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a0005

6.09 The Dynamics of Continental Breakup and Extension

W. R. Buck, Columbia University, Palisades, NY, USA

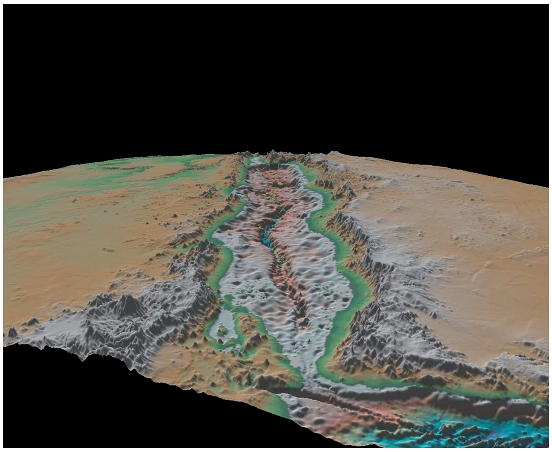
AU2,3 s0005

6.09.1 Introduction

p0005 The earliest ideas about continental drift, which eventually lead to plate tectonics, 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 (b0775 Wegener, 1929). This implies that the continents somehow break apart.

p0010 We now understand that plate tectonics involves the creation of new plates at spreading centers and the return of those plates to the Earth's interior at subduction zones. No terrestrial spreading centers have been active for more than 200 My and spreading centers are constantly being subducted, as the Chile Rise is now being lost beneath South America. Without the development of new spreading centers, plate tectonics might cease. New spreading centers primarily develop either by the splitting of intact plates or when regions of plate convergence begin to extend.

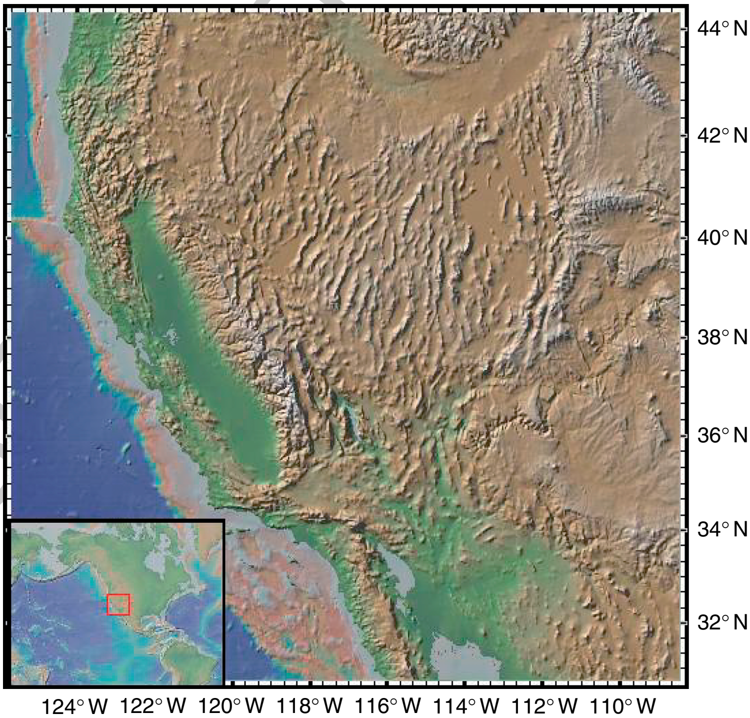
Both plate splitting and extension of convergent regions occur most often in continental, and not p0015



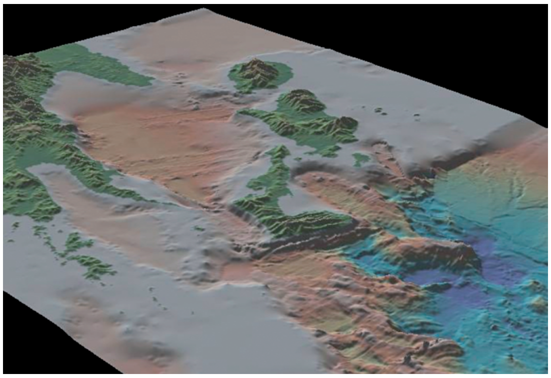
f0005 **Figure 1** Shaded topographic relief image of the Red Sea viewed from the south with a vertical exaggeration of 20. The green to off-white transition marks the boundary between above and below sealevel. The highlands of the Ethiopian Plateau and Yemen Highlands are in the foreground along with the low-elevation Afar Triangle. The Gulf of Suez is just visible in the distance (made with GeoMapApp).

oceanic, lithosphere. Rifts are places where the breakup process either did not or has not yet progressed to seafloor spreading. Thus, they give us a snapshot of the early phase of continental extension. Passive margins (sometimes called rifted margins) are places where extended and/or intruded continent grades the ocean basin formed by postrift spreading, and they give us some constaints on the state of the crust and lithosphere when breakup ended.

Over the 30-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. The pattern of faults, magmatic intrusive and extrusive constructs, uplifted rift flanks, and sediment-filled basins are used to reconstruct what happened during the extension of continents. One of the clearest messages from such observations is that continental rifts form with a variety of geometries, faulting patterns, and subsidence histories. For example, some rifts are narrow, like the Red Sea, (e.g., [b0165] Cochran, 1983a; see **Figure 1**), and some are wide, like the Basin and Range Province (e.g., [b0700] Stewart, 1978; see **Figure 2**). Some areas of apparently narrow rifting, such as metamorphic core complexes, do not subside locally (e.g., [b0180] Coney and Harms, 1984; [b0205] Davis and Lister, 1988; see **Figure 3**), while some rifts, like those in East Africa, form deep basins even with modest amounts of extension (e.g., [b0580] Rosendahl, 1987; [b0250] Ebinger *et al.*, 1989; see **Figure 4**). It has become accepted that the condition of the p0020



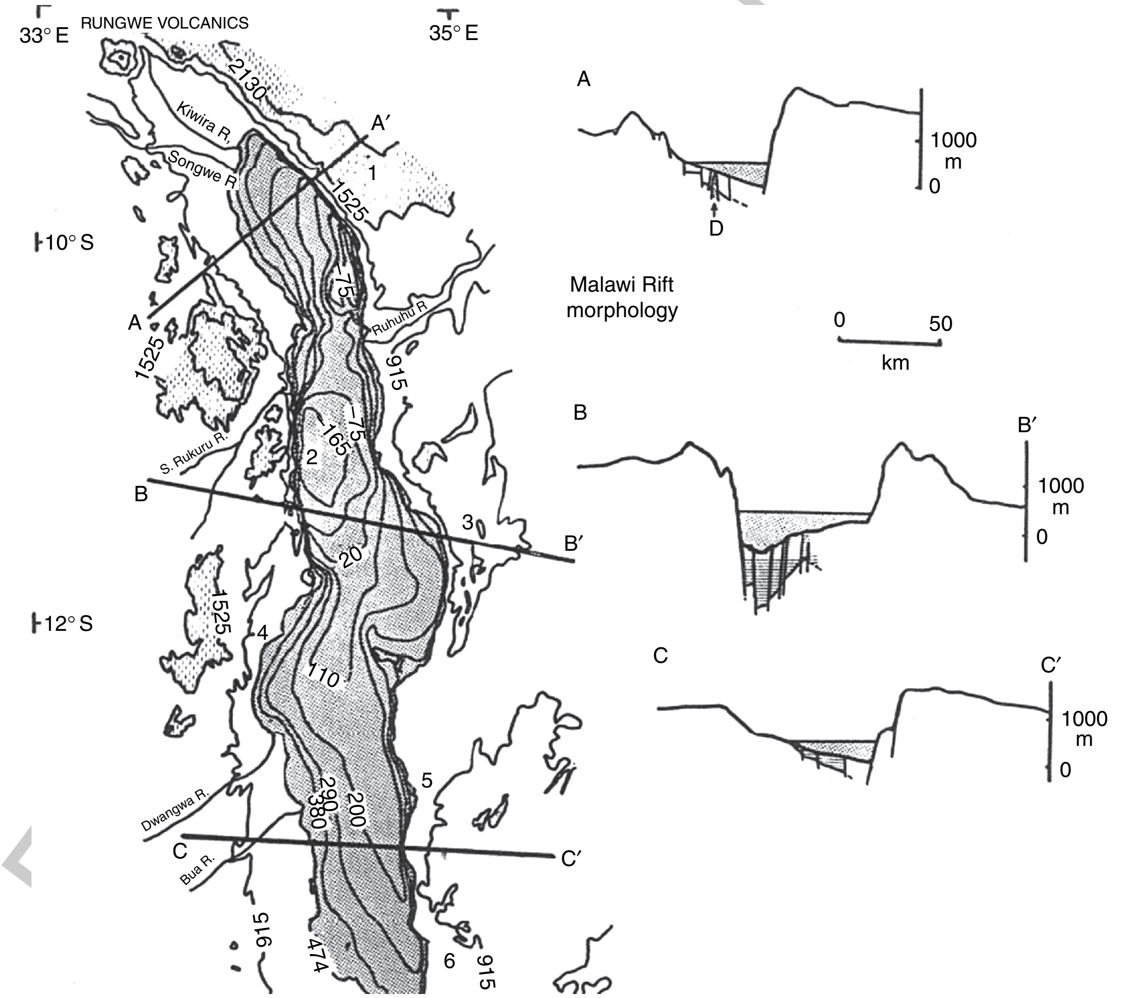
f0010 **Figure 2** Shaded topographic relief image of Western North America shown in plain view. The Basin and Range Province is bounded by the Sierra Nevada on the west and the Colorado Plateau on the east (made with GeoMapApp).



f0015 **Figure 3** Shaded topographic relief image of the D′ Entrecasteau Islands of the Woodlark Basin. These islands are thought to be the youngest continental metamorphic core complexes on Earth (made with GeoMapApp).

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).

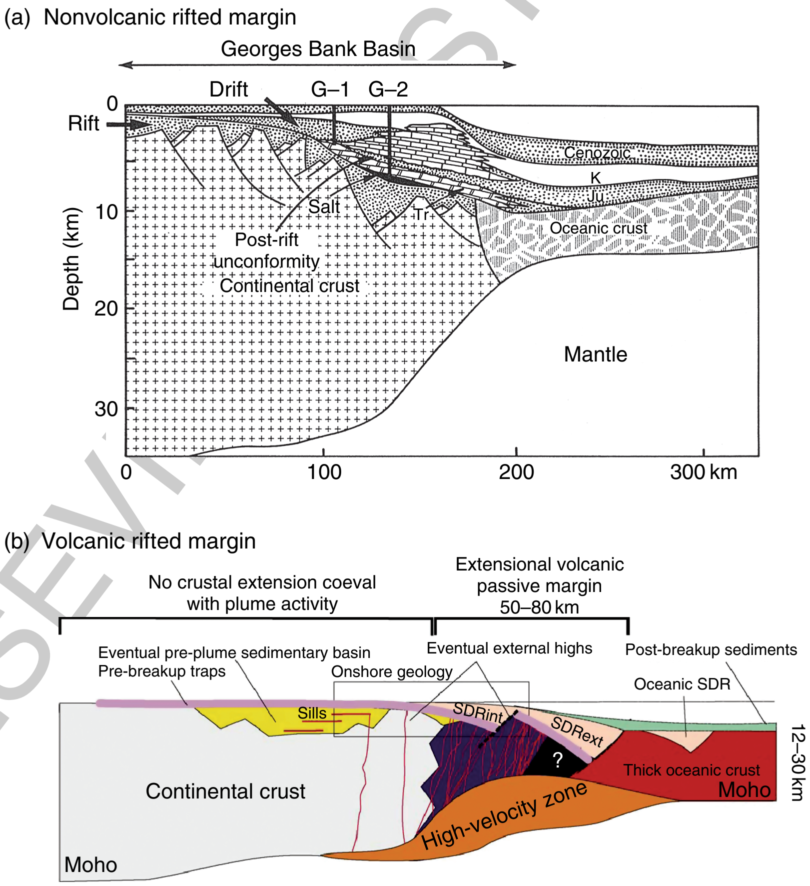
Several decades ago, it was recognized that a few rifts and rifted margins were affected by copius magmatism, such as large parts of the East African Rift or the East Greenland Margin while many others appeared to have been little affected by magmatism, such as the Northern Red Sea or the margin of the US East Coast (Sengor and Burke, 1978). Simple kinematic models of lithospheric 'stretching' (e.g., McKenzie, 1978) reproduce the gross pattern of subsidence across rifts and margins. Thus, most models of rifting, including complex numerical ones, treat p0025



f0020 **Figure 4** Topographic relief and profiles across the Malawi Rift of the East African Rift System (from Ebinger *et al*. (1987)).
AU23 Note the change in polarity of the half-graben comprising the rift.

tectonic stretching and neglect magmatic processes. However, as better data has been collected in rifts (e.g., Ethiopian Rift) and across rifted margins (e.g., AU4 Hinz, 1981; Holbrook *et al.*, 199*) the importance of magmatism has become increasingly clear (**Figure 5**). A number of margins that were formerly thought to be nonvolcanic, like much of the South American–African Margin, were affected by huge volumes of pre- and synrift volcanism and magmatic intrusion. Seismic observations, and in some cases drilling have been used to identify these volcanic packages since they are buried by thick layers of sediment. Also, many nonvolcanic rifts and margins are along-strike continuations of volcanic rifts. This raises the question of whether some of the nonvolcanic sections were affected by magmatism early in their development.

p0030 The issues of the initial state of the lithosphere and the influence of magmatism on rift development are central themes of this chapter. These broad questions necessitate that we investigate the possible meaning of smaller-scale structures such as faults and the mechanics of magmatic intrusion, as well as a plethora of observations concerning rift structure and rifted margin history. Given the great variety of rifts and the wealth of data relating to their formation, it is not surprising that controversies have arisen. In this short chapter only a few of these controversies and only a small fraction of the observations relating to them can be discussed. Among the questions that will be touched on are: (1) Do some extensional (i.e., normal faults) slip with low dip angles (i.e., less than 30°)? (2) Are wide rifts formed by slow extensions or due to the extension of hot, weak lithosphere? (3) Can nonorogenic areas rift without massive magmatic input? Before discussing these controversies an overview of some of the processes that may affect rifting is given.



f0025 **Figure 5** Schematic cross-sections of representative (a) nonvolcanic (from Manspeiserand Cousminier, 1988) and (b) volcanic margins (from Geoffroy, 2005). SDRint and SDRext refer to internal and external seaward dipping reflectors. The high-velocity zone may represent mafic intrusives.

s0010 6.09.2 Processes Affecting the Dynamics of Continental Extension

p0035 A simple way to study the mechanics of rifting builds on kinematic models. The initial pattern of lithospheric extension is specified and the changes in the force needed to continue extension are estimated. If the force decreases, then extension should stay localized where it initiated. If the force increases, then it should be easier for the locus of extension to migrate laterally, resulting in a wide rift. The advantage of this approach is that many parameter combinations can be tested in a short time. The disadvantage is that underlying assumptions and approximations may not always be valid.

p0040 Two properties of the lithosphere control how it deforms in response to applied tectonic forces and to the addition of material such as magma. One is the intrinsic strength or yield stress of the lithospheric material. The yield stress is the applied stress required to produce strain at a given rate. The other property is the density distribution in the lithosphere. The density distribution, including topographic relief, controls the gravitational stresses that must be overcome to deform the lithosphere.

p0045 A number of processes can affect the yield strength and the density distribution of the lithosphere. Models of rifting describe one or more of these processes and how they interact to produce the kinds of rifts that are observed. It is convenient to think of these processes in terms of ones that promote or impede localized deformation of the lithosphere. To do this requires definition of a reference lithospheric state so that the effect of processes can be looked at in light of how much they change the extensional tectonic force needed to continue rifting. The reference lithosphere is taken to be laterally uniform in composition (i.e., crustal thickness) and in thermal structure (so isotherms are horizontal). Since the first infinitesimal increment of deformation does no work against gravity the reference tectonic force for extension involves only the intrinsic strength of the lithosphere.

s0015 6.09.2.1 Tectonic Force for Extension

p0050 Separation of lithospheric plates requires extensional stresses. At any depth those stresses can cause yielding by fault slip, ductile flow, or magmatic dike intrusion, whichever takes the least stress. Here we assume that the reference state does not involve magmatism and so defer discussion of magma intrusion. The difference between the maximum and minimum principal stresses required to cause rock to deform at a given depth, the yield stress, is taken often to be the lesser of two stresses: the stress to produce brittle failure or the stress to cause ductile flow at a specified strain rate.

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 p0055

$$\sigma_1(z) = g\int_0^z \rho_r(z')\,dz' \qquad [1]$$

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 2800 kg m^{-3}; and in the mantle, density is 3300 kg m^{-3}.

At low temperatures and moderate confining pressure, rocks can break on faults. Continued application of stress can cause slip on those faults. If one component of the horizontal stress is the minimum stress (σ_3) then dip-slip faults should form with a normal sense of slip. Following Brace and Kohlstedt (1980), the stress difference ($\sigma_1 - \sigma_3$) needed for normal faulting is estimated under the assumption that cohesionless fractures exist in all directions to accommodate fault slip. Then the minimum stress difference for faulting, the yield stress, is p0060

$$\sigma_f(z) = B(\sigma_1(z) - P_p) \qquad [2]$$

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

If the pore pressure in the rock, P_p, is taken to be hydrostatic then the faulting or brittle yield stress should increase linearly with depth in a constant density lithosphere: p0065

$$\sigma_f(z) = Cz \qquad [3]$$

For crust with a constant density of 2800 kg m^{-3} the gradient of increase of the faulting yield stress with depth, C, is 14.1 MPa km^{-1}. p0070

At high temperature, rocks can flow in response to stress differences without forming macroscopic fractures. For such ductile flow the stress difference for ductile flow, σ_d, and strain rate, $\dot{\varepsilon}$, are found to be related through a flow law: p0075

$$\sigma_d = (\dot{\varepsilon}/A)^{1/n}\exp(E/nRT) \qquad [4]$$

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, $E = 238\,\mathrm{kJ\,mol^{-1}}$, $A = 5.6 \times 10^{-23}\ \mathrm{Pa^{-n}\ s^{-1}}$, and $n = 3.2$; for olivine, $E = 500\,\mathrm{kJ\,mol^{-1}}$, $A = 1.0 \times 10^{-15}\ \mathrm{Pa^{-n}\ s^{-1}}$, and $n = 3$ (Kirby and Kronenberg, 1987).

p0080 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 and radioactive heat production within the crust. For the illustrative examples shown here the thermal conductivity is set to $2.5\,\mathrm{W\,m^{-1}\,°C^{-1}}$ for the crust and $3.0\,\mathrm{W\,m^{-1}\,°C^{-1}}$ for the mantle. The crustal heat production is set to $3.3 \times 10^{-7}\ \mathrm{W\,m^{-3}}$, which contributes $10\,\mathrm{mW\,m^{-2}}$ 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 6(a)** shows yield stress profiles for a moderate heat flow temperature profile, assuming a 30 km thick crust.

p0085 The horizontal force per unit length required to cause tectonic extensional yielding of the entire model lithosphere, F_T, is estimated by integrating yield stress over depth (**Figure 6(b)**). This force depends strongly on the temperature profile and, thus, on the surface heat flow. To extend continental lithosphere with a heat flow of about $40\,\mathrm{mW\,m^{-2}}$, as is seen adjacent to some rifts like the Red Sea (Martinez and Cochran, 1988), may require as much as $30\ \mathrm{Tera\,Nt\,m^{-1}}$ of tectonic force if no magma were intruded.

p0090 A number of factors can affect where extensional strain is concentrated in a continental region. Extension may, and probably does, nucleate in a site of a previous weakness. A prime question for us is whether extensional strain remains concentrated in that initial site or migrates to other places. Here we will not consider propagation of rifting orthogonal to the direction of extension, but we will concentrate on simplified versions of two-dimensional (2-D) models concerned with the vertical plane parallel to the direction of extension.

p0095 AU5 We discuss eight processes that may affect concentration or delocalization of strain during continental. Though this may seem like a bewilderingly large number of things that could affect extension, it should become clear that some of these processes can be insignificant. In an effort to show what may control

