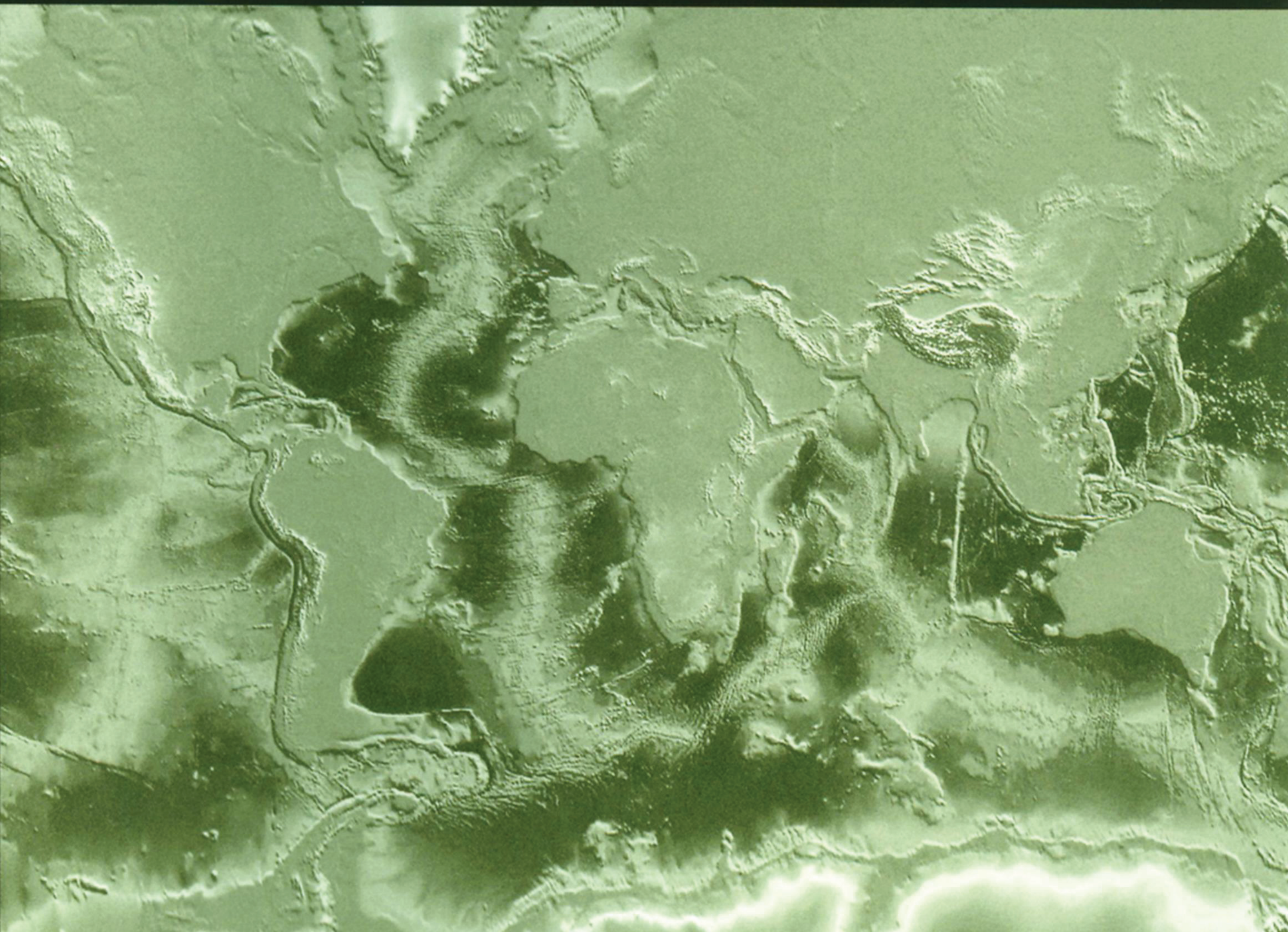
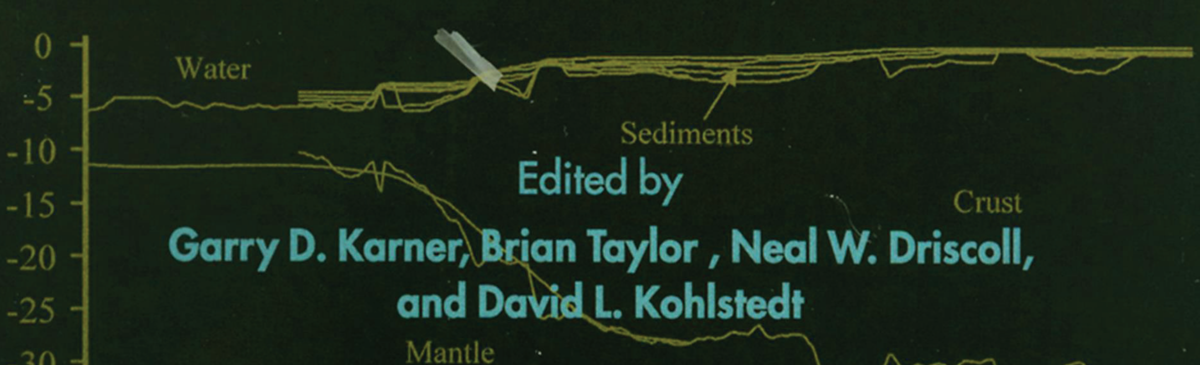
RHEOLOGY AND DEFORMATION OF THE LITHOSPHERE AT CONTINENTAL MARGINS





Edited by
Garry D. Karner, Brian Taylor , Neal W. Driscoll,
and David L. Kohlstedt

Rheology and Deformation of the Lithosphere at Continental Margins

MARGINS Theoretical and Experimental Earth Science Series

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Continental margins are the Earth's principle loci for producing hydrocarbon and metal resources, for earthquake, landslide, volcanic, and climatic hazards, and for the greatest population density. Despite the societal and economic importance of margins, many of the mechanical, fluid, chemical, and biological processes that shape them are poorly understood. Progress is hindered by the sheer scope of the problems and by the spatial-temporal scale and complexities of the processes.

The MARGINS Program (a research initiative supported by the U.S. National Science Foundation) seeks to understand the complex interplay of processes that govern continental margin evolution. The objective is to develop a self-consistent understanding of the processes that are fundamental to margin formation and evolution. The books in the MARGINS series investigate aspects of these active systems as a whole, viewing a margin not so much as a geological entity of divergent, translational, or convergent types but more in terms of a complex physical, chemical, and biological system subject to a variety of influences.

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PREFACE

"Rheology and deformation of the lithosphere at continental margins"

Garry D. Karner, Brian Taylor, Neal W. Driscoll and David L. Kohlstedt

This volume is a collection of papers resulting from presentations made during a four-day short course at the first U.S. MARGINS Theoretical and Experimental Institute (TEI) held January 23–26, 2000. The institute was funded by the National Science Foundation and examined field, laboratory, and modeling constraints on how lithosphere rheology and deformation evolve throughout continental margin evolution. Traditionally, investigations of the rheology and deformation of the lithosphere have taken place at one scale in the laboratory and at entirely different scale in the field; development of an understanding of large-scale processes requires an integrated approach. The long-term objective of the short course and its ensuing publication is to stimulate cross-disciplinary inquiry into the rheology and deformation of the lithosphere. The first day of the short course provided an overview of the setting and nature of deformation at extensional and compressional continental margins. Day two concentrated on: (1) observations supporting, and models explaining, strain partitioning within the crust and lithosphere and (2) numerical and analogue modeling experiments that address the scaling problem of comparing physical experiments with natural systems. Day three focused on laboratory observations related to frictional sliding and crack healing along fault surfaces. Day four was centered on experimental studies of the rheology of crustal and mantle rocks.

The institute significantly influenced the subsequent research objectives and directions of the MARGINS Rupturing Continental Lithosphere (RCL) initiative, which were examined during a two-day workshop that followed the short course. The RCL initiative had as its basic tenet that the mechanisms allowing continental lithosphere to be deformed by weak tectonic forces were not understood, and neither was the manner in which strain was partitioned and magma distributed. These problems were encapsulated by the following themes: (1) the low-stress paradox of lithospheric deformation and (2) strain partitioning of the lithosphere during deformation. A series of papers verified the existence and complexities of the spatial and temporal distribution of strain within deforming lithosphere (chapters 1, 4, and 7: Buck, Davis and Kusznir, and Willett and Pope).

However, the low-stress paradox of lithospheric deformation that figured so prominently in all MARGINS planning documents prior to the TEI was significantly challenged. This paradox relates to the fact that large fault structures (subduction thrusts, major transforms, and perhaps normal detachments) accommodate a major component of strain but move at resolved shear stresses far smaller than those expected to cause failure. In turn, this apparent low-strength property of large faults may be corollary to an even more fundamental issue; namely, the tectonic forces available are insufficient to rupture the continental lithosphere as defined by the integrated yield–stress envelope of the continental lithosphere. Buck (chapter 1) elegantly showed that dike intrusion could reduce the amount of tectonic force required to rift normal continental lithosphere by an order of magnitude below that needed to stretch lithosphere in the absence of dykes.

Active low-angle normal detachments are the extreme case of the weak fault/low-strength paradox. The present debate revolves around whether low-angle faults mapped in such regions as the Whipple and Mormon Mountains of the western United States actually moved at low fault dip angles or moved on high-dipping faults whose footwalls were rotated into the observed field relationships, either in a domino style or by a rolling hinge mechanism. Continental intraplate earthquake focal mechanisms are predominantly related to high-dipping faults. Nevertheless, the megamullion structures of seafloor spreading centers and the geological reconstructions summarized by Axen (chapter 3) for the fault systems of southeastern Nevada, southwestern Utah, and southeastern California appear to require an active period of low-angle normal faulting. The controversy continues.

