Chapter 10

Magma Transport in Dikes

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Chapter Outline

GLOSSARY

brittle A material that has little capacity to deform plastically and is not able to relax stresses that become localized at crack-like defects during an applied deformation. In such a material, such defects will propagate rapidly and spatially localized deformation that is brittle fracture. Such brittle deformation involves the breaking of chemical bonds, whereas plastic and viscous deformation involves microscopic motion of atoms or molecules past one another.

buoyancy Buoyancy is a force induced by the density difference between two bodies. In the case of dikes, the buoyancy force is proportional to the density of the host rock minus the density of the magma. Therefore, this force can be either positive, and drive the magma toward the surface, or negative, and prevent the ascent of the magma. The level at which the buoyancy changes sign is called the level of neutral buoyancy.

buoyancy length The length of a dike at which its buoyancy force is sufficient to result in its upward propagation without the need of additional magma supply from a pressurized reservoir.

constitutive equation A material-specific relation between two physical quantities, such as the response of a material to an externally applied force.

crack A crack is a defect around which atomic bonding is completely broken. This results in a large volume expansion and stress concentration and, hence, a long-range stress field surrounding cracks.

cubic law The volumetric flow rate of a Newtonian viscous fluid through a planar channel of high aspect ratio and that is bounded on both sides by solid walls and scales with the cube of channel thickness. The functional relationship between flow rate and viscosity, pressure gradient, and channel thickness cubed is called the cubic law.

elastic Upon an applied stress a material will deform. If the deformation is instantaneous and recoverable it is called elastic.

feeder dike Dikes that supply magma to the surface during volcanic eruptions.

fracture Interaction among cracks during an applied stress leads to localized deformation that is discernible macroscopically, that is brittle fracture. In other words, brittle fracture starts with the formation of microcracks, which interact to form a macroscopically discernible fracture.

fracture toughness The fracture toughness, specific to the material and opening mode (see mode I definition), is the capacity of the material to resist fracturing when stressed.

The Encyclopedia of Volcanoes. http://dx.doi.org/10.1016/B978-0-12-385938-9.00010-9

The Encyclopedia of Volcanoes, Second Edition, 2015, 215−224

Griffith's theory According to Griffith's theory, an existing crack will propagate if the strain energy released is greater than the surface energy created by the formation of the two new material surfaces.

Hooke's law The constitutive equation for a material that undergoes a recoverable deformation where the strain is linearly proportional to the applied stress.

level of neutral buoyancy (LNB) Theoretically, the level at which buoyantly ascending magma has the same density as the surrounding rock.

linear elastic fracture mechanics (LEFM) Fracture mechanics is the study of the propagation of cracks in elastic materials where the relationship between applied stress and strain is linear. The basic assumptions of LEFM are (1) the presence of crack-like defects; (2) a crack is a free, internal, planar surface with stresses near the crack tip that can be calculated from linear elasticity; and (3) the growth of such cracks leads to the failure, that is, brittle fracture, of the material.

lithosphere The lithosphere can be defined based on seismic, mechanical, or thermal considerations. The seismic lithosphere of the Earth was initially defined as a static unit lying above a low velocity zone for shear waves in the upper mantle, presumably associated with partial melting and weak mechanical behavior, that is, the asthenosphere. The mechanical lithosphere comprises those rocks that remain a coherent part of the plates on geological timescales and deform in an elastic manner, even on long timescales, whereas the underlying asthenosphere undergoes permanent (plastic) deformation. Studies of flexure and deformation due to topographic loads, such as volcanic edifices and mountain belts, have led to a definition of lithosphere as the layer that sustains elastic stresses to support these loads. The thermal lithosphere is taken to be the upper thermal boundary layer of a planet, where heat is transported by conduction, as opposed to the underlying asthenosphere where convection is the dominant heat transfer mechanism. Because the mechanical properties of rocks are strongly temperature-dependent, the mechanically and thermally defined lithospheres are essentially equivalent.

Maxwell model A model for the deformational behavior of viscoelastic materials, which is based on the timescale at which deformation changes from recoverable to nonrecoverable.

mode I There are three ways of applying a force to enable crack propagation. One is a tensile stress normal to the plane of the crack and the two others are shear stresses parallel to the plane of the crack. The mode in which a crack opens under tensile stress normal to the plane of the crack is called mode I, whereas the other two are called mode II and mode III.

Newtonian viscosity If for a viscous material the relationship between applied shear stress and resultant shear rate is linear, the constant of proportionality is called the Newtonian viscosity.

plastic If a solid material undergoes a time-dependent and nonrecoverable deformation under an applied stress it is called plastic. The deformation may be distributed homogeneously throughout the material or it may be localized in a narrow region.

stress intensity factor The growth of a crack under an applied stress is related to the tensile stress acting at the crack tip. For a given opening mode, the relationship between applied stress and crack length is characterized by a constant, called the stress intensity factor.

thermal boundary layer It corresponds to the layer of magma closest to the dike walls where heat transfer is dominated by conduction toward the dike wall, as opposed to along-dike advection by magma flow.

viscoelastic A material that undergoes recoverable deformation on short timescales and nonrecoverable deformation on long timescales is called viscoelastic. The time at which deformation changes from recoverable to nonrecoverable is often referred to as the Maxwell time.

viscous Under an applied stress a viscous fluid undergoes a time-dependent, nonrecoverable deformation that is homogeneous throughout the fluid volume.

yield stress Upon an applied stress a material will deform. If the material will undergo recoverable deformation below some threshold stress value and permanent deformation above this value, this threshold stress value is called the yield stress. After yielding, nonrecoverable deformation is usually referred to as plastic deformation. However, sometimes the terms viscous and plastic deformation are used interchangeably.