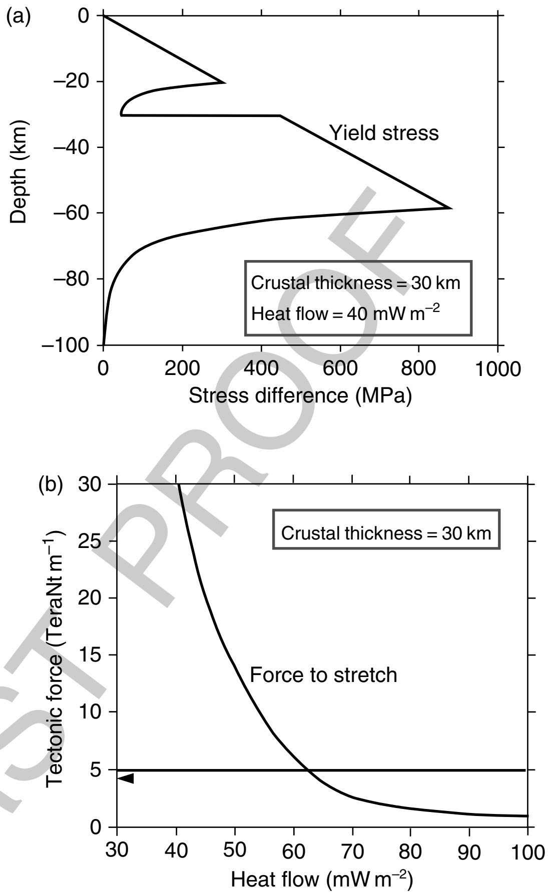


Figure 6 (a) Yield stress profile for diabase rheology crust over olivine rheology mantle for a typical continental thermal profile with an average value of surface heat flow. f0030
(b) Tectonic force for extension (the integral of yield stress with depth) as a function of surface heat flow for 30 km thick crust. The horizontal line gives the approximate maximum tectonic force available to drive rifting.

the importance of each process we derive scaling relations that show what parameters should control the effect of the process. The derivations attempt to show the approximate change in horizontal tectonic force needed for continued rifting due to a particular process. Each process is viewed in terms of how it affects the lithospheric strength or the gravitational force to continue extension. For example, a small amount of lithospheric thinning will reduce the strength of the lithosphere in proportion to the initial strength of the lithosphere times the square of the fractional amount of thinning. Local lithospheric thinning will produce gravitational stresses that also make continued extension easier, but the magnitude of the change in gravitational stresses is typically less than a few percent of the weakening effect. Familiarity with these scaling relations may make it easier to interpret complex numerical models of continental extension in which many parameters may affect the results.

s0020 6.09.2.2 Localizing Processes

s0025 *6.09.2.2.1 Thermal advection due to stretching*

p0100 The most obvious thing that may keep extension focused in one area is the localized thinning of lithosphere as it extends. Extension implies that material points horizontally offset from each other move apart. Conservation of mass requires vertical motion in response to this lateral extension. This implies some upward movement of the material within the lithosphere and so advection of heat. If this heat advection is significantly faster than thermal diffusion, then isotherms at the base of the lithosphere should move up. The lithosphere is thinned and so it is weakened. This process of 'necking' is self-reinforcing. The weaker the thinned or necked area the more concentrated, and perhaps more rapid, the extension. The more concentrated the extensional strain the faster the rift becomes weaker.

p0105 To quantify the necking-related reduction in brittle lithospheric yield strength, ΔF_N, we ignore thermal diffusion and relate lithospheric thinning to extensional strain, ε. The base of the brittle lithosphere is the place where the ductile and brittle yield strengths are equal. The ductile stress depends strongly on temperature, so that when heat is advected upward the depth of the base of the lithosphere moves up. The contribution to the yield force (the integral of the yield strength envelope) due to the ductile part of the lithosphere is a small fraction of the total. Thus, in the interest of simplicity all the strength is taken to be in the brittle part of the lithosphere. Assume that the base of the brittle lithosphere, where $z = H_b$, is marked by an isotherm. The new brittle lithospheric thickness is $(1-\varepsilon)H_b$. Since the brittle lithospheric strength is proportional to the square of the brittle layer thickness the change in strength is

$$\Delta F_N = CH_b^2 - C(1-\varepsilon)^2 H_b^2 \sim 2C\varepsilon H_b^2 \qquad [5]$$

p0110 **Figure 7** illustrates the reduction in strength due to advective lithospheric thinning. In doing this we ignore the effect of strain rate on yield stress, which is treated in the section below on viscous stresses. The key result is that the first-order reduction in strength depends on two times the strain times the initial strength of the lithosphere.

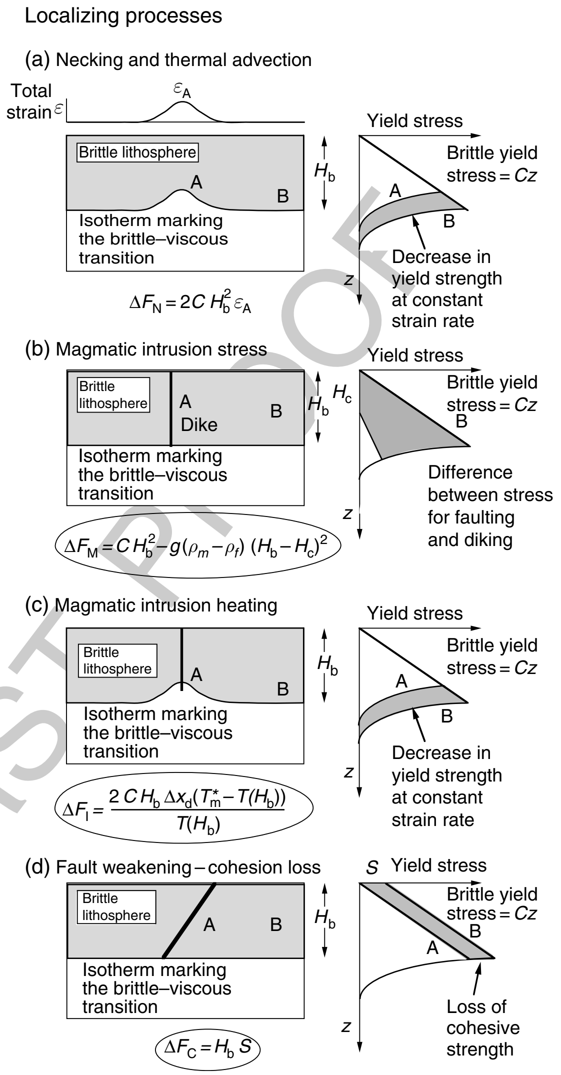


Figure 7 Schematic illustrations of processes that may lead to localization of strain during continental lithospheric extension. Plots to the right show the approximate distribution of yield stress with depth both at the centre of a rift (a) and at an area unaffected by the rifting (b). The difference in the two curves is marked with vertical hatchers and that area is proportional to the change, here reduction, in the tectonic force needed for continued rifting. The scaling of these force changes is given within ovals. See text for further explanation. f0035

6.09.2.2.2 Magmatic intrusion s0030

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 since mantle melting can produce basaltic magma and because more felsic dikes might be too high in viscosity to easily propagate. Dikes should form in planes perpendicular to the least principal stress, σ_3; p0115 for a rift, this should be in vertical planes parallel to the rift (Anderson, 1951). It is assumed that preexisting vertical fractures are prevalent in order 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.

p0120 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 Kerr, 1991). The vertical pressure variation in a static column of fluid magma is related to its density, ρ_f, so for magma emplacement: $\partial\sigma_3/\partial z = \rho_f g$. To specify the level of fluid 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

$$\sigma_M(z) = \sigma_1(z) - g\rho_f z \qquad [6]$$

p0125 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 fluid magma. If the crustal rock density and fluid magma density are taken to be equal, the stress difference for crustal diking is zero. In mantle of density $\rho_m = 3300$ kg m^{-3} the stress difference required for intrusion of fliud magma with density $\rho_f = 2700$ kg m^{-3} increases at a rate of 6 MPa km^{-1} of depth into the mantle. If the ductile mantle is too weak to maintain such stresses, then the magma cannot be emplaced at depth and will be extruded. Then the force to emplace magma is just related to the difference in brittle layer thickness and the crustal layer thickness:

$$F_M = g(\rho_m - \rho_f)\frac{(H_b - H_c)^2}{2} \qquad [7]$$

p0130 The decrease in force needed for rifting if there is enough magma to reach the surface is then:

$$\Delta F_M = F_T - F_M = [CH_b^2 - g(\rho_m - \rho f)(H_b - H_c)^2]/2 \qquad [8]$$

s0035 ### 6.09.2.2.3 Magmatic heat input

p0135 Basaltic magma is hotter than lithosphere so that dikes intrusion advects heat into the lithosphere and can weaken it. Magma also releases a considerable quantity of latent heat when it crystallizes. We can define an effective temperature of magma as T^*_m which is equal to the actual magma temperature plus the latent heat times the specific heat. A reasonable value of the added effective temperature due to latent heat is 300° C so that liquid magma with an actual temperature of 1300° C would have an effective temperature of 1600° C. Since the temperature at the base of the brittle lithosphere is likely to be 600–800° C, the intrusion of magma could heat the rock sufficiently to cause it to flow ductilly. To estimate the reduction in lithospheric thickness on intrusion of a set of dikes of total width Δx_d requires that we specify the width of region of lithosphere that is heated. Taking a conservative estimate that this width of region of intrusion is equal to the lithospheric thickness results in a thinning of the lithosphere of approximately $\Delta x_d(T^*_m - T(H_b))/T(H_b)$ given a linear geotherm in the lithosphere. Thus, the reduction in lithospheric strength due to dike intrusions totaling Δx_d in width is

$$\Delta F_I = 2CH_b\Delta x_d(T^*_m - T(H_b))/T(H_b) \qquad [9]$$

The lithospheric strength here is defined as the strength with no further magmatism. Thus, dike intrusions totaling 10 km wide could very significantly reduce the lithospheric thickness and strength and allow rifting to proceed at moderate levels of forces even if no more magma is supplied. p0140

6.09.2.2.4 Cohesion loss s0040

Within a rift, extension is not smoothly distributed, at least in the near surface, but is concentrated on normal faults. For faults to accommodate so much strain they must be weaker than the surrounding less-deformed rock. One possibility is that faults have lost some or all of the cohesion of the surrounding rocks. p0145

The failure of brittle materials, such as cold rock, has been described by a number of criteria. We assume the Coulomb–Navier criterion which states that failure occurs when the shear stress exceeds cohesion, S, plus a friction coefficient, μ, multiplied by the normal stress. In deriving the brittle yield stress above we followed Brace and Kohlstedt (1980) and others, who adopt this criterion with $S = 0$. p0150

When the yield criterion is reached the rock may break as cohesion is lost. The optimum orientation of faults is controlled by the friction coefficient. Then for cohesion loss, the yield stress on optimally oriented faults is reduced by an amount equal to p0155

$2S/[(1+\mu^2)^{1/2}/2+\mu]$. For μ close to 0.75 the reduction in yield stress is approximately equal to S. Laboratory measured values of rock cohesion, also known as inherent shear strength, range from almost zero for weak sediments to nearly 50 MPa for some igneous rocks ([b0320] Handin, 1966). The existence of large vertical cliffs requires cohesion in large volumes of rock that are of order 10 MPa.

p0160 Assuming that strain weakening results in noncohesive faults then they would be weaker by approximately S. Multiplying this cohesion drop by the thickness of the brittle layer gives the amount of reduction in the yield strength of a fault compared to unfaulted rock. Thus, the reduction in force due to cohesion loss is approximately:

$$\Delta F_{\mathrm{C}} = SH_{\mathrm{b}} \qquad [10]$$

p0165 For a brittle layer thickness that is over 10 km the reduction in strength due to cohesion loss should be fairly small compared to the strength that remains due to rock friction. It is possible that faults develop a lower frictional coefficient or become weaker due to pore pressure increases, though these are difficult to quantify.

s0045 6.09.2.3 Delocalizing Processes

s0050 *6.09.2.3.1 Thermal diffusion*

p0170 Thermal diffusion can lead to thickening and so strengthening of the brittle lithosphere when the lithosphere is out of thermal equilibrium. Lithosphere may be in equilibrium with basal heat flux, which is presumably delivered by mantle convection. Thermal disequilibrium may be produced by several types of extension-related heat advection, such as lithosphere stretching or dike intrusions. Diffusion tends to return isotherms perturbed by advection to their previous configuration.

p0175 It is difficult to derive simple analytic expressions for the rate of lithospheric thickening due to thermal diffusion except for the simplest of boundary and initial conditions. The problem of cooling of a half-space with no heat sources gives a simple result that captures important features of more complex model configurations. For half-space cooling the depth to an isotherm marking the base of the lithosphere increases proportional to $\sqrt{\kappa t}$ where κ is thermal diffusivity (see [b0740] Turcotte and Schubert, 2002). Thus, the velocity of lithospheric thickening is proportional to κ/H_{b}, where H_{b} is the thickness of the lithosphere. In a time interval Δt the lithosphere is thickened by roughly $\Delta t\kappa/H_{\mathrm{b}}$. The yield stress at the base of the lithosphere is CH_{b} and the lithospheric strengthening due to thermal diffusion is that stress times the increase in thickness:

$$\Delta F_{\mathrm{d}} \approx C\kappa\,\varepsilon/\dot{\varepsilon} \qquad [11]$$

where the time interval Δt is replaced by the ration of the strain, ε, and the strain rate of extension, $\dot{\varepsilon}$. For a typical rock diffusivity of $10^{-6}\,\mathrm{m^2\,s^{-1}}$ a million years of cooling would produce an increase in strength of about $5\times10^{11}\,\mathrm{Nt\,m^{-1}}$, which is a small fraction of the strength of lithosphere a few tens of kilometers thick.

6.09.2.3.2 Viscous flow s0055

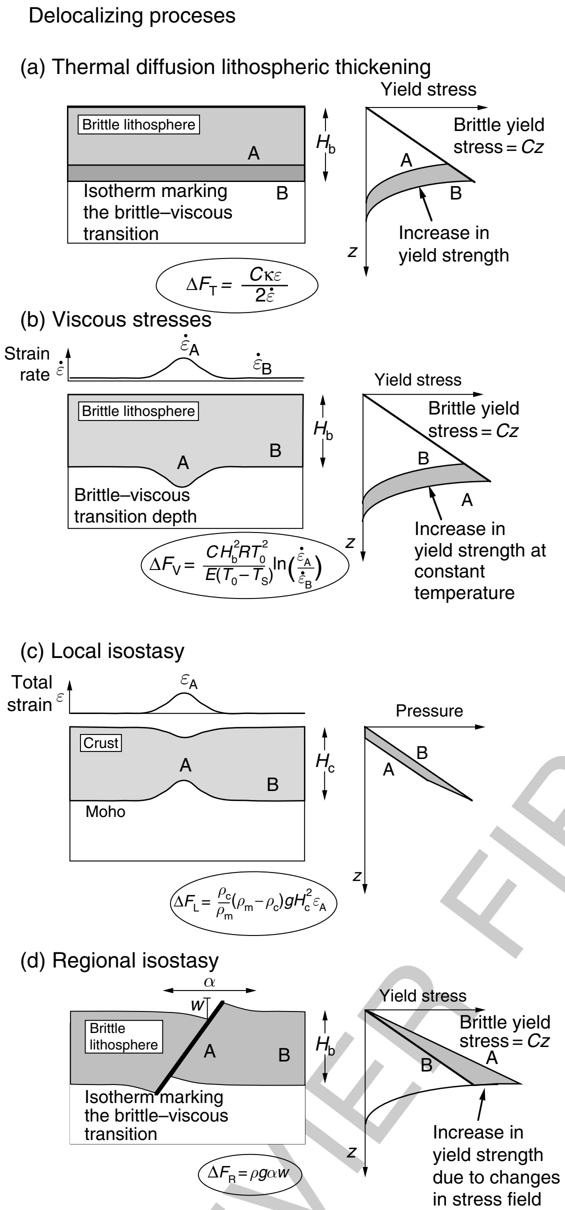
By definition the strength of viscous material is proportional to strain rate, or a strain rate raised to a power. This means that when viscous material strains faster its flow stress is larger. This results in stress being delocalized in a viscous layer. A thought experiment can help illustrate this effect. Assume that the strain rate is greater in one region, as shown in **Figure 8**. Since the viscous flow stresses are greater in the region of high strain rate, the transition between brittle and viscous behavior (sometimes called the brittle–ductile transition) is deeper there. Clearly, where the transition depth is deeper the total yield strength is greater. Thus, the area straining the fastest will be the strongest. p0180

The amount of deepening of this transition and the lithospheric stregthening depends on two things: the contrast in strain rate and the length scale for changes in viscosity. At constant stress the vertical distance Z_e over which the viscosity changes by a factor e (2.72) is related to the temperature gradient in the region of the transition, $\mathrm{d}T/\mathrm{d}z$, and to the temperature there (T_0) as $Z_e = R\,T_0{}^2/(E\,\mathrm{d}T/\mathrm{d}z)$. To get an expression for the related force increase we assume constant stress at the transition depth and then estimate the vertical distance change in that depth for a given contrast in strain rate. Also, we assume a linear temperature gradient through the lithosphere so that $\mathrm{d}T/\mathrm{d}z = T_0/H_{\mathrm{b}}$. Doing this gives a force increase of p0185

$$\Delta F_{\mathrm{V}} = \frac{CH_{\mathrm{b}}^2RT_0^2}{E(T_0-T_{\mathrm{S}})}\ln\left(\frac{\dot{\varepsilon}_{\mathrm{A}}}{\dot{\varepsilon}_{\mathrm{B}}}\right) \qquad [12]$$

where T_{s} is the surface temperature. $\dot{\varepsilon}_{\mathrm{B}}$ and $\dot{\varepsilon}_{\mathrm{A}}$ are the background and local strain rates, respectively.

It is the viscous stress effect that can lead to both folding and boudinage structures in layered rocks p0190



f0040 **Figure 8** Same as **Figure 7** except that the processes could promote delocalization of strain during extension.

(Ramberg, 1955; Biot, 1961; Smith, 1977). It is this viscous delocalization that has been suggested to contribute to the boudinage-like structure of the wide Basin and Range Province of the western United States (Fletcher and Hallet, 1983), as discussed below.

s0060 6.09.2.3.3 Local (crustal) isostasy

p0195 Local isostasy is an idealized description of how lithosphere floats on underlying fluid asthenosphere (Watts, 2001). The term 'local' implies that the surface elevation at a point depends only on the average density of the column of lithosphere below that point. Essentially, the shear stress on vertical planes is taken to be zero. However, horizontal stresses should be continuous across vertical planes. Vertical stresses at a given depth should depend on the topography and average density of material in a column. Thus, where there is topographic relief in local equilibrium there will be nonzero stress differences ($\sigma_1 - \sigma_3$). For example, an elevated area will be in relative tension. Conversely, as low area will be in relative compression as material tries to flow into it.

Localized extension results in crustal thinning. p0200
Because crust is less dense than underlying mantle, local crustal thinning should produce lowered elevations in the center of a rift (e.g., McKenzie, 1978). This puts the center of the rift into relative compression, and this makes continued extension harder.

To estimate the magnitude of this effect we follow p0205
previous workers (e.g., Artemjev and Artyushkov, 1971; Fleitout and Froidevaux, 1983) and assume that the wavelength of crustal thickness variations is large compared to crustal thickness. Then, the increase in the tectonic force for extension due to crustal thinning, ΔF_c, equals the integral over depth of the difference in lithostatic pressure, ΔP. This pressure difference equals $\Delta w \rho_c g$, where Δw is the change in surface elevation between the center of the rift and the adjacent, unrifted area. Here ρ_0 is the density of crust and g is the acceleration of gravity. Topographic relief Δw in local isostatic equilibrium has an amplitude of $\varepsilon_a H_c(\rho_{m\,-}\,\rho_0)/\rho_m$ where ρ_m is the density of the mantle and $\varepsilon_a H_c$ is the amount of crustal thinning at the center of a rift. As long as the strains are not too large then the local crustal buoyancy force, ΔF_L approximately equals $\Delta P H_0$, can be expressed as

$$\Delta F_L = \rho_c g \varepsilon_a H_c^2(\rho_m - \rho_0)/\rho_m \qquad [13]$$

Just as thermal diffusion may act to diminish the p0210
effect of thermal advection during lithospheric necking, lower crustal flow can act to even out crustal thickness variations and so reduce the crustal buoyancy effect. The idea that lateral crustal flow may be important in some areas is discussed in Block and Royden (1990), Buck (1991), Bird (1991).