This same weak fault/low-strength paradox issue was the rationale behind the Ocean Drilling Program drilling (Leg 180) of the Moresby detachment zone in Papua New Guinea, one of the few examples of an active, low-angle ($\sim 30^{o}$) normal fault (Taylor and Huchon 2002). Studies there showed the existence of many meters of talc-chlorite-serpentinite gouge with low coefficients of friction (0.21–0.3; Kopf et al. 2003) within a permeable, porous, and anisotropic fault zone at greater than hydostatic fluid pressures. Scholz and Hanks (chapter 9) effectively dismiss the paradox of the Moresby Detachment in demonstrating that its lock-up angle is consistent with Andersonian failure theory.

The weak fault/low-strength paradox has become entwined with the elastic thickness controversy in which earthquakes in midplate settings rarely occur below 40 km depth, indicating that the physical and chemical conditions prevailing in deeper rocks do not permit them to deform by brittle failure. In support of this observation, the elastic thickness of the continents inferred from free-air gravity, Bouguer gravity, and topography data is typically less than 40 km and less than the local depth to the Moho. In contrast, estimates of flexural loading of the lithosphere require elastic conditions to prevail to depths of 40 to 100 km over time periods of many millions of years. Hence the controversy: how is it possible for the Earth to support loads elastically at great depth and over long periods when the crust fails seismically at shallow depth and at short periods? Directly linked to this controversy is the viability of the yield-stress envelope for continental lithosphere. For many years, laboratory measurements of high-temperature creep of rock-forming minerals has been used to infer that crustal minerals should deform more readily than olivine at the same temperature. This led to the "jelly sandwich" image of a brittle upper crust, a potentially weak ductile lower crust, and a stronger upper mantle. Topography and the distribution of deformation near the Earth's surface concur with this image, at least for regions like the Basin and Range Province and Tibet. To what extent does a jelly sandwich simulate the rheology of continental lithosphere? Jackson (chapter 2) introduces a contentious idea suggesting that the strength of the continental lithosphere resides in its seismogenic layer, which is contained wholly within the crust, and that the continental lithospheric mantle is characterized by a wet rheology and thus is relatively weak. Willett and Pope (chapter 7), via a series of finite-element modeling experiments for the regional and intensive compressional deformation of continental lithosphere (bivergent orogenic edges and orogenic plateaus), offer important insights into the actual rheological behavior of the lithosphere.

In a set of related papers, Ruff and Hyndman (chapters 5 and 6, respectively) characterize the rheology of the zone between interacting converging plates, the seismogenic zone, which is defined by the spatial extent of earthquakes. Their intent is to define the processes controlling the updip and downdip rupture limits of the seismogenic zone. In this environment, the updip fault rheology appears to be dominated by temperature, which in turn controls the onset of seismic behavior via the dehydration of stable sliding smectite clay to stick-slip chlorite/illite, either in overlying sediments or within the fault zone gouge. The downdip limit of the seismogenic zone appears to be a function of the temperature dependence of the slip characteristics, for example, from stick-slip to stable sliding, in the fault zone material and the composition and thus rheology of the material in the overriding plate. Ruff (chapter 5) also attempts to define the controls on the various depths to the seismogenic limit within continental interiors, which seems to require more than just a temperature control.

Having a weak lithospheric mantle appears to be consistent with the laboratory studies reported by Xu et al. (chapter 10) and Evans et al. (chapter 11). Xu et al. investigated the role of melt on the anelastic and plastic properties of partially molten rocks as well as the effect of deformation on the distribution of the melt phase. The melt phase provides short-circuit diffusion paths or melt-rich bands, which aid in the relaxation of stress concentrations. The melt-rich bands are zones of low viscosity and high permeability, which act on a geologic scale to produce a marked anisotropy in seismic properties in addition to profoundly influencing the style of deformation. The link between magmatic processes and lithospheric strength is further explored by Evans et al., who show that rock strength decreases significantly when even a small amount of melt is present. In contrast, Chester et al. (chapter 8) describe the details of the porosity and permeability structure of large-displacement, strike-slip fault zones of the San Andreas system. The damaged zone and fault core are composed of very fine-grained, altered fault rocks in which the relatively permeable damage zone acts as a conduit for fluid flow along the fault and the low-permeability fault core serves as a barrier for cross-fault flow; the fault zone at least within the upper crust is conducive to fluid flow.

As with any professional publication, the quality of the final product is a strong function of the expertise and dedication of reviewers within the earth science community. The volume editors would like to acknowledge the following people for their unselfish donation of time and critical reviews of the various chapters: Rick Allmendinger, Atilla Aydin, Chris Beaumont, Roger Buck, Fred Chester, Reid Cooper, Tim Dixon, Rebecca Dorsey, Georg Dresen, Bob Engdahl, Laurel Goodwin, Greg Hirth, Bill Holt, John Hopper, Roy Hyndman, Barbara John, Dan Lizarralde, Steve Mackwell, Simon Peacock, John Platt, Leigh Royden, Carolyn Ruppel, Ernie Rutter, Dale Sawyer, Chris Scholz, Rick Sibson, Eli Silver, Joann Stock, Olaf Svenningsen, Uri ten Brink, Harold Tobin, Doug Wiens, Colin Williams, and Teng-fong Wong. The volume editors would also like to acknowledge the efforts of Andreas Aichinger and Steffi Rausch of the Hawaii MARGINS Office in organizing the Snowbird MTEI and the dedication, tenacity, and patience of Joan Basher of the Lamont MARGINS Office in bringing this publication to closure. The efforts and contribution from all these people are much appreciated.

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Rheology and Deformation of the Lithosphere at Continental Margins

CHAPTER ONE

Consequences of Asthenospheric Variability on Continental Rifting

W. Roger Buck

Introduction

The earliest ideas about continental drift (Wegener 1929) were based on the observation that the eastern coasts of North and South America matched the shape of the western coasts of Europe and Africa. This implies that the continents somehow split apart. Plate tectonics holds that continental breakup involves rifting the entire lithosphere, the cold outer layer of the earth that is too strong to flow along with the deeper interior.

During the thirty plus years since the acceptance of plate tectonics, much effort has been made to characterize rifts and rifted margins and understand the processes affecting them. One of the clearest messages from such studies is that continental rifts form with a variety of geometries, faulting patterns, and subsidence histories. For example, some rifts are wide, like the Basin and Range Province, and some are narrow, like the Red Sea, (e.g., England 1983). Some areas of apparently narrow rifting, such as metamorphic core complexes, do not subside locally (e.g., Coney and Harms 1979; Davis and Lister 1988), whereas some rifts, like those in East Africa, form deep basins even with modest amounts of extension (e.g., Rosendahl 1987; Ebinger et al. 1989). It has become accepted that the condition of the lithosphere at the time of rifting, its thermal structure and crustal thickness, can have a profound effect on the tectonic development of a rift (e.g., Sonder et al. 1987; Braun and Beaumont 1989; Dunbar and Sawyer 1989; Buck 1991, Bassi 1991).

There has been far more work on the effect of variations in lithospheric, as opposed to asthenospheric conditions, on the evolution of continental rifts. There are several good reasons for this lithospheric emphasis. It is easier to constrain lithospheric conditions by characterizing the geologic history and the geophysical structure of a rift, and the heat flow in adjacent areas. Further, it has taken time to work out the physics of lithospheric stretching and how processes can vary for different initial conditions and rates of extension. Also, lithospheric stretching models have been very successful at explaining many features of rifts and passive margins (e.g., McKenzie 1978).

A growing number of observations have not been explained by lithospheric stretching models. The most obvious observation involves magmatism. Many, possibly most, margins seem to be affected by massive magmatic intrusion and volcanic outflows, even before the onset of faulting and subsidence that mark stretching (e.g., Sengor and Burke 1978). As more data are collected, more margins appear to be "volcanic" (see White and McKenzie 1989; van Wijk and Cloetingh 2002). For example, the North American East Coast was once regarded as a prime example of nonvolcanic passive margin (e.g., Steckler and Watts 1981). Now, seismic data for the offshore area and geologic mapping onshore indicate that as much magma and lava was emplaced along the East Coast as for any margin (Holbrook and Keleman 1993). Models of the opening of the South Atlantic emphasize the effect of lithospheric stretching and detachment faulting (e.g., Etheridge et al. 1989; Lister et al. 1991), but massive piles of volcanic flows are inferred for the South American margin (Hinz 1981, White and McKenzie 1989). Thick volcanic layers are also seen on the 2,000-km-long Greenland margin (e.g., Mutter et al. 1988). The earliest rifting stage of the Red Sea is marked by massive flood basalts at the southern end of the Red Sea (e.g., Menzies et al. 1992) and dike intrusion in the north (Pallister 1987), yet models of Red Sea rifting usually ignore magmatic effects (e.g., Steckler 1985; Wernicke 1985; Martinez and Cochran 1988; Buck et al. 1988; Chery et al. 1992).

There are at least three major problems with "tectonic extension" models that ignore the effects of magmatism and flow of melt-depleted asthenospheric. They can be described in terms of three paradoxes.