1. INTRODUCTION

Dikes are the principal mode of magma transport within and through planetary **lithospheres**. Dikes represent the critical link between regions of melt production and regions of melt accumulation, such as magma chambers. In some cases, dikes propagate to a planet's surface, resulting in fissure eruptions. In other cases, dikes may give way to more spatially localized conduits at shallow depths, supplying magma to individual volcanic vents. This chapter provides a summary of the mechanics and

FIGURE 10.1 Aerial view of a 8-km-long dike that radiates from Ship Rock, New Mexico, USA, a volcanic plug that rises 550 m above the valley floor. Three prominent dikes radiate out from Ship Rock (Nature/Universal Images Group/Getty Images).

dynamics of dike propagation and magma transport through dikes.

2. OVERVIEW

Dikes and sills are sheet-like bodies of igneous rock that were emplaced as magma within preexisting rock. They are typically of the order of 1−10 m in thickness and of the order of 1−100 km in lateral extent (Figure 10.1). Dikes are formed by fracturing the host rock, due to magma injection (Figure 10.2). Their propagation is oriented preferentially in the direction perpendicular to the minimum total compressive stress of the surrounding rock, but may also follow preexisting planes of weakness, such as bedding planes or preexisting **fractures**.

The formation and propagation of dikes may allow the rapid transport of magma without extensive solidification due to cooling. Estimates of dike propagation speeds, deduced from seismic observations, are of the order of meters per second. In order for dikes to propagate, magma needs to exert sufficient stress at the dike tip to overcome the fracture strength of the rock. Furthermore, magma pressure within the dike needs to exceed the stresses exerted by the surrounding rock on the dike walls, in order to keep the dike from closing. Pressure gradients within the magma also need to balance **viscous** resistance, in order to allow magma to flow into the dike as it grows in length. Processes that may result in the required magma pressures may include the change in volume associated with melting, compaction of partially molten rock, magma **buoyancy** relative to the surrounding rock, and in the case of dikes emanating from inflated magma reservoirs, the **elastic** energy stored in the reservoir's wall rock, commonly known as reservoir overpressure.

Theoretical treatments of dike propagation therefore have to consider the fracture mechanics at the dike tip, magma flow within the dike, deformation of the surrounding wall rock, as well as cooling of the dike (Rubin, 1995). We will define several characteristic pressure scales, in order to facilitate the distinction of dominant stress balances for dike propagation. End-member cases for propagating dikes are (1) dikes of finite volume that are "self-propagating" due to magma buoyancy; (2) dikes that propagate within partially molten rock and are supplied with melt from the surrounding porous wall rock; and (3) dikes that are connected to a magma reservoir and are supplied by magma from within the reservoir.

3. DEFORMATIONAL BEHAVIOR OF MATERIALS

Upon an applied stress a continuous material will deform. For a fluid of **Newtonian viscosity,** the resultant deformation is nonrecoverable. For a linearly elastic solid the deformation is instantaneous and recoverable. If the solid deforms elastically below some threshold stress, but yields to nonrecoverable deformation above this **yield stress** it is called **plastic**. Nonrecoverable deformation can also be **brittle** and localized, which involves the breaking of chemical bonds and is called fracture.

3.1. Elastic Deformation

The **constitutive equation** for linear elastic deformation is

$$\sigma = M\epsilon, \tag{10.1}$$

where σ is the applied stress, M is the elastic constant, and ϵ is strain. This relation is also called **Hooke's law**. There are five elastic constants. For example, Young's modulus, E, applies to the case of uniaxial stress

$$\sigma = E\epsilon \tag{10.2}$$

and the shear modulus, μ, in the case of shear stress

$$\sigma = \mu\epsilon. \tag{10.3}$$

The relationship between E and μ is given by

$$E = 2\mu(1+\nu), \tag{10.4}$$

where ν is the Poisson's ratio.

3.2. Viscous or Plastic Deformation

The constitutive equation for linear viscous deformation is

$$\sigma = \eta\dot{\epsilon}, \tag{10.5}$$

where η is the Newtonian viscosity and $\dot{\epsilon}$ is the strain rate. For many natural materials, the relation of stress to strain rate is not linear and constitutive models for such non-Newtonian rheology often relate stress to strain rate via a power law.