Isostatic topography in a rift can be caused by p0215
temperature differences. However, the thermal buoyancy force change due to lithospheric stretching is usually much smaller than that due to crustal

thinning. This is because density changes due to a temperature change of 500° C are 10 times smaller than the density difference between crust and mantle. The thermal buoyancy force change scales with the thermal lithospheric thickness squared. Thus, when the lithosphere is thick, thermal buoyancy effects can be of larger magnitude than the crustal buoyancy force change because the depth extent of temperature anomalies can be much greater than the depth to Moho, as noted by Turcotte and Emmerman (1983) and Le Pichon and Alvarez (1984). Thermal and crustal buoyancy are of opposite sign for a rift.

s0065 *6.09.2.3.4 Regional isostasy*

p0220 Fault offset results in stress changes around and on the fault. Forsyth (1992) argues that those stress changes should inhibit continued offset on the fault. Eventually it could be easier to break a new fault in adjacent lithosphere than for continued slip on an original fault.

p0225 **Figure 8**(**d**) illustrates this effect. As described in a later section the scaling between the force and the vertical deflection is directly proportional to the horizontal wavelength of the response of the lithosphere to vertical loads, represented by the flexural parameter, α. It also and is approximately:

AU6

$$\Delta F_{\mathrm{R}} = g\rho_{\mathrm{c}}\alpha w \quad [14]$$

where w is the vertical offset across the fault.

s0070 *6.09.2.3.5 Additional effects*

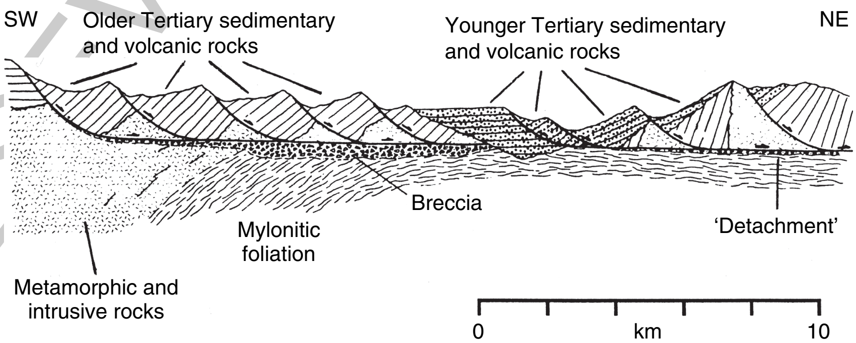
p0230 Other factors may have a considerable effect on the geometry of extension, but they are even harder to quantify than the factors discussed here. For example, erosion and sedimentation may reduce both the local isostatic (crustal buoyancy) effect and the regional isostatic effect due to fault offset. This could therefore diminish those delocalizing effects (Burov and Cloetingh, 1997). However, it is difficult to quantify the rate of erosion or to specify either the distribution of sediment or its density.

The rest of this chapter is concerned with the several controversies that continue to exercise researchers in the field of continental extension. Each of these controversies can be cast in terms of a pair of end-member concepts and the title of the following sections emphasizes these extreme possibilities. p0235

6.09.3 High-Angle vs Low-Angle Normal Faults s0075

One of the most exciting and contentious areas of work on extensional tectonics in the last 20 years involves the interpretation of subhorizontal normal faults (**Figure 9**). These 'low-angle normal faults' were first recognized in continental metamorphic core complexes (e.g., Coney and Harms, 1984) but now have been seen on the ocean floor near some mid-ocean ridges (e.g., Tucholke *et al.*, 1998, 1997). The excitement centers around whether the faults actively slipped at low dip angles (<30°) or at greater dips. Under the tenets of classical fault mechanics, normal faults in the brittle upper crust should initiate at dips greater than 45° and should be active at dips of no less than 30° (Anderson, 1942; Byerlee, 1978; Sibson, 1985). p0240

Normal faults appear to fall into two distinct populations. High-angle normal faults dip at an angle of >30° and are offset <10 km (Vening Meinesz, 1950; Stein *et al.*, 1988). Low-angle normal faults dip between 30° and subhorizontal, and some appear to have accommodated horizontal throws as much as 50 km or more (e.g., Wernicke, 1992; Davis p0245



f0045 **Figure 9** Interpretive cross-section of the Whipple Mountains metamorphic core complex, thought to be a site of large magnitude extension with no local subsidence. The detachment fault appears to have accommodated tens of kilometers extension and is not in a subhorizontal position. From Davis GA (1980) Problems of intraplate extensional tectonics, Western United States. In: *Continental Tectonics*, pp. 84–95. Washington, DC: National Academy of Science.

and Lister, 1988; Axen *et al.*, 1990; Yin and Dunn, 1992; John and Foster, 1993). Before discussing ideas for the origin of low-angle normal faults the question of how topographic relief is produced by offset of high-angle normal faults will be surveyed.

s0080

6.09.3.1 Rift Shoulder Uplift

p0250 Active rifts are primarily identified by their characteristic topographic relief, with a depression or rift valley surrounded by elevated flanks or shoulders (**Figures 1** and **5**).

AU7 p0255 Geologic and geophysical data indicate that the shoulder of a rift is typically footwall of a major, rift-bounding normal fault. Vening-Meinesz (1950) considered how normal fault offset might produce topographic relief. He assumed that the load was related to a fault cutting low-density crust floating on the asthenosphere. In this formulation the force supporting uplift depends on the crustal thickness, h_c, and the density contrast between crust and mantle. To quantify the model, imagine that two floating blocks cut by a fault are moved apart and denser fluid upwells to fill the space between them. If the blocks are crust with density ρ_c and the fluid mantle has a density ρ_m, then the triangular parts of the blocks will be subject to buoyancy forces related to their shape. The upward force on the triangle of the footwall block is given by

$$V_0 = \frac{g h_c^2 \rho_c(\rho_m - \rho_c)}{2\rho_m \tan\theta} \qquad [15]$$

where g is the acceleration of gravity, h_c is the thickness of the crust, and θ is the dip of the fault cutting the blocks. Applying this load to the end of a thin elastic plate floating on the mantle would produce a vertical deflection of the plate of:

$$w(x) = \frac{2V_0}{\rho_m g \alpha} e^{-x/\alpha} \cos\frac{x}{2} \qquad [16]$$

where α is the flexural parameter of the flexed plate (e.g., Turcotte and Schubert, 2002). Neglecting the distance from the point of application of the load and the edge of the region with full crustal thickness the maximum deflection occurs at $x = 0$ and equals

$$w(0) = \frac{h_c^2 \rho_c(\rho_m - \rho_c)}{\alpha \rho_m^2 \tan\theta} \qquad [17]$$

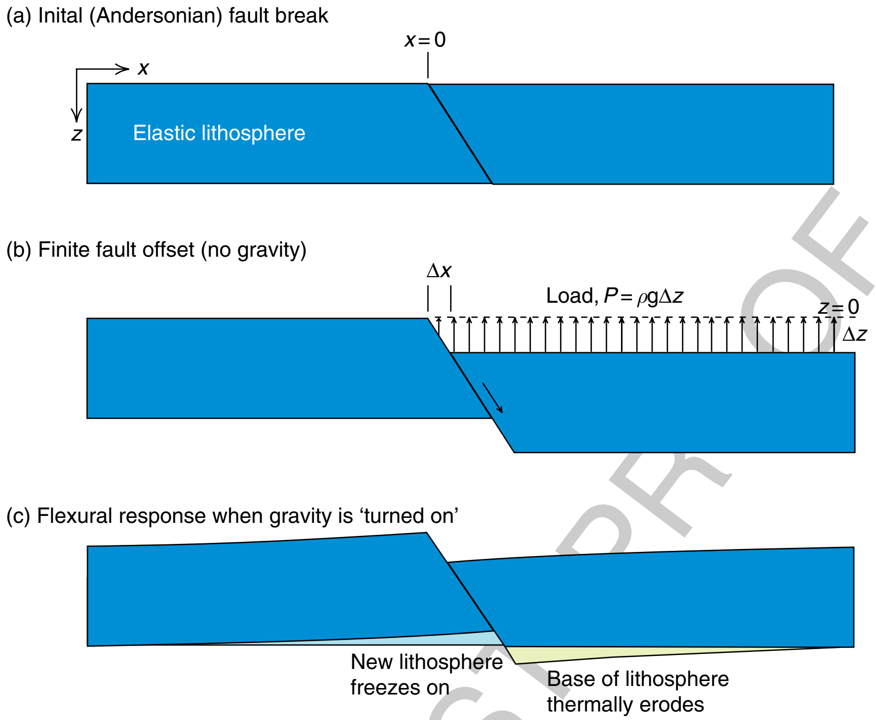
p0260 For a typical (i.e., moderate heat flow) continental region the model appears to work. For such a region it is reasonable to assume $h_c = 30$ km, $\alpha = 60$ km, $\rho_c = 2900$ kg m^{-3}, $\rho_m = 3300$ kg m^{-3}, so the uplift of the footwall given by eqn [3] is 1.6 km. Even though the load of sediments filling the hanging wall basin would reduce the footwall uplift, the model fit is in reasonable agreement with the observed uplift magnitude for continental rifts.

This model does not work for oceanic rifts, since oceanic crust is much thinner and denser than continental crust. However, the topographic relief at mid-ocean ridges is comparable to that seen in continental rifts. Taking $h_c = 6$ km, $\alpha = 12$ km and $\rho_c = 3000$ kg m^{-3} with other values held the same, eqn [17] predicts maximum footwall uplift of 0.16 km. This is far smaller than that observed for oceanic rifts. p0265

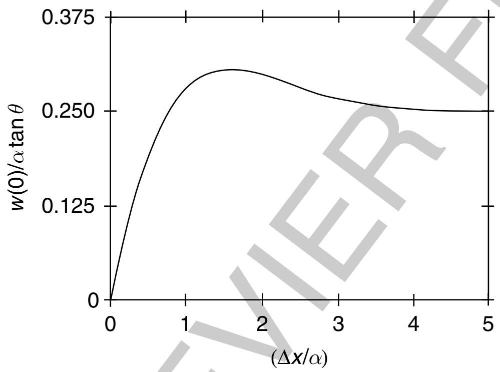
Vening-Meinesz (1950) was right that the loads produced by normal fault offset should be supported regionally, but he did not consider the load produced by the offset of the Earth's surface. This load is even more important than the load caused by offset of the Moho, which was treated by Vening-Meinesz (1950). The slip of a normal fault should deflect the surface of the Earth and create topographic relief even if the crustal thickness is zero and the lithosphere has the same density as the asthenosphere. p0270

Basic rock mechanics (Anderson, 1951) predicts that shear displacement should be easier than opening displacement at depth in the Earth. The top and bottom sides of a normal fault should remain in contact as the fault shears. The offset of the fault pushes the footwall block up and the hangingwall block down. Flexing in response to the loads should result in a curved fault so it is not immediately clear how to formulate a thin plate flexural approximation to the effect of fault offset. Weissel and Karner (1989) formulated the problem by conceptually turning off gravity when the fault offsets half the Earth's surface down relative to the other side of the fault. The offsets can be considered loads equal to the vertical offset times gravity times density. When gravity is 'turned on' these loads deflect the lithosphere, resulting in a topographic pattern as shown in **Figure 10**. Braun and Beaumont (1989) used numerical models of extension of a 2-D viscous-plastic layer to show that local extensional thinning of the layer (necking) could produce reasonable uplift of rift shoulders. They also interpret their results in terms of a two-stage process in which stretching would produce a topographically low basin without gravity, but with gravity it uplifts rift shoulders (see **Figure 10**). Braun and Beaumont (1989) also noted that such uplift might explain the 'breakup unconformity' seen at many rifted margins (**Figure 11**). p0275

Using the Weissel and Karner (1989) approach an analytic solution for the deflection of the surface can p0280



f0050 **Figure 10** Schematic of a way to treat the effect of normal fault offset on topographic relief. (a) shows an ideally oriented fault and (b) shows how one side of the fault would move down in the absences of gravity. The magnitude of the load, P, would exist if gravity were 'turned on'. (c) shows the flexural response to that load.



f0055 **Figure 11** The results of an analytic calculation (eqn [1]*) of the uplift of the footwall breakaway of a normal fault as a function of the horizontal offset, Δx, of the fault. The offset is normalized by the flexure parameter, α, and the vertical uplift is normalized by α tan θ, where θ is the fault dip angle.

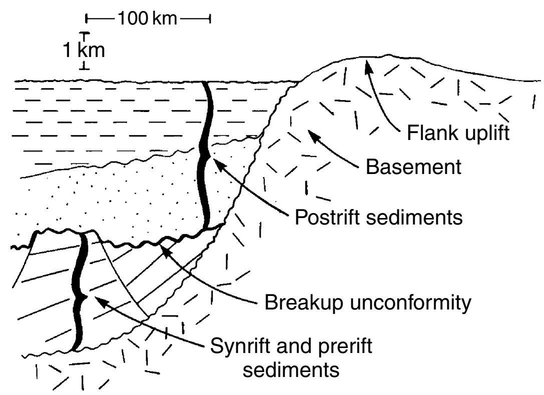


Figure 12 Illustration of typically observed rift flank uplift and rift margin syn- and postrift sediments with an intervening 'breakup unconformity'. From Braun J and Beaumont C (eds.) (1989) Contrasting styles of lithospheric extension: Implications for differences between basin and range province and rifted continental margins. extensional tectonics and stratigraphy of the north Atlantic margins. *American Association of Petrololeum Geologist Memoir* 46: 53–79. f0060

be derived assuming thin elastic plate flexure theory. For the geometry in which a normal fault drops the hangingwall down (**Figure 12**) the load distribution is

$$V(x) = \begin{cases} \rho g \tan\theta x, & 0 < x < \Delta x \\ \rho g \tan\theta x_0, & x > \Delta x \end{cases} \qquad [18]$$

where $x = 0$ is the position of the intersection of the fault with the surface on the footwall (the breakaway) and $x = \Delta x$ is the intersection of the fault with the surface on the hangingwall. The vertical deflection at position x due to a localized load V at position x' is

$$w(x) = \frac{V(x')\rho g}{4\alpha} e^{-x/\alpha}\left(\cos\frac{(x'-x)}{\alpha} + \sin\frac{(x'-x)}{\alpha}\right) \quad x' \geq x \qquad [19]$$

p0285 The maximum uplift of the footwall occurs at $x = 0$ and this can be calculated by integrating eqn [19] from $x = 0$ to $x = \infty$:

$$w(0) = \frac{\alpha \tan\theta}{4}\left\{e^{-\Delta x/\alpha}\left(\sin\frac{\Delta x}{\alpha} - \cos\frac{\Delta x}{\alpha}\right) + 1\right\} \qquad [20]$$

p0290 **Figure 12** plots the value of the breakaway uplift versus horizontal fault offset. The uplift increases with offset up to the point where fault offset $\Delta x = (\pi/2)\alpha$ then diminishes slightly. The maximum uplift is $\sim 0.3\alpha \tan\theta$. For a fault dip angle of 45° and a flexure parameter $\alpha = 10$ km, the maximum uplift given by eqn [6] is about 3 km. This is on the high-end of the range of uplifts seen for oceanic or continental rifts. Several things can limit the development of topographic uplift across a normal fault, including sedimentary loading of the down-dropped side of a fault to produce the kind of half-graben basin seen for many continental rifts (e.g., Stein *et al.*, 1988).

s0085 6.09.3.2 Low-Angle Fault Development and Stress Rotation

p0295 On the basis of geologic mapping of inactive faults, many authors have contended that slip has occurred along some normal faults with dip angles <30° (Wernicke, 1981; Davis and Lister, 1988; Miller and John, 1988). There is no clear evidence, however, for low-angle normal faults that are active today. Focal mechanism studies indicate the orientation of seismogenic faults. Information such as aftershock locations or the relation between surface fault breaks and hypocenters is needed to determine which nodal plane is the fault plane. Well-constrained fault-plane solutions indicate that most seismogenic normal faults dip at angles >30° (Jackson, 1987; Thatcher and Hill, 1991), although at least one has been interpreted as a low-angle fault (Abers, 1991).

p0300 Several authors have suggested that normal faults could initiate with low dips if unusual loads reoriented the tectonic stress field itself (Spencer and Chase, 1989; Yin, 1989; Parsons and Thompson, 1993). Under the right set of regional or local loading conditions, the principal stresses might be rotated to a configuration at least geometrically compatible with low-angle normal faulting under the assumption that faulting occurs at an angle of approximately 30° to the maximum principal stress, a well-established result of Mohr–Coulomb fracture mechanics. These stress rotation models have a strong intuitive appeal. They tie a ubiquitous and puzzling feature of the Basin and Range Province, regional detachment faulting, to conditions known or strongly suspected to have existed there at the onset of extensional deformation: orogenic loading, ductile flow below the brittle layer, and widespread calc-alkaline magmatism. They also demonstrate that unusual boundary conditions can alter stress orientations.

The papers advocating initiation of normal faults at low dip angles (Spencer and Chase, 1989; Yin, 1989; Parsons and Thompson, 1993) did not address the question of whether the magnitudes of the reoriented stresses would allow regional low-angle normal faulting under geologically realistic conditions. Wills and Buck (1995) carried out simple analyses designed to test this aspect of several stress-field rotation models. Their results show that the areas at which these models predict low-angle normal fault development are the least favorable places for fault slip to occur. They also quantified the magnitude of spatial variations in cohesive strength and pore pressure required to initiate slip on low-angle normal faults, variations that are implausible. p0305

6.09.3.3 Fault Rotation s0090

An alternative to slip on low-angle normal faults is that the upper parts of some actively slipping high-angle normal faults rotate to shallower dips. Spencer (1984) first suggested that the isostatic response to offset of a normal fault would tend to decrease the dip of the fault. However, Spencer (1984) confined his discussion to the rotation of active low-angle faults. Hamilton (1988) and Wernicke and Axen (1988) argued that large rotation of high-angle faults is consistent with the structures seen in two different extensional settings. Buck (1988) also argued that rotation could explain low-angle fault structures, and calculated the flexural response of lithosphere to the loads caused by the offset of a high-angle normal fault. This model produced realistic low-angle fault geometries only when (1) the lithospheric yield strength was finite and (2) when the offset of the model fault was about twice the lithospheric thickness. p0310

To get the inactive, up-dip parts of model normal faults to rotate to a low-angle orientation Buck (1988) assumed that large offsets, relative to the lithospheric thickness, could occur on a high-angle normal fault to produce low-angle fault structures. A major question is why such large offsets might occur on some high-angle normal faults and not on others. p0315 s0095 ### 6.09.3.4 Large Offset of Normal Faults

p0320 The offset of a dip-slip fault produces topography and so changes the stresses around the fault. For a high-angle normal fault, the topographic relief should build up quickly as the fault is offset. Vening-Meinesz (1950) recognized this and was among the first to suggest that the stress changes related to normal fault offset could result in new faults being formed. Offset of an elastic layer by slip of one fault produces maximum bending stresses at about a flexural wavelength from the fault. Vening-Meinesz (1950) assumed the next fault would break where the bending stresses at the surface were maximally extensional, and result in a graben. This assumption is reasonable, since the yield stress (the stress needed to break and slip on a fault) is minimum at the surface.

p0325 A different approach to analyzing the effect of normal fault offset on stresses was suggested by Forsyth (1992). In contrast to Vening-Meinesz (1950) he ignored the direct effect of bending stresses on promoting layer breaking and instead estimated the increase in the average regional tectonic stress in a layer due to the buildup of fault related topography. Forsyth (1992) noted that Anderson's theory for normal faulting is only valid for infinitesimal fault slip because it only considers the work done overcoming friction on the fault surface. The initial orientation of a fault requiring the least regional stress is the one that dissipates the least friction on the fault per unit of horizontal displacement. When a fault is displaced, work is done in the bending of the lithospheric plate cute by the fault. To estimate this work, Forsyth (1992) approximated the lithosphere as a perfectly elastic layer floating on an inviscid substrate. The deflection of the layer caused by fault offset is estimated by using the thin-plate flexure equation. To do this analytically, the fault is treated as a vertical boundary cutting the lithosphere. The topographic step across the fault is the horizontal fault throw, Δx, times tan θ, where θ is the fault dip.

p0330 Because of the work done bending the lithosphere, it takes extra horizontal tensional stress to keep the slip occurring on the fault. Forsyth (1992) found that the extra horizontal stress increases linearly with Δx depends on $\tan^2 \theta$. He estimated that after only a few hundred meters of slip on a typical high-angle fault, it is easier to break a new fault rather than to continue slip on the original fault. On a low-angle fault, the extra horizontal stress needed for continued motion builds up more slowly with offset. Forsyth (1992) suggested that an initially low-angle normal fault could build up much more offset than a high-angle fault.

