Effect of lithospheric thickness on the formation of high- and low-angle normal faults

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ABSTRACT

It is accepted that extension of a homogeneous brittle layer should produce high-angle normal faults. The rotation of the upper parts of such a fault to a low dip requires fault offset that is greater than the local lithospheric thickness. New model calculations indicate that a fault cutting lithosphere thicker than 10–20 km can build up only a few kilometres horizontal throw before a new fault replaces it. Such a small offset leaves the fault little rotated from its initial high dip angle. For thinner lithosphere, the fault offset may be much larger, leading to rotation of the upper, inactive parts of a high-angle fault to a low dip angle. In continents, this can happen where thin upper-crustal lithosphere is decoupled from mantle lithosphere by weak lower crust. The calculations suggest that normal faults begin with high dip angles and only in areas of very thin lithosphere can they be offset enough to produce low-angle fault structures. This prediction is consistent with the occurrence of low-angle faults in areas with higher than normal heat flow.

INTRODUCTION

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; Weissel and Karner, 1989). Low-angle normal faults dip between 30° and subhorizontal, and some appear to have accommodated horizontal throws as much as 50 km or more (Davis and Lister, 1988).

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). The simple Mohr-Coulomb theory predicts that normal faults should form with dips between 60° and 75° for internal rock-friction coefficients in the experimentally determined range of 0.6–0.8 (Byerlee, 1978).

An alternative hypothesis to that of 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. In Buck (1988), I also argued that rotation could explain low-angle fault structures, and I 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.

In Buck (1988), I assumed that large offsets could occur on a high-angle normal fault to produce low-angle fault structures. Here, I consider the mechanical reason that such large offsets might occur on some high-angle normal faults and not on others.

PREVIOUS WORK ON NORMAL-FAULT MECHANICS

In a general theory for faulting, Anderson (1942) gave an explanation for the formation of high-angle normal faults. According to two-dimensional Mohr-Coulomb theory, fault slip occurs when the shear stress τ satisfies the equation

$$\tau = \mu\sigma_n + \tau_o, \tag{1}$$

where μ is the coefficient of friction, σ_n is the normal stress, and τ_o is the cohesion (Jaeger and Cook, 1979). Anderson assumed that two of the principal stresses in the earth are aligned in the horizontal plane and the third is vertical. For normal faulting, the maximum compressive stress, σ_1, is vertical. The optimum orientation is the one that minimizes the difference between σ_1 and the minimum compressive stress σ_3 required to satisfy equation 1. The optimum fault dip is $45° + ½ \tan^{-1}\mu$. For $\mu = 0.6$, the optimum fault dip is ~60°.

Sibson (1973) used Anderson's approach to calculate the magnitude of the stress required for slip on a normal fault. For slip on a cohesionless fault, the ratio of σ_1 to σ_3 must equal a constant determined by the fault dip and the friction coefficient. For an optimally oriented fault, the constant has a minimum value that is termed k_μ, as given in Figure 1. Thus, slip can occur on a properly oriented fault when $\sigma_3 = \sigma_1/k_\mu$. For $\mu = 0.6$, the constant k_μ is 3.12. Figure 1A gives the expression for the average horizontal stress in a layer required for slip on an existing normal fault assuming that σ_1 equals the average overburden, $\rho gh/2$, where ρ is the layer density, g is the acceleration of gravity, and h is the thickness of the lithospheric layer. Figure 1C gives the expression for the stress needed for creation of a new fault in a layer with a cohesion of τ_0 and an internal coefficient of friction of μ.

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 cut 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 $\tan\theta$, where θ is the fault dip as shown in Figure 1.

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 and for a fixed value of Δx depends on $\tan^2\theta$. He estimated that after only a few hundred metres 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