- The "Tectonic Force" Paradox. It may take more force to extend thick lithosphere than is available. Stretching models imply faulting of cold upper mantle under rifts in normal lithosphere, but deep earthquakes are not observed in such settings (e.g., Maggi et al. 2000). Magmatic accommodation of extension may be needed to explain rifting in areas of low-to-normal heat flow.
- The "Extra Subsidence" Paradox, also known as the "Upper Plate" Paradox (Driscoll and Karner 1998). Observations indicate that some margins subside more than is predicted by simple models, prompting development of kinematic models with fairly complex patterns of assumed lithospheric stretching (Royden and Keen 1980; Wernicke 1985; Driscoll and Karner 1998). Dynamical models produce such strain patterns only when special patterns of preexisting weakness are assumed (e.g., Dunbar and Sawyer 1989).
- The "No Magma" Paradox. Some margins are amagmatic, even where the crust is highly attenuated or mantle is present at the ocean floor (e.g., at the Iberia Margin, Whitmarsh et al. 1990). This is surprising because stretching implies lithospheric thinning, causing vertical advection and pressure release melting of normal asthenosphere. The melt should be emplaced at or near the surface.

This chapter makes a case that, in addition to variability in lithospheric properties, asthenospheric variability may also be needed to explain observed features of rifts and rifted margins. Specifically, this chapter considers ways that melting of extremely hot (possibly plume related) mantle may resolve the paradoxes of continental rifting listed previously. The three major sections of the chapter deal with a suggested resolution for each paradox.

Magma-Assisted Rifting of Thick Continental Lithosphere (The "Tectonic Force" Paradox)

Both simple analytic or semianalytic rifting models (e.g., McKenzie 1978; Buck 1991) and more complex numerical simulations (e.g., Braun and Beaumont 1989; Bassi 1991; Dunbar and Sawyer 1989) assume that the average stress or tectonic force required to initiate rifting is available. This may not be true, however, for rifting of thick, strong lithosphere in the absence of basaltic magmatism. Several authors have estimated that the tectonic forces likely to be available for rifting is in the range of 3–5 TeraNt/m (Forsyth and Uyeda 1975; Solomon et al. 1980). The tectonic force needed for amagmatic extension of initially thick lithosphere may be up to an order of magnitude greater than that available (Kusznir and Park 1987; Hopper and Buck 1993; and discussion in the next section). Intrusion of basaltic dikes can allow lithosphere to separate at much lower levels of tectonic force than possible without the dikes.

Areas of initially thin lithosphere should rift at relatively low levels of tectonic force. Consistent with this, some areas of high heat flow and initially thick crust, such as the North American Basin and Range Province, did seem to start extending with little or no basaltic volcanism. Models neglecting magmatism do predict the general patterns of observed extensional strain inferred for such areas (Buck 1991). It should be noted that these "hot" weak areas are not typical of continents. The effects of orogenesis, especially thickening of radiogenic crust, have been suggested as a way of heating regions such as the Basin and Range and the Aegean Sea extensional provinces (e.g., Sonder et al. 1987). But, in areas of low-to-normal heat flow, the earliest phase of rifting is often accompanied by basaltic magmatism.

Morgan (1971) noted that huge volumes of flood basalt extruded in many areas of continental breakup. He suggested that such magmatism was associated with active mantle plumes, which somehow triggered rifting. Since that time much work has been done on estimating how plumes could produce extensional stresses by causing regional uplift (Sengor and Burke 1978; Spohn and Schubert 1982; Bott 1991). Great effort has also been made modeling plume temperatures and the magma volumes that result from active plumes or "plume head" upwelling (Richards et al. 1989; Griffiths and Campbell 1990; Hill 1991), and from passive upwelling related to the stretching and thinning of lithosphere (White and McKenzie 1989). However, the mechanical effects of large-scale magma intrusion on lithosphere-scale rifting remain largely unquantified.

This section will suggest the association of magmatism and rifting in areas of low-to-normal heat flow is not coincidental, but that magmatic intrusion may allow rifting to proceed given available tectonic forces. Specifically, it is suggested that: (1) considerable extensional tectonic force may be needed to cause lithosphere-cutting dike intrusion rather than extrusion of large magma volumes; (2) such dike intrusion would accommodate extension plate separation at as little as one tenth the tectonic force needed for amagmatic rifting; (3) after a short period of magma-assisted rifting (from 1 to 10 million years [m.y.] for a reasonable range of extension rates), initially thick lithosphere could be heated and weakened sufficiently to continue extending at a low-force level even without continued intrusion; and (4) the uplift and subsidence patterns of several major rifts and continental margins are more consistent with magma-assisted rifting than with simple lithospheric stretching models. A simple numerical model for estimating temperature and strength changes during rifting with dike intrusion is described.

Tectonic Force for Extension

Separation of lithospheric plates requires extensional stresses at a rift. At any depth those stresses can cause yielding by fault slip, ductile flow, or dike intrusion, whichever takes the least stress (figure 1.1). The extensional state of stress is approximated using the usual assumption that the vertical or z-direction is the largest principal stress and equals the lithostatic stress (Anderson 1951) given by

$$\sigma_1(z) = g\int_0^z \rho_r(z')\,dz' \tag{1.1}$$

where g is the acceleration of gravity and ρ_r is the density of rock in the lithosphere. In the crust the assumed density is 2,800 kg/m^3; and in the mantle, density is 3,300 kg/m^3.

Dikes are magma intrusions with a thickness much smaller than their width or length. Molten basalt is assumed to be the material filling rift-related dikes, because mantle melting can produce basaltic magma and more felsic dikes might be too high in viscosity to propagate easily. Dikes should form in planes perpendicular to the least principal stress, σ_3; for a rift, this should be in vertical planes parallel to the rift (Anderson 1951). It is assumed that preexisting vertical fractures are prevalent to avoid the complications of fracture mechanisms (e.g., Rubin and Pollard 1987). However, the extra stress needed to break open dikes in unbroken rock should be limited by the rock tensile strength, which would make a small contribution to the tectonic forces estimated here. Neglected also are the viscous stresses associated with the flow of magma in a dike, since the goal is to estimate the minimum stress difference (defined as $\sigma_1 - \sigma_3$, where σ_1 is the maximum principal stress) required to have magma stop and freeze at a given depth in a dike.

Before freezing, magma in a dike can cease moving up or down when the static pressure in the dike equals the horizontal stress at the dike wall (Lister and

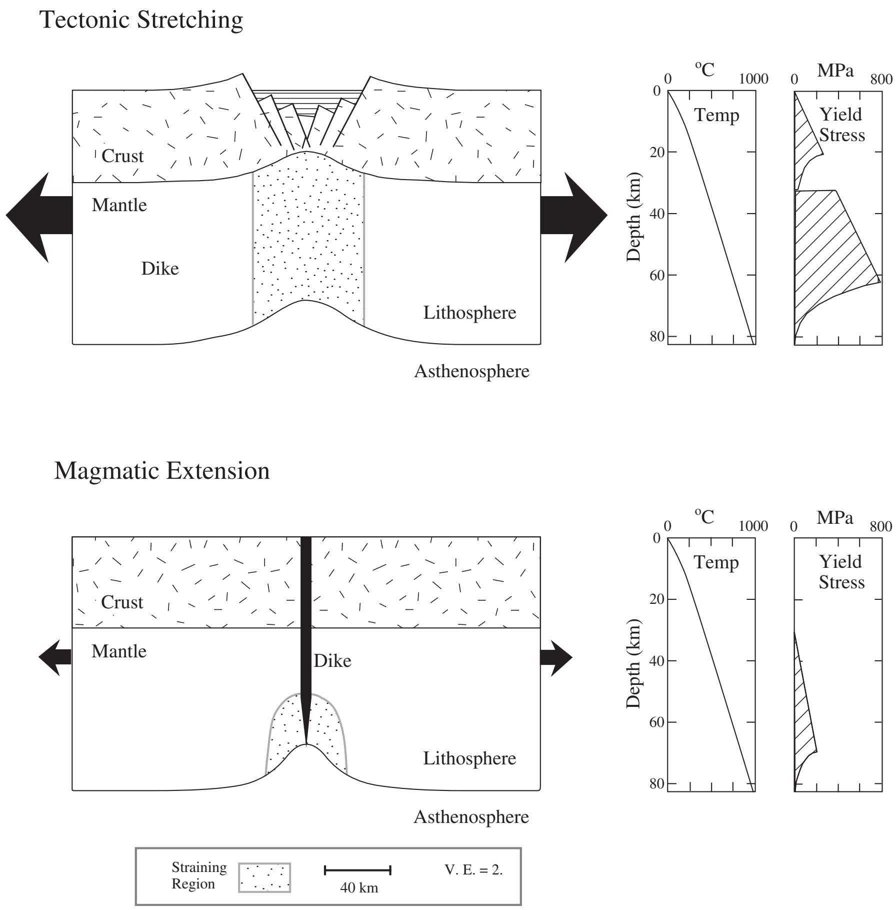
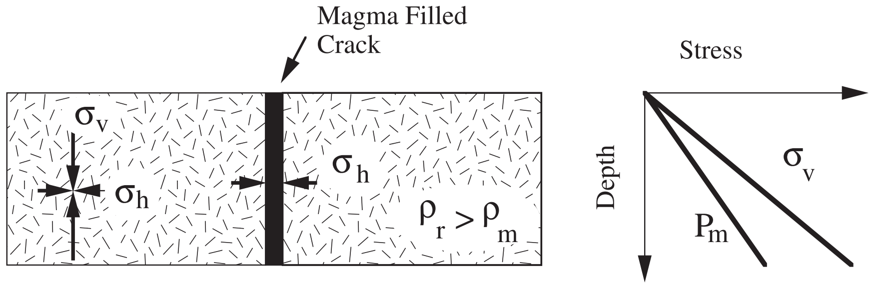


Figure 1.1 Schematic of the difference between extension of thick lithosphere with and without magmatic intrusion. Note the large difference in the yield stress, the stress difference needed to get extensional separation of two lithospheric blocks.