Buck (1993) used the approach of relating work building topography to tectonic forces suggested by Forsyth (1992), but he modified the way fault dip was considered (**Figure 13**). As in the description of topographic relief produced by fault offset, described above, the nonvertical dip of the fault significantly reduces the topography-related work to continue fault displacement. Another important change in formulation was inclusion of the effect of finite yield strength on bending stresses. A Mohr–Coulomb plate will bend more easily than a purely elastic plate, since bending stresses cannot exceed the yield stress. Inclusion of finite yield stress in this model radically lowers the size of the tectonic force increase due to fault related topography (as shown in **Figure 14**). Using reasonable values for friction and cohesion p0335

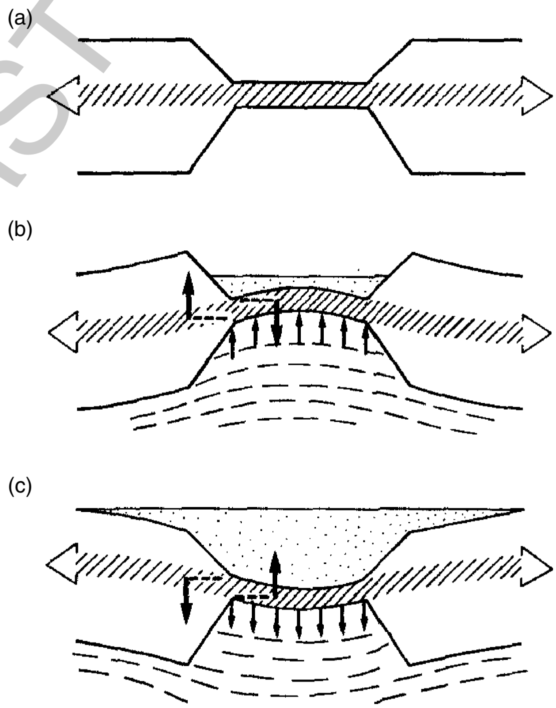
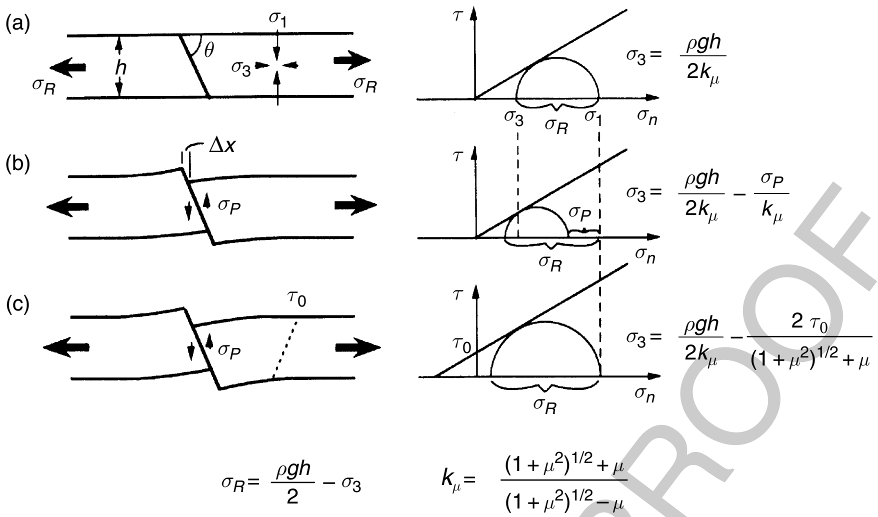


Figure 13 Schematic of the mechanism of (a) lithospheric necking and (b) resultant uplift caused by the gravitational response to the lowering of the surface due to plate stretching. (c) illustrates how continued sediment input (combined with thermal subsidence) can lead to subsidence of once-elevated rift flanks. After Braun J and Beaumont C (eds.) (1989) Contrasting styles of lithospheric extension: Implications for differences between basin and range province and rifted continental margins. extensional tectonics and stratigraphy of the north Atlantic margins. *American Association of Petrololeum Geologist Memoir* 46: 53–79. f0065



f0070 **Figure 14** Illustration of the effect of plate bending stresses on the stress state required for slip on a normal fault. (a) shows the initial Andersonian stress state for slip on a cohesionless, optimally oriented fault. (b) shows the lowering of the minimum stress, σ_3, required for slip when the plate bending stress can be considered to reduce the vertical stress by an amount σ_P. (c) shows the regional stress difference needed to break a new fault with a shear strength τ_0. Here μ is the friction coefficient. From Buck WR (1993) Effect of lithospheric thickness on the formation of high-and low-angle normal faults. *Geology* 21: 933–936.

for a 10 km thick layer, the Buck (1993) model predicts possibly unlimited fault offset. For a thicker layer the fault offset may be limited to an amount smaller than the layer thickness.

s0100 ### 6.09.3.5 2-D Models of Fault Formation and Offset

p0340 The thin-plate approximation used in the studies described above is clearly not valid for large fault offset. To be confident of internal consistency in normal fault evolution models require treatment of the 2- or 3-D stress and strain field. Analog models are useful for simulating the early, small-offset stage of fault development (e.g., Tirel *et al.*, 2006; Withjack and Schlische, 2006; Brun *et al.*, 1994; Corti *et al.*, 2003; AU8 Tron and Brun, 198*; Supak *et al.*, 2006), but cannot easily simulate the thermally controlled strength field evolution likely to affect large offset faults. Ideally, models could follow the development of faults in an extending, 3-D, viscous–elastic–plastic layer. Given the numerical cost of such models groups have first AU9 developed 2-D models (**Figure 15**).

p0345 Early studies numerically simulated normal fault offset in 2-D cross-sections of elastic layers, assuming that the fault offset varies slowly in the third dimension (e.g., Melosh and Williams, 1989; King and Ellis, 1990). These studies assumed a pre-existing weak fault embedded in a purely elastic layer and solved for the

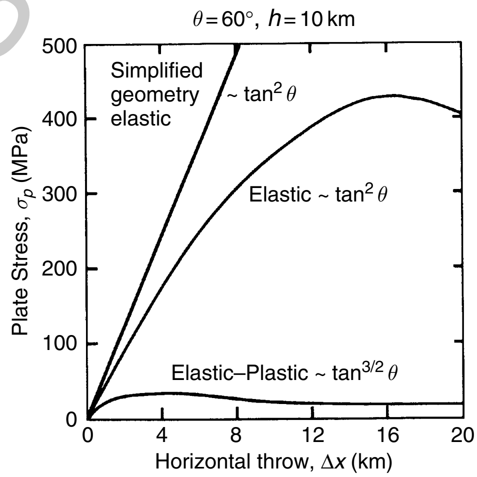
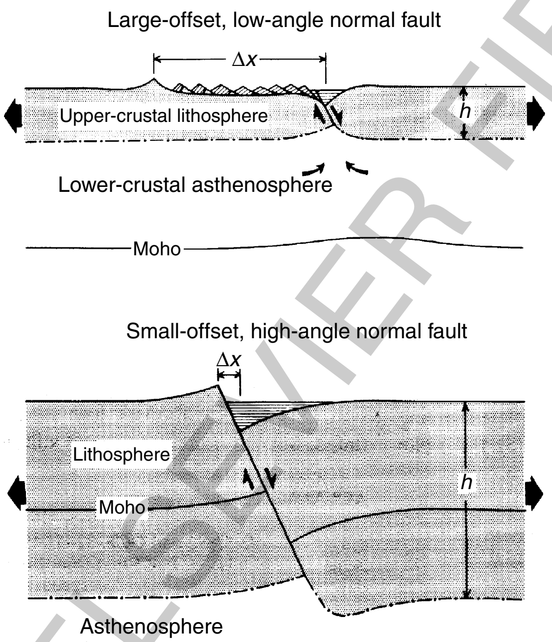


Figure 15 Results of analytic and thin plate flexure numerical calculation of the change of plate stress with horizontal fault offset for a floating brittle layer 10 km thick and a 60° dipping normal fault. The straight line is from Forsyth (1992) while the elastic curve includes the effect of finite fault dip. The inclusion of finite yield strength of the brittle layer greatly reduces the magnitude of the plate bending stress. From Buck WR (1993) Effect of lithospheric thickness on the formation of high-and low-angle normal faults. *Geology* 21: 933–936. f0075

topographic relief and stress changes around the offset faults. Given the potential importance of the finite brittle yield strength (here described as plastic deformation) in controlling how lithosphere can bend in response to fault offset, there has been great effort to include plastic deformation in numerical models. Braun and Beaumont (1989) and Bassi (1991) treat the lithosphere as a viscoplastic layer but were not concerned with localized fault development. Behn *et al.* (2002) and Huismans *et al.* (2005) allowed strain rate or strain-dependent weakening that lead to localized zones of concentrated deformation (model 'faults'). Poliakov and Buck (1998) adapted a numerical treatment of viscous–elastic–plastic deformation to normal fault development. They showed that a sequence of high-angle faults might form and accommodate extension at a simple model mid-ocean ridge AU10 structure (**Figure 16**). In the Poliakov and Buck (1998) formulation the faults weaken as a function of their offset related strain, up to a maximum amount of weakening.

p0350 **Figure 17** shows results of 2-D numerical experiments of extending an elastic–plastic layer that stays nearly uniform in thickness while a fault forms due to strain-dependent cohesion loss (see Section 6.09.2). The fault offset produces topography very much like that predicted by eqn [19] except that the wavelength of deformation appears to evolve in the early stage of model fault offset. The maximum fault produced topographic relief, shown in **Figure 18**, also follows the pattern predicted by the simple analytic expression derived above (eqn [20]).

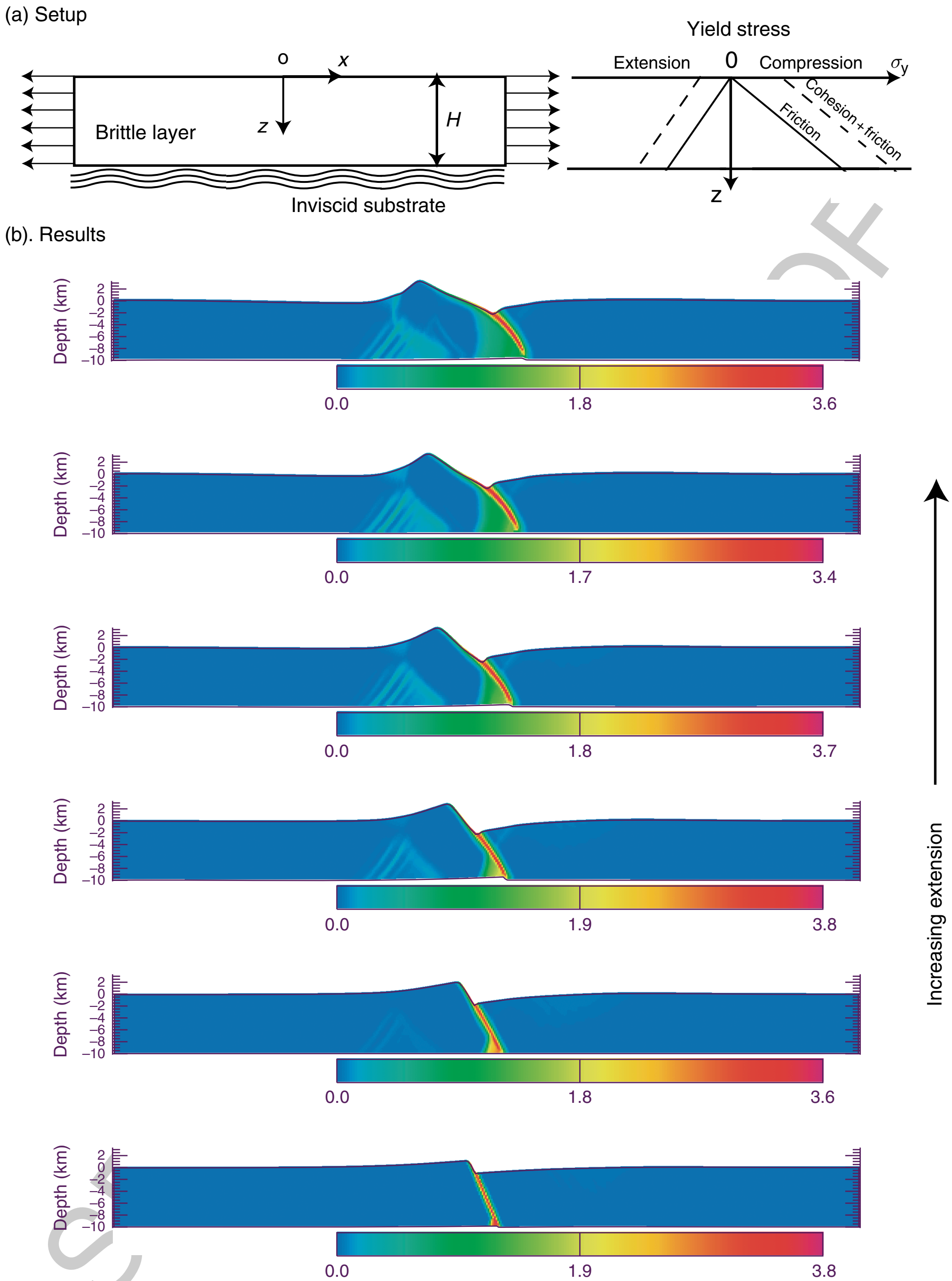


f0080 **Figure 16** Cartoons of possible implications of the calculations of plate bending stresses given in **Figure 15**. Because plate bending stresses scale with the square of plate thickness while the cohesive strength of the plate scales linearly with thickness a noncohesive fault in a thick layer could accrue very large offset. A noncohesive fault cutting a thick cohesive layer might not achieve large offset before another fault replaced it.

Numerical experiments also have addressed the question of fault offset and rotation. Lavier *et al.* (1999, 2000) used an elastic–plastic formulation and investigated how much fault weakening is needed to get large offset of a normal fault. The base of the extending layer was kept at a constant depth to simulate cooling of asthenosphere pulled up as the footwall of the fault moved out and up. **Figure 19** shows the results of one model experiment that produced a large offset fault. The inactive, up-dip part of the footwall side of the fault was rotated into a flat lying position. Lavier *et al.* (2000) showed large offset faults could only for when the fault weakening was AU11 above a minimum level, but that the rate of fault weakening with strain had to be within bounds. If the maximum amount of fault strain weakening is independent of the thickness of the brittle layer being extended then large offset faults will only happen when a layer is thinner than a given thickness. **Figure 20** illustrated the force changes due to normal fault offset estimated for thin and thick layers based on the numerical models. Only for a thin layer do we expect to get a large offset fault. **Figure 21** (from Lavier and Buck, 2002) shows two model cases of normal fault offset in extending viscous–elastic–plastic layers where the thermal structure is allowed to vary due to advection and diffusion of heat. It shows that a thin layer could develop a large offset fault while multiple, basin-bounding faults develop in a thicker layer. This is consistent with the observation that large offset normal faults are only seen in areas of higher than average heat flow where one expects thinner than average lithosphere. p0355

6.09.4 Pure versus Simple Shear Rifting s0105

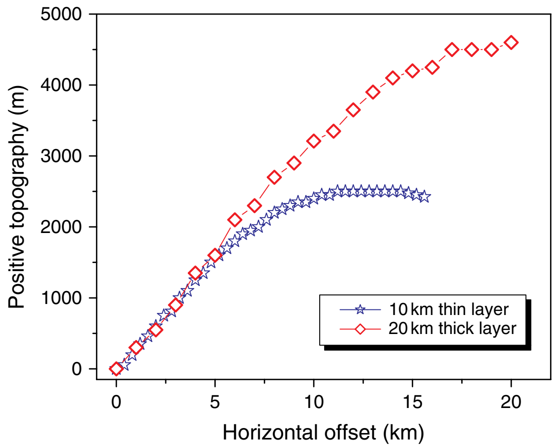
Rift valleys are partly filled with sediment and rifted margins are characterized by syn- to post-rifting sedimentary units that may be more than 10 km thick. Such thick piles of sediment can only be accommodated if the rifted region subsides by several kilometers. The sediments act as a load that pushes down the surface and amplifies the amount of subsidence that would occur with no sedimentary infill. If a rift basin is narrow compared to the flexural p0360



f0085 **Figure 17** (a) Setup for a numerical model of extension of a floating brittle Mohr–Coulomb layer with a single normal fault seeded at the center of the model layer. Grid spacing is 250 m through this 10 km thick layer and a cohesion of 20 MPa is reduced to 2 MPa over a fault offset of 1500 m. (b) Results of numerical calculation showing the development of plastic (brittle strain) and surface topography. Note that the model is similar in setup to the conceptual model pictures in **Figure 10**.

wavelength of lithospheric response, then the load is regionally compensated. As noted in the last section, sediment loading can pull down the rift shoulders as it fills in the rift basin. If the rift is wider than the flexural wavelength, then the sediment can be treated as a locally compensated load.

To illustrate local compensation of sediment consider the simple case of an air filled rift basin with an p0365



f0090 **Figure 18** Results of the uplift of the footwall breakaway point as a function of horizontal fault offset. The 10 km layer case is pictured in **Figure 17** while the results of a 20 km thick layer with similar setup are also shown. The numerical results fit very well by the analytic estimate shown in **Figure 11** for a flexure parameter of about 0.8 times the layer thickness and for a fault dip of ~50°.

initial depth of d_i. Imagine that gravity is 'turned off' as the basin is filled with sediment. When gravity is turned on the sediment load pressure is the fill depth, d_s $(=d_i)$, times the sediment density, ρ_s, times the acceleration of gravity, g. This load will push down the surface and, if we assume that sediment inflow keeps the basin filled to sea level, then the basin will continue to deepen until the sediment load is 'compensated'. Local compensation implies that the crust and mantle lithosphere floats on fluid mantle asthenosphere. This means that the weight of a column of crust and mantle to an arbitrary 'depth of compensation' must be constant. As sediment fills the hole it displaces mantle which is denser than sediment. Equating the column weight before and after basin fill, the ultimate basin fill depth d_s is related to the unfilled depth as

$$d_s = d_i\,\rho_m/(\rho_m - \rho_s) \qquad [21]$$

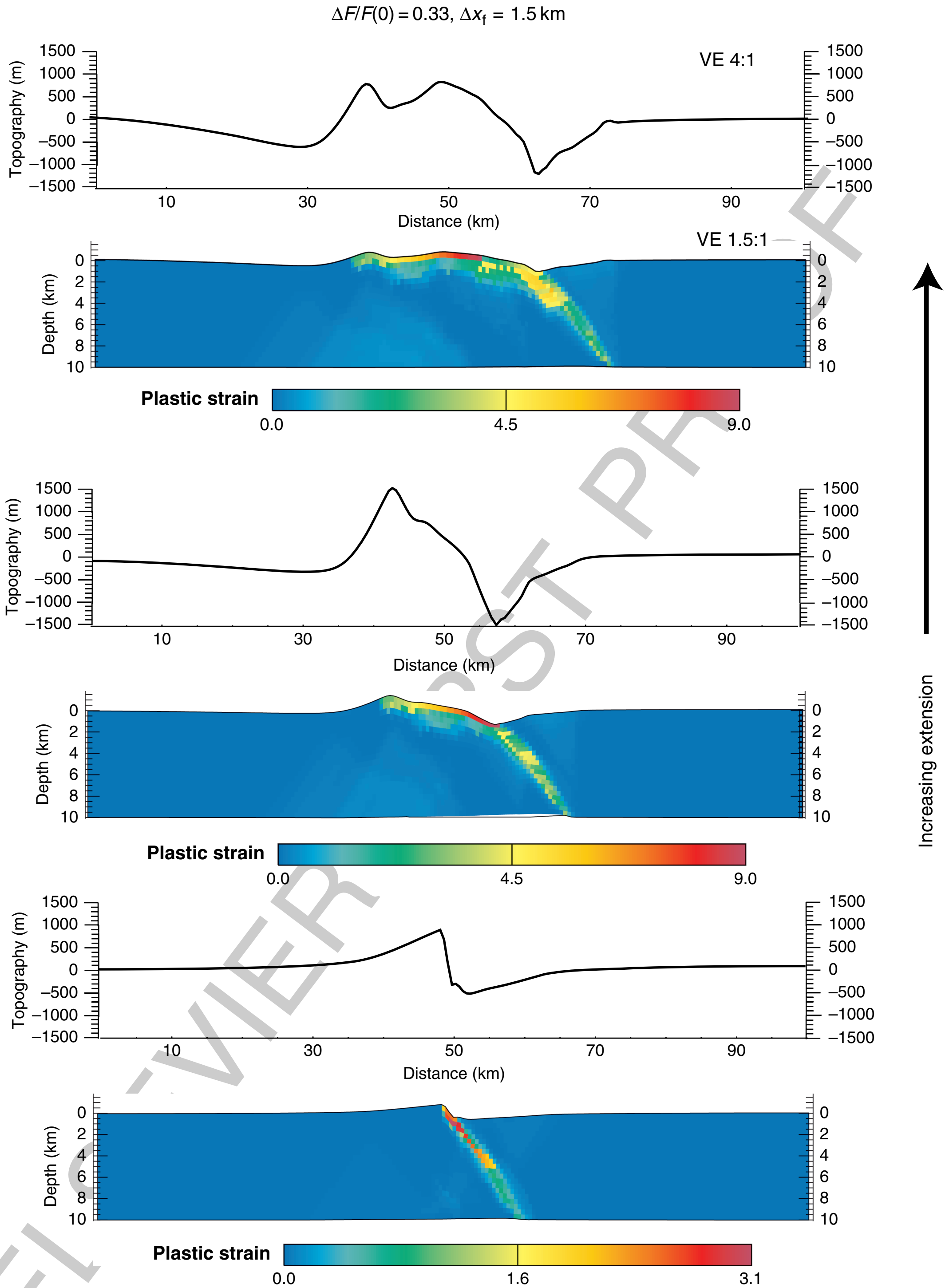
where ρ_m is the density of mantle. Taking $\rho_s = 2600$ kg m^{-3} and $\rho_m = 3250$ kg m^{-3}, a 1 km deep air filled basin would be filled with locally compensated sediment to 5 km. Deep holes might be likely to be filled in by water before being filled with sediment, but this intermediate step does not change the results of this simple analysis.

p0370 To get thick sedimentary packages requires a depression or hole. Wells drilled into rifted margins showed two interesting things about the thick sedimentary sequences. The sediments were deposited over a long time (tens to hundreds of million years) and the infilling sediments were generally deposited at shallow water depths in a lake or in an ocean basin (Sleep, 1971; Steckler and Watts, 1978). Since compensation should occur on a timescale of thousands of years, these observations imply that the hole the sediments were filling had subsided with time.