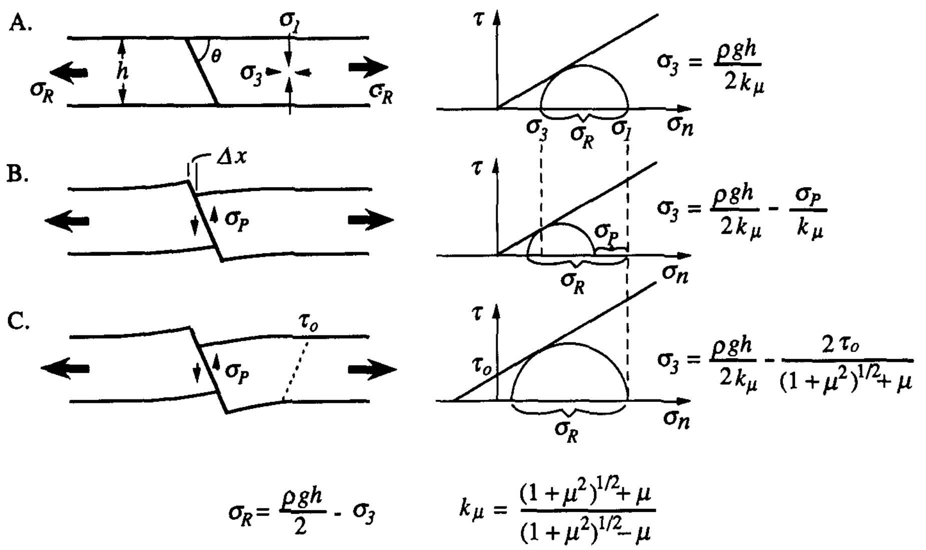

Figure 1. Stress required for (A) slip on undisplaced normal fault, (B) continued slip on fault with finite horizontal offset Δx, and (C) creation of new fault in layer with cohesive shear strength τ_0. Average vertical stress σ_1 is average overburden, $\rho gh/2$, except near the offset fault where it is reduced by σ_P. Lithospheric layer thickness denoted by h; σ_R is regional stress. Coefficients of friction on fault and of internal friction for layer are taken to equal μ. Mohr circle diagrams illustrate relations between stress components for these three situations.

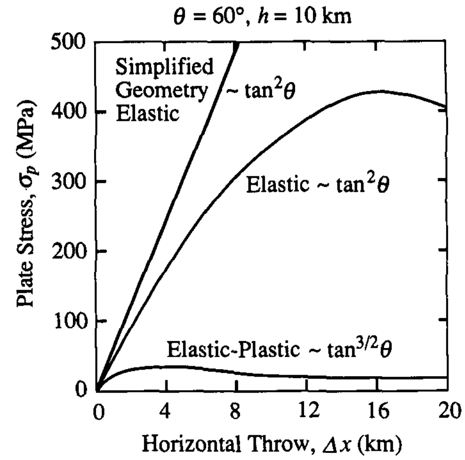

Figure 2. Results of three different kinds of model calculations of plate-bending stress, σ_P, as function of horizontal throw on normal fault dipping at 60° through 10-km-thick layer. Line labeled "Simplified Geometry Elastic" is analytical result from Forsyth (1992). Curve marked "Elastic" is from numerical calculation with more realistic geometry, but with infinite yield stress in layer. "Elastic-Plastic" curve is for diamond-shaped yield-stress envelope with maximum yield stress of 90 MPa at 5 km depth. Dependence of σ_P on dip angle θ is indicated.

with offset. Forsyth (1992) suggested that an initially low-angle normal fault could build up much more offset than a high-angle fault.

NEW MODEL FORMULATION

The normal-fault model formulated here uses the assumptions of Anderson (1942) regarding the orientation of principal stresses and the ratio of maximum to minimum stress required for fault slip. It also follows Forsyth (1992) in considering the effect of finite offset of a normal fault. However, instead of a vertical topographic step across the fault, in this model the topography includes the dipping fault at the surface as was done by Weissel and Karner (1989), Stein et al. (1988), and Buck (1988). In addition, the lithosphere is taken to have a finite yield strength, rather than being perfectly elastic.

Consider the lithosphere to be a cohesive elastic-plastic layer of thickness h cut by a single high-angle fault. This layer floats on an inviscid fluid substrate of the same density as the layer (2800 kg/m^3). The offset of the normal fault bends the lithosphere as indicated in Figure 1B. The bending of lithosphere reduces the vertical stress acting on the fault by an amount σ_P. If one likens the lithosphere adjacent to a fault to a diving board, then σ_P is the stress on the end causing the board to bend. The primary stress change should be in the vertical direction when the deflection of the plate is small compared to the flexural wavelength for the bending. To keep the analysis simple, we assume that this is true even if the plate deflections are large. To allow continued slip on that fault, σ_3 must be reduced, and so the regional stress σ_R is increased by an amount σ_P/k_μ (see Fig. 1B). The work done bending the plate reduces the normal stress on the fault, thus reducing the work done against friction on the fault.