Kerr 1991). The vertical pressure variation in a static column of magma is related to its specific weight, ρ_m, so for magma emplacement: $\partial\sigma_3/\partial z = \rho_m g$. To specify the level of magma pressure, it is assumed that dikes always cut to the surface, where the pressure is zero. In that case the stress difference required for dike emplacement is

$$\sigma_m(z) = \sigma_1(z) - g\rho_m z \tag{1.2}$$

where ρ_m is the magma density, taken to equal 2,700 kg/m^3. Clearly, with these simplifications, the stress difference for magma to allow extensional separation between blocks of lithosphere depends only on the density difference between the lithosphere and magma (figure 1.2). Since the crustal density and magma density are taken to be equal, the stress difference for crustal diking is zero. In a mantle of density 3,200 kg/m^3, the stress difference required for intrusion increases at a



$$\sigma_v(z) = \rho_r\, g\, z$$

$$\sigma_h(z) = P_m = \rho_m\, g\, z$$

Figure 1.2 The stress distribution when extensional separation of two lithospheric blocks is accommodated by magmatic intrusion. Here the lithospheric rock density is ρ_r, greater than the magma density ρ_m. The horizontal stress σ_m equals the pressure in the dike, while the vertical stress is taken to be the overburden.

rate of 5 MPa/km of depth into the mantle. If the mantle is too weak to maintain such stresses, then the magma cannot be emplaced at depth and will be extruded.

At high temperature, rocks can flow in response to stress differences without forming macroscopic fractures. For such ductile flow the stress difference, σ_d, and strain rate, $\dot{\varepsilon}$, are found to be related through a flow law:

$$\sigma_d \;=\; (\dot{\varepsilon}/A)^{1/n}\, \exp(E/nRT) \tag{1.3}$$

where T is absolute temperature, R is the universal gas constant, E is activation energy (e.g., Goetze and Evans 1979), and A is a constant for given material. The ductile yield stress depends on the composition of rock as well as temperature. Dry anorthite rheology is assumed for the crust and a dry olivine rheology for the mantle. For anorthite, $E = 238$ kJ mol^{-1}, $A = 5.6 \times 10^{-23}$ Pa^{-n} s^{-1}, and $n = 3.2$; for olivine, $E = 500$ kJ mol^{-1}, $A = 1.0 \times 10^{-15}$ Pa^{-n} s^{-1}, and $n = 3$ (Kirby and Kronenberg 1987).

Following Brace and Kohlstedt (1980), the stress difference needed for normal faulting is estimated under the assumption that cohesionless fractures exist in all directions to accommodate fault slip. The yield stress for faulting is

$$\sigma_f(z) \;=\; B(\sigma_1(z) \;-\; P_P(z)) \tag{1.4}$$

where $B = 2f/[(1 + f^2)^{1/2} + f]$, where f is the coefficient of friction. Assuming $f = 0.85$, which is the average friction coefficient for a wide range of rocks (Byerlee 1978), makes the constant $B = 0.8$. The pore pressure in the rock, P_p, is taken to be hydrostatic.

To estimate the stress difference for extension (the yield stress) as a function of depth, z, we must specify the temperature profile through the lithosphere. This is done by assuming temperatures are in steady state with a constant heat flow from below, radioactive heat production within the crust, and a given heat flow at the surface. The thermal conductivity is set to 2.5 W m^{-1} °C^{-1} for the crust and 3.0 W m^{-1} °C^{-1} for the mantle. The crustal heat production is set to 3.3×10^{-7} W m^{-3}, which contributes 10 mW m^{-2} to the surface heat flow for a 30-km-thick crust. The mantle heat flow is adjusted to provide a given surface heat flow for a specific crustal thickness.

Figure 1.3 shows yield stress profiles for a moderate heat-flow temperature profile, assuming a 30-km-thick crust. The dashed lines show the model yield stress if no magma is available to accommodate extension; the solid line shows the situation if enough magma is available to just reach the surface. If an intermediate amount of magma were supplied in this conceptual model, it would be emplaced at depth, while extension near the surface was accomplished by faulting (e.g., Rubin and Pollard 1987). In some sense the dashed profile for amagmatic stretching can be seen as an upper limit on the yield stresses and the solid line as a lower limit.

The horizontal force per unit length required to cause extensional yielding of the entire model lithosphere, F_{ys}, is estimated by integrating yield stress over depth (figure 1.4). This force depends strongly on the temperature profile and, thus, on the surface heat flow as well as the magma supply. To extend continental lithosphere with a heat flow of about 40 mW/m^2, as is seen adjacent to some rifts like the Red Sea (Martinez and Cochran 1988), may require as much as 30 TeraNt/m of tectonic force if no magma were intruded. Extending the same lithosphere with copious magma may take almost an order of magnitude less force.

Two situations lead to extrusion of magma. First, if the tectonic force is too small to allow lithosphere-cutting dikes, then magma should be extruded along with dikes of small lateral extent. In that case, magma-assisted rifting cannot occur. This

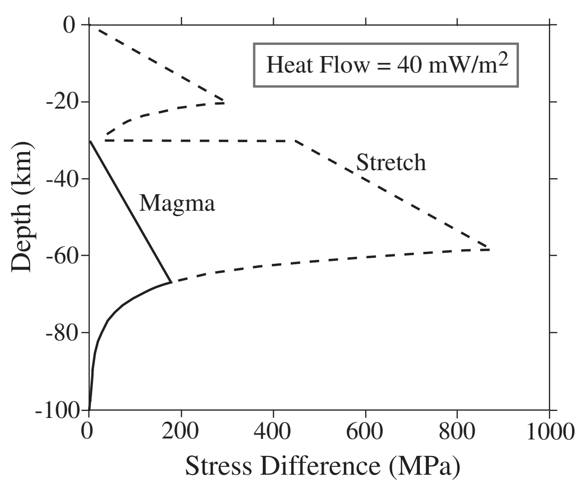


Figure 1.3 Example of yield stresses for a strain rate of 10^{-14} s^{-1} for 30-km-thick crust with a thermal profile derived (as described in the text) for a surface heat flow of 40 mW/m^2. The solid line shows the stress difference for magmatic rifting and the dashed line shows the yield stress for tectonic stretching.

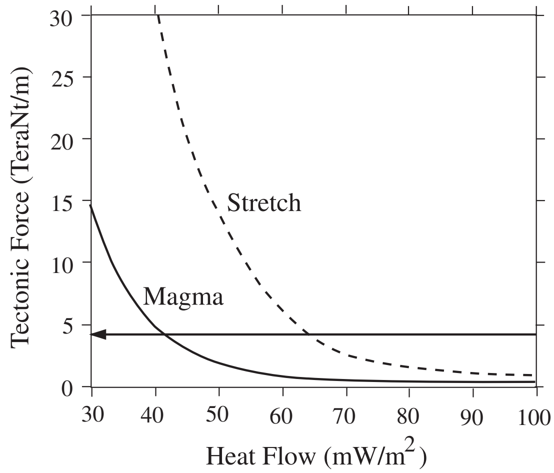


Figure 1.4 Tectonic force for extension either with or without magma as a function of the surface heat flow for a crustal thickness of 30 km. The tectonic force is the result of integrating yield stress envelopes such as those shown in figure 3. The horizontal bold line is the estimated value of plate extensional driving forces.

may be the situation for most ocean island basalts and some continental flood basalts, such as the Columbia River flood basalts. The other interesting case is if the tectonic force is great enough for large-scale diking, but the rate of extension requires less magma than is supplied. In this case, extrusion should occur on top of an area of rifting, as may have occurred in many areas discussed by White and McKenzie (1989), such as the rifting of Greenland from Norway, Madagascar from India, the East Coast of North America from Africa, South Africa from Antarctica.

The amount of basaltic magma available to facilitate rifting may vary with distance along some rifts. There is strong seismic and geodetic evidence from the active extensional plate boundary in Iceland that dikes propagate at least 60 km from central volcanoes (Einarsson and Brandsdottir 1980). These dikes can be intruded at depth with surface normal faulting and no accompanying extrusion of lava (Trygvasson 1984). During dike intrusion sequences there appears to be significant extrusion near the central volcano, while there is no extrusion far from the volcano (Trygvasson 1984). It is possible that a similar pattern occurs on a larger scale for some continental rifts. For example, the rifting of Arabia from Africa coincides with copious volcanism along the southern Red Sea coast (Menzies et al. 1992) while there is on-land evidence of a few dikes and little volcanism in the northern Red Sea region (Pallister 1987). More dikes may have intruded at depth in the northern Red Sea than made it to the surface.

Magma-Assisted Rift Evolution

To begin to relate these ideas about magma emplacement to observed characteristics of rifts and margins, a simple model is used to estimate the time evolution of rift-zone temperature, crustal composition, and lithospheric strength given a large flux of basaltic magma. The key assumption is that magma is intruded as dikes only at depths where the lithosphere is strong enough to hold the magma down for a given density structure. Magma is intruded at a temperature of 1,200°C and the latent heat of fusion adds another 300°C effective initial temperature. Also, it is assumed that magma is intruded at the rift center where the lithosphere is thinnest. Thus, the models differ from those of Royden et al. (1980), who investigated the thermal effect of an arbitrary distribution of intruded magma into stretching lithosphere and did not consider evolution of strength.

In these simple two-dimensional thermal models we are not concerned with whether the magma comes from below the rift or by lateral flow along the rift. The axis of the rift is considered to be a line of divergence so that lithosphere moves horizontally away from the line of dike intrusion (figure 1.5). Here, the lithosphere is defined as any material at a depth where initial rift temperatures were less than 1,200°C. Between the depth where lithospheric stresses are large enough for magma emplacement and the base of the lithosphere, we assume that plate separation occurs by distributed pure shear. The width of the pure shearing region is taken to equal its thickness.