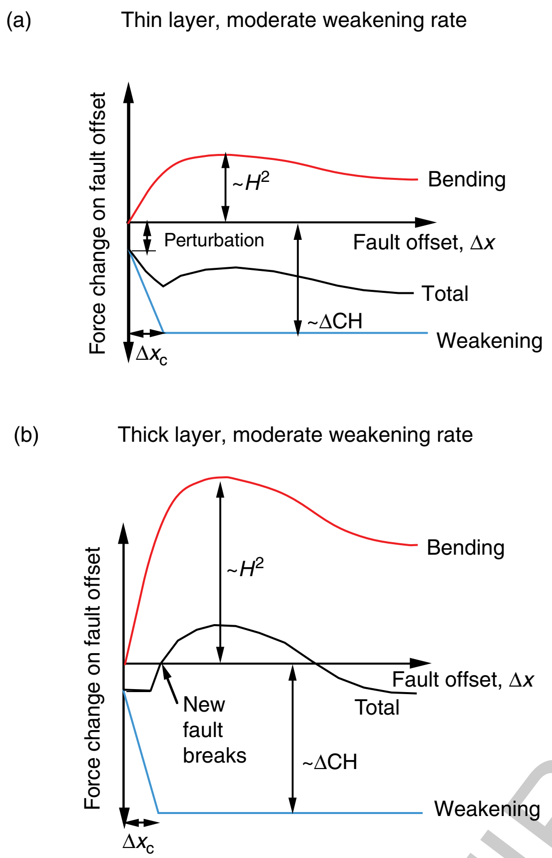
To understand the mechanism by which rift basins p0375
and rifted margins subside we need to understand subsidence of ocean basins. One of the most striking features of the ocean basins is the smooth pattern of depth increase with distance from the mid-ocean ridges. The depth of seafloor increases with the square root of its age (Parsons and Sclater, 1977). This is a clear confirmation of plate tectonics, which holds that the oceanic plates are formed when a hot mantle and magma upwells at a spreading center and then cools as they move away (see **Figure 22**). Since most plates move laterally much AU12
faster then they thicken by cooling (except very close to the spreading center) this cooling may be approximated by the cooling of a half-space of hot material. Lithospheric material contracts and so becomes denser as it cools. Thus, as the lithosphere cools and shrinks it will float lower on the asthenosphere. Combining half-space cooling with thermal contraction and local compensation predicts subsidence that is proportional to the square root of lithospheric age (e.g., Langseth and Taylor, 1967; McKenzie, 1967; Parsons and Sclater, 1977).

Sleep (1971) noted that the rate of subsidence of p0380
rifted margins also is related to the square root of sediment age. This suggested that rift margin subsidence might be controlled by lithosphere cooling down from a hot state at the time of rifting. However, there is still the problem of how to get a initially hot region to cool. Sleep reasoned that if continental lithosphere, with a surface close to sea level, were heated and thinned thermal expansion would float the surface above sea level. The surface would gradually subside back to sea level and no marine sediments could fill in. Something had to happen to lower the surface. The idea favored by Sleep (1971) was erosion of the uplifted surface. If a layer of low-density crust (compared to mantle) were removed then the surface could subside below sea level as cooling proceeded. However, the sediments that would be eroded off the elevated region are not seen.

Several authors (Artemjev and Artyushkov, 1971; p0385
Salveson, 1978; McKenzie, 1978) suggested that thinning of the lithosphere by pure shear extension created the thermal input needed to explain the later thermal subsidence. Because there is no active



f0095 **Figure 19** Results of a numerical model of extension of a floating brittle Mohr–Coulomb layer with a single-seeded normal fault. The model is similar to that shown in **Figure 17** but the calculated amount of extension is large enough to allow the footwall of the fault to rotate so that abandoned parts of the fault rotate to, and even past, horizontal. The cohesion loss of the fault was a function of strain with a decrease of one-third of the initial layer brittle yield strength occurring linearly with fault offset up to 1.5 km. From Lavier L, *et al*. (2000) Factors controlling normal fault offset in an ideal brittle layer. *Journal of Geophysical Research* 105(B10): 23431–23442.



f0100 **Figure 20** Schematic explanation of the reason that thin layers might allow unlimited normal fault offset while a thicker layer would not. The force change due to bending is compared to the assumed (or model input) fault weakening amount. If the summed force change is negative then the fault can continue slip (a) while if the force change becomes positive the fault should be replaced by another fault.

asthenospheric heating of the lithosphere in these models, they are termed 'passive'. The crustal thinning allows subsidence below sea level, or the original level of the continent, and sediment accumulation.

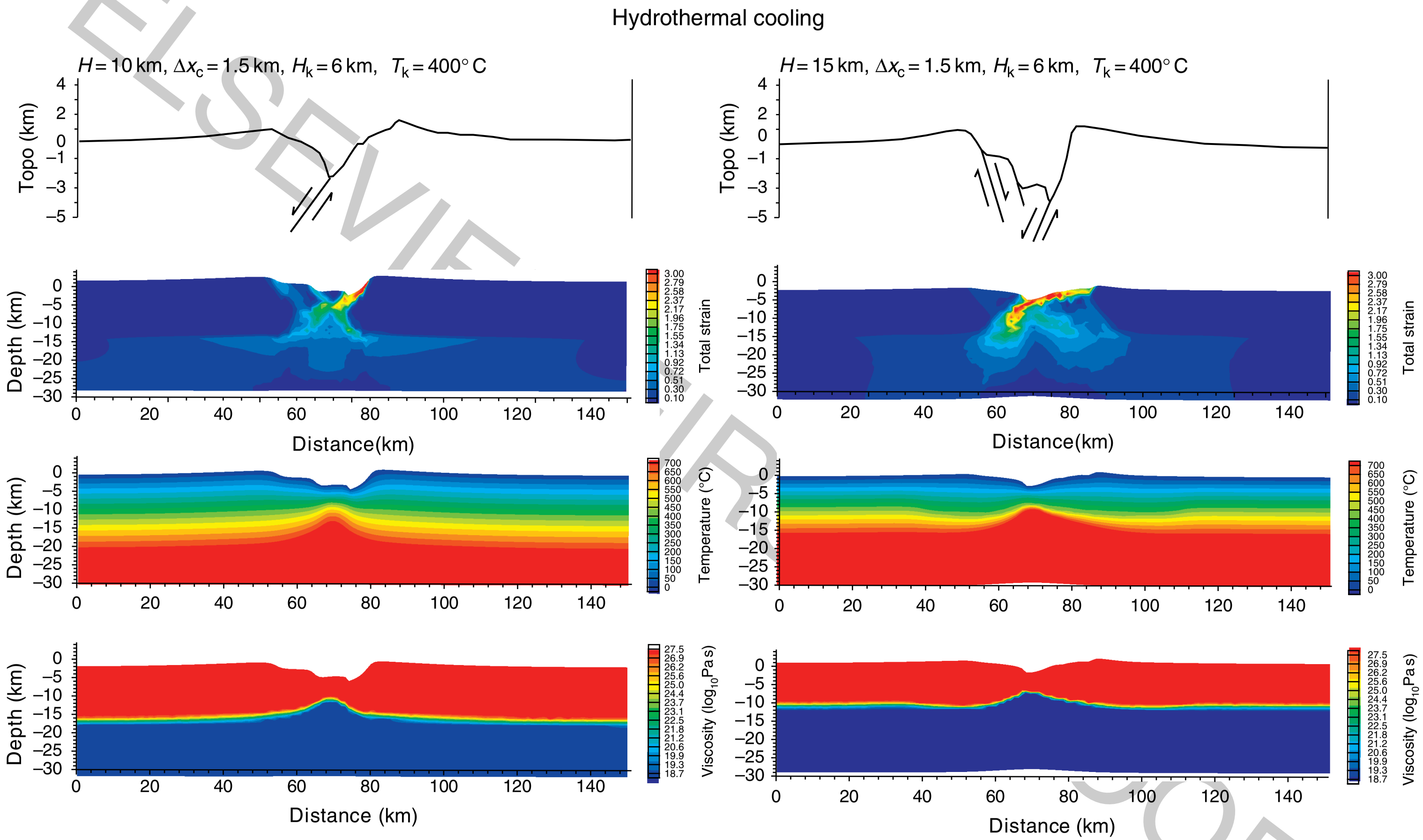
p0390 [b0500] Mckenzie (1978) quantified the effects of uniform pure shear extension or 'stretching' of the entire lithosphere. Stretching should thin both the crust and thermal lithosphere (see **Figure 23**). Local compensation of crustal thinning gives 'tectonic subsidence' on the timescale of the thinning. Lithospheric thinning causes immediate uplift followed by slow 'thermal subsidence'. In McKenzie's stretching model the rift is approximated as a rectangular region of extenional pure shear. Pure shear describes homogeneous thinning of an entire block of material by a stretching factor β given by the ratio of the initial to final thickness ($\beta > 1$ for extensional thinning).

If the initial thermal gradient in the lithosphere is given by dT/dz, instantaneous pure shear thinning will result in steepening of the gradient to $\beta dT/dz$. After rifting, time-dependent conductive cooling of the lithosphere eventually results in re-equilibration of the thermal gradient to its initial value. The stretching model is useful since it gives a 1-D description of the effect of lithospheric extension and so is easy to implement mathematically. [b0500] McKenzie's (1978) model neglects lateral heat flow and heat loss during the finite time duration of extension but matches the main characteristics of the subsidence histories of many margins (see **Figure 24**). p0395

The instantaneous uniform pure shear extension model was further generalized by Jarvis and McKenzie (1980) to include the effects of vertical heat loss during a finite duration extension event. They were able to model the heat flow and subsidence through a finite period of lithospheric stretching and thinning and the subsequent evolution after the end of the extension event. In general, the heat flow and subsidence history differed little from the instantaneous extension model, provided the time during which extension occurs is not too long. For a lithosphere thinned to half of its original thickness, for example, they found good agreement with the instantaneous case provided the duration of extension is less than 15 Ma. p0400

Though the stretching model gives a simple explanation of the approximate pattern of subsidence seen at most rifts it fails to explain the details of the subsidence. Where constraints on subsidence are good, the ratio of thermal subsidence to tectonic subsidence is greater than predicted by the stretching model (e.g., [b0585] Royden and Keen, 1980; [b0215] Davis and Kuszner, 2004; see **Figure 25**). The problem is even worse if the effects of lateral heat conduction are added to the model ([b0160] Cochran, 1983b; [b0015] Alvarez *et al.*, 1984). The syn-rift subsidence is augmented owing to the additional lateral heat loss, and the post-rift subsidence is diminished compared to the 1-D model. p0405

It appears that extra lithospheric thinning is needed compared to a standard stretching model. This observation led to the proposal of a two-layer stretching model ([b0585] Royden and Keen, 1980) in which the subcrustal lithosphere is decoupled from the crust and can be extended by a greater amount than the crust (see **Figure 26**). Although this model provides a simple way to specify the amount of lithospheric heating independently of the amount of crustal extension, it lacks a physical mechanism for this process. p0410



f0105 **Figure 21** Results of numerical model calculations for extension of thin and thick brittle lithosphere. For both cases, the rheology is viscous–elastic–plastic with the viscous strength depending strongly on temperature. The evolving temperature field is computed under the assumption that hydrothermal circulation cools the shallow part of the lithosphere. Extension of the hotter, thinner lithosphere leads to a single, large-offset fault. With extension of the cooler thicker layer multiple normal faults. From Lavier and Buck, 2003.

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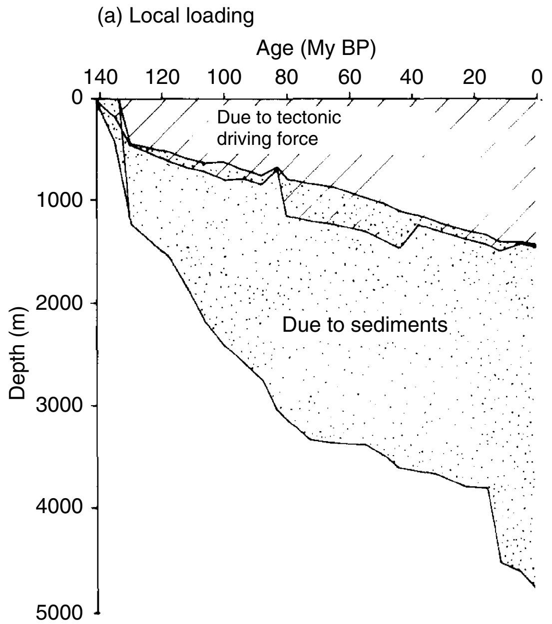
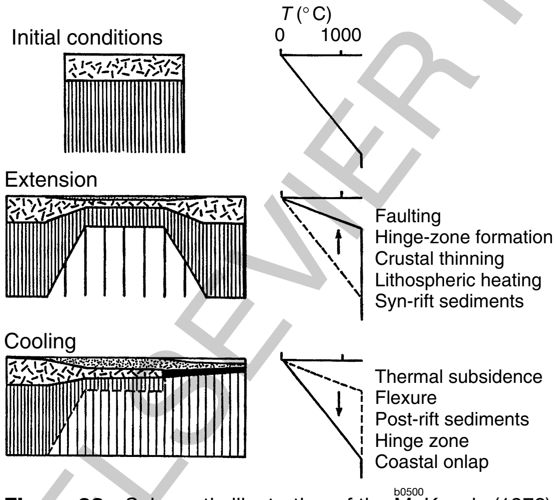


f0110 **Figure 22** Subsidence data for the margin of the western Mediterranean showing that subsidence was not driven just by sediment loading. Estimated components of subsidence with time due to the isostatic response to sediment loading compared to the subsidence that would occur without sediment infill, here labeled as being due to an (unspecified) tectonic driving force. From Watts AB and Ryan WBF (1976) Flexure of the lithosphere and continental margin basins. *Tectonophysics* 36: 25–44.



f0115 **Figure 23** Schematic illustration of the McKenzie (1978) lithospheric stretching model. From Steckler *et al.*, 1988. AU26

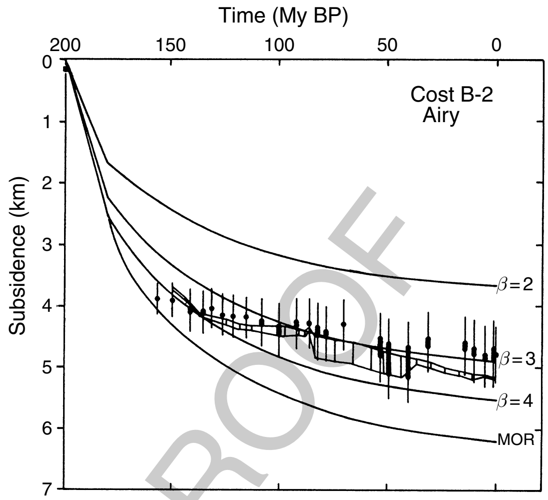


Figure 24 Comparison of subsidence data for the COST-B2 well on the US East Coast with predictions of the stretching model with different stretching factors, β. From Steckler *et al.*, 1988. f0120 AU27

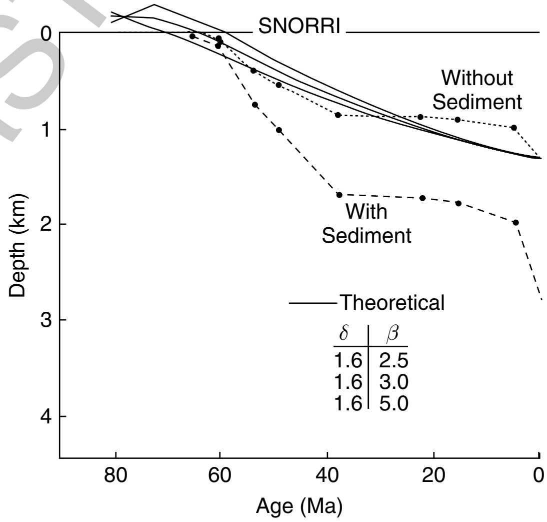


Figure 25 Subsidence data (dashed and dotted lines) for a well on the Labrador continental margin. Solid lines are predictions of a model with greater stretching of the lithosphere (factor β) compared to stretching of the crust (by a factor δ). From Royden L and Keen CE (1980) Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth and Planetary Science Letters* 51: 343–361. f0125

p0415 Another way to explain the observed increase in thermal relative to initial subsidence is to horizontally offset the locus of crustal versus lithospheric thinning. This is key a feature of the 'simple shear model' of lithospheric extension that was advanced by Wernicke (1985). One side of the rift is viewed as the upper plate and the other is the lower plate of gently dipping normal fault or detachment (see **Figure 27**). Offset of the detachment would lead to very asymmetric syn- and post-rift subsidence. On

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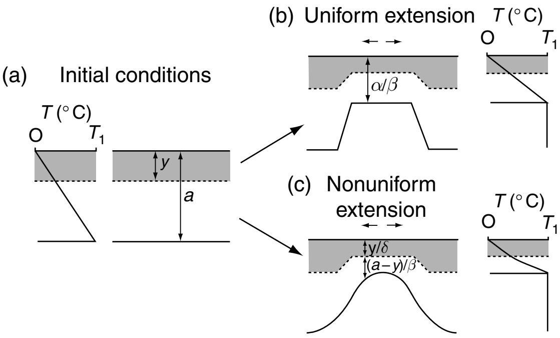


f0130 **Figure 26** Schematic of the 'two-layer' stretching model
AU28 in which the lithosphere is stretched by a different from the crust. From Royden L and Keen CE (1980) Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth and Planetary Science Letters* 51: 343–361.

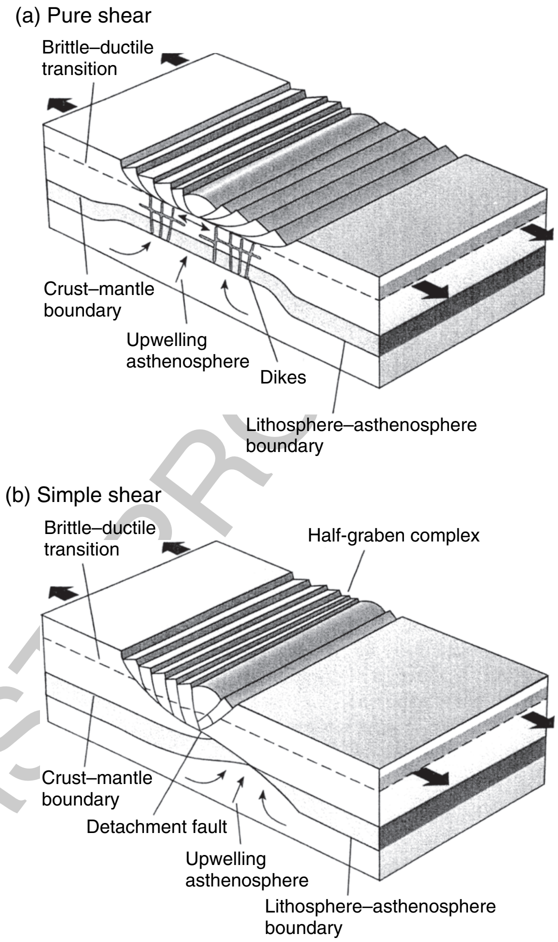


Figure 27 Schematic of structures formed by either pure shear or simple shear lithospheric extension. From Lister GS and Davis GA (1989) Models for the formation of metamorphic core complexes and mylonitic detachment terranes. *Journal of Structural Geology* 11: 65–94. f0135

the upper plate side areas of little crustal thinning
AU13 overly places where the lithosphere is greatly thinned. Thus, the upper plate side could be a place of little initial tectonic subsidence and large magnitude thermal subsidence, fitting the general pattern of observed rifted margin subsidence.

p0420 Several geological and geophysical observations have suggested the importance of simple shear extension in the crust which gives rise to asymmetric structures. Low-angle normal faults exposed at the Earth's surface have been traced to mid-crustal levels
AU14 using seismic reflection techniques. COCORP seismic lines in the eastern Basin and Range show low-angle normal faults that penetrate the upper and middle crust ([b0010]Allmendinger *et al.*, 1987). The Bay of Biscay, which has been considered a classic example of a pure shear margin ([b0220]de Charpel *et al.*, 1978), is interpreted by [b0445]Le Pichon and Barbier (1987) to show evidence of crustal-scale simple shear along a detachment. Metamorphic core complexes show that rocks can come up from mid-crustal levels apparently in association with extensional shear zones ([b0210]Davis, 1983). The observation of synthetically dipping normal faults over broad areas (e.g., the southern Basin and Range; e.g., [b0700]Stewart, 1978) has been taken as evidence for simple shear ([b0790]Wernicke, 1981).

p0425 Strong topographic and volcanic asymmetries exist across some rifts and conjugate passive margins, among them the Red Sea Rift ([b0785]Wernicke, 1985), East African Rift ([b0060]Bosworth, 1987) and the Southeast Australian–Lord Howe Rise conjugate margin (Lister *et al.*, 1988). The direct observations of simple shear extension are confined to the crust. However, the observation of normal faults extending to great depth in the crust combined with observations of topographic asymmetries across rifts have led to the suggestion that normal faults and/or ductile shear zones may extend through the entire lithosphere ([b0790]Wernicke, 1981, [b0785]1985; [b0470]Lister *et al.*, 1986). Such lithospheric detachment models imply that much of the deformation in an extending region occurs as simple shear rather than pure shear.