The model lithosphere is elastic when bending stresses are less than a specified yield stress. When bending stresses equal the yield stress, plastic deformation occurs. This is the same approach used to model the bending of oceanic lithosphere at subduction zones (e.g., McAdoo et al., 1978). The commonly accepted form of the yield-stress envelope (Brace and Kohlstedt, 1980) shows yield stress increasing linearly with depth to the point where high-temperature ductile flow strength equals brittle strength. At greater depth, the yield strength decreases as the temperature increases. A diamond-shaped envelope is used to approximate this form, as in Buck (1988), because it simplifies the model calculations. The compressional and extensional yield stresses at a given depth are taken to be equal. The slope of the yield stress with depth is 18 MPa/km for the results shown in the figures, although cases with different slopes were calculated. The Young's modulus of the layer is 5×10^{10} Pa, and the Poisson's ratio is 0.25.

The finite strength of the plate means that its effective rigidity, D, depends on the curvature of the plate. Large curvature reduces the effective rigidity and the flexural wavelength. The effective elastic plate thickness is proportional to $D^{1/3}$. For small deflections of a plate, the effective plate thickness is equal to the plate thickness h, but when the curvature is large, the effective thickness may be reduced by as much as a factor of ten. The curvature of the plate can vary with horizontal position. Therefore, a finite-difference approximation to the thin-plate flexure equation is used which incorporates lateral variations in plate rigidity (see Buck, 1988).

The vertical stress change on the fault, σ_P, caused by plate bending is calculated in two ways. The vertical load over the fault area supported by plate bending is roughly $\partial/\partial x(D\partial^2 w/\partial^2 x)$, where x is horizontal distance and w is the vertical deflection of the plate (Turcotte and Schubert, 1982). Dividing this load by the horizontal projection of the fault, $\tan\theta/h$, gives σ_P. Alternatively, we can use the fact that the load supported by plate bending is equivalent to the derivative of the work done bending the plate with respect to the vertical fault throw. Work is done storing potential energy in topography, storing elastic energy in the plate, and dissipating energy during plastic yielding of the plate. Numerical calculations of the energy changes give the more consistent estimates of σ_P and so are used here.

RESULTS

Fault geometry and layer yield strength affect the calculated stress changes near a normal fault (Fig. 2). Here, the model lithosphere is 10 km thick, and the dip of the active part of the fault is 60°. The top two curves are for elastic plates with infinite yield stress throughout the layer. The case considered by Forsyth (1992) with a step in

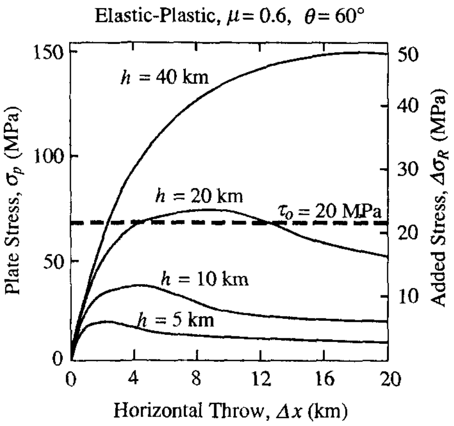

Figure 3. Results of model calculations showing that maximum value of σ_P scales with layer thickness. On right side is added regional stress needed to continue slip on fault (σ_P/k_μ) for μ = 0.6. Dashed line shows added stress needed to break new fault in layer with τ_0 = 20 MPa.