3.3. Viscoelastic Deformation

In addition many materials behave like an elastic solid on short timescales, but viscously over longer times. The most well-known constitutive model for such **viscoelastic** deformation is the **Maxwell model**, where the total strain rate is the sum of an elastic strain rate and a viscous strain rate, with the latter becoming increasingly dominant at times greater than or equal to the Maxwell time.

3.4. Brittle Fracture

Under application of a stress, the fragmentation of a solid volume into two or more parts is referred to as fracture. Although plastically deformable solids can undergo

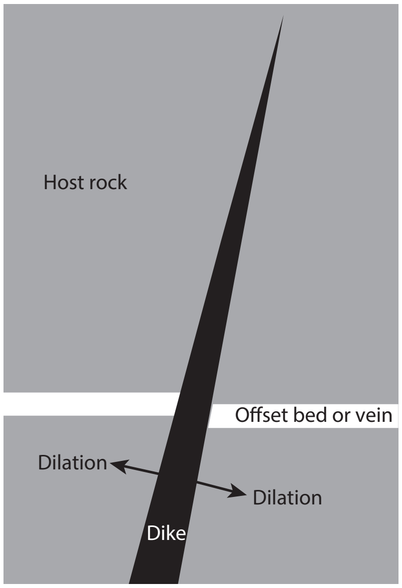


FIGURE 10.2 Schematic illustration of a dike with wall displacements that correspond to dilation by magma injection, as indicated by the offset of preexisting planar features. This type of fracture wall displacement is called opening mode or mode I displacement.

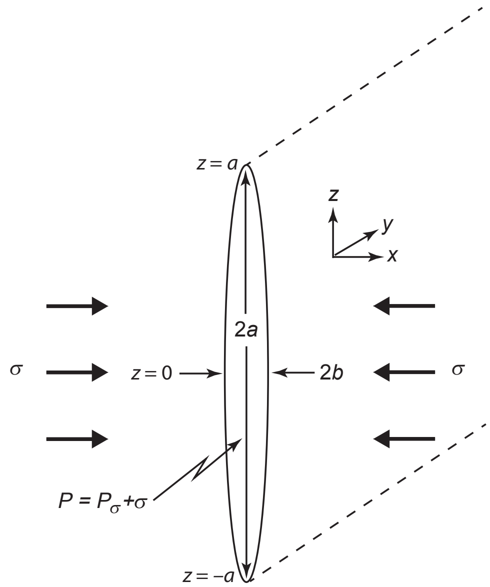


FIGURE 10.3 Geometry of an elastically walled dike of infinite extent in the y direction. The dike is of height $2a$, width $2b$, and has an internal pressure of P that exceeds the stress in the surrounding rock is σ by P_σ.

ductile fracture, we will focus, herein, on brittle fracture. The brittle fracture of a solid body typically involves the accumulation of damage, resulting in the nucleation of **cracks** or voids and finally, their growth. On a macroscopic level, the mechanical properties of solids tend to be limited by their imperfections, also known as defects. Linear elastic fracture mechanics (LEFM) provides a conceptual framework that is based on crack propagation from such defects, in order to asses a material's **fracture toughness**. Because the in situ stresses at depth are never tensile, dike formation and propagation requires that magma pressure exceeds the least principal stress by some critical amount in order for a tensile fracture to grow.

4. LINEAR ELASTIC FRACTURE MECHANICS

LEFM provides a framework for quantifying the stress required to propagate a fracture, through the use of a parameter known as the fracture toughness.

4.1. Elastic Pressure

Consider a two-dimensional crack (Figure 10.3) of length $2a$ and subject to a uniform excess internal pressure P_σ, relative to a uniform remote stress σ. This crack will have a half thickness of

$$b(z) = \frac{(1-\nu)P_\sigma}{\mu}\sqrt{a^2 - z^2} \quad |z| < a, \tag{10.6}$$

where z is the vertical coordinate. From the pressure required to open such an idealized crack by $2b$ at $z=0$, one can define a characteristic elastic pressure as

$$P_e \sim \frac{\mu}{(1-\nu)}\frac{b}{a}. \tag{10.7}$$

4.2. Fracture Pressure

According to Griffith, an existing crack will propagate if the strain energy released is greater than the surface energy created by the formation of the two new crack surfaces. The practical application of **Griffith's theory** is difficult, in part because of the challenges posed by measuring quantities such as surface energy. Instead material fracture is conventionally treated within the framework of LEFM, which postulates the existence of microcracks within any real solid volume. Because stress becomes concentrated at the tips of these microcracks they can grow within a process zone located at the fracture tip, resulting in the opening and propagation of a macroscopic fracture.

As a consequence of these material imperfections, actual fracture strengths are considerably less—by about one to two orders of magnitude—than the theoretical cohesive material strength, also called cleavage stress. Moreover, the fracture of a material can be predicted in terms of the tensile stress acting in the crack tip region.

