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FLEXURAL ROTATION OF NORMAL FAULTS

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Abstract. A conceptual model is proposed for the generation of low-angle normal faults in Metamorphic Core Complexes. The model is based on three assumptions: (1) the isostatic response to normal fault motion is of regional extent; (2) when a fault segment is significantly rotated from the optimum angle of slip, relative to the crustal stress field, it is replaced by a new planar fault oriented in the optimum direction; and (3) the fault cuts the entire upper crust and fault motion always nucleates in the same region at the base of the upper crust. The stress field is considered to be uniform through the crust and the regional isostatic response of the crust to loads is computed using the thin plate flexure approximations. Active low-angle normal faults are difficult to reconcile with rock mechanics theories, earthquake focal mechanism studies, and geochronologic results indicating rapid cooling of core complex rocks. The model does not require active fault slip on low-angle faults. The flexural response to normal faulting is shown to be significantly affected by anelastic behavior of the crust and by loading due to sedimentation. The anelastic response to large bending stresses results in significant reduction in the effective elastic rigidity of the upper crust. This can explain the observed short wavelength of topographic response to normal fault loads. The model results in: (1) a nearly flat-lying abandoned normal fault (or "detachment") below slices of upper plate rocks and sedimentary infill; (2) a strong contrast in metamorphic grade across the abandoned "detachment"; (3) rapid movement of lower plate rocks from midcrustal depths to shallower depths. These results are qualitatively in agreement with geologic observations for core complexes.

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INTRODUCTION

There has been much recent discussion about the origin of "Metamorphic Core Complexes." Faults within these complexes appear to have undergone a normal sense of motion but are gently dipping or subhorizontal [e.g, Profett, 1977; Davis, 1980; Miller et al., 1983]. This is particularly interesting since seismically active normal faults are generally observed to be dipping at steeper angles (>30°) [e.g., Jackson, 1987].

More than 20 core complexes have been recognized in the Basin and Range province of western North America, all of Tertiary age [Crittenden et al. 1980]. Several characteristics of metamorphic core complexes are illustrated in Figure 1. The fundamental relationship in core complexes is that low-grade or unmetamorphosed rocks overlie high-grade metamorphic "core" rocks [Davis, 1983]. The boundary between rocks of contrasting metamorphic grade, which is often subhorizontal, has been variously called a "low-angle normal fault," a "detachment," or a "decollement." Although these terms imply that fault motion took

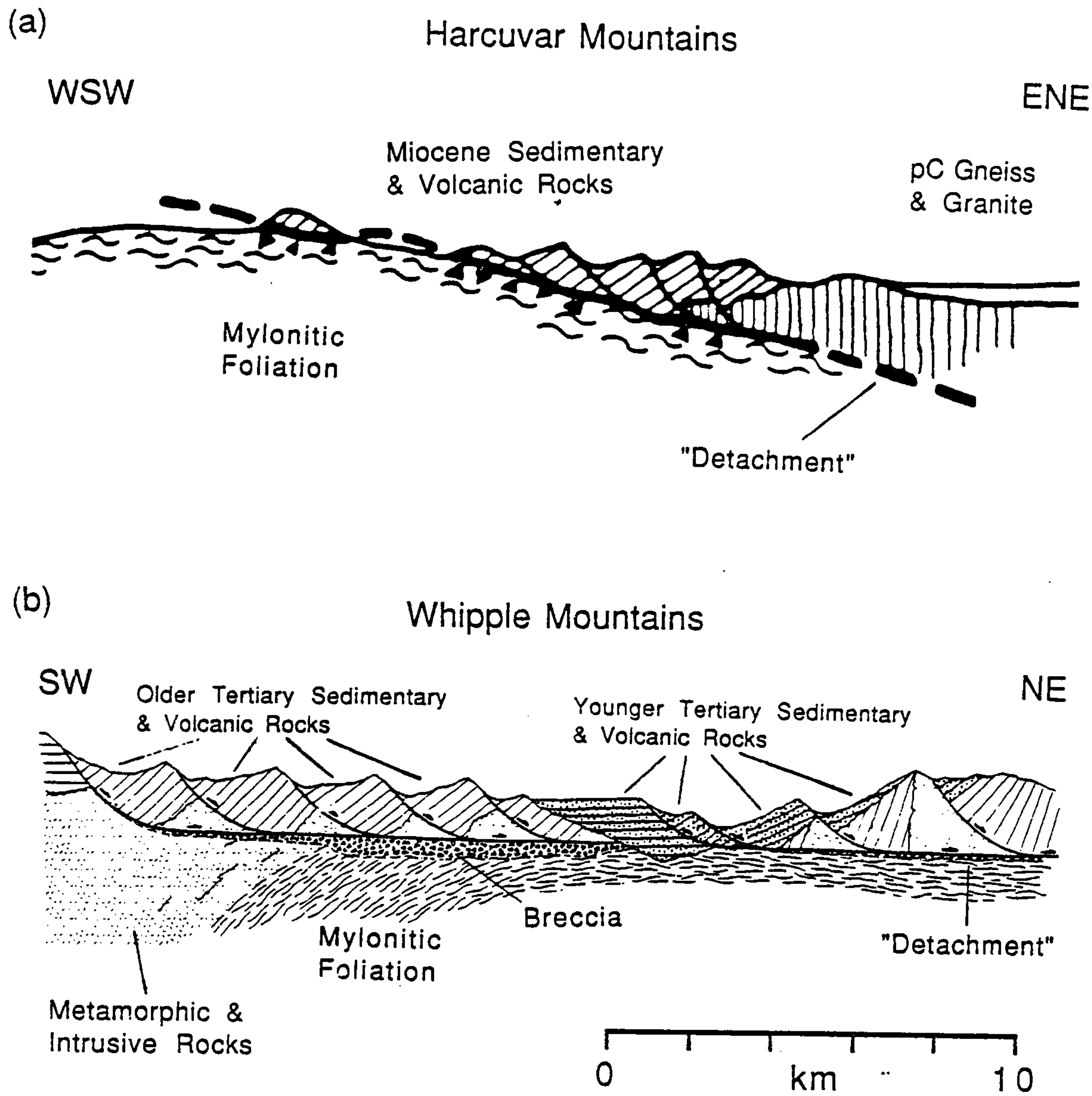

Fig. 1. Interpretative cross sections of two metamorphic core complexes showing features common to many core complexes: (a) the Harcuvar Mountains in Arizona [after Rehrig and Reynolds, 1980] and (b) the Whipple Mountains in California with inferred doming removed [after Davis, 1980].

place on a nearly flat-lying fault, there is considerable debate on the origin of such structures. Here the term "detachment" is used in a nongenetic sense to describe the major discontinuity in a core complex.

