. Because of the inherently higher T of the ductile host rock, relative to that in the shallower crust, more chemical interaction occurs between country rock and magma, creating broad, compositionally complex border zones and contact aureoles (discussed later).

Tectonically Created Room. The notion that magma emplacement is "structurally controlled" in orogenic belts has been popular for decades. Dilatant fault zones where rock has been broken up (brecciated) are favorable sites for magma emplacement. Bends along an active fault (Figure 9.30) provide a potential void for magma emplacement. Hinge zones of actively forming folds may also be a local site where magma might find space.



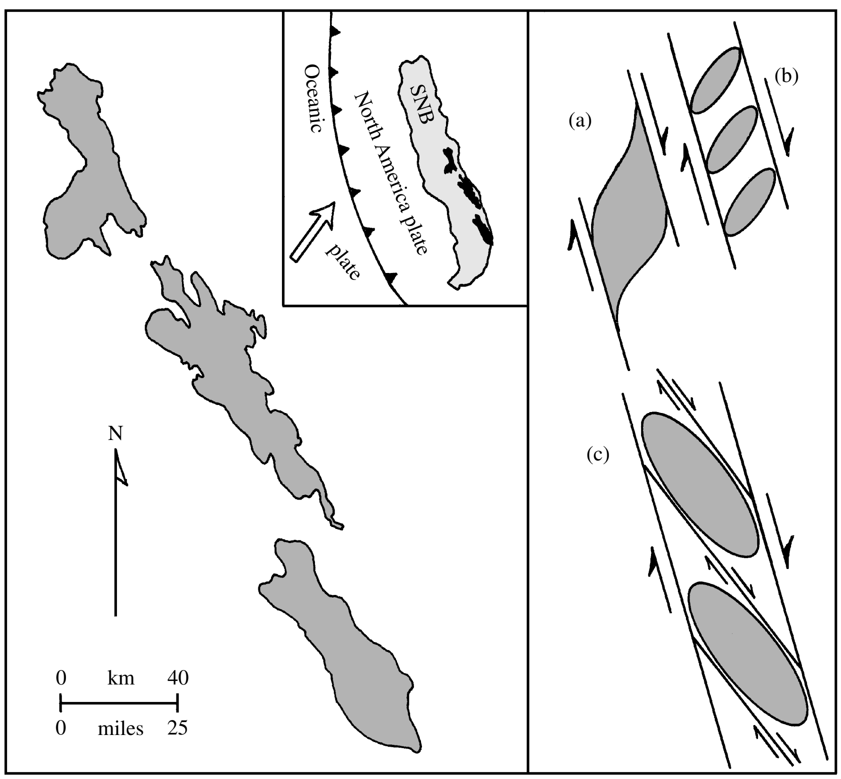
9.30 Space for magma intrusion created by jogs, or changes in orientation, in a fault. As movement occurs, a potential void into which magma may be emplaced is created. The fault may be of any type—normal, reverse, or strike-slip.

Emplacement of magmas into the continental crust in subduction zones would seem to be difficult because of the compressive state of stress in this tectonic regime. Yet, paradoxically, this setting is where voluminous granitic batholiths have been emplaced during broadly concurrent plate convergence. Lithospheric plates usually converge obliquely, rather than everywhere perpendicular, a characteristic that actually creates large-scale possibilities for magma emplacement (Bouchez et al., 1997). A case in point is the relative convergence direction of the North American and oceanic plates to the west when the Mesozoic Sierra Nevada batholith (Figure 9.16) was forming. The right-lateral component of motion by the oceanic plates along the plate juncture created shear zones tens of kilometers long within the batholithic terrane. Domains of extension in this overall compressional/strike-slip tectonic regime can develop between adjacent subparallel shears and can provide potential room for magma emplacement (Figure 9.31).