Particularly good data to test the predictions of the simple shear model come from studies of the Northern Red Sea. The asymmetric uplift and volcanism bordering the Red Sea is well known and has been cited as evidence for a through-going lithospheric detachment ([b0785]Wernicke, 1985). The kinematic history of opening of the northern Red Sea is well constrained by geological and geophysical data. Although rifting in the Red Sea began around p0430 the end of the Oligocene–Early Miocene ([b0165]Cochran, 1983a), evidence from the Gulf of Suez and Gulf of Aqaba ([b0675]Steckler *et al.*, 1988a) suggests that in the northern Red Sea, most of the extension has occurred within the last 19 My and that it continues to the present. The young age of this rift is particularly important for discriminating between models of its formation because many of the effects are transient and will not be observable tens of millions of years after rifting.

p0435 Geophysical fieldwork in the northern Red Sea provided excellent heat-flow coverage of this area ([b0170]Cochran *et al.*, 1986; [b0490]Martinez and Cochran, 1988). To compare the predictions of the simple shear model to this data [b0130]Buck *et al.* (1988) used a numerical technique that solved for 2-D conductive and advective heat transport through time. Simple shear extension of the lithosphere was modeled as occurring along a straight shear zone and the topographic response to simple shear could be described in terms of local isostasy.

p0440 The long wavelength topographic asymmetry across the Red Sea, which has been cited as evidence for simple shear extension of the lithosphere, was not matched by any of the simple shear model cases. The observed high heat flow anomalies in the Red Sea require a large component of pure shear lithospheric extension centered under the region of maximum crustal extension. In contrast, at the plate separation rate of the northern Red Sea, simple shear extension of the lithosphere along a shallow (<30°) dip detachment is ineffective in reproducing the observed heat-flow anomalies. Only a narrowing region of pure shear extension can satisfy the width of the rift, and the peak observed heat-flow values of $\sim$300 mW m^{-2}.

p0445 Other, older continental margins have been studied in terms of whether various kinematic extensional models can explain their structure and subsidence patterns. However, as conjugate margins have been studied it seems that both rifted margins subside the way this model predicts for the upper plate margin. This has been described by [b0240]Driscoll and Karner (1998) as the 'upper plate paradox'.

s0110 6.09.5 Wide versus Narrow Rifts

p0450 It has long been recognized that extension of continents can result in either narrow or wide rifts (e.g., [b0265]England, 1983). For narrow rifts, such as the Rhinegraben ([b0365]Illies and Greiner, 1978), the Gulf of Suez (Steckler *et al.*, 1988b), the East African Rift System (e.g., [b0250]Ebinger *et al.*, 1989), the Baikal Rift (e.g., [b0845]Zorin, 1981), and the Rio Grande Rift (e.g., [b0520]Morgan *et al.*, 1986), the region of intense normal faulting is on the order of 100 km wide. For wide rifts, the type example of which is the Basin and Range Province of western North America, the region of significant normal faulting is as much as 800 km wide ([b0315]Hamilton, 1987). The Basin and Range is characterized by small lateral gradients in crustal thickness, while narrow rifts may show large lateral gradients in thickness of crust and in topography.

Most rifts are relatively narrow (**Figures 1** and **4**) and this is generally ascribed to the weakening effect of thermal advection. The lithosphere can be thinned by stretching or heated by injection of magma. Since hot crust and mantle materials flow at lower yield stresses (eqn [4]) a hotter region will be weaker. Extension should stay concentrated in the region of weakest lithosphere. The zone of extensional strain should remain confined to a region that is about as wide as the lithosphere is thick. p0455

If a region of focused extension should become effectively stronger than the surrounding region then one might expect the locus of extension to migrate to the weaker regions. If this continued then a region of rifting that is wider than the lithospheric thickness might develop. Three delocalizing processes have been suggested as ways to get wide rifts: (1) thermal diffusion strengthening, (2) viscous strain-rate strengthening, and (3) the change in gravitational stresses due to crustal buoyancy. The scaling relations derived in an earlier section help us to evaluate the potential of each of these mechanisms to produce a wide rift. p0460

6.09.5.1 Slow Rifting and Thermal Diffusion s0115

[b0265]England (1983) used the thin-layer approximate to investigate whether wide rifts could occur due to thermal diffusion strengthening. He considered the evolution of continental yield strength during extension and noted that, if the mantle is much stronger than the crust, then very slow extension could lead to a strength increase in the extending area. The basic idea is that weak crust is thinned by stretching and isostatically 'replaced' by intrinsically stronger mantle. Diffusive cooling has to be significant to have this compositional strengthening trump the advective weakening effect. To get this effect to produce strengthening during extension [b0265]England (1983) had p0465

to assume that mantle is much stronger than crust and that extension rates were very slow.

p0470 According to England (1983), as well as other workers (Kusznir and Park, 1987; Sonder and England, 1989), the diffusive strengthening effect explains the occurrence of wide regions of continental extension like the Aegean Sea and the Basin and Range Province. However, observations do not seem to confirm this prediction since some wide rifts did not extend at a slower velocity than did some narrow rifts.

p0475 Sawyer (1985) questioned the validity of the England (1983) analysis since it did not consider the effect of finite brittle yield strength of the crust and mantle. The yield strength in the England (1983) model only involved the viscous yield strength. Sawyer (1985) has questioned whether the crust–mantle strength contrast is likely to be as large as assumed by England (1983). At a range of temperatures likely to be found at the base of the crust, the ductine yield strength of crust can be thousands of megapascals lower than that of the mantle. Sawyer (1985) noted that the brittle failure stress may be independent of rock type so that the contrast in crust and mantle strength is limited. When even conservative brittle yield strength values are included it greatly lowers the threshold extension rates that would lead to lithospheric strengthening and so widening of a rift. Also, if the crust and mantle are dry, then the viscous yield strength of the crust may not be significantly different from that of the mantle (Kohlstedt *et al.*, 1995; Mackwell *et al.*, 1998).

p0480 Naturally, the importance of diffusion relative to advection of heat depends strongly on the rate of deformation. Scaling arguments can give a good idea of the strain rates that could produce extensional strengthening. The force for continued extension is reduced due to advective thinning by roughly $2\varepsilon C H_{\mathrm{b}}^{2}$. The best case to make the slow-strain-strengthening mechanism work is when the brittle layer thickness H_{b} is close to the crustal thickness H_{c}. The force for continued extension is reduced due to advective thinning by $\sim 2C\kappa(\varepsilon/\dot{\varepsilon})$. A ratio of the advective and diffusive force changes should be less than 1 for cases where the diffusive effect dominates. This occurs when $\dot{\varepsilon} < \kappa/H_{\mathrm{c}2}$. For $H_{\mathrm{c}} = 40$ km and $\kappa = 10\text{–}6$ m^2 s^{-1} the maximum estimated strain rate is 10–15 s^{-1}. This is an extremely low extensional strain rate. To get such a low strain rate the side of a 40 km wide zone of extension would have to be separating a velocity of less than 1 mm yr^{-1}! This scaling shows that the diffusive strengthening effect becomes more important for thinner crust, assuming that the base of the crust is hot enough for the crust there to be ductile and so weaker than the mantle. Still, White (2004) suggests that England's (1983) mechanism might explain the abandonment of many intracratonic rifts that seem to have extended at very low strain rates.

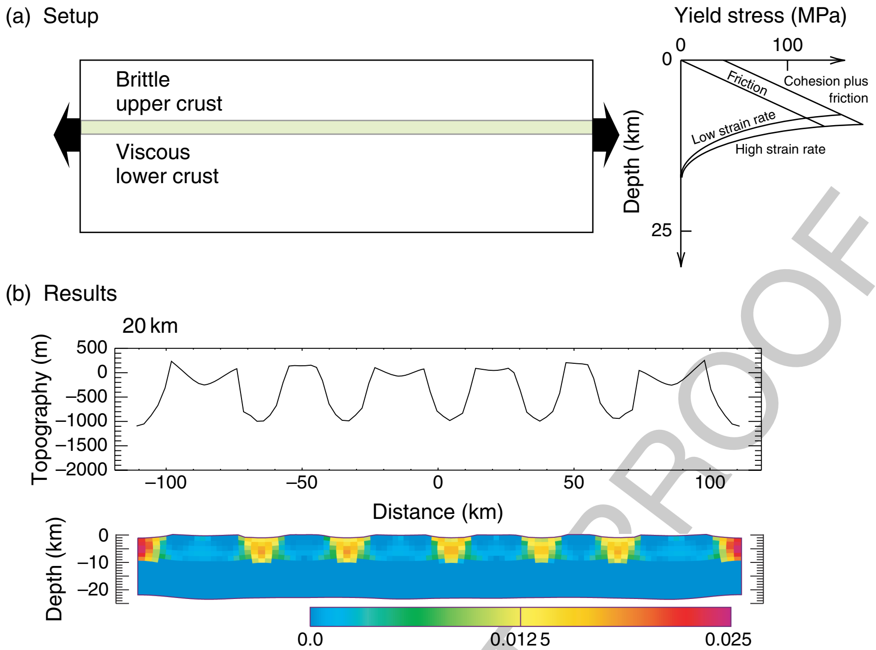
6.09.5.2 Viscous Stresses s0120

Another possible effect to explain how wide rifts form relates to the stresses needed for yielding by viscous flow. The ductile yield stress in a representative flow law (eqn [4]) depends on strain rate. Thus, any area extending at a high rate will extend at higher stresses. This idea prompted Bassi (1991) to suggest that wide rifts could be a product of viscous stress resisting localization of deformation. p0485

It is difficult to consider using simple scaling relations to look at competition between viscous strain-rate strengthing and other effects. The difficulty is that the viscous effect is independent of strain and most weakening (localizing) effects increase with increasing strain. To analyze the effect of viscosity on deformation a linear stability approach is often used. The linearization of the model equations, implicit in these calculations, means that these results are strictly applicable only to the earliest phase of extension when strains are small. These models predict only two forms of extension of an initially uniform lithosphere, both of which could be termed a wide rift (e.g., Fletcher and Hallet, 1983; Zuber and Parmentier, 1986; Martinod and Davy, 1992). The lithosphere extends either in a uniform pure shear mode, or with laterally periodic variation in the rate of extension. This second type of extension, often called 'lithospheric boudinage', is suggested to explain the development of a series of basins and ranges in a broad region of continental extension. p0490

Linear stability analysis does not predict a narrow rift mode of extension. Since many rifts, like the Rhine graben and the Red Sea are narrow, one wonders what situations lead to their development as compared to wide rifts like the Basin and Range. Zuber and Parmentier (1986) and Bassi (1991) show that a large initial perturbation in lithospheric thickness or strength might cause localization in a single narrow rift, as opposed to a sequence of basins and ranges. p0495

The effect of viscous stresses on the distribution of extension in finite amplitude (as opposed to linear stability) numerical model has been treated by a number of workers (e.g., Bassi, 1991; Huismans *et al.*, 2003, 2005; Buck *et al.*, 2003; Nagel and Buck, 2006). **Figure 28** shows the distribution of strain in a series p0500



f0140 **Figure 28** (a) Numerical setup and (b) results for a model in which the viscosity of the lithosphere forces a distributed pattern of extension. Color scale indicated brittle plastic strain in a series of local basins that are separated by higher ranges. After Buck WR, *et al*. (2003) A numerical model of lithospheric extension producing fault bounded basins and ranges. *International Geology Review* 45(8): 712–723.

of basins and ranges in a model with viscous effects driving delocalization. The viscous effects depend both on the viscosity variation with depth at the base of the lithosphere and on the extent to which viscous material weakens as a functions of strain or strain rate. In no models with large viscous delocalization do we get model large offset faults (e.g., b0535 Nagel and Buck, 2006). Thus, the existence of localized large-offset strike slip and normal faults argues against the importance of viscous delocalization.

s0125 6.09.5.3 Local Isostatic Crustal Thinning

p0505 The gravitational stresses related to local thinning of the crust could lead to wide rifting. To do this, the stretching-related gravitational delocalizing effect has to be larger than the localizing (weakeing) effect of lithospheric advective thinning. The scaling relations derived in Section 6.09.2 give an indication of what controls these two effects. The force for continued extension is reduced due to advective thinning by roughly $2\varepsilon C H_b^2$. For reasonable friction coefficients and hydrostatic pore pressures in the crust the brittle yield constant $C \sim 14 \times 10^3\,\mathrm{Pa\,m^{-1}}$. For a brittle layer thickness $H_b = 20\,\mathrm{km}$ and a strain $\varepsilon = 10\%$, this scaling estimate is for weakening of $\sim 5 \times 10^{11}\,\mathrm{Nt\,m^{-1}}$. AU15

p0510 As discussed in the 'local isostasy' section, the local isostatic gravitational force change for crustal stretching is $2\varepsilon(\rho_c/\rho_m)\ (\rho_m - \rho_c) g H_b^2$. For crust 50 km thick with density $2800\,\mathrm{kg\,m^{-3}}$ overlying mantle with a density of $3200\,\mathrm{kg\,m^{-3}}$ a 10% strain would produce a force increase of $\sim 10^{12}\,\mathrm{Nt\,m^{-1}}$. Thus, these simple scalings indicate that the wide rift mode should occur when the continental crust is thicker and hotter than normal. Extension of lithosphere with normal crustal thickness and average heat flow gives a narrow rift because the advective weakening dominates.

p0515 The main question about the crustal buoyancy mechanism for wide rifting concerns the hot lower crust. The mechanism should only produce wide rifting if the lower crust has to be weak, but not too weak. To understand this we need to consider how flow of the lower crust will tend to fill in any area of crustal thinning. Crustal thinning lowers pressures in the crust where the crust is thinned. Surrounding crust will pour into this region if any part of the crust is weak enough to flow. The lower crust should be the hottest part of the crust and so, if there is little or no compositional stratification of the crust, the lower crust should have the lowest effective viscosity. Hot, low-viscosity lower crust may flow into the region of crustal stretching; thus, reducing the crustal thickness variations needed for buoyancy resistance to continued extension. If the lower crust is extremely hot and weak, then this model predicts a 'core complex mode' of extension, in which the locus of upper-crustal extension remains fixed in one place while the lower crust thins over a broad region.

p0520 To more quantitatively access this idea for wide rifting Buck (1991) combined simple calculations of four of the processes discussed in the earlier section on rift processes: (1) local isostatic crustal buoyancy, (2) lower crustal flow, (3) thermal advection, and (4) thermal diffusion. Extension was assumed to be concentrated in a single region of pure shear stretching and the width of the region equaled the lithospheric thickness. The strain rate of stretching was fixed and changes in the vertical thermal structure of the straining region were computed using a 1-D heat equation. Changes of crustal thickness and buoyancy were calculated using a 1-Dl thin channel equation to describe advective crustal thinning and lower-crustal flow. After a small amount of extension the change in force was computed. If the force decreased the extension was considered to remain in the initial location while if the force increased the extension was taken to migrate. Migration of extension would eventually produce a wide rift. Because of the simplifying approximations of the model, millions of parameter combinations could be tested using modest amounts of computer time. Thus, maps of the wide rifting mode could be plotted in terms of crustal thickness and model heat flow for different crustal rheologies and assumed extension rates (**Figure 29**). The model predicts that wide rifting could result from crustal buoyancy for reasonable rheologies. The range of conditions for model wide rifts is broadly consistent with observations (**Figure 30**).

p0525 The possibility that wide rifting cold be controlled by crustal buoyancy has been suggested by more complete thin-layer calculations (Hopper and Buck, 1996) and by fully 2-D thermo-mechanical calculations (e.g., Christensen, 1992). For all these approaches the wide rift and the 'core complex' modes are only predicted for thick, hot crust.

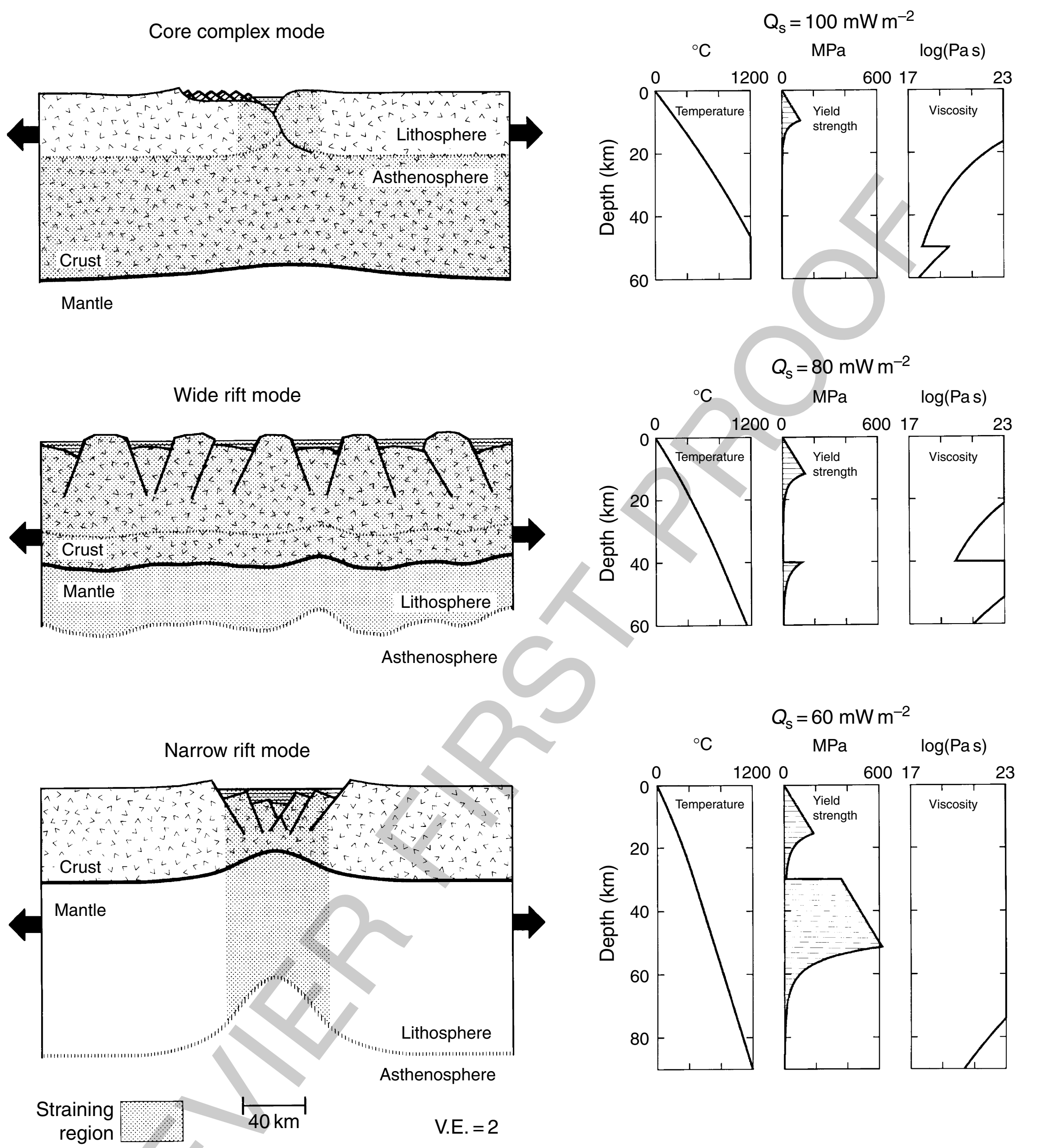
p0530 As noted in Section 6.09.1 the main place where crust is thickened and heated is in orogens; thus, it is not surprising that most wide regions of rifting like the Basin and Range Province and the Aegean occur in regions where the crust was orogenically thickened. Indeed, it appears that extension driven by collapse of high mountains can occur even while convergence is still occurring across those mountains (see Dewey, 1988; Willet and Pope, 2004). Wide rifts may just be the result of extension of particularly wide orogenic belts where the crust has become fairly uniformly thick and hot as for the Tibetan Plateau (Royden *et al.*, 1997).