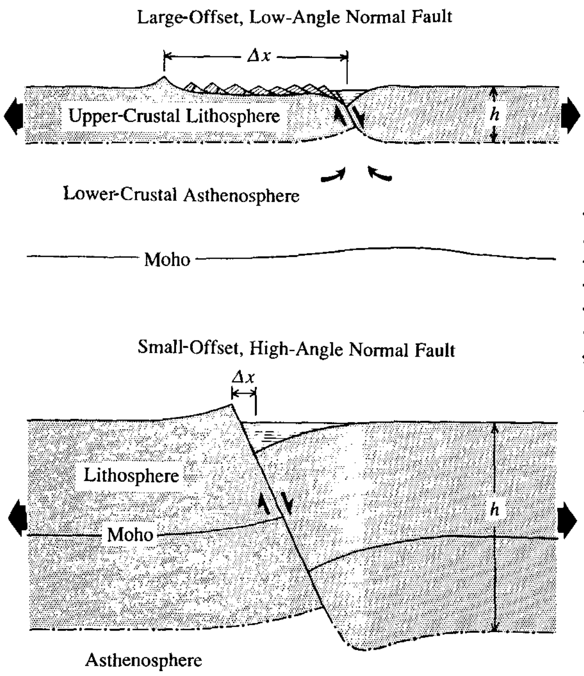

Figure 4. Illustration of model result that lithospheric thickness, h, may control horizontal throw on a normal fault, Δx. Offset on a normal fault cutting a thin layer can be large enough to cause rotation of part of fault to a flat position, whereas for a fault cutting thick lithosphere, Δx is strictly limited. Top drawing is based on results in Buck (1988) in which sediment (indicated by parallel lines) fills in fault-bounded basin.

topography across the fault predicts a linear increase in σ_P with Δx. For the models with a more realistic geometry, σ_P does not increase without limit, but reaches a maximum value. The bottom curve shows that plastic yielding drastically decreases the maximum value of σ_P and the horizontal offset required to reach it compared with a purely elastic plate. This effect is easy to understand qualitatively because plastic yielding limits the maximum moment supported by the plate and lowers the flexural wavelength.

Results for the elastic-plastic model are shown for different layer thicknesses in Figure 3. The maximum value of σ_P scales with thickness of the plate. Though not shown in the plots, the value of σ_P at a given value of Δx scales with fault dip as $\tan^{3/2}\theta$. These numerical results are summarized in the following relation:

$$\sigma_P|_{\max} = C\rho g h \tan^{3/2}\theta, \qquad (2)$$

where the estimated value of the dimensionless constant C is 6×10^{-2} for the parameters used here. The magnitude of the yield stress was also varied while the envelope was kept diamond shaped. For a set plate thickness, the maximum value of σ_P varies as the square root of yield stress at the middle of the plate.

The added regional stress needed for continued fault slip (σ_P/k_μ) is shown on the right side of Figure 3. The dashed line in that figure shows the change in horizontal stress that will allow a new fault to break for cohesive shear strength τ_0 = 20 MPa. Assuming cohesion is not dependent on depth, the same amount of added horizontal tension will allow a new fault to develop for any layer thickness. However, the maximum amount of extra stress needed to continue slip on a fault scales with layer thickness. When $h < \sim 18$ km, for this set of parameters, the stress needed to continue slip on an existing fault never exceeds the stress needed to produce a new fault. A fault cutting a layer thicker than 18 km should have slip of only a few kilometres before the stress needed to continue slip exceeds the stress needed to produce a new fault.

DISCUSSION

A key question about low-angle normal faults is whether they originate with a high or a low dip angle. On the basis of the present analysis, I suggest that normal faults first have slip with steep dips and that those that can accommodate large horizontal offset are rotated to a low dip angle. The only normal faults that can be offset a large amount are those that cut through relatively thin lithosphere. The offset on a normal fault formed in thin lithosphere is not limited by the stress changes associated with lithospheric bending. As offset increases, the top part of the fault rotates to a lower angle and is exposed at the surface. A normal fault formed in thick lithosphere would accumulate only a few kilometres of horizontal throw before it is replaced by a new fault. Significant fault rotation cannot occur before it is abandoned. Figure 4 shows the contrasting normal-fault structures that can form in thick and thin lithosphere. The cutoff in lithospheric thickness between these behaviors depends on poorly known physical parameters such as lithospheric cohesion, yield strength, and the friction coefficient on a fault. However, laboratory estimates of those parameters put the cutoff lithospheric thickness in the range of 10–20 km.

The lithosphere should be thinner than typical continental crust for large-offset, low-angle normal faults to develop according to this model. This implies that the formation of continental low-angle normal faults requires hot, weak lower crust that decouples the upper-crustal lithosphere from the mantle lithosphere (see Fig. 4). The heat flow must be greater than about 100 mW/m^2 for the upper-crustal lithospheric thickness to be less than about 15 km, according to laboratory estimates of rock rheology (e.g., Brace and Kohlstedt, 1980). For the lower crust to flow fast enough that upper-crustal faults are not affected by mantle deformation requires crust >40–50 km thick, even with a heat flow of 100 mW/m^2 (see Buck, 1991).