A normal sense of offset motion on the detachment is inferred from strain indicators and structural relations [e.g., Davis et al. 1983, 1986]. Upper plate rocks, those above the detachment, have typically undergone only brittle deformation while lower plate rocks have experienced considerable ductile deformation [e.g., Miller et al. 1983; Dokka et al., 1986]. The lower plate rocks frequently show mylonitization, indicating that they were deformed in a temperature range where quartz flowed but where feldspars fractured. High-temperature rock mechanics experiments show that this could occur between about 400° and 500 °C [Caristan, 1982]. The lower plate mylonites often show overprinting of brittle deformation on the ductile fabric [e.g., Crittenden et al., 1980].

Geochronologic studies of lower plate rocks from several core complexes indicate rapid uplift of midcrustal rocks [Davis, 1988]. Dallmeyer et al. [1986] use $^{40}Ar/^{39}Ar$ dating of lower plate rocks from the Ruby Mountain metamorphic core complex in eastern

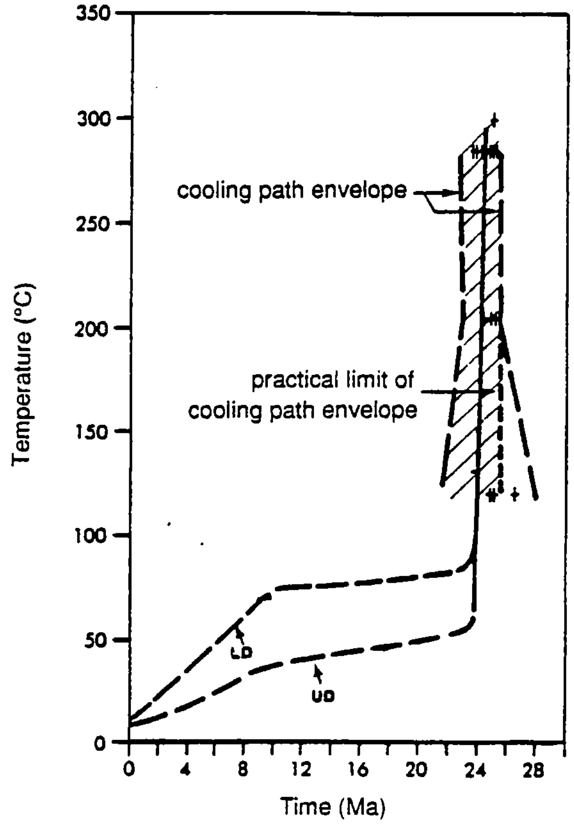

Fig. 2. Temperature-time curve inferred for rocks within (UD) and below (LD) the Ruby Mountains "detachment" in Nevada, based on fission track dating [after Dokka et al., 1986].

Nevada to infer rapid cooling from above 500°C to below 300°. In a fission track study of mylonitic rocks from the same area, Dokka et al. [1986] find that cooling from about 300°C to below 70°C occurred between about 25 and 23 Ma [see Figure 2]. If the geothermal gradient is assumed to have been 25°C/km, then this would give a rate of upward motion of greater than 4 km/m.y. Mylonites from the Whipple Mountain core complex in California have also been studied using $^{40}Ar/^{39}Ar$ and fission track methods [Davis, 1988; Dokka and Lingrey, 1979]. These studies indicate upward rates of motion for mylonites below the Whipple Mountain detachment which are as great or greater than for the Ruby Mountain detachment.

Pseudotachylite, a rock which is thought to be formed only during earthquake faulting, has been found in association with detachments. This has led some authors to claim that low-angle fault slip may be seismogenic [John, 1987].

Core complexes have had a great influence on discussions of extension, since they are found in one of the few accessible areas which appear to have been extended by large amounts. Continental margins, which are also thought to be greatly extended, are typically buried under deep piles of sediment. Mid-ocean ridges are the major areas of plate divergence, but because of the difficulty in working in deep water they are are not well mapped on a fine scale. Also, there is little or no erosion of mid-ocean ridges, making it difficult to infer cross sectional relations among faults.

The basic problems to be explained in core complexes are twofold: (1) how to produce a subhorizontal detachment surface separating rocks of very different metamorphic grade and (2) how to rapidly bring mylonitic rocks from 10–15 km depth up close to the surface. In this paper a new model for the origin of detachments and associated structures is proposed. Before developing this model a brief review of other ideas for the origin of detachments is given.

PREVIOUS MODELS

Several models have been proposed to explain observations in metamorphic core complexes. They fall broadly into two categories, according to how mylonites are brought to the surface. The proposed mechanisms for this are: (1) motion on a high-angle fault or set of high-angle faults and (2) motion on a low-angle normal fault. For reasons discussed below, 30° is used here as the cutoff between low-angle and high-angle faults.

In the first model the detachment is assumed to mark a brittle-ductile transition which starts out at a depth of 7–10 km depth in the crust [Proffett, 1977; Rehrig and Reynolds, 1980; Miller et al., 1983; Gans et al., 1985]. The crust above the detachment deforms as a set of closely spaced parallel normal faults, initially slipping at a high angle. As the faults slip, they must be rotated to a shallower dip simply to conserve volume. The detachment is considered to be the surface where faults end and more ductile (pure) shear begins. When the set of upper plate faults has been rotated to a shallow angle so that they can no longer slip, a new set of high-angle faults cuts them.

This model is in accord with simple rock mechanics theory and seismic observations discussed below. However, it is difficult to produce the sharpness of the contact between metamorphic grades which is observed. It is difficult to understand why the detachment should continue to be a boundary between areas deforming differently as it is brought close to the surface. If the detachment surface began moving in the depth range of the brittle-ductile transition, then, as it moves up, it should move out of the brittle-ductile transition. This is true even at extension rates of several cm/yr (see thermal calculations of Buck et al. [1988]). Even if the detachment were to remain a boundary between faulting and more distributed pure shear strain, there should be a gradational contact between metamorphic grades across the detachment. Simple shear along the upper plate set of faults and rotation of the faults is equivalent to pure shear on a gross scale [Jackson, 1987]. Pure shear of a crustal column will lead to thinning of that column, but rocks from different levels will not be juxtaposed. Finally, to pull a flat-lying detachment surface up from the bottom of the upper crust and keep it relatively continuous would require extremely uniform extension of the upper crust. Active faults observed today in extending regions tend to be associated with topography which reflects movement along widely separated faults. Twenty-five kilometers is a typical spacing between faults in the Basin and Range [Fletcher and Hallet, 1983]. To bring a detachment up from depth would require a much closer spacing of faults [only a few kilometers] and extremely uniform slip on these faults. A special mechanism is needed to account for the close spacing of primary faults required by this model.

The second model assumes that the detachment surface was an active low-angle normal fault [Wernicke, 1981]. This does explain the sharp contact between rocks of very different metamorphic grade across a fault surface. It is also in accord with observations of large lateral offsets between structural markers across a detachment [e.g., Davis, 1988]. In this model the faults in the upper plate (see Figure 1) above the detachment are considered to be due to structural collapse of the hanging wall of a low-angle fault. The major problem with this model is that low-angle normal fault motion is not favored in an extending brittle layer [e.g., Jaeger and Cook, 1959] and that low-angle normal fault earthquakes are seldom observed seismically [Jackson and McKenzie, 1983; Jackson, 1987].