9.4.3 The Intrusion–Host Rock Interface

Aspects of magma emplacement are recorded in the interface, or **contact,** between the intrusion and its host rock. Smaller intrusions, such as thin, rapidly cooled dikes, transfer heat but not matter across the intrusive contact and in this sense are closed systems. Larger, longer-lived intrusions tend to be open in that matter—chiefly volatile fluids—as well as heat move across the contact. Also, mechanical work of deformation can be done on the country rocks by the intrusive magma (Figures 9.5, 9.19, 9.21, 9.27, 9.28, 9.29). The **contact aureole** of perturbed country rock and the thermal, chemical, and deformational gradients recorded in it tell of emplacement processes and provide chronologic information (Paterson et al., 1991). Likewise, records of similar gradients may be evident within the margin of the intrusion, and these can provide additional insights into emplacement processes.

Contacts and Border Zones. The simplest interface is the **sharp contact.** On a particular scale of observation,



9.31 Possible tectonic control on making room for plutons within the Sierra Nevada batholith. Map on the left shows three composite plutons (shaded) comprising the Cathedral Range intrusive sequence emplaced at 92–81 Ma. Smaller-scale map in upper middle of figure shows position of the three intrusive masses (black) within the larger Sierra Nevada batholith (SNB; Figure 9.16), which was created during mostly oblique convergence (arrow) of oceanic plates beneath the North American continental plate during the Mesozoic. Panel on right shows three possible ways (a, b, c) in which local extensional domains (shaded ellipses) might be created in the right-lateral strike-slip regime in the continental plate during oblique plate convergence; these extensional domains could create room for magma emplacement. (c) Extensional regions bridging between secondary right-lateral shear zones seem to correspond most closely in shape and orientation with the Sierra intrusive masses. (Redrawn from Tikoff and Teyssier, 1992.)

the interface between magmatic and host rock can be pencil line–thin (Figure 9.32). Intrusion of magma against significantly cooler, usually shallow crustal, host rock leads to rapid conductive and perhaps advective cooling along a pronounced thermal gradient. Chemical and thermal interactions between magma and host rock are minimal in a narrow contact aureole. Grain size in the magmatic rock generally diminishes toward the contact. Sharp contacts without grain size reduction might reflect a more dynamic intrusive margin where flowing magma swept away the initial chilled contact material.

For many, generally larger intrusions, the interface between magmatic and country rock is gradational over several to as much as hundreds of meters. For such a **border zone,** significant thermal, chemical, and physical interaction occurred between magma and host rock. One type of physical interaction is pervasive injection of dikes and sills into country rock that is initially relatively cool and brittle (Figures 9.4, 9.5, and 9.33a), which allows dislodged fragments to be stoped into the magma. Or hotter country rock permeated by magma can be assimilated or heated sufficiently to melt partially, forming a contaminated border zone (Figure 9.33b, c).