The initial temperature field varies only in the vertical direction, with the profile derived from the same steady-state model parameters described earlier. Given the flow field and temperature structure we compute the time evolution of the temperature field using a standard finite difference scheme (e.g., Buck et al. 1988). As the temperature field changes, the ductile yield stresses should change, and so should the depth range where magma can be emplaced.

Specifying the depth of the transition from dike intrusion to ductile flow depends on estimating the ductile yield stress. Estimating this stress is not straightforward, since it depends on the temperature and strain rate fields. The common assumption used to estimate yield stresses is that the strain rate is uniform with depth and over the region of extension. This situation may never really be obtained in a rift and certainly should not be the case when there are lateral temperature

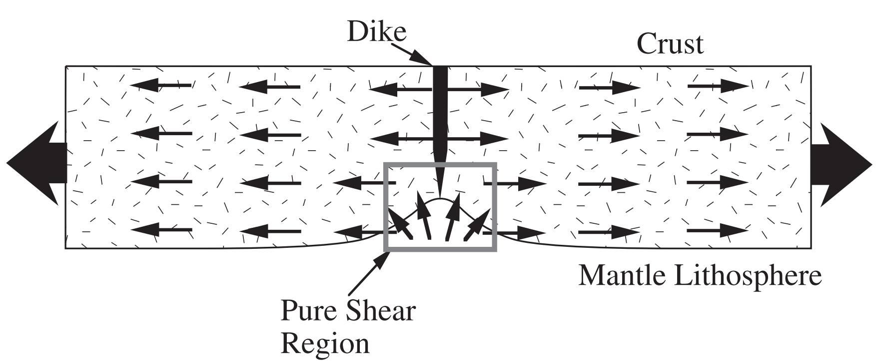


Figure 1.5 Illustration of the velocity field assumed for a thermal model of magma-assisted rifting. Pure shear is assumed in the region where ductile stresses are too low to allow magma emplacement.

variations; instead the hottest area should be the weakest and should thus strain the fastest.

Short of solving the full two-dimensional equilibrium equations for elastic, plastic, and viscous deformation, the following approximations are made. The stress at any depth is taken to be constant. For thermally activated creep and laterally varying temperatures, the strain rates must be greatest where the temperatures are greatest. For simplicity, all components of the strain rate tensor are neglected except the horizontal normal one, and so $\dot{\varepsilon}$ in equation (1.2) is replaced with $\dot{\varepsilon}_{xx}$. The rifting velocity must equal the integral of $\dot{\varepsilon}_{xx}$ across the rift, so the stress difference required to give the assigned rifting velocity u_r for a given temperature field $T(x,z)$ is

$$\sigma_d(z) = \left| \frac{u_r}{2A\int_0^{W/2} \exp\left(\frac{E}{RT(x,z)}\right)dx} \right|^{\frac{1}{n}} \tag{1.5}$$

where W, the width for integration, equals the lithospheric thickness.

Using the stress difference from equation (1.5) to calculate the maximum depth of dike intrusion, the flow field is adjusted as described previously. Temperature changes due to extensional flow in the lower lithosphere and dike intrusion at shallower levels are computed at every time step. At any time, the yield stress can be integrated over depth to estimate the tectonic force needed to continue extension.

Results for Magma-Assisted Rifting

We are interested in how rifting and magma intrusion weakens the lithosphere, since we want to consider whether a rift may continue to extend for a given regional tectonic force even if the magma supply is reduced. Therefore, through a calculation in which magma supply is sufficient to be intruded at all possible depths, we calculate the yield stress as if the magma were suddenly "shut off."

Figures 1.6–1.8 show results from one calculation of the evolution of magma-assisted rifting. Here, the half-rifting velocity was taken to be 1 mm/yr, the initial surface heat flow, $Q_s = 40$ mW m^{-2}, and the crustal thickness was 30 km. Figure 1.6 shows the model isotherms after 10 m.y. of extension. At the start of rifting, magma was emplaced down to ~60 km depth; by the end of the calculation magma was being emplaced only within the crust, because the mantle was too hot and weak to retain magma.

Figure 1.7 shows yield-stress profiles calculated at million-year intervals. The integrated yield stresses through time, both with and without magma supply, are shown in Figure 1.8. After approximately 5 m.y. into the calculation, the tectonic force for amagmatic extension has dropped to the level required to initially achieve

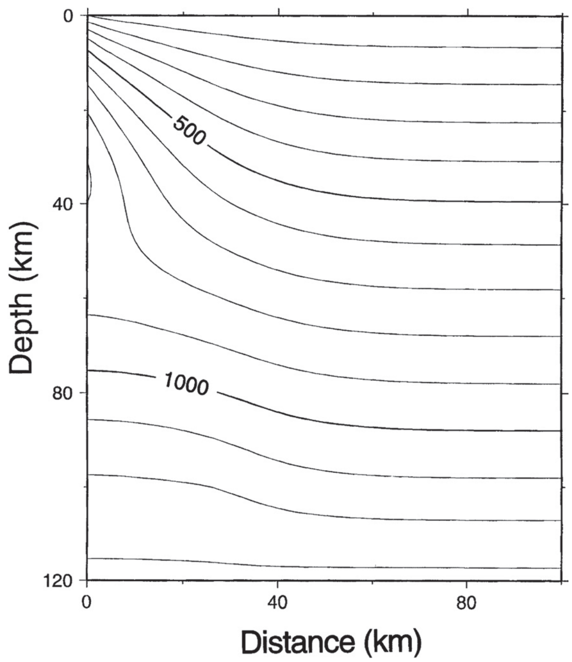


Figure 1.6 Contours of temperature across half a rift zone after 10 m.y. of magmatic extension at a rate of 0.1 cm/yr. The initial crustal thickness was 30 km and the initial surface heat flow was 40 mW/m^2.

magmatic extension. The amagmatic (stretching) force is only weakly dependent on rifting velocity, so when an area has sufficiently weakened, the rate of extension might well increase. In contrast, the rate of magmatic extension depends mainly on the rate of magma supply.

It should be noted that only end-member models have been considered: either no magmatic intrusion, or enough to intrude at all depths where the lithosphere is strong. Other possibilities clearly exist. If the tectonic force is intermediate between the amount needed for these end members, then magma should be intruded over a reduced depth range, with strain at shallow depths accommodated by either elastic deformation, fissure opening, or fault offset (Rubin and Pollard 1987). Also, the portion of the lithosphere stretching tectonically might increase as the lithosphere thins.

The main potential objection to the idea that magma intrusion may be needed to allow rifting thick lithosphere is that our estimate of the stretching strength, based on the approach of Brace and Kohlstedt (1980), may be too high. There is evidence that for large-offset thrust faults, and possibly strike-slip faults, the stress for faulting the shallow lithosphere may be much less than the frictional stress assumed here. Usually, high pore fluid pressures are taken to be the cause of low stress brittle deformation (e.g., Hubbert and Rubey 1959). However, it is not clear whether areas of stretching can be characterized by low-stress brittle deformation, since pore pressures should be decreased by extension.

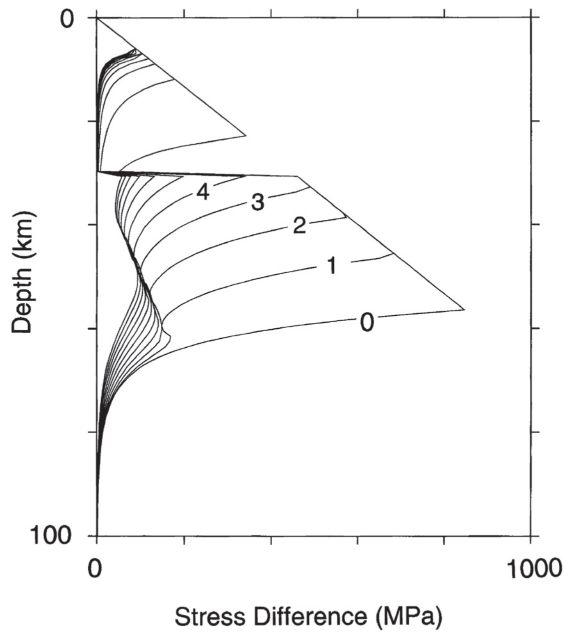


Figure 1.7 The tectonic yield stress as a function of depth for 1 m.y. time steps in the calculation illustrated in figure 6. The curves show the yield stresses needed for continued extension if the magma supply were suddenly cut off at the given time.

Comparison with Observations

Various effects of magmatic intrusion into rifts may be observable. The most direct observation would involve seismic imaging of basaltic bodies at depth. This is particularly challenging if there was little volcanism and most magma was intruded at tens of kilometers depth. A related challenge is that the seismic velocity of basalts may be only slightly greater than that for typical continental crust (see Kelemen and Holbrook 1995).

Another important effect to consider is the pattern of subsidence and uplift across rifts. Magmatic accommodation of extension should result in less subsidence than tectonic stretching of continental lithosphere, as discussed next.

The most promising places to test magmatic rifting models may be at the distal ends of young rifts and margins that are clearly affected by magmatism. The younger the rift or margin the better the chance that the early magmatic and tectonic history can be resolved. Therefore, we focus most of the present discussion on the northern sections of the 2,000-km-long Red Sea rift.