Simple fault mechanics theory [Anderson, 1951] predicts that the preferred orientation for active normal faults is for dips between 45° and 70° from the horizontal. This theory is based on the assumption that the maximum compressive stress σ_1 is vertical and the minimum compressive stress σ_3 is horizontal. This assumption is in accord with stress measurements made in boreholes around active faults [e.g., Zoback and Healy, 1984]. Sibson [1985] points out that for a typical static rock friction coefficient of 0.75, frictional normal slip failure cannot occur on planes dipping less than 37°. Bradshaw and Zoback [1988] consider the effect of pore pressure on frictional sliding of normal faults. Even with overpressuring, they find that for slip to occur on faults dipping less than 20° requires frictional sliding coefficients close to zero. Overpressuring should be difficult to achieve on a normal fault, since any secondary normal faults in the upper plate should provide an escape path for fluids.

Jackson [1987] reports on a worldwide survey of fault plane solutions for large continental normal faulting earthquakes. He shows that the overwhelming majority of such earthquakes nucleate in the depth range 6-15 km on faults dipping between 30° and 60°. Among all earthquakes studied whose faulting was observed to break the ground surface, allowing one of the nodal planes of the fault plane solution to be unambiguously chosen as the fault plane, no event with a dip less than 30° was observed.

The active low-angle normal fault model has difficulty explaining the geochronologic data from core complexes discussed above. Figure 3 shows that a fault dipping at 10° would have to slip at a rate of at least 2.5 cm/yr to explain the rate of cooling of lower plate rocks inferred from fission track studies. The present-day rate of extension of the entire width of the Basin and Range province is estimated from geodetic measurements to be less than about 1 cm/yr [D. E. Smith, personal communication, 1988]. To match the geochronologic data, several times the extension rate for the entire Basin and Range would have to be concentrated on a single low-angle fault.

In a study which did not specifically look at the

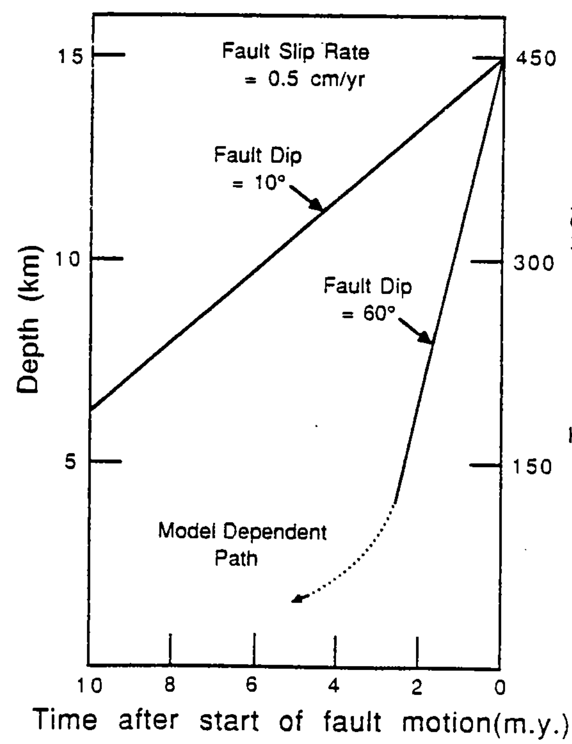

Fig. 3. Depth-time and temperature-time curve for rocks just below an active fault which is slipping at 0.5 cm/yr for two different dip angles. Temperature is related to depth by assuming a constant geothermal gradient of 30 °C/km. See text for discussion.

problem of metamorphic core complex origin a different normal faulting model is considered. Buck et al. [1988] calculate the effects of assuming local isostatic compensation of topography produced by movement on a planar normal fault. This results in lower plate (footwall) rocks, originally adjacent to the active normal fault, being exposed at the surface. Mylonites formed by strain at midcrustal levels would be carried to the surface if this model were correct. The essentially flat fault surface is "abandoned" in that it no longer moves when it is rotated into a horizontal position. For a high normal fault dip angle (>30°) a moderate slip rate on the fault would cause lower plate rocks to be uplifted and cooled at a rate compatible with geochronologic data discussed above (see Figures 2 and 3). However, the local compensation model does not explain how upper plate rocks could end up on top of an abandoned detachment surface. It also ignores the fact that there is no reason to expect zero flexural strength of the crust in an extending region.

CONCEPTUAL PHYSICAL MODEL

The results of Buck et al. [1988] prompted the present study, which is a first attempt to consider the possible effects of regional isostatic response to large offsets on a normal fault. The model is based on three assumptions:

1. The isostatic response to normal fault motion is of regional extent.

2. If a fault segment is significantly rotated from the optimum angle of slip relative to the crustal stress field, then a new planar fault oriented in the optimum direction will replace it.

3. The fault cuts the entire upper crust (Figure 4) and continues to cut through the same area of the base of the upper crust. Fault motion is always considered to nucleate in the same region at the base of the upper crust.

These assumptions have a basis in observations and theory. First, based on understanding of the materials that make up the crust, one would expect that the crust should have a finite flexural strength [e.g., Goetze and

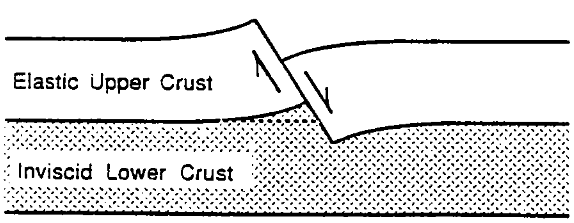

Fig. 4. Schematic of a normal fault cutting the entire upper crust. The base of the upper crust is always at a fixed depth, as shown by the dashed line.

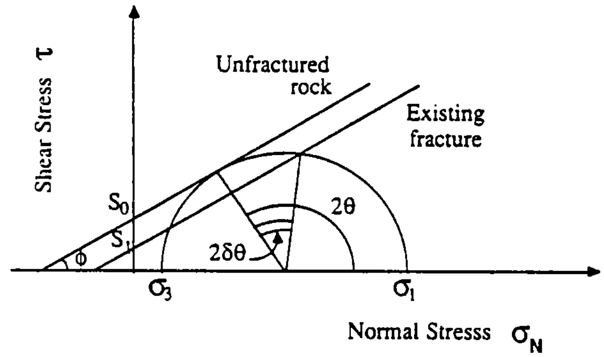

Fig. 5. Plot of shear stress τ versus normal stress σ_n for frictional sliding on a fault (Byerlee's law) and for fracture (Coulomb failure) of rock. The Mohr's circle shows that when the angle of the fault surface to the minimum principal stress has rotated $\delta\theta$ from the optimum value, θ, then failure of the rock can occur. Here it is assumed that the mimimum stress is horizontal.