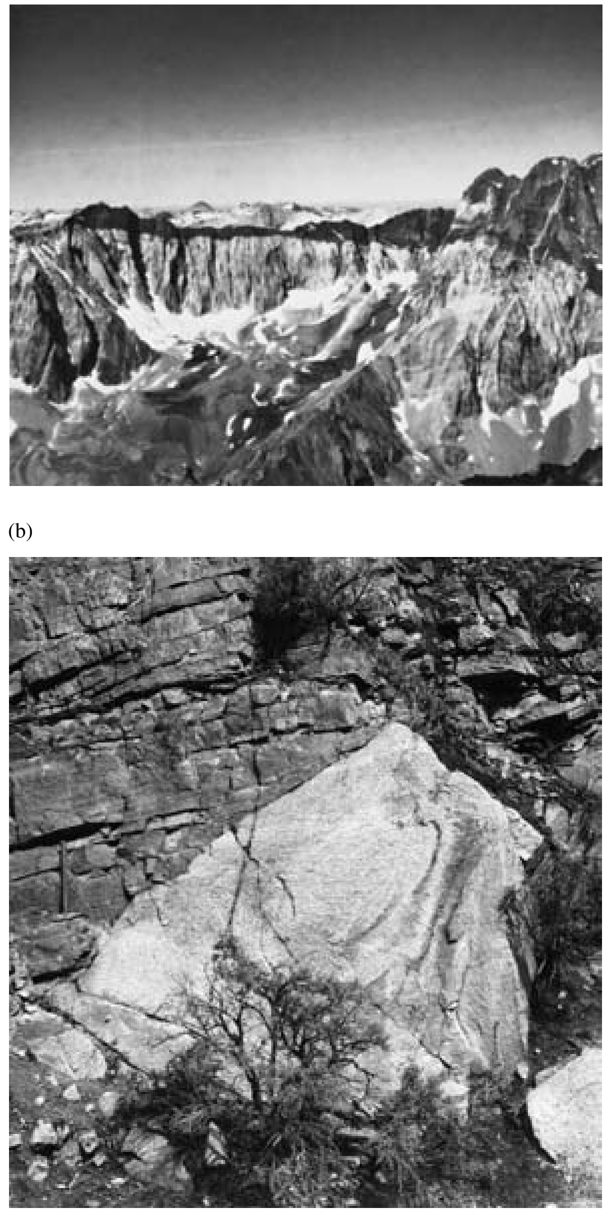
The variety of magma-country rock interfaces is virtually limitless. Wide variations that can be seen in a single intrusion reflect contrasts in thermal, chemical, and rheologic processes and properties.

Chronology of Magma Emplacement Relative to Tectonism. As challenging as elucidating the process of magma emplacement is determining the time of intrusion relative to episodes of regional tectonism, such as characterize orogenic zones at convergent plate boundaries. Was the intrusion pretectonic: Was the magma emplaced prior to deformation of the rocks in a broad region around the pluton? Or was the intrusion syntectonic, as the plutons in the alleged Sierra Nevada shear zones of Figure 9.31? Or was the intrusion posttectonic, so that structures in the wall rocks are cut by the intrusion? Answers to these questions are often crucial in working out the chronologic evolution of orogenic belts because intrusive rock is commonly the material best suited to isotopic dating.

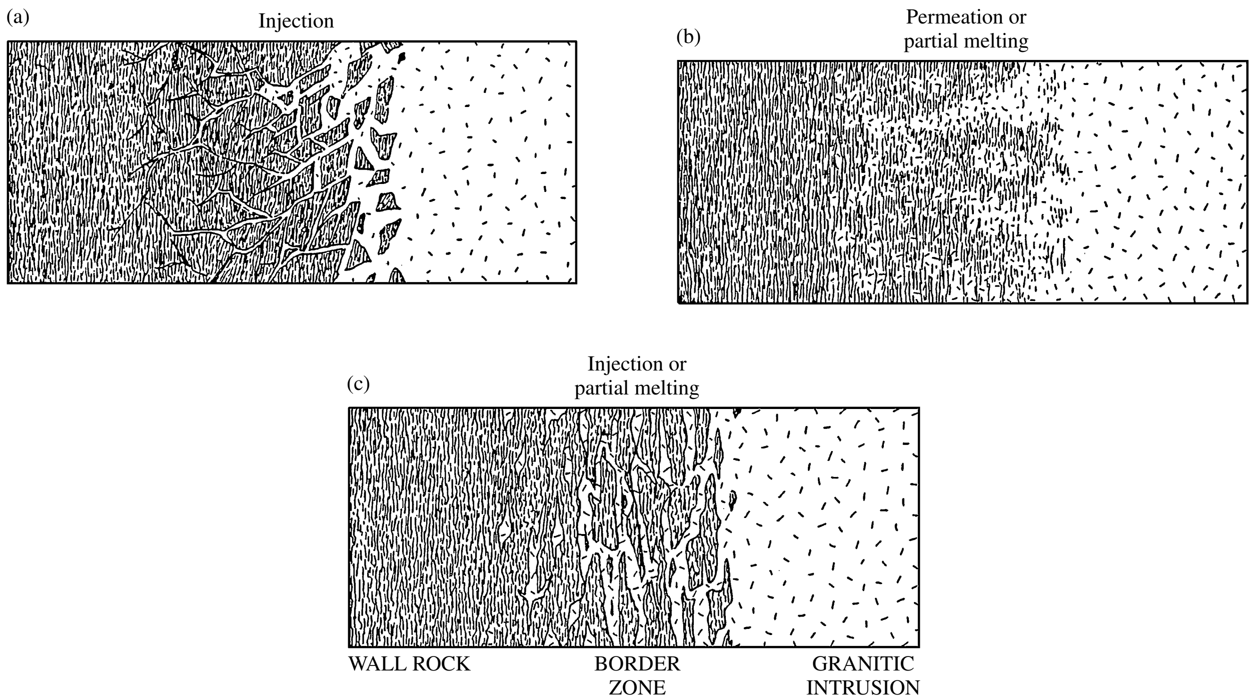
Determination of the relative age of forcefully emplaced intrusions is hindered by the fact that they deform aureole rocks and are themselves deformed, especially in their outer, more rigid margin. It must be determined whether deformation is localized in and around the intrusion or is of a regional character in which the pluton participated. A general difficulty in determining chronology of emplacement stems from the nature of the pluton itself. The lack of mechanically contrasting layers, such as bedding in sedimentary rocks and foliation in metamorphic, precludes development of folds that manifest deformation. Other possible effects of deformation must be sought.