Fault offset must be greater than the thickness of the lithosphere to get the abandoned sections of the fault rotated to a low dip angle. As a continental normal fault slips, it pulls up lower crust, which cools to form new upper crust. In areas of rapid fault offset, the lithosphere will be thinned owing to vertical advection of heat during extension, but some lower crust should still cool to form new upper-crustal lithosphere. As the lower-plate side of the fault is exposed at the surface and rotated, a continuation of the fault must cut through the newly added part of the lower plate.

Forsyth (1992) came to the opposite conclusion about the initial dip of low-angle normal faults. He held that large-offset normal faults had to originate along low-angle, preexisting weaknesses. He was forced to this conclusion because he calculated that regional stress increases rapidly and without limit as a function of horizontal fault displacement (see Fig. 2). The conclusions in both this paper and in Forsyth (1992) are based on estimates of the change in regional stress σ_R required to keep one normal fault slipping. We agree that bending of the lithosphere by fault offset causes this stress to increase. Although I follow Forsyth's innovative approach to this problem, I find a much smaller increase in σ_R with fault offset as illustrated in Figure 2. In addition, I find that σ_R reaches a maximum value after a relatively small fault offset. The difference is due to three factors included in the present calculations: (1) the horizontal offset of the fault at the surface, (2) the reduction in normal stress on a fault caused by plate-bending stress, and most important, (3) the finite yield strength of the lithosphere.

Observations are consistent with the prediction that lithospheric thickness controls whether high- or low-angle normal faults develop. The areas where low-angle normal faults occur are characterized by high heat flow and large crustal thickness. The best studied low-angle normal faults are found in core complexes in the Basin and Range province of the western United States. These core complexes typically formed early in the extensional history of that region, when the crust was thicker than the present value of about 40 km (Davis and Lister, 1988). During the formation of core complexes, the heat flow was likely to have been at least as high as its present range of 80–100 mW/m^2 (Lachenbruch and Sass, 1978). Core complexes have been mapped in other recently collapsed orogenic belts, such as the Aegean (Lister et al., 1984), where the heat flow is currently high. High geothermal gradients at the time of core complex development are indicated by the common association of plutonism synchronous with low-angle fault development (e.g., Hill et al., 1992). The lack of a topographic depression in core complexes is often taken as evidence of weak lower crust at the time of extension (Gans, 1987; Block and Royden, 1990). Many core-complex faults are not associated with preexisting fault structures (e.g., Davis and Lister, 1988), and this argues strongly against the conclusion of Forsyth (1992).

Another area where the lithosphere may be thin enough for large-offset, low-angle normal faults to develop is at mid-ocean ridges. Low-angle normal-fault structures have been mapped on the Mid-Atlantic Ridge south of the Kane Fracture Zone (Karson et al., 1987). However, dike intrusion may accommodate most extension at spreading centers as it does on Iceland, where normal-fault offsets do not exceed 80 m (Gudmundsson, 1992).

More complex models of faulting and lithospheric bending are needed to determine whether normal faults with large offsets could be rotated and continue to slip at lower dip angles. For slip at very low dips (less than 30° for $\mu = 0.6$), the horizontal stress must be tensional, and such a stress state is unlikely in the earth (Sibson, 1985). Naturally, if the effective coefficient of friction of the fault is somehow reduced during offset, then low dip angles on active segments of faults are possible without a state of tension. Effects such as rotation of principal stresses, thermal changes, and thermal buoyancy could be incorporated in future models. These effects may alter the lithospheric thickness at which there is a transition between small-offset and large-offset faults, but the transition should still occur.

ACKNOWLEDGMENTS

Supported by National Science Foundation grants OCE-92-17892 and OCE-92-00945. I thank Agust Gudmundsson, Pall Einarsson, Leigh Royden, and Michelle Kominz for constructive reviews, and the Science Institute of the University of Iceland for providing a place to write about this problem even in a land with no low-angle normal faults. Lamont-Doherty Contribution 5103.

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Manuscript received February 8, 1993
Revised manuscript received June 15, 1993
Manuscript accepted July 5, 1993