Evans, 1979]. Extensional stresses should reduce the flexural rigidity of the crust, as described below and by Bodine et al. [1981]. High crustal temperatures should lead to lower flexural strength. The Basin and Range is underging extension, and heat flow there is about twice that of average continental crust [Lachenbruch and Sass, 1977]. However, analysis of the correlation of gravity and topography in the Basin and Range by Bechtel and Forsyth [1987] indicates a finite flexural strength for that region.

As noted earlier, fault mechanics theory predicts that faults have a range of optimal orientations to the principal stresses. According to the Mohr-Coulomb criterion [Coulomb, 1773], shear failure will occur once the shear strength and internal friction of a material are overcome [Jaeger and Cook, 1959]. Rock mechanics experiments show that when a fault is rotated away from the optimal angle for shear, frictional sliding along that fault eventually becomes more difficult than creating a new fault through unfractured rock (see Nur et al. [1986] for further discussion). Laboratory studies show that the shear stress τ required for motion on an existing fracture is controlled by the cohesive strength S and the normal stress σ_n acting on the fracture [e.g., Byerlee, 1978]:

$$\tau = S + \mu\sigma_n \tag{1}$$

where μ is the coefficient of friction. This relation is plotted on a Mohr circle diagram in Figure 5 for two values of the cohesive strength. S_1 is the cohesion for an existing fault and S_0 is for unfractured rock. As long as a normal fault remains close to the optimum dip angle for frictional sliding θ, which for this example is about 60°, no new faults will be cut.

However, if the fault rotates so much that the Mohr circle comes to intersect the line representing shear stress required to cut a new fracture, a new fault will be cut. The amount of rotation, $\delta\theta$, required for the cutting of a new fault depends on three values: (1) the difference between the cohesive strength of the unfractured rock, S_0, and the cohesive strength of preexisting faults, S_1; (2) the coefficient of friction $\mu=\tan\phi$, and; (3) the absolute magnitude of the horizontal and vertical stresses (σ_1 and σ_3). The angle of rotation is

$$\delta\theta = \frac{1}{2}\cos^{-1}\left[1 - \frac{2\,(S_0 - S_1)\cos\phi}{(\sigma_1 - \sigma_3)}\right] \qquad (2)$$

For the example shown in Figure 5 this rotation is about 15°. The value of μ and S_1 for an existing fracture should be relatively independent of rock type [Byerlee, 1978]. The cohesive strength of unfractured rock, however, depends strongly on rock type. Therefore $\delta\theta$ must be considered a variable in the model. For most estimates of μ for crustal rocks the preferred angle of slip is about 60° [Sibson, 1985].

The idea that a normal fault could cut the entire upper crust and not continue through the lower crust is also based on theory and observation. The seismically determined focal depth of continental normal faulting earthquakes are strongly clustered in the range 6-15 km [Jackson, 1987]. Gabbros and diorites, which are thought to constitute much of the lower crust, should flow at lower crustal temperatures and pressures [Caristan, 1982]. At geologic strain rates, little deviatoric stress could be stored in lower crustal rocks. Fault motion at midcrustal depths may produce mylonites. A combination of grain size, mineral assemblages, and crystalographic fabrics is likely to make a midcrustal mylonite zone weaker than the surrounding rock [e.g., White, 1976]. This may cause fault motion to nucleate repeatedly in a localized area. Also, localized intrusion of magma may make a particular area of crust weaker than other areas. The correlation in time and space between nonbasaltic volcanism and extension within core complexes has been noted by several authors [e.g., Glazner and Bartley, 1984; Gans, 1987].

NUMERICAL MODEL

Quantification of this conceptual model is a very difficult task. The stresses and strains within crust undergoing large normal fault offset may be quite complex. Recent investigations in rock mechanics suggest that the crust may be divided into several regions each with a unique flow law [Scholz, 1987]. Each type of deformation is described by a different set of parameters determined by laboratory measurements. Extrapolation from laboratory strain rates to geologic strain rates is not well quantified [Goetze and Evans, 1979]. Uncertainties in the average composition of the crust may lead to large uncertainties in estimates of the average strength of parts of the crust [e.g., Kusnir and Park, 1987]. Offset of a primary fault may lead to motion on other faults. These secondary faults may not be easily described using continuum mechanics. Erosion and sedimentation act as loads deforming the crust. It is appropriate to first consider simplified calculations for crustal deformation. If the results of this study are considered interesting then more complex and rigorous models should be constructed.

Approximations

Two major approximations are made in formulating numerical calculations. First, the stress field is considered to be uniform through the crust with the maximum principal compression vertical. Second, the regional isostatic response of the crust to loads is computed using the thin plate flexure approximations [e.g., Timoshenko and Woinowsky-Krieger, 1959].

A uniform stress field is assumed so that the Mohr-Coulomb failure criteria, described above, can be used to predict when an active fault will be abandoned and replaced by a new fault. Stresses due to topography, plate bending, and secondary faults may contribute to a nonuniform stress field. These effects make the determination of a preferred fault orientation extremely complex.

Many authors have modeled regional compensation by treating the lithosphere as a thin elastic plate floating on an inviscid fluid [e.g., Vening Meinesz, 1941; Walcott, 1970; Watts, 1978]. Vening Meinesz [1950] used thin plate theory to approximate the response of the lithosphere to normal faulting. Several different approximations are implied by the the thin plate theory [Timoshenko and Woinowsky-Krieger, 1959]. ~~Among them are that the plate is thin compared to the wavelength of its deformation, vertical displacements are small compared to the flexural wavelength, vertical normal stresses within the plate are small compared to plate bending stresses, and the plate is perfectly elastic.~~ Small vertical displacements are required for the derivation of a simple differential relation between loads p(x) and vertical displacements w(x):

$$\nabla^2(D\,\nabla^2 w) + \rho g w = p \qquad (3)$$

where D(x) is the flexural rigidity of the lithosphere, ρ is the crustal density, g is the acceleration of gravity, and $\nabla^2 = \partial^2/\partial x^2 + \partial^2/\partial y^2$ [Timoshenko and Woinowsky-Krieger, 1959]. This relation depends on the neglect of membrane stresses and strains and on the use of the small angle approximation. Equation (3) is solved for points on a uniformly spaced grid using an implicit numerical method.

Comparisons between predictions of thin plate theory and more general theories of elastic plate bending have been made by several workers. Comer [1983] solved the general problem of the response of an elastic plate of finite thickness to harmonic loads, assuming small vertical displacements. His results show that the thin plate results are accurate approximations of thick plate results when the flexural wavelength is greater than about 6 times the elastic plate thickness. Savage and Gu [1985] used the thin plate approximation in modeling a normal fault using the same geometry assumed here. They show that the surface deflection approximated by thin plate theory compares well with the deformation predicted by a two-dimensional solution for small displacements of an elastic layer over a viscoelastic half-space [Rundle, 1982]. The length of time after faulting was chosen to be long compared to the viscous relaxation time of the half-space so that it would approximate the behavior of an inviscid layer.