(a)

(b)

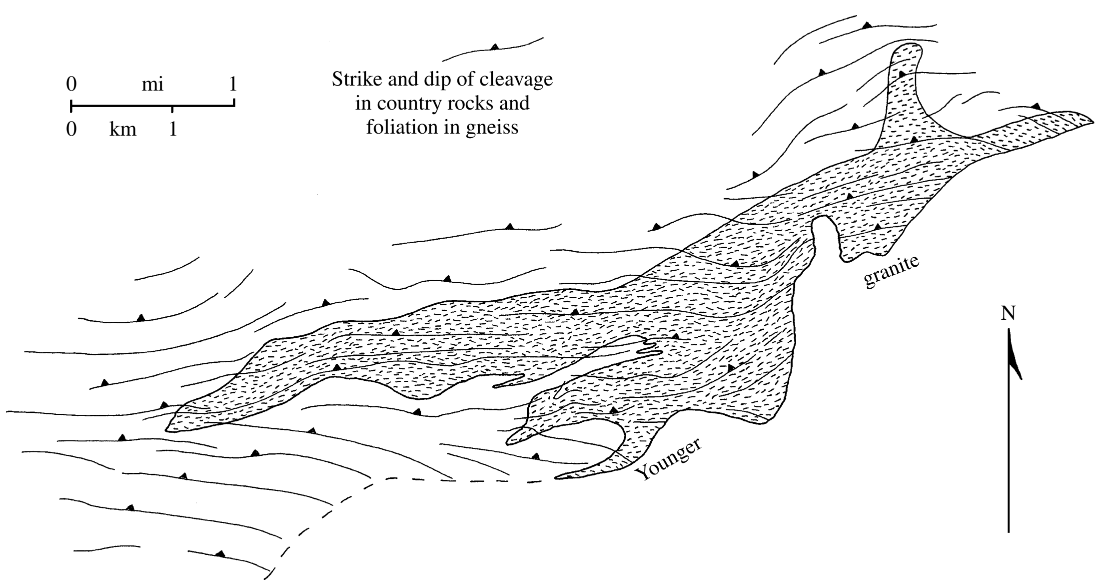


9.32 Sharp intrusive contacts of magma against host rock. (a) Panoramic view of contact between Taboose Pass septum of dark-colored metamorphosed sedimentary rock (exposed on ridge and peaks) against lighter-colored Jurassic granite (in lower slopes) in the Sierra Nevada batholith, California. The contact with the septum (or screen) appears horizontal but actually dips steeply between the granite and Cretaceous granodiorite (on back side of ridge). Width of field of view is about 3 km. (Photograph provided by John S. Shelton and information by Clifford A. Hopson.) (b) Outcrop view of a **sharp, discordant contact** between granodiorite and overlying darker-colored, layered quartzite. Note hammer for scale.