6.09.6 Dikes versus Stretching to Initiate Rifting s0130

Areas of initially thin lithosphere should rift at very low levels of tectonic force. Further, some areas of high heat flow and initially thick crust, such as the North American Basin and Range Province, start extending with little or no basaltic volcanism. Models neglecting magmatism do predict the general patterns of observed extensional strain inferred for such areas. It should be noted that these 'hot' weak areas are not typical, but may require some kind of preheating. The effect of orogenesis, especially thickening of radiogenic crust, has been implicated as a way of heating regions such as the Basin and Range and the Aegean Sea extensional provinces (e.g., Sonder *et al.*, 1987). However, in areas of low-to-normal heat flow, the earliest phase of rifting is often accompanied by basaltic magmatism. p0535

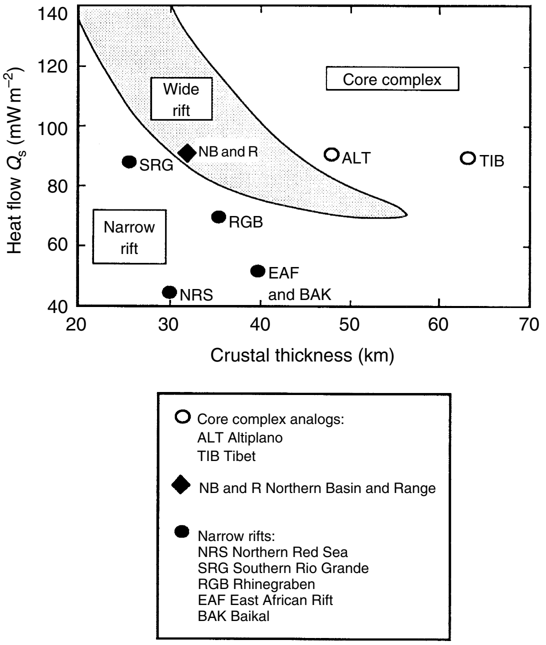
Though it has long been known that many rifts are associated with volcanism (**Figure 31**), a fundamental change in our understanding of rifted margins has been driven by new observations concerning the volume and flux of magmatism associated with those margins. There has been a gradual realization that most rifted margins are temporally and spatially associated with very large magma input. Seismic data has been one of the key inputs that have lead to a change from emphasis from passive lithospheric stretching to active magma-assisted rifting. It was the seismic imaging of 'seaward dipping reflectors' along many margins that led to increased interest in the role of magma during rifting. p0540

Many rifts and rifted margins are associated with large igneous provinces, but the igneous provice, where there are thick layers of subareal extruded basalt account for a small fraction of the length of the rift. Seaward-dipping reflectors are seismically bright layers that dip ocean-ward on many rifted margins such as parts of the South Atlantic Margin (Hinz, 1981) and the North Atlantic Greenland Margin (e.g., Holbrook *et al.*, 2001). Drilling into the units has confirmed that these are basalt flows (Larsen *et al.*, p0545



f0145 **Figure 29** Illustration of model results predicting different forms of extension depending on the initial lithospheric thermal structure and crustal thickness. Both the core complex model and the wide rift mode are cases where warm crust is driven by gravity (local isostatic forces) to deform over a region that may be wider than the lithospheric thickness. From Buck WR (1991) Modes of continental lithospheric extension. *Journal of Geophysical Research*, *B*, *Solid Earth and Planets* 96(12): 20161–20178.

1993). These dipping relectors are thought to form the same way tilted basalt flows form on Iceland where flows from a rift center load and depress the surface isostatically. As the rift widens the center of magmatic fissure eruptions moves away from the marging, resulting in the kinds of pattern schematically shown in **Figure 32**. The imaging of seaward-dipping reflectors and deeper high-velocity units that may be mafic intrusives (see **Figure 5**) indicates that many more margins may be considered to be 'volcanic' than was previously thought. It is estimated that many millions of cubic kilometers of basalt is intruded and extruded along some rift systems (e.g., b0330 Holbrook *et al.*, 2001) (**Figure 33**). AU16



f0150 **Figure 30** Predicted modes of extension as a function of crustal thickness and surface heat flow compared to data for regions showing those modes of extension, or in the case of Altiplano and Tibet having the conditions that might lead to core complex formation. From Buck WR (1991) Modes of continental lithospheric extension. *Journal of Geophysical Research*, *B*, *Solid Earth and Planets* 96(12): 20161–20178.

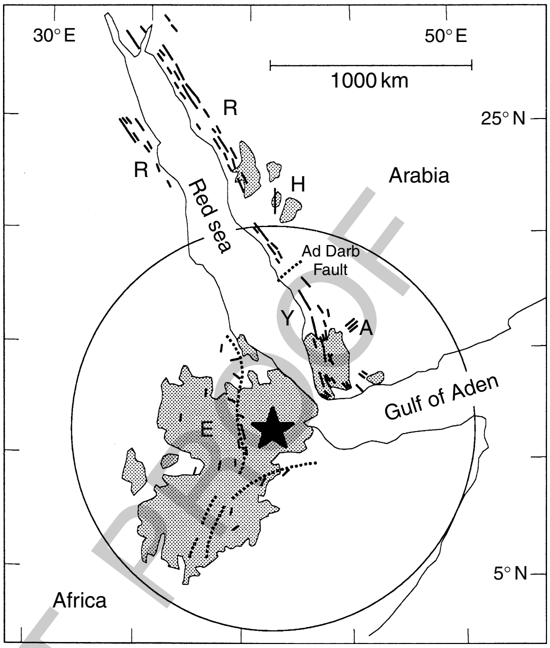


Figure 31 Map of the distribution of the Ethiopian flood basalts and related dikes along the Red Sea. From Ernst RE and Buchan KL (1997) Giant radiating dyke swarms; their use in identifying pre-Mesozoic large igneous provinces and mantle plumes. In: Mahoney JJ and Coffin MF (eds.) *Geophysical Monograph: Large Igneous Provinces*; *Continental*, *Oceanic*, *and Planetary Flood Volcanism*, pp. 297–333. Washington, DC: American Geophysical Union. f0155

p0550 There are several models for the origin of magma that affects volcanic margins. A key controversy is whether the magmatism caused the rifting or vice versa. The idea that rifting causes volcanism comes from considering the way lithospheric stretching advects hot mantle upward and so can lead to melting of that passively flowing mantle (e.g., White and McKenzie, 1989). However, several lines of evidence support the idea of some kind of active mantle upwelling to trigger rifting. First, volcanism is often spread regions that does not include apparently significant rifting and therefore stretching (**Figures 34** and **35**). Dikes appear to radiate from a central location that may be the locus of plume upwelling (**Figures 35** and **36**). Also, the volcanism seems to predate rifting in many rifts (**Figure 37**).

AU17

p0555 Rifts and passive margins associated with large igneous provinces tend to be nearly straight. No rift is perfectly straight, but if a great circle can pass through part of every segment of a rift, then that rift could be said to be straight. For many rifts, like the Aden Rift, the small-scale segments of the rift do not parallel the coastline or the border faults of the rift

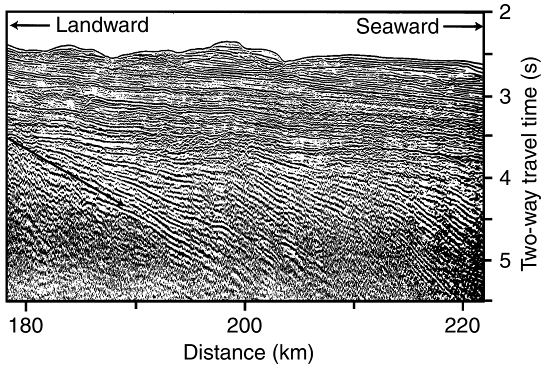
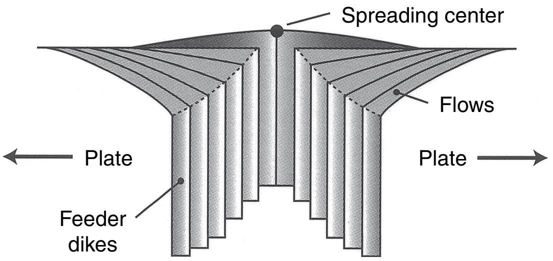
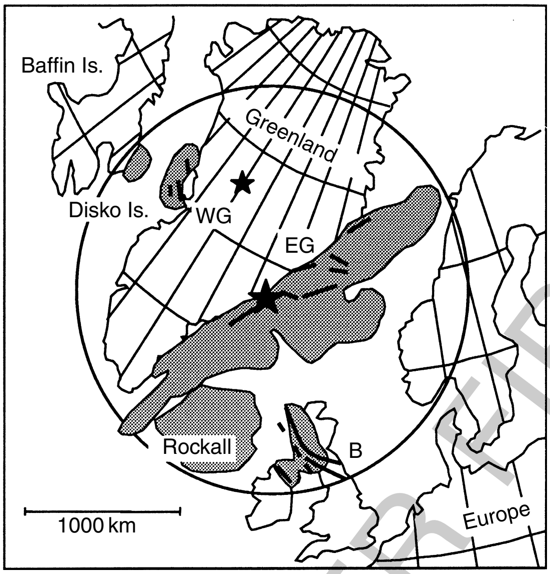


Figure 32 Example of a multichannel seismic profile with arrow showing the prominent seaward-dipping reflectors along the East Greenland Margin. From Hopper JR, *et al*. (2003) Structure of the SE Greenland margin from seismic reflection and refraction data; implications for nascent spreading center subsidence and asymmetric crustal accretion during North Atlantic opening. *Journal of Geophysical Research*, *B*, *Solid Earth and Planets* 108: 22. f0160



f0165 **Figure 33** Illustration of the isostatic model of formation of seaward-dipping reflectors as basalt flows coming out of a rift spreading center. After Palmason G (1980) A continuum model of crustal generation in Iceland: kinematic aspects. *Journal of Geophysical Research* 47: 7–18.



f0170 **Figure 34** Distribution of basalts and dikes for East Greenland (EG), West Greenland (WG), and the British Tertiary Basalt Provinve (B). From Ernst RE and Buchan KL (1997) Giant radiating dyke swarms; their use in identifying pre-Mesozoic large igneous provinces and mantle plumes. In: Mahoney JJ and Coffin MF (eds.) *Geophysical Monograph: Large Igneous Provinces; Continental, Oceanic, and Planetary Flood Volcanism*, pp. 297–333. Washington, DC: American Geophysical Union.

(e.g., Leroy *et al.*, 2004). This is not surprising since the stress orientation and opening direction can change after the initial phase of opening.

p0560 The Red Sea is the clearest example of a straight rift with the coastlines showing only slight deviations from a great circle over 2000 km and the axis of spreading and rifting is even straighter than the coast. The South Atlantic Margins of South America and Africa, associated in space and time with the Parana Flood Basalts

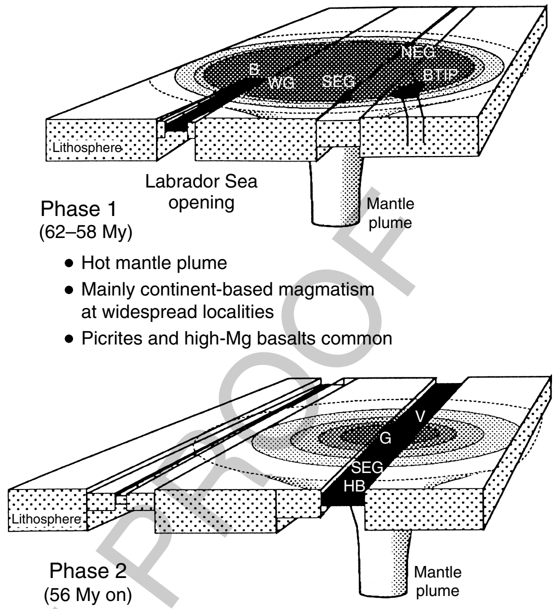


Figure 35 Illustration of the idea that plumes may have fed magma to West and East Greenland during the formation of that large igneous province and those rifted margins. From Saunders AD, *et al*. (1997) The North Atlantic igneous province. In: Mahoney JJ and Coffin MF (eds.) *Large Igneous Provinces*; *Continental*, *Oceanic*, *and Planetary Flood Volcanism*, vol. 100, pp. 45–93. Washington, DC: American Geophysical Union. f0175

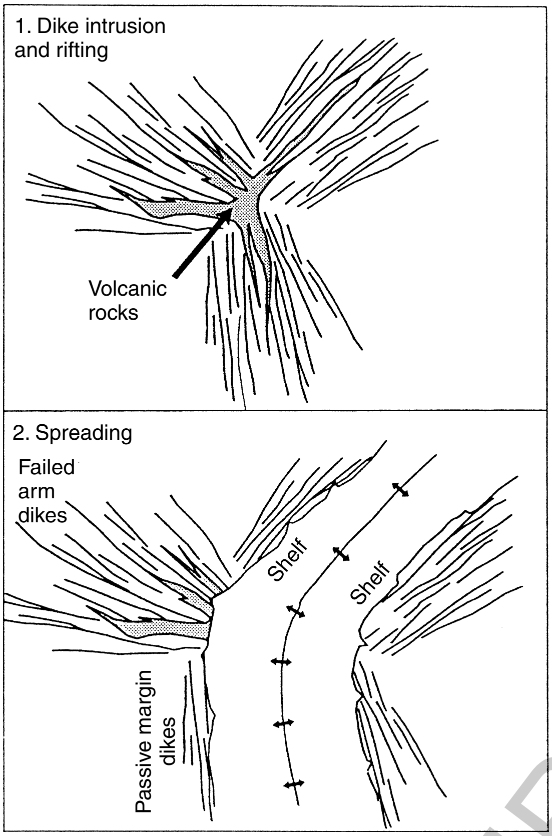
(Hinz, 1981), are another very straight margin. The dike intruded and volcanic covered North Atlantic Margins of Greenland (e.g., Holbrook *et al.*, 2001) and the conjugate European margins are also remarkably straight on a scale of thousands of kilometers.

A magmatic rift does not have to be volcanic. The p0565
Northern Red Sea and the Gulf of Suez lack evidence of massive syn-rift volcanism (forming many-kilometer-thick seaward-dipping seismic relector packages) that have been documented along much AU18
of the South Atlantic and Greenland Margins (Hinz, 1981; Mutter and Zehnder, 1988). However, synrift dikes striking parallel to the rift are seen as far north as the Gulf of Suez and valley filling, synrift volcanics are common there as well (Patton *et al.*, 1994).

In this section simple estimates of the effect of p0570
copious magma on the force to rift will be made and compared to the estimated force available for rifting.

6.09.6.1 Force Available for Driving Rifting s0135

Many authors have discussed ways that plate tectonic p0575
and plume-related processes could produce relative tension at a rift (e.g., Forsyth and Uyeda, 1975;



f0180 **Figure 36** Illustration of how giant radiating dike swarms are often seen to relate to the geometry of subsequent rifting. From Ernst RE and Buchan KL (1997) Giant radiating dyke swarms; their use in identifying pre-Mesozoic large igneous provinces and mantle plumes. In: Mahoney JJ and Coffin MF (eds.) *Geophysical Monograph*: *Large Igneous Provinces; Continental*, *Oceanic*, *and Planetary Flood Volcanism*, pp. 297–333. Washington, DC: American Geophysical Union.

Solomon *et al.*, 1980), and these approaches give similar estimates of rift-driving forces. The simplest of these approaches relates to the uplift that may occur over a region of abnormally hot mantle. The radiation of rift branches away from the uplifted plateau areas of Ethiopia and Yemen are consistent with the extensional driving force being related to the uplift of that region (e.g., Sengor and Burke, 1978; Ernst *et al.*, 1995).

p0580 For uplift due to a low-density root, the extensional force scales with the magnitude of the uplift, e, and the depth of the low-density compensation layer, d (Spohn and Schubert, 1982; Bott, 1991). The force is roughly $e\rho_{\mathrm{m}}gd$, where ρ_{m} is the mantle density and g is the acceleration of gravity. For a root of uniform density, hot mantle between 100 and 200 km, d would be 150 km. Then $\sim$1 km of uplift compensated at 150 km depth gives $\sim 5 \times 10^{12}\,\mathrm{Nt\,m^{-1}}$ of rift driving force. Higher elevations and deeper compensation depths are possible, but the force may be spread over a longer rift than the width of the uplifted area. Thus, the average level of rift force is likely to be less than about $5 \times 10^{12}\,\mathrm{Nt\,m^{-1}}$.

6.09.6.2 Force Needed for Tectonic Rifting s0140

Recall from Section 6.09.2 of this chapter that the tectonic yield strength depends on the thermal structure of the lithosphere and to a lesser extent on the composition of the lithosphere. To compare the force for tectonic extension to that required for magmatic extension it is sufficient to ignore the possibility that crust is weaker in terms of ductile flow than mantle. To the extent that we can ignore the contribution of ductile yield stress to the lithospheric strength and if we take the density of the lithosphere constant with depth we can get a simple estimate of the tectonic force for rifting: p0585

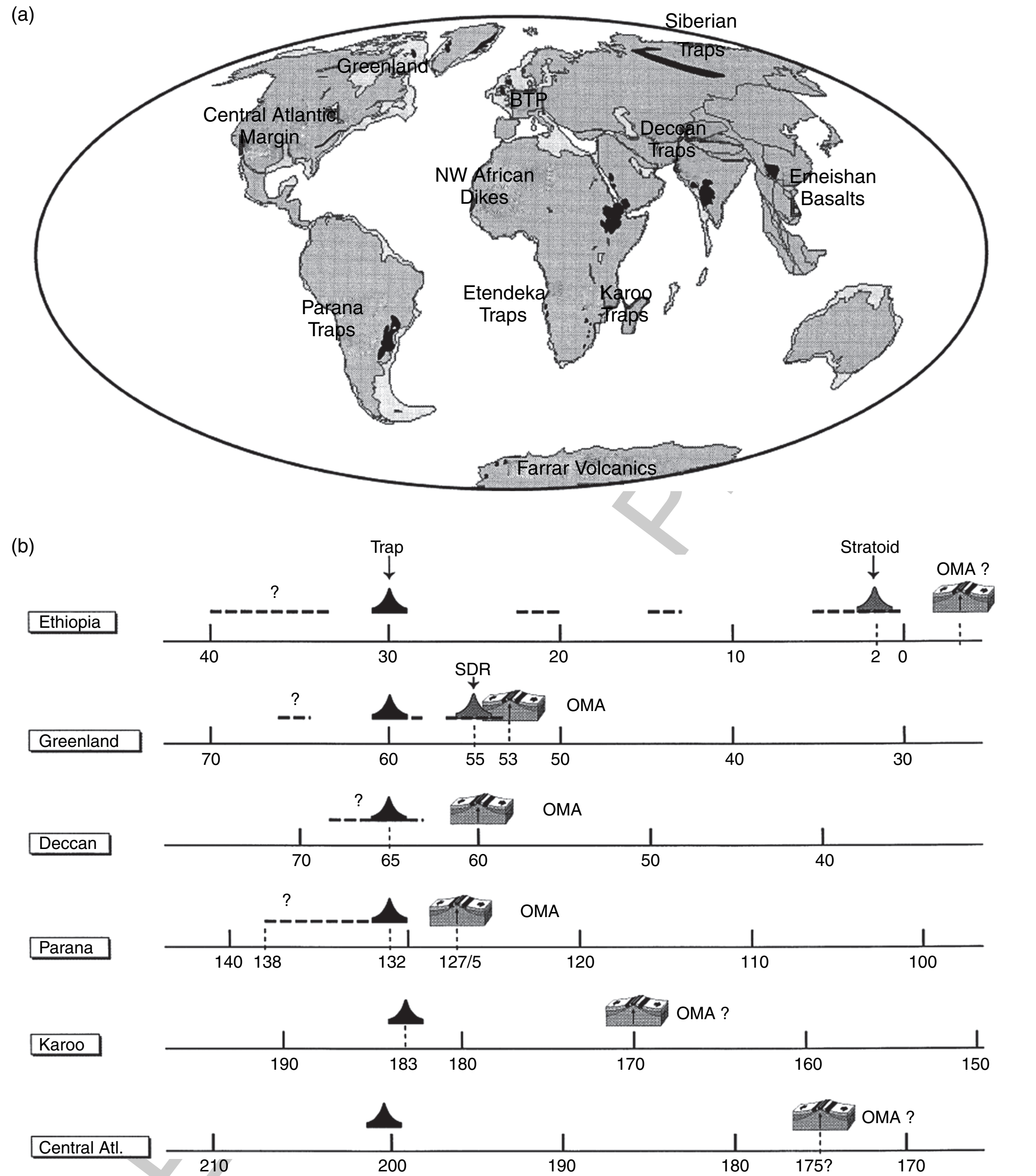
$$F_{\mathrm{T}} = \frac{CH_{\mathrm{b}}^{2}}{2} \qquad [22]$$

where H_{b} is the thickness of the brittle lithosphere and C is a constant defined in eqns [2] and [3], and is approximately $1.4 \times 10^{4}\,\mathrm{Pa\,m^{-1}}$. Neglecting hydrostatic pore pressures increases the value of C, and so the estimated force, by about 50%, and taking a mantle density for the lithospheric density increases it a further 10%. Assuming the creeping part of the lithosphere contributes to the tectonic force also increases this estimate, but for crust creeping more easily than the mantle the tectonic force is reduced by an amount that depends on the crustal thickness, rheological constants, and thermal state (e.g., Brace and Kohlstedt, 1980; Sawyer, 1985; Kusznir and Park, 1987). For the purpose of making simple comparisons with the force needed to open a magmatic rift in this chapter eqn [4] is sufficient.