Under pure tensile failure (**mode I**) this stress is proportional to the **stress intensity factor**, defined as

$$K = Y\Delta P_\sigma \sqrt{\pi a}, \tag{10.8}$$

where Y is a constant that depends on the crack opening mode and geometry, and for the geometry under consideration has a value of 1. The minimum value of K at which fracture occurs or a crack propagates can be determined experimentally and is known as the fracture toughness, K_c. Laboratory measurements at atmospheric conditions indicate that $K_c \sim 1$ MPa m$^{1/2}$ and that it increases with pressure, although the actual fracture toughness under field conditions remains poorly resolved and the range of plausible values is from about 1 MPa m$^{1/2}$ to 100 MPa m$^{1/2}$. The characteristic fracture pressure scale is thus defined as

$$P_f \sim \frac{K_c}{\sqrt{a}}. \tag{10.9}$$

5. MAGMA FLOW IN DIKES

Most treatments of magma flow within dikes approximate the flow as one-dimensional, laminar, and incompressible. In the case of a vertically oriented dike, the volumetric flow rate is

$$Q = 2bw\bar{u}_z = -\frac{w}{3\eta} b^3 \frac{dP'}{dz}, \tag{10.10}$$

where $\bar{u}_z$ is the average magma velocity in the z-direction, η is the magma viscosity, P' the nonhydrostatic magma pressure, and w is the dike breadth, which is the dimension perpendicular to the direction of propagation (z) and to the width (x) of the dike. Equation (10.10) yields the characteristic viscous pressure scale

$$P_\eta \sim \frac{\eta a Q}{b^3 w} \sim \frac{\eta a \bar{u}}{b^2}. \tag{10.11}$$

Furthermore, using the average magma velocity, $\bar{u}_z$, together with mass balance results in an equation for the change in dike width

$$\frac{db}{dt} + \frac{d}{dz}(b\bar{u}_z) = 0. \tag{10.12}$$

6. DIKE FORMATION AND PROPAGATION

6.1. Dikes within Melting Zones

The isotopic disequilibria found in mid-ocean ridge, Ocean Island and island arc basalts require average melt ascent rates of 1–50 m yr^{-1} (Kelemen et al., 1997; Turner and Bourdon, 2011). It is thought that these high ascent rates and isotopic disequilibria are most likely a consequence of channelized porous flow by reactive melt transport within the asthenosphere. However, there are other processes that have the potential to enhance rates of melt transport in melting zones, such as the formation of melt-filled veins during deformation of the partially molten rock (Kohlstedt and Holtzman, 2009).

To what extent dikes contribute to melt transport within melting zones, however, remains unclear (Kelemen et al., 1997; Weinberg, 1999). In the case of highly viscous felsic melts, the flow of melt into the dike from the surrounding partially molten rock is very slow and it has been suggested that a viable alternative could be that upward heat transfer may facilitate pervasive melt migration through ductile deformation, rather than elastic fracture (Weinberg, 1999).

Dike growth in partially molten peridotite could be a consequence of excess pore pressure due to the volume increase associated with melting and compaction (McKenzie, 1984). Melt could thus be forced into preexisting melt pockets and exceed the rock fracture toughness, thus facilitating fracture propagation (Rubin, 1998). If dike growth in partially molten peridotite is treated as a purely elastic process, it is found that the upward propagation and width of dikes are primarily controlled by the balance between the rate of porous inflow from the surrounding rock and upward flow of melt within the dike. Because the melt flux into the dike is, to first order, dependent on the ratio of permeability to melt viscosity the porous influx for felsic dikes is rate limiting, making dike propagation difficult. For asthenospheric dikes the balance between porous inflow and upward propagation results in theoretical dike widths of the order of 10^{-2} m (Rubin, 1998). Such dikes, if they exist, would be too thin to propagate within the lithosphere without freezing. Consequently, lithospheric transport of asthenospheric melts requires that dikes are fed by large accumulations of melt, perhaps at the top of the melting column or from a channelized tributary network (Weinberg, 1999).

6.2. Buoyancy-Driven Dike of Finite Volume

Vertically oriented dikes of constant volume can "self-propagate," once they reach a critical length, called the **buoyancy length**, denoted as a_b. At this length, the balance between the body force exerted on the magma within the dike and the lithostatic force against the dike walls causes the dike tip to propagate upward. At the same time the tail is being squeezed shut, thereby displacing the magma upward. The stress due to buoyancy scales as the characteristic buoyancy pressure

$$P_\rho \sim \Delta\rho g a, \tag{10.13}$$

where $\Delta\rho = \rho_r - \rho$ is the difference between host-rock density, ρ_r, and magma density, ρ. It is balanced by the fracture pressure, P_f (Eqn (10.9)). Equating these two pressure scales one obtains a scaling relation for the buoyancy length, that is, the half length required for a "self-propagating" dike, given by

$$a_b \sim \left(\frac{K_c}{\Delta\rho g}\right)^{2/3}. \tag{10.14}$$

An important implication of this relation is that for reasonable geological conditions, that is, $\Delta\rho \sim 100$ kg m^{-3} and $K_c \sim 1$ MPa m$^{1/2}$, once a magma-filled fracture has formed and grown to a length of the order of 100 m, the fracture resistance of the rock to further propagation is negligible.