9.33 Idealized **border zones** between magma intrusion and host rock. These zones can range from less than a meter to hundreds of meters in thickness.

Determination of the relative chronology of pluton emplacement and regional tectonism is based on fabric and field relation criteria (Paterson et al., 1989). The strongest, but not the only, evidence for **pretectonic pluton emplacement** is an intrapluton foliation that parallels country rock foliation and is independent of the pluton margin, locally cutting across it at high angles. An example of a pluton and its country rock that were deformed together so as to develop a solid-state foliation continuous through both is shown in Figure 9.34. Evidence for **posttectonic pluton emplacement** can be found in the "cookie-cutter" Mount Ascutney



9.34 **Pretectonic pluton.** The original porphyritic granitic rock of the Saint Jean du Gard pluton north of Paris, France, is now a metamorphic augen gneiss in which the foliation is continuous with cleavage in the slate country rocks. Thus, both country rocks and pluton were deformed and metamorphosed as a single rock mass and possess similar anisotropic fabric. (From De Waard, 1950.)

intrusions (Figure 9.22), which are sharply discordant to country rock foliation and, therefore, were intruded after development of foliation in the metamorphic wall rocks. **Syntectonic pluton emplacement** can be the most difficult to ascertain. The most convincing occurs where the pluton contact is subparallel and continuous with the foliation in the country rocks beyond that of the immediate thermal contact aureole. Magmatic minerals, perhaps defining an internal foliation in the pluton, such as an igneous lamination (Section 7.9), have the same isotopic age as minerals defining the country rock foliation.

SUMMARY

Magma ascent through the lithosphere is a thermal-mechanical phenomenon governed by buoyant driving forces and viscous resistive forces related to the rheology of magma and lithospheric rock and to their contrast in density. Other governing factors include tectonic regime, state of stress, and stratification of the lithosphere with respect to strength and brittle versus ductile behavior. Energy for ascent is ultimately provided by gravity and internal heat in the Earth.

"Density-filtered" mantle-derived basaltic magmas can underplate and stagnate near the base of the lower-density continental crust. These buoyantly blocked magmas provide heat that can drive partial melting of the crust. Magma fractionation and assimilation of felsic country rock may reduce the density of evolved magmas, allowing farther ascent.

Subvertical basaltic dike swarms are crustal-scale features that testify to copious magma transport from a mantle source through the crust. In extensional tectonic settings, magmas follow self-generated extensional cracks oriented perpendicular to the least horizontal principal stress. If vertically continuous, overpressured basaltic magma can be driven upward tens of kilometers, even through less dense rock material. In compressional regimes, basaltic magmas are less able to rise all the way to the surface. Regional, far-field states of stress can be modified in the local environment of a central magma intrusion, allowing magma to intrude as cone sheets and radial dikes. Dike transport of magma is a trade-off between rate of flow and rate of solidification through conductive cooling; slowly flowing viscous magma dissipates heat to the wall rock and becomes immobile. Because distance of magma transport through a dike is proportional to the fourth power of its width and inversely proportional to magma viscosity, low viscosity basaltic magma can readily move a long distance through dikes a meter or so wide. In contrast, much more viscous granitic magma requires wider dikes through which to move.

Volatile-rich mafic to ultramafic magmas probably begin their ascent through the lithosphere via self-induced fractures. As overpressured magma follows the upward-propagating crack, it may be arrested if the supply is exhausted or continue as a fluid-oversaturated magma column that "drills" its way upward and finally may breach the surface, where it creates an explosive maar crater underlain by a diatreme.