6.09.6.3 Force Needed for Magmatic Rifting s0145

As long as the magma is less dense than the rock, it intrudes then some extensional stress has to be applied to keep the magma from rising to the surface and extruding. Only when the magma is 'kept down' by this stress difference will the dike open at depth and so allow plates to move apart (**Figure 38**). Neglecting any stress needed to open a crack, as p0590

AU30



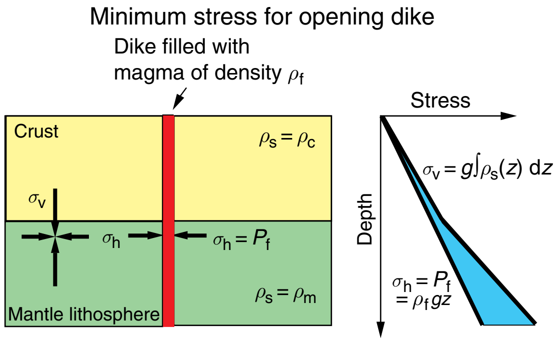
f0185 **Figure 37** (a) The distribution of large igneous provinces over the last 300 My. (b) shows the temporal relation between the major (trap) phase of vocanism in these provinces compared to the onset of ocean-floor spreading (as seen by oceanic magnetic anomalies, OMA). From Courtillot *et al.*, 2001.

AU29

discussed earlier, the minimum stress difference needed to open a magma filled dike is

$$\sigma_{\mathrm{M}}(z) = \int_0^{H_{\mathrm{b}}} g(\rho_{\mathrm{s}}(z) - \rho_{\mathrm{f}})\,\mathrm{d}z \qquad [23]$$

p0595 Equation [23] is valid only when the fluid magma density is less than or equal to the average lithospheric density. For example, for constant lithospheric density that is less than the magma density an extensional force would be required to pull dense magma up from the asthenosphere and magma could not rise to the surface and the limits of integration would have to be changed. Such cases are not likely to be important for lithosphere thicker than a



f0190 **Figure 38** The stress distribution for extensional separation of two lithospheric blocks by a vertical magmatic intrusion (a dike). Here the solid lithosphere has a density, ρ_s, that is greater than the fluid magma density, ρ_f. The yellow region represents crust with a density, ρ_c, close to that of basaltic magma in dikes and the green represents mantle lithosphere with a density, ρ_m, greater than that of the crust. The horizontal stress, σ_h, equals the pressure in the dike, P_f, while the vertical stress, σ_v, equals the overburden pressure. The area between these two stresses (highlighted in blue) equals the force that must be applied to keep the dike open. From Buck WR (2006) The role of magma in the development of the Afro-Arabian rift system. In: Yirgu G, Ebinger CJ, and Maguire PKH (eds.) *The Afar Volcanic Province within the East African Rift System*, vol. 259, pp. 43–54. London: Geological Society.

few kilometers and for basaltic magma densities, so they are not discussed further.

p0600 The extensional tectonic force to open a magma-filled dike through denser lithosphere with a thickness h_l is

$$F_M = \int_0^{H_b} \sigma_M(z)\,dz \qquad [24]$$

p0605 Consider the simple case that the densities of the crust and mantle, ρ_c and ρ_m, respectively, are constant with depth. If the entire crust is brittle down to its base at $z = H_c$ and the thickness of mantle lithosphere is H_m (so the total lithospheric thickness $H_b = H_c + H_m$), then

$$F_M = g(\rho_c - \rho_f)\frac{H_c^2}{2} + g\left[(\rho_c - \rho_f)H_c + (\rho_m - \rho_f)\frac{H_m}{2}\right]H_m \qquad [25]$$

p0610 The density of continental crust is not likely to be very different from the density of basaltic magma. If that is true then it takes no force to open a dike in the crust, but it still takes considerable force to open a dike into the mantle lithosphere. For $\rho_c = \rho_f$ we get

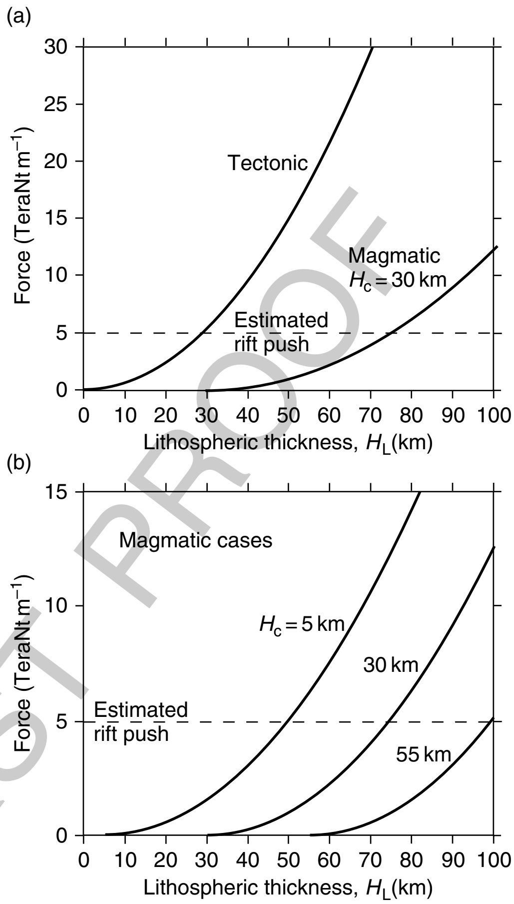
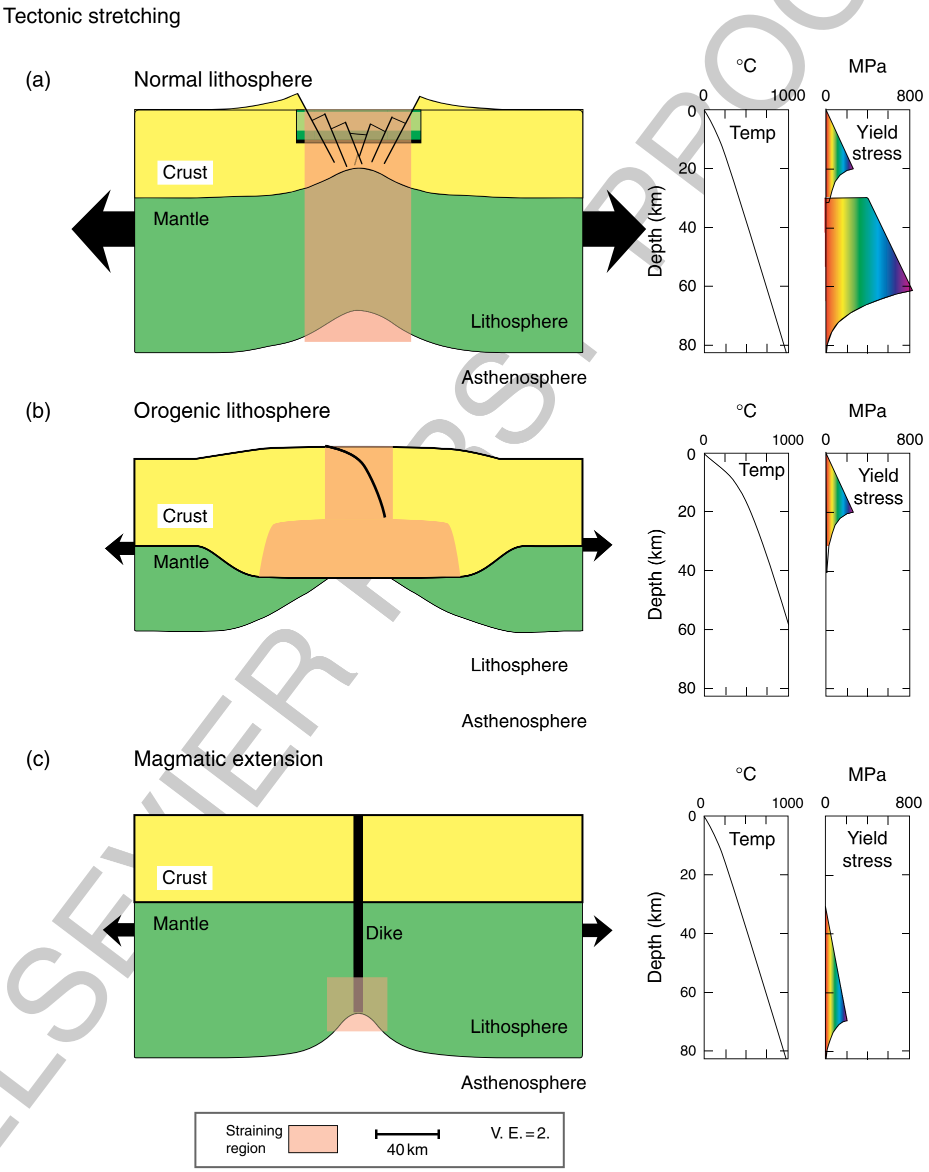


Figure 39 Illustration of estimated force required for lithospheric extension for either tectonic or magmatic rifting as a function of lithospheric thickness. Equation [22] was used to compute the tectonic force and eqn [7] was used to compute the magmatic force. (a) shows the forces as a function of lithospheric thickness assuming a 30 km thick crust. The dashed line is an estimate of the force available to drive lithospheric extension. (b) shows how the crustal thickness affects the estimated magmatic force. It is the mantle lithospheric thickness, which equals the total lithospheric thickness minus the crustal thickness, that controls this force. From Buck WR (2006) The role of magma in the development of the Afro-Arabian rift system. In: Yirgu G, Ebinger CJ, and Maguire PKH (eds.) *The Afar Volcanic Province within the East African Rift System*, vol. 259, pp. 43–54. London: Geological Society. f0195

the further simplification that gives eqn [7] (given in an earlier section).

For reasonable density values we plot this estimate of the force to dike through the entire lithosphere versus the lithospheric thickness (**Figure 39**) and we compare this to the quadratic relation between tectonic force and lithospheric thickness for tectonic stretching (eqn [22]). For a density contrast between p0615 solid mantle and fluid magma of 500 kg m^{-3} (based on mantle density of 3250 kg m^{-3} and magma density of 2750 kg m^{-3}) it would take 4×10^{12} Nt m^{-1} to dike trough a 40 km thick lithospheric mantle layer. To dike through 100 km of mantle lithosphere would take 2.5×10^{13} Nt m^{-1} and this is considerably more force than is likely to be available to drive rifting. The term magmatic rifting will be used to describe lithospheric extension that is aided by dike intrusion and the force for magmatic rifting is taken to be the force to open dikes through the lithosphere.

p0620 Dikes may not open through the entire thickness of the lithosphere if insufficient magma is available, and for such cases the force required to rift would be intermediate between the force for tectonic rifting and the force for magmatic rifting. **Figure 40** shows how different these forces are likely to be as a function of lithospheric thickness. The tectonic force



f0200 **Figure 40** Schematic of stresses needed for extension of continental lithosphere. Note the large difference in the yield stress, the stress difference needed to get extensional separation of two lithospheric blocks. Tectonic extension of (a) normal continental lithosphere should require very large yield stresses and correspondingly large tectonic extensional forces. (b) Orogenically lithosphere should have a higher geotherm and much lower lithospheric strength. (c) Magmatic intrusion may allow extension of normal lithoshere at modest tectonic force levels.

depends on the square of the whole thickness of the lithosphere (times $\sim$6000 Nt m^{-3}), while the magmatic force depends on the square of the mantle lithospheric thickness (times $\sim$2500 Nt m^{-3}). For reasonable driving force levels, only lithosphere thinner than $\sim$30 km thick should rift tectonically (i.e., in the absence of magmatic dike intrusion). For a normal continental crustal thickness of 40 km, the base of the lithosphere could be as deep as $\sim$80 km and still allow magmatic rifting at reasonable force levels. **Figure 40** shows that lithosphere with thicker continental crust should rift magmatically with much less force than for lithosphere with thin crust.

p0625 Rifts intruded by large volumes of basaltic magma should subside little during the intrusion phase of extension. If sufficient dike intrusion occurs to heat the lithosphere appreciably then its strength could be lowered to the point where tectonic extentions can proceed at moderate stress levels. Such combined magmatic and tectonic extension could give a rifted margin with less tectonic subsidence and more thermal subsidence than predicted by a McKenzie stretching model. As noted earlier this is the pattern of subsidence most commonly seen at rifted margins.

s0150 6.09.6.4 The Meaning of Rift Straightness

p0630 The straightness of many rifts may indicate that they were initiated by magmatic dike intrusions. Lithospheric extension can be accomplished by either dike opening or fault slip. Both dikes and faults form in response to stress difference in the lithosphere. Faults form and slip where the shear stress is great enough to overcome the strength of the material and allow brittle deformation. Faults change orientation where the stress orientations change or where there are strength variations. So, rifts where faults accommodate the extension do not have to be straight. They can curve where the stresses change or 'side-step' into weaker areas.

p0635 Dikes are narrow magma filled tension cracks (e.g., [b0475] Lister and Kerr, 1991) and as such the plane of the dike must be orthogonal to the least principal stress in a strong, brittle layer. When the least principal stress is horizontal the dike is vertical and these are the kinds of dikes discussed here. The key to the straightness of dikes may be that dikes, unlike faults, need a source of magma and a connection to that magma source. The open part of the dike is the conduit connecting the source area to the tip of the dike. If dikes are fed from a distributed magma source below a brittle layer then the dikes do not have to be straight except on the scale of the layer thickness. One way to produce dikes that are straight over lateral distances longer than the brittle layer thickness is if the magma source is localized. The dikes can only remain connected to the source if they are straight.

If the magma does come from a central source then the dike cannot propagate if it loses connection to the magma supply because the connection is the open, unfrozen, straight dike behind the propagating tip. If a dike tip were to step laterally away from the plane of the open dike by more than the width of the dike it would no longer be fed magma. Dikes exposed in ophiolites and bordering rifts are typically about a meter wide (e.g., [b0750] Varga, 2003), so very small offsets are viable. p0640

Straight dikes have been observed propagating from a central magma chamber along a subaerial segment of the Mid-Atlantic Ridge in Iceland. In 1975 an episode of approximately 15 dike intrusion and magma extrusion events began, with the longest dike propagating 70 km from the Krafla central volcano ([b0725] Tryggvason, 1980). Seismic and geodetic measurements unequivocally show that the dikes are sourced from a central magma chamber that subsided while the dike was propagating ([b0260] Einarsson, 1991). Dikes on the flanks of active volcanoes, like those propagating down the east rift zone of Kilaeua Volcano in Hawaii, are seen to be straight (e.g., [b0150] Cervelli *et al.*, 2002). Ancient dikes in the MacKenzie Dike Swarm of Canada are straight and traceable for thousands of kilometers ([b0275] Fialko and Rubin, 1999). p0645

The fact that most mid-ocean ridge segments are straight and nearly orthogonal to the spreading direction may reflect dikes fed from central magma chambers. New data on ridge segments suggests that at the slowest spreading rates there are nonmagmatic segments that are not straight. [b0235] Dick *et al.* (2003) note that the slowest spreading centers, such as the Gakkel Ridge in the Arctic and oblique sections of the Southwest Indian Ridge, show an alteration of volcanic segments and nonvolcanic ones. Peridotite samples dredged from the surface of the nonvolcanic segments indicate that the mantle there is stretched with little or no input of magma. When these nonvolcanic segments are along oblique sections of the ridge, the usual pattern of ridge segments orthogonal to transform faults is not seen. Such segments are cut by numerous faults that trend oblique to the spreading direction. p0650

s0155 6.09.6.5 The Distance of Dike/Rift Propagation

p0655 Dike propagation may be controlled by a dauntingly large number of thermal and mechanical processes. Among them are the pressure and flux of magma coming out of a source region, the viscous resistance to magma flow along the body of the dike and into the tip region, elastic stresses in lithosphere, and the freezing of magma (e.g., Lister and Kerr, 1991; Rubin, 1995). Although the details of dike mechanics are controversial (e.g., Delaney and Pollard, 1982; Fialko and Rubin, 1999; Ida, 1999), there is no doubt that the distance of dike propagation should be related to the supply of magma. The volume of magma intruded into the thick lithosphere along a several-thousand-kilometer-long rift has to be massive.

p0660 The size of a magma chamber should play a major role in determining how much magma can be supplied to dikes as they propagate. The magma pressure is likely to be reduced as volume is extracted from a magma chamber. Buck *et al.* (2006) argue that lithosphere-cutting dikes stop when the 'driving stress' (defined as the difference between magma pressure and tectonic stress orthogonal to the dike) becomes too small, either because the dike slows and the tip freezes, or because the driving stress is not enough to break open a new section of dike. The larger and shallower the magma chamber the smaller the pressure drop on extraction of a given volume of magma. Large magma chambers should supply large amounts of basalt to a propagating dike and so could be necessary to the production of long dikes.

p0665 Simple thermal arguments would suggest that the size and depth of a magma chamber should correlate with the flux of magma coming into a region. The greatest known fluxes of magma occur during the geologically short periods when large igneous provinces form. The volumes are up to several million cubic kilometers and the time interval of high rate magma output is 1 My or less (Courtillot and Renne (2003), and references therein). Very large, near-surface magma chambers are likely to form during periods of high melt flux from localized mantle upwellings. Magma chambers the size of large gabbroic layered intrusions found near the centers of some large igneous provinces could feed dikes propagating thousands of kilometers. For example, the Skaergaard layered intrusion of East Greenland is estimated to have a volume of $\sim$300 km^3 (Nielsen, 2004), sufficient to fill a dike 2000 km long, 50 km high, and 3 m thick.

The regions where the mantle lithosphere is very thick may not rift even if copious, high-pressure basaltic magma is present (see **Figure 39**(**b**)). Recall that extrusion of magma on the surface limits the maximum magma pressure. Cratonic regions, where the lithosphere may be well over a hundred kilometers thick (Jordan, 1975; Venkataraman *et al.*, 2004), and old oceanic lithosphere, where the mantle lithosphere may be $\sim$60 km thick (e.g., Wiens and Stein, 1984) should be too thick to rift. p0670

Dikes and associated magmatic rift propagation can stop either because the magma pressure gets too low, or the stress is not sufficient to open a dike through the lithosphere. For the Afro-Arabian Rift System it may be the lack of extensional stress sufficient to open dikes through very thick mantle lithosphere that limits dike propagation. Previous workers have noted that changes in lithospheric strength may have controlled the termination of the Red Sea Branch (e.g., Steckler and ten Brink, 1986) or affected the structure of the East African Branch (e.g. Rosendahl, 1987). Those workers were concerned with the difference in tectonic strength of the lithosphere. It is possible that similar arguments may apply for magma-assisted rifting. It may be the increase in force for magmatic rifting related to mantle lithosphere thickening that limits rift propagation. p0675

This is suggested by the observation that the Northern and Southern Branches of the system end close to regions of thick mantle lithosphere. The Northern Red Sea Branch ends close to the Mediterranean Sea where old oceanic lithosphere may be too thick to be cut by dikes. p0680

6.09.7 Conclusions and Future Work s0160

The questions discussed here are certainly not settled and new observations and models are clarifying the arguments. Several other major problems of continental extension have not been discussed here. For example, what controls the time interval between major rifting events? What controls the length of rifts? Why are fault patterns at rifts so variable? New observations and improved numerical models should allow us to address these and other rift-related questions. p0685

More geologic and geophysical data is needed on rift and margin structure and history. The amount and timing of magmatism be constrained. To understand the fault structures forming margins we need better constraints on actual distribution of faults and p0690 AU19 their offset histories. It appears that we are still in the discovery phase of finding out what rifts are like.

p0695 Continuum numerical models are good for studying some of the problems of large-scale rift structure. New techniques need to be developed to deal with. A great challenge is to combine models of highly localized processes such as fault zone evolution and dike propagation with larger regional scale models of continental extension. On the other end of the spectrum detailed rift models need to be embedded into global-scale numerical models to test ideas about the ways large-scale convective processes may be linked to the formation of continental rifts.

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Author's Contact Information

AU1 **W. R. Buck**
Lamont-Doherty Earth Observatory of Columbia University
Palisades
NY 10964
USA