The Gulf of Suez is one of the best-characterized recent continental rifts in an area of low (~40 mW/m^2) heat flow. This rift is a part of the Red Sea rift system that ceased most extension when the Aquaba-Dead Sea oblique rift/transform developed 12–13 Ma (LePichon and Gaulier 1988). Dikes intruded this region beginning about 35 Ma (Pallister 1987; Dixon et al. 1989). However, rapid subsidence and rift shoulder uplift did not begin until after 25 Ma (Jarrige et al. 1990;

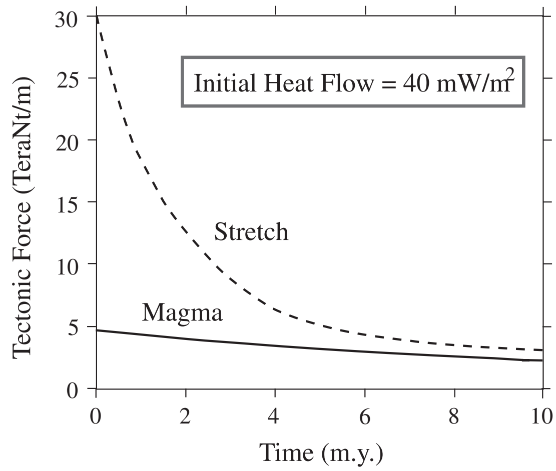


Figure 1.8 Evolution of the extensional force needed for tectonic and magmatic extension in the model of magmatic extension shown in figure 6. The curve labeled tectonic can be thought of as the force needed for continued extension if the magma supply were suddenly cut off. The curve labeled magmatic assumes that enough magma to form dikes that reach the surface is present at all times.

Omar et al. 1989; Omar and Steckler 1995). The magma-assisted rifting model may explain some observed features for the northern Red Sea and the Gulf of Suez rifts, long considered good examples of passive, essentially amagmatic, rifting (Steckler 1985; Martinez and Cochran 1988).

Seismic Structure

One would expect that a similar amount of basalt might have been intruded during the early phase of northern Red Sea rifting as may have intruded into the Gulf of Suez. The northern Red Sea has undergone much more extensional widening than the Gulf of Suez and so shows far greater average subsidence (Martinez and Cochran 1988). Any intrusives may be harder to image in the Red Sea because of the greater bathymetric relief and the greater thickness of salt in that region (Gaulier et al. 1988). Also, one would be searching for a rather subtle difference in seismic velocity and crustal thickness structure. The Gulf of Suez may never have reached the phase of large-magnitude tectonic subsidence that shaped the present day Red Sea. For the less extended Gulf of Suez, the difference in the crustal structure predicted by a pure tectonic stretching model, as opposed to a magma-assisted model, might be more readily resolved.

The calculation illustrated in figures 1.6 through 1.9, constructed with the early history of the Gulf of Suez and northern Red Sea in mind, allows rifting at low initial tectonic stresses. According to the model presented here, the average crustal thickness might have changed little during the rifting that produced the Gulf of Suez. This may seem contradictory to the observation of large tectonic faults that

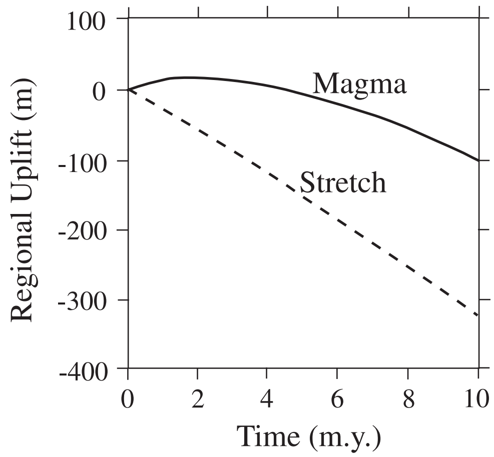


Figure 1.9 Comparison of the predicted average regional isostatic elevation changes with time for two rift models. Density changes are taken to affect the elevation over a region of width equal to the initial thermal lithospheric thickness. The solid line is for the model of magma-assisted rifting described in the text, and the dashed line is for a model of pure shear necking over a region as wide as the initial thickness of the lithosphere.

account for kilometers of near-surface brittle stretching. However, the intrusion of basalt may have occurred at greater depths, even well into the mantle at the start of rifting. If we consider the intruded basalt to be part of the crust, then during the early phase of magma-assisted rifting the average regional crustal thickness could increase. As the lithosphere is weakened by intrusive heating, a greater proportion of extension might have been accommodated by tectonic stretching, producing the observed slip on faults.

Subsidence/Uplift

Tectonic stretching (without magmatism) also may change the average density and so the elevation of a region. Continental lithosphere can be thought of as hot mantle replacing crust and cold mantle. Thinning the compositionally low-density crust causes subsidence, whereas thinning the thermally dense lithosphere causes initial uplift. The total initial and long-term effect of stretching typical continental lithosphere should be regional subsidence (McKenzie 1978). Regional, rather than local, elevation must be considered, since the lithosphere maintains finite strength during rifting.

Solidified basaltic magma is less dense than mantle, so basalt intrusion can affect the average density and the isostatic elevation of a region. The emplacement of large quantities of basalt into a rift can accommodate extension with little or no crustal thinning. In fact the intrusion of basalt into the mantle can effectively thicken the crust. So dike intrusion can lessen the initial amount of subsidence or even lead to regional uplift.

Figure 1.9 compares the average isostatic elevation through time for magma-assisted rifting with that predicted by a standard stretching model. The uplift or subsidence is calculated assuming that at 0°C crust and basalt both have a density of 2,800 kg/m^3, while mantle has a density of 3,300 kg/m^3 at the same temperature. The temperature field computed during rifting is related to the density field using a thermal expansion coefficient of 3.5×10^{-5} °C^{-1}. Density changes related to crustal thinning, basalt intrusion, and temperature changes are integrated over depth, $D = 150$ km, and over a 100-km-wide region of the center of the rift. Decreases or increases in the weight of the rift region cause uplift or subsidence, respectively, because the region is taken to float on hot mantle asthenosphere with a density, ρ_a, of 3,285 kg/m^3. Formally, the elevation change equals $D\Delta\rho/\rho_a$ where $\Delta\rho$ is the average density change of the rift region. Figure 1.9 shows potentially observable differences between the tectonic stretching and the magma-assisted rifting models.

Some continental margins such as the Bay of Biscay seem to fit the general predictions of the stretching model (LePichon and Sibuet 1981). However, Royden and Keen (1980) showed that the subsidence history recorded in wells on the passive margin off the Canadian East Coast do not fit the McKenzie (1978) stretching predictions. These data require less initial tectonic subsidence (related to crustal thinning) relative to the long-term thermal subsidence. The magma-intrusion model gives less tectonic subsidence than the stretching model, as shown in figure 1.9. The subsequent thermal subsidence (not shown here) for comparable amounts of extension is affected little by the magma intrusion. Thus, this model predicts subsidence patterns that are consistent with the general trend of the data analyzed by Royden and Keen (1980).

Steckler (1985) showed that the Gulf of Suez does not match the predictions of the tectonic stretching model. He analyzed the tectonic subsidence in the rift and the surrounding rift-shoulder uplift and found that the average present-day regional elevation is close to zero: the volume of the uplifted rift shoulders approximately equals the volume of the subsided gulf basin, after corrections are made for loading of basin sediments (see figure 1.10). Tectonic stretching should produce long-term average regional subsidence. Steckler (1985) explained the lack of such subsidence in terms of a convective input of heat. However, the lack of large magnitude regional subsidence across the Gulf of Suez is consistent with the injection of significant quantities of magma into the rift in the early stages of extension.

Other failed rifts may have extended when basaltic magma was being intruded. One candidate for magma-assisted rifting is the Mesozoic Dnieper-Donets Basin of southern Ukraine, where well data indicate very little tectonic phase subsidence, but large-magnitude thermal subsidence (Starostenko et al. 1999)

Magmatic input may be necessary for the active rifting seen in several areas of presumed thick lithosphere, including the Rhinegraben, the Baikal Rift, the Rio Grande Rift, and along parts of the East African Rift. The Rhinegraben cuts a region of northwest Europe characterized by normal heat flow, averaging about 40 mW m^{-2} (Illies and Greiner 1978). As argued previously, such heat flow may indicate very large lithospheric strength in extension. Volcanism is contempora-

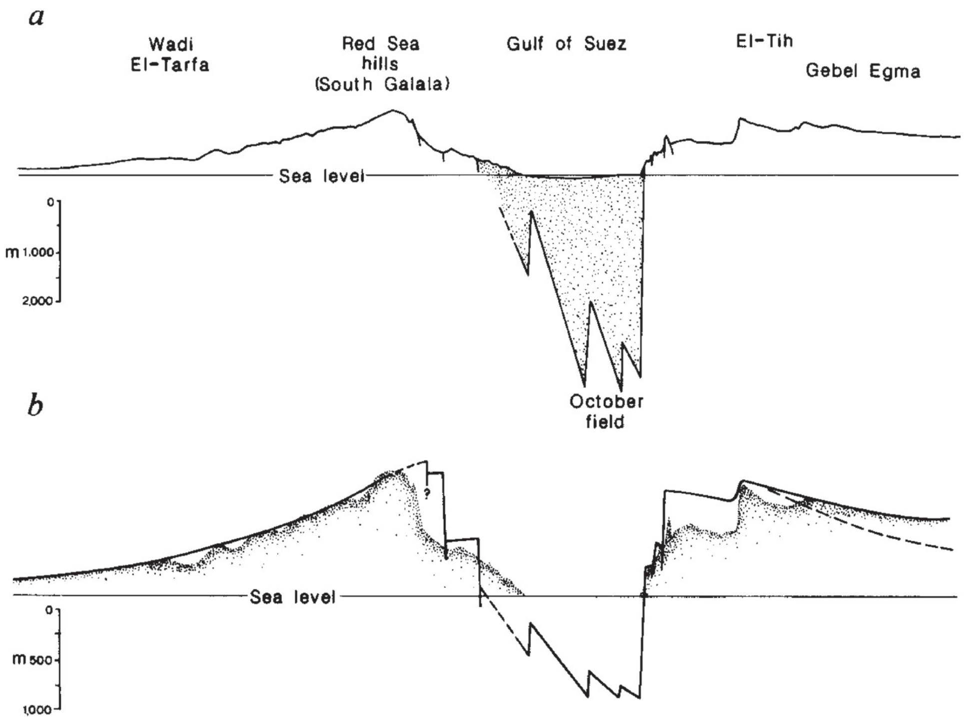


Figure 1.10 (a) Topography and basement relief for a transect across the central part of the Gulf of Suez Rift. (b) Shows the topography modified by the effect of flexural sediment unloading (from Steckler 1985). This shows that the net subsidence averaged across the rift is close to zero, since the uplifted flanks nearly balance the down-dropped center of the rift. The average elevation of the rift plus flanks is positive, but this may reflect the fact that the region outside the flanks has a positive elevation of ~500 m.

neous with the rifting along the Rhinegraben (Illies and Greiner 1978). The region around Baikal, another Cenozoic rift, is also about 40 mW m^{-2} (Morgan 1982). Seismic surveys across this Siberian rift show evidence for a ~10-km-thick layer at the base of the rifted crust with a seismic velocity consistent with basaltic "underplating" (Zorin 1981). The East African rift and the Rio Grande Rift also cut areas with near-normal heat flow (Morgan 1982) and parts of these rifts are characterized by recently active volcanoes (Ebinger et al. 1989, Mohr 1992).