Thin plate theory has been applied to geologic problems even in cases where some of the thin plate approximations are not rigorously valid. The bending stresses at subduction zones are sufficiently large that part of the lithosphere may deform anelastically [e.g., Goetze and Evans, 1979; Chapple and Forsyth, 1979]. Thus the approximation of perfect elasticity breaks down. When anelastic behavior is treated, the rigidity of the lithosphere can be shown to depend on plate curvature (e.g., Bodine and Watts [1979]; and discussion in the next section). When variations in rigidity as a function of curvature are included, agreement between observed and predicted topography at subduction zones is good [McAdoo et al., 1978; Chapple and Forsyth, 1979; Bodine and Watts, 1979; McNutt and Menard, 1982].

In the laccolith intrusion problem vertical displacements are large compared to flexural wavelength. Pollard and Johnson [1973] and Jackson and Pollard [1988] use thin plate theory to treat the bending of sedimentary rock layers by the intrusion of a laccolith. They consider data for Mt. Holmes in the Henry Mountains, Utah. where the inferred topographic offset is 1.2 km. The wavelength of bending is approximately of 5 km. In this case, several of the thin plate approximations are not strictly valid such as perfect elasticity, the small-angle approximation, and the neglect of membrane stresses. However, the simple theory provides good quantitative agreement with the shape of deformed beds [Jackson and Pollard, 1988].

The problem of the response to large offset of a normal fault has recently been treated by several authors. C. P. G. King et al., (The growth of geological structures by repeated earthquakes 1; Conceptual framework, submitted to *Journal of Geophysical Research*, 1988) and R. C. Stein et al., (The growth of geological structures by repeated earthquakes 2; Field examples of continental dip slip faults, submitted to *Journal of Geophysical Research*, 1988; hereinafter referred to as RCS [1988]) use the formulation of Rundle [1982] to model the vertical deflection of elastic upper crust to fault motion and thin plate flexure approximations to describe its response to sediment loading. They model fault offsets as great as 5 km and require effective flexural rigidities as low as 2 km to match the observed shape of crustal layers. J. K. Weissel and G. D. Karner, (Flexural uplift of rift flanks due to tectonic denudation of the lithosphere during extension, submitted to *Journal of Geophysical Research*, 1988) use thin plate flexure to model normal fault offset. They successfully model both oceanic and continental normal faults. The treatment of the normal fault problem used by all of these authors is very similar to that used in this paper.

In the present model the thin plate theory is not rigorously justified since finite rotations are required for abandonment of faults. It is the purpose of this study to make only qualitative comparisons with data on the geometry of core complex structures. One should not view the numerical results as quantitative predictions of the conceptual model. Rather, this is the first step in going beyond the "artist's conception" phase of interpreting a conceptual model.

Model Formulation

The model is described by three steps which are repeated in time. These steps are illustrated in Figure 6. The first step is the offset of a high-angle normal fault due to a relative separation of the two sides of the fault. Several reference frames can be adopted without changing the model results. For simplicity a reference frame is chosen in which the footwall is fixed. The hanging wall side of the normal fault moves parallel to the fault and downward, as shown in Figure 6a. This is the same frame of reference used by Vening-Meinesz [1950] and Savage and Gu [1985]. The next step is the isostatic response to the fault offset. The treatment of the flexural effect of normal faulting is given by the thin plate equation (equation (3)). The flexural response causes rotation of the fault at shallow depths while deeper parts of the fault remain straight [Figure 6b]. The final step is the cutting of a new upper part of the normal fault at the optimum dip angle, as given by Mohr-Coulomb theory (Figure 6c).

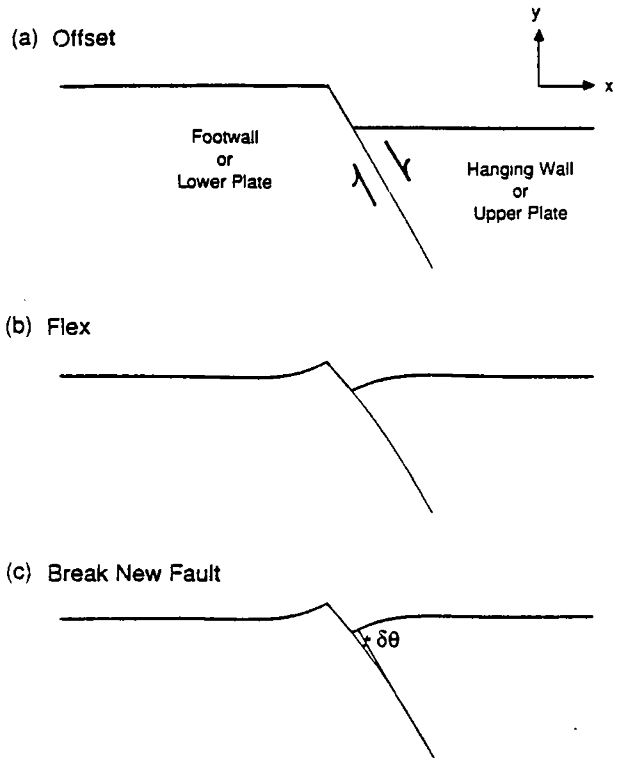

Fig. 6. Model formulation. (a) shows offset along a high-angle normal fault; (b) shows flexural response to the topographic load caused by the offset. The flexure causes rotation of the high angle normal fault over a flexural wavelength. In (c) the rotated fault is abandoned and a new extension of the fault, which is rooted in the lower crust, is cut through the top of the upper crust.

The fault offset represents a load which must be isostatically compensated. In the chosen frame of reference the hanging wall moves down parallel to the fault during offset. The downward component of offset at any position x is $\delta y(x)$. The load, $p(x)$, due to this offset is a negative load since it tends to pull the surface up. For the hanging wall side, $p(x)$ is given by

$$p(x) = -\delta y(x)\rho\, g \qquad (4)$$

where ρ is the crustal density. Note that when the surface is horizontal the fault offset results in $p[x]$ being the same for all x positions. When the surface is not horizontal, the load due to fault offset is a function of x position.

The thin plate flexure equation (3) is used to describe vertical displacements due to loads $p(x)$. Flexure tends to spread the displacements over a wide region even when the load is concentrated. The wavelength of the response depends on the flexural rigidity D. The rigidity D can be related to the effective elastic thickness of the lithosphere T_e:

$$D = \frac{ET_e}{12(1-v^2)} \qquad (5)$$

Young's Modulus E is taken to be 5×10^{10} N/m^2 and Poissson's ratio v is 0.5. The term effective elastic thickness T_e [e.g., Bodine et al., 1981] is used since the physical properties of the lithosphere may vary with depth. T_e is the thickness of a perfectly elastic plate, with given elastic constants, which has the same rigidity as the lithosphere. The wavelength of the flexural response to a point load is about $5(T_e)^{3/4}$ in kilometers for T_e in kilometers. In this paper the only parameter used to describe the flexural response of the crust will be T_e, with the understanding that it is related to D through equation (5).