For a dike of volume, V, Taisne et al. (2011) found that during the early stages of dike propagation the dike increases in width. Once the dike has reached a width of

$$w_f \sim \left(\frac{\mu}{1-\nu}\frac{V}{\Delta\rho g}\right)^{1/4}, \tag{10.15}$$

obtained from a balance between P_e and P_ρ, the increase in dike width becomes negligible, due to the fracture toughness of the solid. Instead, the dike will predominantly increase in length, at a rate governed by a balance between P_η and P_ρ, and at a velocity of $\overline{u} = a/t$. Its length thus scales as

$$a \sim \left(\frac{\Delta\rho g V^2}{\eta w_f^2} t\right)^{1/3}. \tag{10.16}$$

Because the velocity of melt flow depends on the square of dike width (Eqn (10.10)), it becomes virtually impossible for dike tail to completely close and some fraction of melt will become "trapped" within the tail. Consequently, the net volume of melt within the dike above the tail decreases with time (Figure 10.4) and the dike decreases in thickness until it eventually stalls.

6.3. Dike Connected to a Magma Source

For a dike that remains connected to regions of subsurface magma accumulation, that is, magma reservoirs, the volume of melt within the dike will increase over time. Magma flow into the dike may be driven by a combination of buoyancy and reservoir pressure, with the latter due to elastic deformation of the reservoir's host rock. Assuming that on the timescale of dike propagation the rock can be treated as elastic and that the magma is incompressible, the relationship between the change in magma pressure, relative to the ambient stress within the surrounding wall rock, is linearly proportional to the change in volume of stored magma, ΔV_{ch} within a reservoir of volume, V. The resultant pressure scale is

$$P_{ch} \sim \mu \frac{\Delta V}{V}. \tag{10.17}$$

Changes in magma storage arise due to an imbalance between volumetric rate of magma supply to the chamber, Q_s, and magma volumetric flow rate into the dike, Q. The resultant rate of change in chamber pressure is

$$\frac{dP_{ch}}{dt} \sim \frac{\mu}{V}\left[Q_s(t) - Q(t)\right]. \tag{10.18}$$

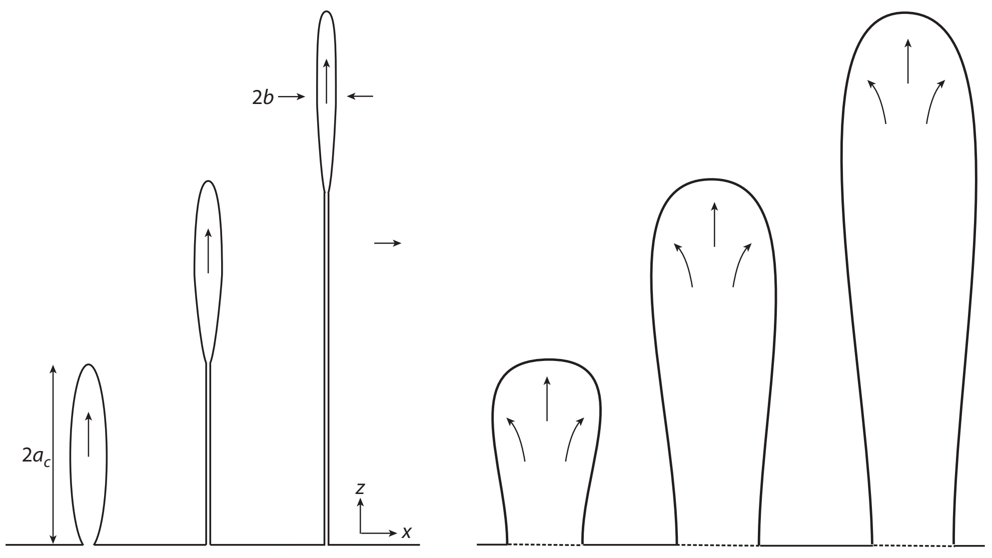


FIGURE 10.4 Schematic illustration showing the shape of a propagating dike of finite volume. As the dike ascends a small volume of magma remains trapped in the tail, which does not close completely. Consequently, the head of the dike decreases in thickness as it rises (Taisne et al., 2011).