Dynamic Subsidence of Passive Margins (The "Extra Subsidence" or "Upper Plate" Paradox)

Many margins show more subsidence after the early "tectonic" phase than is predicted by uniform pure shear stretching of typical crust and mantle lithosphere (see Lister et al. 1986; Driscoll and Karner 1998). For example, analysis of deep-well data for the Atlantic margin of Canada shows extra "thermal phase" subsidence after a phase of assumed tectonic subsidence (Royden and Keen 1980).

Observations similar to those described here led several workers to suggest that the geometric pattern of lithospheric extension is significantly more complex than uniform pure shear. Royden and Keen (1980) proposed a "two-layer" stretching model in which the mantle lithosphere stretched more than the crust. An alternative model to explain subsidence with little near-surface extension (stretching) is the "simple shear" model (e.g., Wernicke 1985). The idea is that a lithosphere cutting low-angle fault or shear zone accommodates much of the strain during rifting. Vertical sections through parts of the side of the rift above the shear zone, called the upper plate, would experience little crustal thinning but large amounts of mantle lithosphere thinning. The simple shear model had the added appeal that it could explain the topographic asymmetry seen across many conjugate margins (e.g., Lister et al. 1986). To explain a subsidence event for the Exmouth Plateau, off N.W. Australia, that does not seem to involve upper crustal extension, Driscoll and Karner (1998) proposed an extreme variant on the simple shear model. To explain the Exmouth subsidence they called for several hundred kilometers of offset between upper crustal thinning and lower crustal/mantle lithospheric thinning.

Numerical models of lithospheric stretching that treat the evolution of mechanical strength during rifting (e.g., Braun and Beaumont 1989; Bassi 1991; Chery et al. 1992) tend to show fairly symmetric patterns of deformation that are similar to the necking pattern seen for laboratory necking of metal rods. The necking strain predicted by such dynamical numerical models is similar to that predicted by kinematic pure shear models, if the width of pure shear necking equals the thickness of strong lithosphere. Dynamical numerical models that predict very asymmetric strain patterns generally assume preexisting laterally offset regions of crust and mantle strength (e.g., Dunbar and Sawyer 1989). Dynamical models with strain weakening applied to extension of initially symmetric lithosphere do produce asymmetric fault patterns, but typically only on the scale of the upper crust (e.g., Buck and Poliakov 1998; Lavier et al. 2000). If highly asymmetric lithospheric deformation is common during rifting, then one must assume large-scale and large-magnitude prerift asymmetric weak zones in the lithosphere.

In at least one site where simple shear lithospheric stretching was suggested to explain rift asymmetries, subsequently collected data contradicted that suggestion. The strong topographic asymmetry across the Red Sea rift, with the Arabian side ~500 m higher than the Egyptian side, had made this rift system a prime example of a possible simple shear rift (Wernicke 1985). Cochran et al. (1993) did detailed transects of heat-flow measurements across the northern Red Sea, an ~20 million year old rift that has opened ~100 km. Thermal models of pure shear stretching and simple shear extension were compared with the data (Buck et al. 1988; Martinez and Cochran 1988). This showed that the pure shear models fit the heat-flow and subsidence data in the rift. The simple shear models fit neither the heat flow nor the asymmetric topography of that rift.

Another potential problem with the simple shear model is that in some cases both sides of a conjugate margin look like "upper plate margins" in that they show more long-term subsidence than can be explained in terms of the estimated local crustal stretching (Driscoll and Karner 1998). This has been dubbed the "upper plate paradox" by Driscoll and Karner (1998).

Driscoll and Karner (1998) conducted a detailed study of a particularly clear example of subsidence that cannot be explained by uniform pure shear lithospheric stretching: the Exmouth Plateau. They used seismic reflection lines and deep-well data to look at the tectonic and subsidence history of the ~300-km-wide Exmouth Plateau margin. The plateau was tectonically extended and faulted during the Middle Jurassic. This rifting event did not result in seafloor spreading but did produce observable subsidence, because sediment filled in the area to about sea level. Another rifting event affected the plateau during the early Cretaceous, and seafloor spreading commenced adjacent to the western side of the plateau. The interior of the plateau shows no evidence of tectonic extension, because the sediments deposited since the first rifting event are not faulted. However, the plateau subsided at least a kilometer and, where loaded by new sediments, subsided even further.

To explain the large plateau subsidence that begins during seafloor spreading adjacent to the plateau, Driscoll and Karner (1998) assumed upper crustal thinning on the west side of the plateau and lower crustal thinning under the entire plateau. In their kinematic model, strain was distributed through the shear flow within the presumably weak lower crust. Many workers have argued that weak lower crust can flow under areas of high heat flow like the Basin and Range Province of western North America (e.g., Gans 1987; Block and Royden 1990; Bird 1991). In the model of Buck (1991) the change from core complex style extension to Basin and Range style extension is due to a decrease in the rate of lower crustal flow. The rate of lower crustal flow is related to its viscosity, which should be a function of temperature. High-temperature crust should have a low viscosity and would easily flow.

Other areas where the lower crust may flow easily are high-elevation plateaus like the Altiplano and Tibet. The correlation between elevation and crustal flow is probably not a coincidence. These plateaus are high because the crust is thick. The base of thick crust can be very hot even with moderate heat flow and so temperature gradients. The active part of the Basin and Range has an average elevation of nearly 2 km while the crustal thickness there is only about 30 km (~average or below average for continents). The higher-than-normal elevation is likely to reflect very high temperatures below the region. This is consistent with the very high surface heat flow (~100 mW/m^2) (e.g., Lachenbruch and Sass 1978). Thus, it is not surprising that the lower crust in the Basin and Range is hot and therefore can flow.

It is much harder to explain how the lower crust of the Exmouth Plateau was hot enough to shear easily when rifting and seafloor spreading occurred. The main problem is that the region was close to sea level at that time (Driscoll and Karner 1998). The crustal thickness then, according to these authors, would have been

~5 km thicker than its present average thickness of 20–25 km. To explain the sea level elevation would require very low temperatures below the plateau and so very low thermal gradients. It is therefore hard to understand how the lower crust under the Exmouth Plateau would have been hot enough to shear at high rates during a rifting event. In the following section, it is hypothesized that the lateral flow of low-density mantle asthenosphere may cause the "extra" subsidence seen on some margins.

Assume that a local area of thin lithosphere exists in the vicinity of a large volume of anomalously hot mantle (figure 1.11). The hot mantle might have

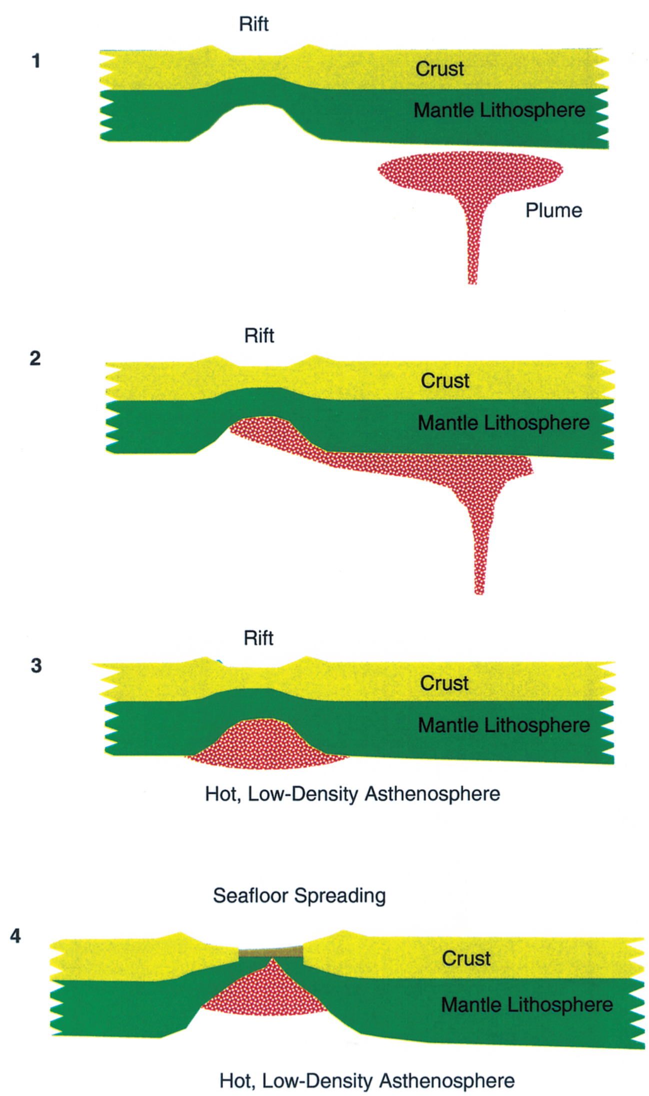


Figure 1.11 Schematic of the pooling of low-density, plume-related asthenosphere under an area of thin lithosphere.

been delivered to the shallow asthenosphere by a plume or plume head (e.g., Campbell and Griffith 1990), though its genesis is of little consequence to this discussion. The hot mantle should be much lower in density than surrounding "normal" asthenospheric mantle both due to thermal expansion and possibly due to depletion on partial melting (Oxburgh and Parmentier 1977). This hot asthenosphere should then pond beneath the area of thin lithosphere. The hot, low-density asthenosphere displaces the dense asthenosphere below the thin lithosphere, resulting in uplift of that area.