Derivation of strains through the upper crust requires a corollary approximation. [redacted]

When fault offset occurs, the normal fault should be rotated as shown in Figure 6. If small fault displacements are considered, then both the constant stress field and thin plate approximations are valid. In this model it is assumed that the response to large fault displacements is qualitatively similar to the results for small displacements. Large fault offsets are required to produce fault rotations which allow part of the fault to be abandoned.

[redacted]

RESULTS

Constant Flexural Rigidity

[redacted] calculation for extension with [redacted] of 0.5 km are [redacted] Model results for different [redacted] figure. In the

[redacted]

The effect of changing the effective flexural rigidity of the crust is illustrated in Figure 8. The plots show the faulting geometry for effective elastic thicknesses of 1.0 km, 0.5 km, and 0.25 km. All lengths in the calculated fault geometries scale approximately with the T_e to the 3/4 power. This scaling is not surprising since the wavelength of the flexural response depends on $T_e^{3/4}$. The scaling would be exact if the depth of nucleation were not at a finite depth. [redacted]

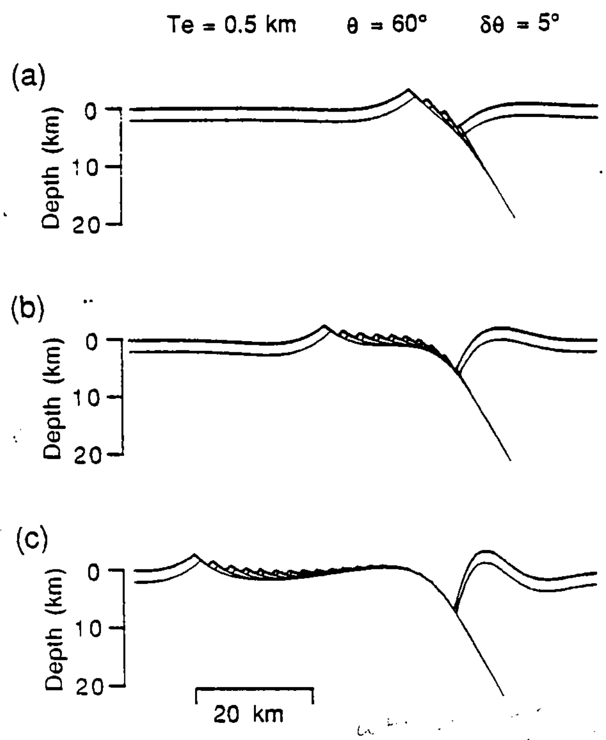

Fig. 7. Topography and positions of active and abandoned faults resulting from a calculation with a fault angle θ of 60°, an effective flexural rigidity of 0.5 km, and a rotation angle δθ of 5°. A line at 2 km depth is also plotted. Horizontal offsets of approximately (a) 15 km, (b) 30 km, and (c) 60 km are shown. There is no vertical exaggeration.

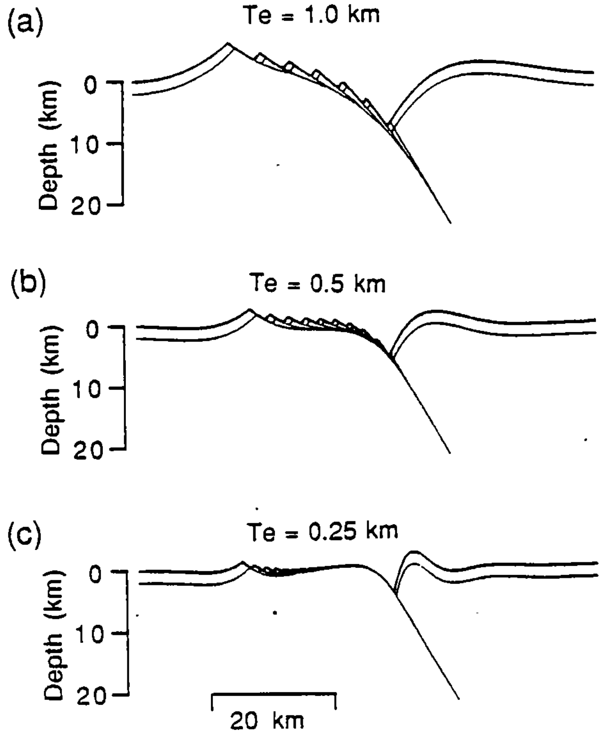

Fig. 8. Same as Figure 7 for different flexural rigidities: T_e = 1 km, T_e = 0.5 km, T_e = 0.25 km.

Variable Flexural Rigidity

The strong bending of the elastic lithosphere should lead to a lowering of T_e. Based on laboratory studies of the deformation of minerals which make up the crust and upper mantle [e.g., Goetze and Evans, 1979], bending stresses should cause part of the elastic upper crust to depart from perfect elasticity. This should occur because stresses within the plate are limited by the strength of the lithosphere. [redacted]

Using the concept that lithospheric stresses are limited by a yield stress envelope, a relation between plate curvature and plate rigidity can be derived [Chapple and Forsyth, 1979; Bodine and Watts, 1979; McNutt