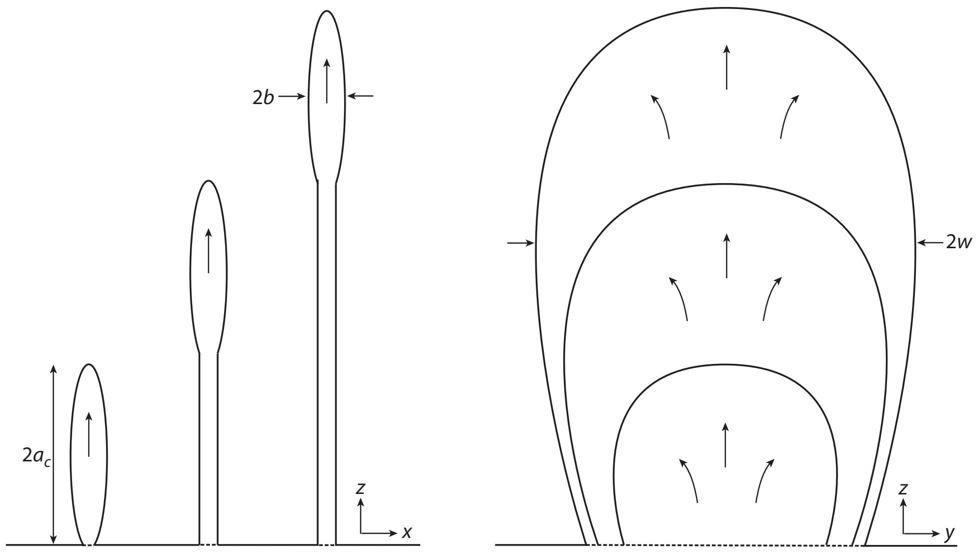


FIGURE 10.5 Schematic illustration of a dike fed by a pressurized magma reservoir, propagation in the steady-state regime (modified after Menand and Tait (2002)).

Typically, the tensile strength of the surrounding wall rock is of the order of 1–10 MPa, implying that excess magma chamber pressures will be of that magnitude (Gudmundsson, 2012). The relative importance of P_{ch} and P_ρ depends on dike length, dike orientation, and magma density, relative to the surrounding rock. Menand and Tait (2002) showed that for a constant P_{ch}, dikes will initially propagate in a regime where P_ρ is negligible and the dominant balance is between P_{ch} and P_f. In this initial transient regime dike propagation is radial with both a and w increasing (Figure 10.5). For vertically oriented dikes, also called **feeder dikes** because they typically feed surface eruptions from magma reservoirs, the flow eventually transitions to a balance between P_ρ and P_f in the bulbous dike tip and between P_ρ and P_η in the thinner dike tail (Figure 10.5). This transition occurs at a critical length, a_c, which is defined by this balance and the thickness of the dike tail can be estimated from the balance between P_ρ and P_η as (Lister and Kerr, 1991)

$$b_t \sim \left(\frac{\eta Q}{w \Delta \rho g}\right)^{1/3}. \tag{10.19}$$

The volumetric flow rate of magma into an incipient dike is inversely proportional to magma viscosity and proportional to the dike thickness cubed (Eqn (10.10)). Consequently, there is an expected time delay of the order of hours to days for basalts and years for rhyolites before an incipient dike tip becomes sufficiently pressurized to propagate (McLeod and Tait, 1999). Once a dike has exceeded a length of a_c, an important implication of the above results is that for vertically oriented dikes connected to a source reservoir, the negligible resistance of rock to fracture will allow them to propagate until they either reach a level of neutral buoyancy (LNB) or until the magma within the dike freezes. The rate of propagation will be controlled by the viscous resistance to flow of magma into the dike tip and Eqn (10.19) implies that dikes of highly viscous felsic magma will be considerably thicker than mafic dikes and/or propagate a lower speeds.

7. THE FATE OF DIKES

7.1. The Level of Neutral Buoyancy

For buoyantly ascending feeder dikes, the flow adjusts so that the local viscous pressure gradient, dP_η/dz is balanced by local hydrostatic pressure gradient, dP_ρ/dz, in order to transfer the volume flux, Q, from below. As a result, the dike thickness varies with depth according to the balance between elastic stress, P_e, and excess internal pressure, P_σ. The balance between P_η and P_ρ in theory requires that magma will not ascend beyond the regional **LNB**, the depth above which dP_ρ/dz changes from values of less than zero to greater than zero (Figure 10.6). In the bulbous region near the tip, however, P_η can be neglected and the pressure balance is between P_e and P_ρ. In order to accommodate the volume flux, Q, the head of the dike will inflate thereby increasing P_e and allowing an overshoot beyond the LNB by slow propagation of the dike tip into the upper layer where magma is negatively buoyant (Taisne and Jaupart, 2009).

Eventually, because of magma accumulation and local stress buildup, magma will spread laterally along the LNB, either as vertically oriented dikes or horizontally oriented sills, depending on the stress field. Repeated intrusions may result in the formation of crustal hot zones, wherein

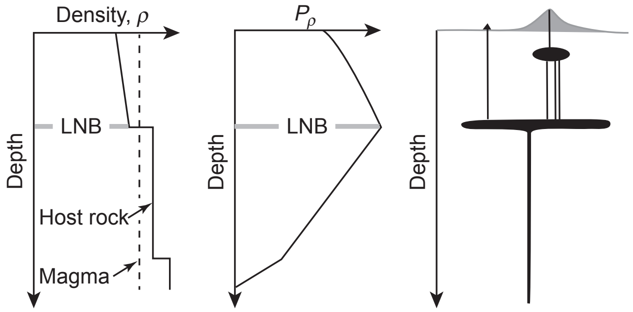


FIGURE 10.6 Schematic illustration of the density ρ (left) and buoyancy pressure P_ρ (center) for a feeder dike at the scale of a density stratified lithosphere (adapted from Lister and Kerr, 1991). The level of neutral buoyancy (LNB) is defined by the change in sign of dP_ρ/dz. At the LNB magma may accumulate and differentiate in "hot zones," from where shallow crustal reservoirs are supplied.

fractional crystallization and assimilation allow magma to evolve toward more felsic compositions and sufficient buoyancy to ascend to shallower levels.