To estimate the amount of uplift produced by such ponding we need to know the density and thickness of the hot mantle. The elevation, e_0, will be related to the normal mantle density, ρ_{NM}, the anomalously hot mantle density, ρ_{AM}, as

$$e_0 = (H_L - H_R)(\rho_{NM} - \rho_{AM})/(\rho_{AM} - \rho_W) \tag{1.6}$$

(1.6) assumes the elevation is submarine with a water density, ρ_W, and that $(H_L - H_R)$ is the thickness of the region of anomalous low-density mantle. H_L equals the thickness of the lithosphere around a region with thinner lithosphere of thickness H_R (figure 1.12).

One may get some idea of the uplift produced by hot asthenosphere by looking at the anomalous depth of midocean ridges that are affected by mantle plumes. A prime example is the Reykjanes Ridge south of Iceland. The crustal thickness is about 8 km (Ritzert and Jacoby 1985), about average for midocean ridges, but the water depth is only about 1000 m (e.g., Talwani et al. 1971). The usual depth to ridges is about 3,000 m (e.g., Small 1998). Thus, we can attribute 2,000 m of elevation to low-density asthenosphere below the northern Reykjanes Ridge. If the hot, depleted mantle layer were 200 km thick, then it would have to be just 23 kg/m^3 less dense than normal mantle to explain the anomalous depth of the Reykjanes Ridge. This assumes the density of normal mantle $\rho_{NM} = 3{,}300$ kg/m^3 and $\rho_W = 1{,}000$ kg/m^3. Such a layer thickness is not inconsistent with recent models of plume-ridge interactions (e.g., Sleep 1990; Ribe et al. 1995; Ito et al. 1996). If the layer of hot asthenosphere is thinner, its density has to be lower to explain the ridge depth.

Part of the density anomaly is likely to be due to depletion of the hot mantle caused by pressure release melting on ascent. The material density will become progressively smaller as more melt is extracted at shallower depths of melting (e.g., Oxburgh and Parmentier 1977; Klein and Langmuir 1987). Thus, even if material pooled beneath an area of thin lithosphere cools, it can still remain lower in density, and so positively buoyant, compared with normal mantle. If the area of thin lithosphere is rifted, the buoyant asthenosphere could pour into the area of extremely thin lithosphere where seafloor spreading is beginning (figures 1.11 and 1.12).

The lateral flow of low-density asthenosphere causes subsidence in the region where it was pooled, because it is replaced by mantle of normal density. The rate of subsidence is related to the velocity at which rifting or seafloor spreading occurs, u_p, and on the width of the lithosphere (figure 1.12). It also depends on the

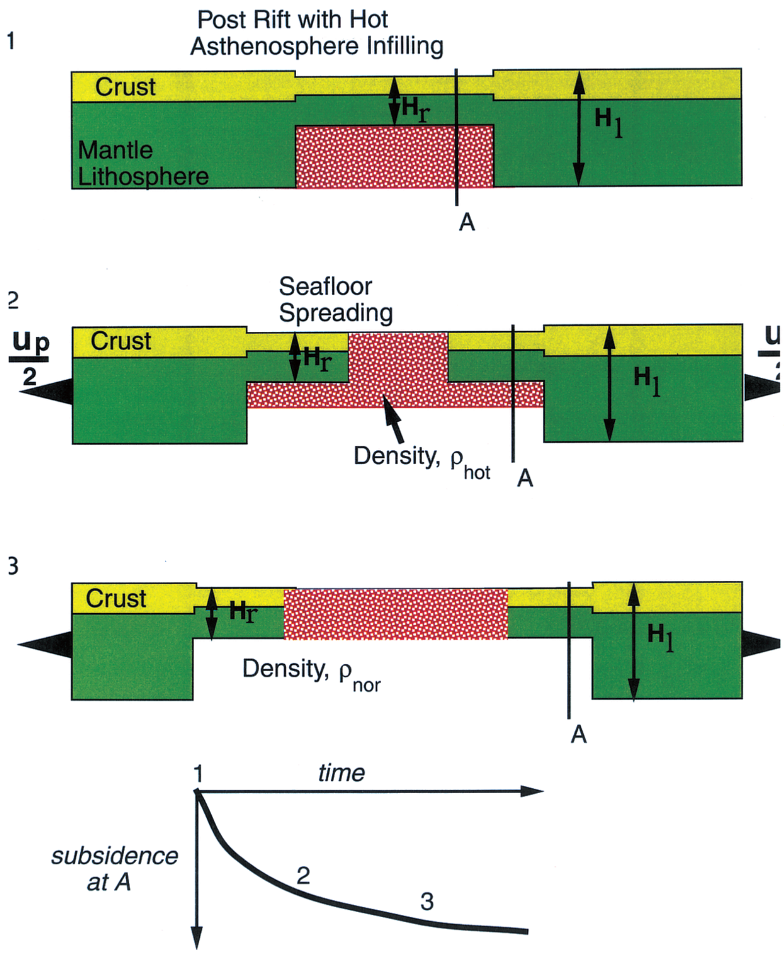


Figure 1.12 Geometry assumed to calculate the effect of flow of a low-density asthenospheric layer on subsidence of a rift.

depth to the base of the ponded layer, H_L, and of course, on the density contrast with normal mantle. The local isostatic subsidence (the change in elevation of a point in the region of pooled lithosphere) due to outflow is given by

$$\frac{e(t)}{e_0} = \left[\left(\frac{1}{1+u_p t/W}\right) - \frac{H_r}{H_L}\right] \Big/ \left(1 - \frac{H_R}{H_L}\right) \tag{1.7}$$

Figure 1.13 shows the predicted "extra subsidence" due to flow of low-density asthenosphere. The maximum subsidence equals e_0 and depends on the initial thickness of the depleted layer and on the density contrast. The time needed for the entire depleted layer to flow out from under the initially thin lithosphere, and for the related subsidence to cease, equals $(W/u_p)[(H_L/H_R) - 1]$. For $(H_L/H_R) = 3.33$, $W = 100$ km, and $u_p = 1$ cm/yr, the time for total outflow is about 23 m.y. This case is one of the two shown in figure 1.13. The other case had an initially

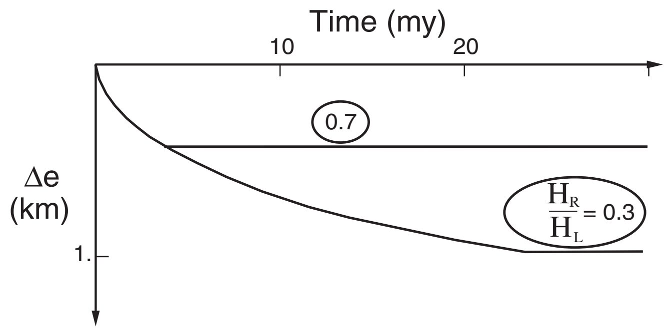


Figure 1.13 Results of analytic model calculation of dynamic subsidence using equation (1.7). The two curves correspond to cases with the same model parameters except that the ratio of the initial depth to top and bottom of the layer, H_R/H_L, are 0.3 and 0.7, as labeled. The initial width of thin lithosphere, W, is 100 km; the plate-spreading velocity, u_p, is 1 cm/yr; $H_L = 150$ km; and the density difference between anomalous and normal mantle is 25 kg/m^3.

thinner depleted layer and so subsided less and took less time to have all depleted material flow out.

Viscous Flow of Depleted Layer

Up to now, it has been assumed that flow in the depleted layer is fast enough to keep the base of the layer flat. One would expect that the flow rate is limited by the viscosity of the layer. If the viscosity were extremely great then the layer would not flow out in a geologically observable time. The pressure-driven flow in a broad, relatively thin viscous layer can be approximated by one-dimensional channel flow, making it easy to relate asthenospheric viscosity to the time for layer thinning.

The viscous layer thins as it flows into the space created by seafloor spreading (figure 1.12). To first order, the thickness with time and distance from the site of seafloor spreading obeys a diffusion relation. The effective diffusivity for the thinning of the layer can be estimated in the way often done for flow-related thinning of viscous lower crust (see Bird 1991; Buck 1991). It is assumed that the top and bottom boundaries of the layer can be described as "no slip." This is probably reasonable for the top boundary, but may not describe the bottom boundary where underlying mantle may flow easily. Changing the bottom boundary condition to free slip would decrease the estimated time for flow and thinning by a factor of four. Since an order-of-magnitude estimate of parameters is needed here, more complex boundary conditions will not be considered.

The flow is taken to be driven by pressure gradients that arise due to local isostatic compensation of the lateral density variations associated with layer thickness variations. Then the effective flow diffusivity is