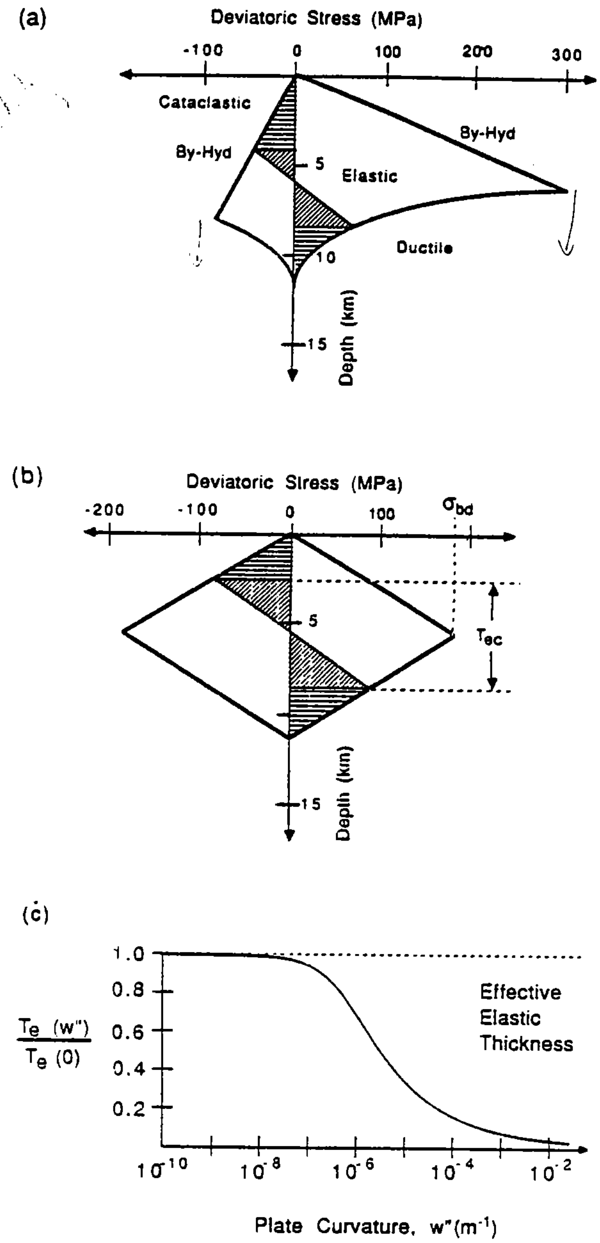

Fig. 9. (a) Yield stress envelope showing the stress which causes rock failure as a function of depth for a given temperature structure and strain rate within the crustal lithosphere. (b) a simplified yield stress envelope used to calculate the approximate dependence of the effective flexural rigidity and the curvature of the lithosphere given in the Appendix. (c) the relation between the effective elastic lithospheric thickness T_e as a function of plate curvature, w" for the yield stress envelope shown in (b) as given by equation (A5). T_e is normalized by its value for zero curvature.

and Menard, 1982]. For the purpose of this paper the yield stress envelope can be approximated by a diamond-shaped envelope, as shown in Figure 9b. This simplified shape allows the derivation of an analytic relation between curvature and effective elastic thickness of a plate T_e. Since crustal composition and thermal state are uncertain, this treatment is warranted. The Appendix gives this derivation. Figure 9c shows the variation of the effective elastic thickness T_e and curvature w" for the yield stress envelope shown in Figure 9b. When curvatures are greater than 10^{-6} m^{-1}, the plate rigidity goes down markedly.

The lowering of rigidity by plate bending is incorporated into the calculation of the response to normal fault offset. After a fault offset of 200 m the vertical response to the new load is calculated using equation (3). The curvature at each point is calculated and the local reduction in T_e is given by equation (A5) (in the Appendix). The grid spacing is taken to be 200 m. To insure numerical stability, the values of T_e are averaged using Gaussian interpolation with a half width of 1 km. The width of the Gaussian distribution used for interpolation has little effect on the model results as long as it is greater than the minimum flexural wavelength resolved by the grid spacing but not more than several times this distance. Using the new distribution of T_e, the response to fault offset loading is recomputed. For each offset the distribution of T_e and curvature is iterated to convergence (until changes in vertical offset are less than 40 m between iterations). [redacted]

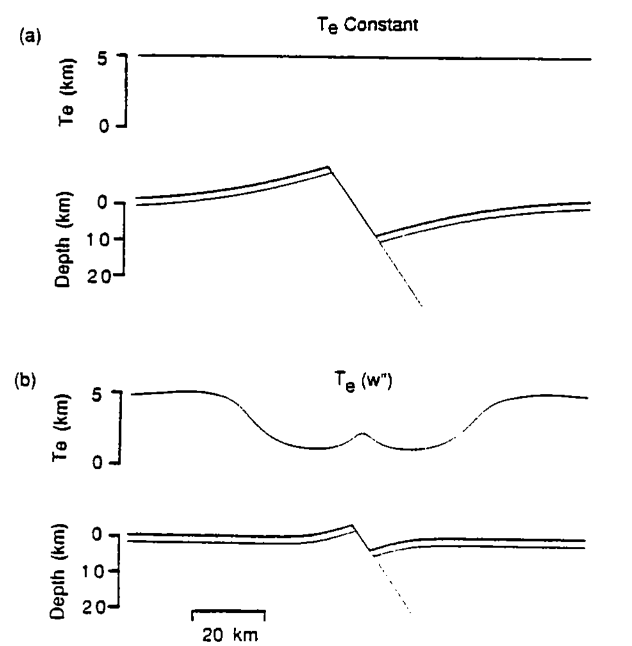

Fig. 10. Two treatments of offset on a single fault required to produce 5° of rotation of the fault. (a) shows results with constant flexural rigidity and (b) shows that when T_e is taken to be a function of plate curvature a much smaller fault offset is required to produce 5° of rotation of the fault.

[redacted]

The results of one calculation with repeated abandonment of fault segments are shown in Figure 11 with $T_e = 5$ km for crust of zero curvature. T_e drops to very low values in the region adjacent to active faulting. This has the effect of reducing the model topographic relief since much smaller fault offset is required to produce a given rotation of the active normal fault. [redacted]

Calculation with Sedimentation

Large topographic slopes should lead to erosion and structural collapse of the hanging wall block. Erosion rates depend on rock type and environmental factors in a complex way. Collapse of the hanging wall block by faulting is also difficult to model. Both processes should act to fill in the topographic low over the

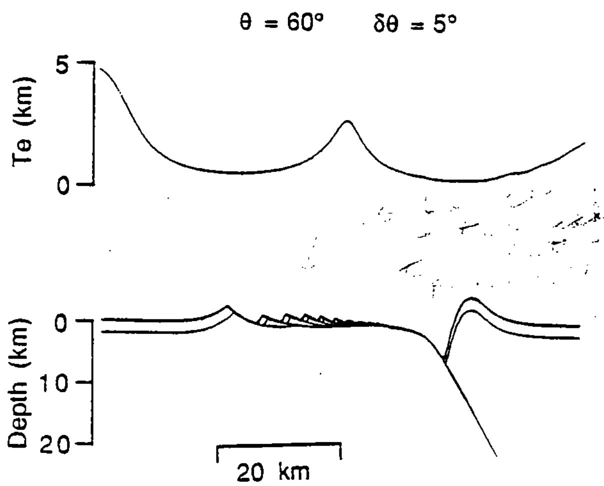

Fig. 11. Topography and positions of active and abandoned faults from a calculation in which the rigidity depends on curvature of the upper crust. Active fault dip $\theta = 60°$ and $\delta\theta = 5°$. The upper line shows the effective elastic lithospheric thickness T_e for positions across the fault system. $T_e = 5$ km for areas with zero curvature.