7.2. Ambient Stress Field

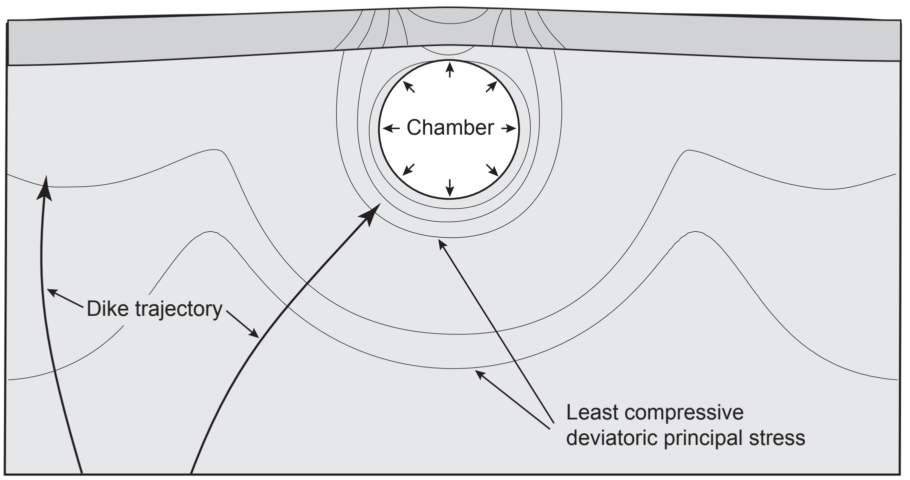
Dikes preferentially intrude in the plane perpendicular to the direction of minimum compressive stress. This behavior can, however, be modulated in the presence of significant variations in fracture toughness of the surrounding rock, for example, due to stratification (Figure 10.7; Maccaferri et al., 2010). It is thus frequently observed that dikes invade such planes of weakness or preexisting fractures, perhaps becoming sills or incipient magma reservoirs in the process. However, if rock fracture is negligible in the overall force balance, intrusion along preexisting structures may offer negligible mechanical advantage and dike orientations will be predominantly determined by the regional and/or local stress fields, with the latter potentially changing abruptly if there is strong stratification. Dikes may thus become arrested, if the local stress field is unfavorable to dike propagation, and magma supply from depth is insufficient for magma accumulation and breakthrough (Gudmundsson, 2006).

The upper crustal stress field gets further modified due to the presence of volcanic edifices at the Earth's surface. As the volcanic edifice grows in size, it generates an increasingly significant compressive stress field that may impede vertical dike propagation, especially for less evolved, more dense magmas (Pinel and Jaupart, 2000). The result may be predominant magma storage at depth. By the same token, for large and shallow reservoirs the edifice load may produce tensile stresses surrounding the magma chamber, which reach a maximum at the top of the chamber. Consequently, dike formation directly beneath the summit may be favored, resulting in the development of a central vent system. In detail, the distribution of compressive and tensile stresses surrounding a magma chamber at depth beneath a volcanic edifice depends on edifice size (height and diameter), as well as chamber size and depth.

7.3. Thermal and Viscous Effects

Thermal effects of magma ascent are perhaps most important during basaltic eruptions, which often are associated with dikes that form linear fissures at the Earth's surface. Whereas small eruptions typically cease within less than a day, larger eruptions tend to become localized into individual vents after sometime. These vents may perhaps become reactivated during subsequent eruptions. This localization is a consequence of the strong dependence of flow rate on dike thickness (Eqn (10.10)) and ensuing thermal and viscous feedbacks (Figure 10.8). Because most of the flow becomes channeled to the thicker sections of the dike, the rate of magma cooling as the magma ascends will

FIGURE 10.7 Schematic illustration of stress field (contours are schematic and not drawn at equal stress increments) associated with an idealized inflated magma reservoir and overlying rock strata (adapted from Gudmundsson, 2006; Karlstrom et al., 2009). Note that dikes within some zone will be "captured" by the magma chamber, as well as the effect of stratification on the stress field.