Slow ascent of viscous granitic magma through ductile country rock as buoyant diapirs kilometers in diameter is an alternative mechanism to dike transport. The more or less spherical shape of the ascending diapir minimizes viscous drag because of minimal surface area relative to volume. This, in turn, maximizes buoyant lift and retention of thermal energy. Diapirs are most likely in the ductile, hotter lower crust, where wall rocks can readily be thermally "softened" to facilitate rise before stagnation at higher cooler crustal levels, where increased country rock strength arrests continued ascent. Subsequent diapirs are likely to ascend more easily in the thermally conditioned wake of preceding ones.

Magma is finally emplaced as an intrusion or pluton as a result of increased viscosity and loss of mobility upon cooling, insufficient buoyancy to carry it higher, or increased strength of the country rocks. Emplacement processes are most properly considered in three dimensions. However, the pluton–country rock system is usually observable only in surface exposures of generally less than a kilometer of surface relief. Careful and critical observation of field relations, fabric, and rock and mineral compositions is required to interpret the specific processes involved in magma emplacement for a particular pluton accurately. The relative contributions of different emplacement processes are variable, even for a particular pluton. Laccoliths and some other crustal plutons create room by upward forceful deflection of the overlying roof rocks. Stoping is favored in fractured shallow crustal rock but even with assimilation of country rock does not create new room for magma, therefore permitting only a transfer in position between these masses. The absence of country rock xenoliths in an intrusion does not necessarily rule out stoping and assimilation; piecemeal stoped blocks may have been completely assimilated into the magma, or foundered roof slabs may have sunk below the level of exposure. Some plutons appear to have forcefully pushed aside their ductile wall rocks as the magma ballooned laterally. Local extensional tectonic environments may create room for magma emplacement.

Criteria for determination of the time of magma emplacement relative to regional tectonism are based on fabric and field relations near the pluton margin. Such determinations can be crucial in unraveling complex orogenic chronology.

CRITICAL THINKING QUESTIONS

9.1 Discuss the role of buoyancy in the movement and stagnation of magma in the lithosphere.

9.2 Discuss the origin and consequences of magma overpressure. How is it a factor in ascent and emplacement of magma? How does high versus low overpressure control different mechanisms of intrusion? How does overpressure move magma to the surface, where it can erupt?

9.3 Contrast a local state of stress near a central intrusion and a regional state and the processes through which these control the ascent and emplacement of magma.

9.4 Contrast how different sheet intrusions are emplaced.

9.5 Compare and contrast ascent of magma from its source by diapirism and by diking with respect to thermomechanical factors and rheology of magma and country rock.

9.6 Characterize mechanisms of emplacement and making of room for intrusive magma, indicating specific evidence for each mechanism.

9.7 Discuss the so-called room problem for large intrusions: whether it is real or fictive and why.

9.8 Indicate how the relative age of regional deformation and processes of intrusion can be determined in features of an intrusion and its country rock.

9.9 In Figure 9.7a, propose a reason why most of the dikes in the Mackenzie swarm are oriented approximately north-south.

9.10 A long-standing observation is that the most common intrusive rock is granitic and the most common extrusive is basaltic. Propose reasons why this might be so.

PROBLEMS

9.1 If the change in melt density with depth is taken into account (Problem 8.9), how does this affect the magnitude of the excess magma pressure in Figure 9.2? Explain fully and indicate any assumptions made.

9.2 Derive the expression for magma overpressure, or hydraulic head, $\Delta\rho gz$. Indicate any assumptions made.

9.3 What is the velocity of basalt magma rising through a 1-m-wide dike if $\eta_{magma} = 100$ Pa s and $(\rho_{crust} - \rho_{magma}) = \Delta\rho = 0.1$ g/cm^3? (*Answer:* 3.3 m/s.)

9.4 Some 2-m-wide basalt dikes in Iceland are as much as 30 km long. All other factors being the same, including magma supply rates, a 6-m-wide dike in the Mackenzie swarm in Canada could be how long? (*Answer:* 2430 km.)