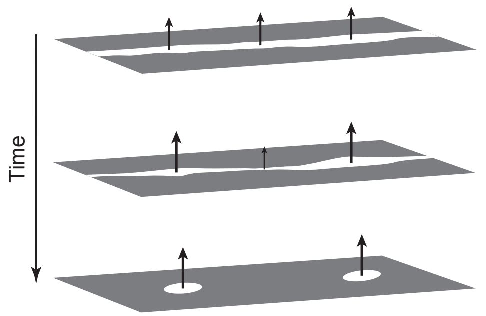


FIGURE 10.8 Schematic diagram illustrating flow localization due to feedbacks between dike thickness, magma flow rate, magma cooling, and magma viscosity (adapted from Bruce and Huppert, 1989). Because magma flow rate is slower in narrower sections of the dike, the ratio of advective heat transport to conductive magma cooling is smaller, resulting in cooler magma, higher viscosity, and, hence, a further decrease in flow rate, as well as magma solidification against dike walls. Ultimately, this may result in the localization of flow into conduit-like pathways, or in the case of eruptive fissures into the localization of the activity into individual vents.

be lowest there and perhaps even result in the melt-back of dike walls. In contrast, sections that are of small thickness will have lower flow rates and faster rates of magma cooling. This will result in increased magma viscosity, which is strongly dependent on temperature, and perhaps magma solidification against the dike walls (Figure 10.9).

Magma cooling and/or solidification and melt-back are a consequence of the conduction of heat from the flowing magma into the dike walls. The degree to which magma cools, perhaps resulting in solidification, depends on the balance between along-dike advective heat transport by the flowing magma (which depends on the magma velocity) and the across-dike diffusion of heat to the conduit walls (which depends on the gradient in temperature across the magma within the **thermal boundary layer** close to the dike walls). The ratio of these two competing heat transfer mechanisms leads to the following dimensionless parameter (Bruce and Huppert, 1989)

$$\Pi \sim \frac{b^4 \Delta P}{\kappa \eta a^2}, \tag{10.20}$$

where $\Delta P/a$ is the average pressure gradient driving magma flow within the dike and κ is the thermal diffusivity. For small Π magma cooling will be dominant and the dike will close by the positive feedback between viscous resistance to magma flow and solidification (Wylie et al., 1999). Cooler magma will have a higher viscosity, which will reduce magma velocity, all else being equal (Figure 10.10). Hence, Π will decrease further, implying a further decrease in advective heat transfer through the dike, relative to conductive heat loss to the dike walls. The result will be flow localization, that is, sections of the dike with small Π tend to close, whereas sections with large Π remain open, perhaps even increasing in width and magma flux due to melt-back of dike walls. The latter occurs when Π is sufficiently large and the magma ascends fast enough for the advective heat flux to maintain a high magma temperature and steep thermal gradients in the magma closest to the dike walls.

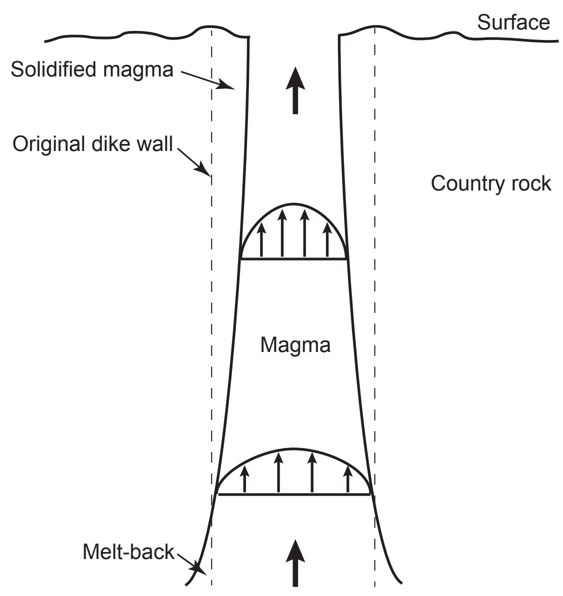


FIGURE 10.9 Schematic diagram illustrating the evolution of a two-dimensional dike undergoing widening due to melt-back near its bottom and narrowing due to solidification of magma against conduit walls (adapted from Bruce and Huppert, 1989).

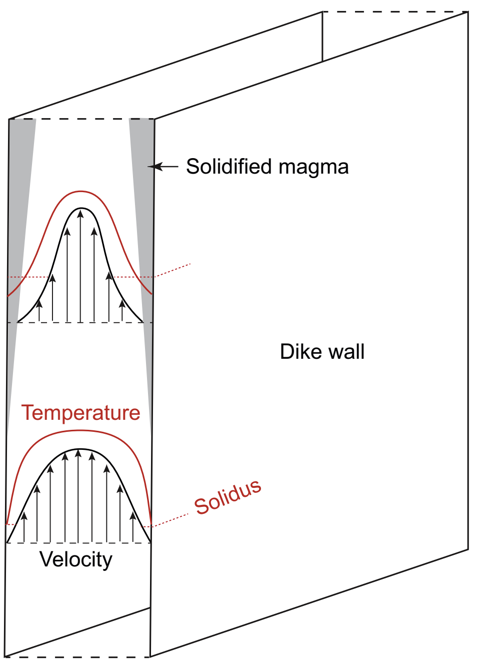


FIGURE 10.10 Schematic diagram illustrating the evolution of the velocity and temperature profiles for a magma with a temperature-dependent viscosity. As the magma cools against the conduit walls, viscosity increases. Consequently, velocity decreases, which reduces the advective heat transport, resulting in a positive feedback between cooling, increasing viscosity, and decreasing velocity (e.g., Wylie and Lister, 1995).

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