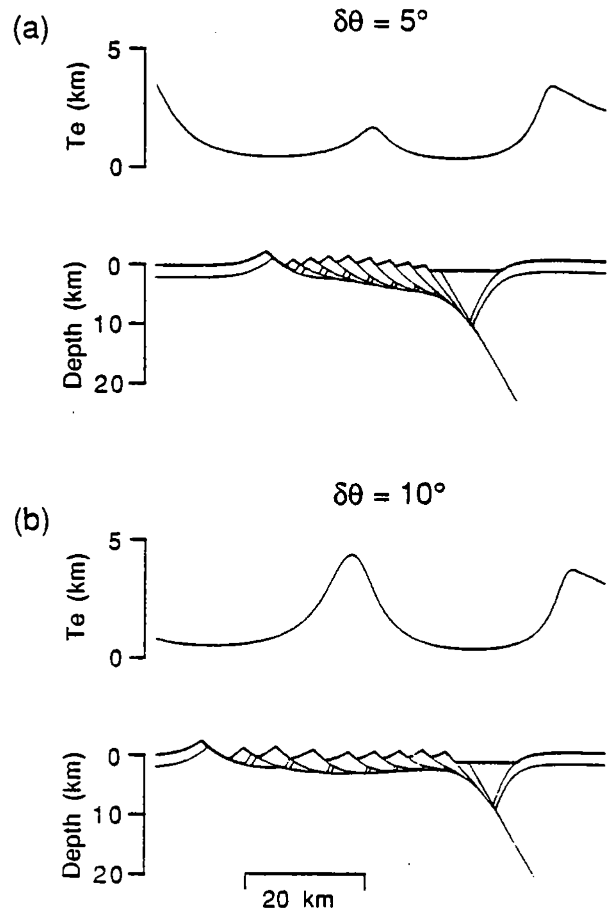

Fig. 12. Results of variable rigidity calculations in which sediment has been added to fill in all levels below 1000 m depth. Two cases of assumed angle of fault rotation before fault abandonment are shown: $\delta\theta = 5°$ and $\delta\theta = 10°$.

actively slipping fault. In the interest of simplicity, only the effect of sediment loading is treated here. Sediment is made to fill in all depths below a depth of 1 km. [redacted] local plate rigidity is again taken to be a function of plate curvature through equation (A5). Figure 12 shows the results of two calculations with sediment loading added to the normal fault loading. The calculations differ only in the amount of fault rotation before abandonment $\delta\theta$.

[redacted]

DISCUSSION AND CONCLUSIONS

Several features of metamorphic core complexes are qualitatively reproduced by this model. The model predicts (1) a nearly flat-lying abandoned normal fault below slices of upper plate rocks and sedimentary infill, (2) a strong contrast in metamorphic grade across this abandoned "detachment," and (3) rapid movement of lower plate rocks from midcrustal depths to shallower depths.

In the flexural rotation model the major slip motion takes place on a high-angle fault. Mylonites could be formed by strain in the midcrust and carried rapidly upward in the footwall of a high-angle fault. For example, the model shown in Figure 12b predicts that material from 15 km depth would be brought to within 3 km of the surface with 10 km of extension. Observations in core complexes show that the material below the detachment (at 0-5 km depth) originated in the middle of the crust (at 10-15 km depth) and was brought up in as little as 2 m.y. [Dokka et al., 1986; Davis; 1988]. Extension rates as low as 0.3 cm/yr could do this. Active high-angle fault motion is consistent with seismic observations of normal fault earthquakes. It is also consistent with the occurrence of earthquake generated pseudotachylites being associated with detachments.

The results of calculations with sediment infill suggest that such loads may be an important control on detachment geometry. Structural collapse of the hanging wall block would also act to fill in the topographic low over the actively slipping fault, as would volcanics. All infilling material should contribute to the wedges which bury the detachment. The infill would be essentially unmetamorphosed, while the rocks below part of the detachment would be highly metamorphosed. One can speculate that differences between core complexes may be due to different erosion rates and to the resistance of the hanging wall to collapse.

A low effective elastic thickness of the crust is required to produce a geometry of abandoned faults which are of about the same dimensions as observed for core complexes. Extremely high crustal temperature gradients could lead to such low elastic rigidity. A better explanation is that plate bending stresses and the finite yield strength of the crust lower rigidities. RCS (1988), in studying high-angle normal faults with displacements up to 5 km, also require low flexural rigidities. The relation between plate bending and rigidity can also explain their results (see Figure 10).

A major question about metamorphic core complexes is why they are only found in a relatively small number of places. The present-day extension of the Basin and Range primarily seems to produce high-angle normal fault structures [Fletcher and Hallet, 1983; Jackson, 1987]. These calculations do not address this problem, but the model results do have implications for it. In this model, crustal extension is accommodated by lower crustal material being pulled up and cooled. Were the Moho pulled up by repeated fault motion it might become energetically more favorable for faulting and extension to occur in an adjacent area of unthinned crust [Artyushkov, 1973]. Viscous flow in the lower crust would tend bring material into the area of crustal thinning and flatten Moho. The rate of flow in the lower crust is directly proportional to viscosity there. The lower crust must flow sufficiently in the time between major faulting events to maintain a relatively flat Moho. High crustal temperatures could lower the viscosity enough for this to happen [e.g., Kusnir and Park, 1987]. The seismically determined Moho in the Basin and Range is nearly flat [Allmendinger et al., 1987], indicating that the lower crust can flow fast relative to other areas where there is measurable Moho topography. When core complexes were formed, the local crustal temperatures may have been greater than today. Several workers have noted the correlation of nonbasaltic volcanism with extension in core complexes [Glazner and Bartley, 1984; Gans, 1987]. This type of volcanism may reflect melting of the lower crust. Temperatures high enough to cause melting should make the average viscosity of the lower crust quite low. Thus extension might be concentrated in the area of a single fault when the crust there is anomalously hot.

New geologic data could be used to test predictions of the flexural rotation model. This model predicts that fault blocks are formed sequentially, becoming younger toward the hanging wall. In the alternative model of rotating sets of high-angle faults [e.g., Miller et al. 1983] the fault blocks are formed contemporaneously. The active low-angle model [Wernicke, 1981] makes no prediction of timing of block faulting. Radiometric dating of volcanic strata formed at the time of active fault motion in core complexes may allow determination of the sequence of faulting. In the Death Valley area, which may be a recent core complex, the sequence of block faulting is consistent with the flexural rotation model [J. D. Walker, personal communication, 1988].

More work is needed to test whether the approximations used here reasonably describe how the crust responds to normal faulting. Two-dimensional numerical calculations of the effect of large offsets of a normal fault in a crust with brittle and ductile regions must be done to confirm these results. The calculations will be complex and depend on many physical parameters. However, the similarity between predictions of this simple model and field observations should be enough to warrant further work on the isostatic response of the crust in extension.

APPENDIX

Considering a simplified yield stress envelope as shown in Figure 9b, a relation between curvature of a lithospheric plate and its effective elastic thickness is derived following Bodine et al., [1981] and McNutt [1984]. The effective rigidity of the lithosphere, D, is related to the bending moment of the plate, M, through the curvature of the plate:

$$D(x) = \frac{M(x)}{w''(x)} \tag{A1}$$

The moment is calculated by integrating the horizontal stresses σ_{xx} through the plate:

$$M(x) = \int \sigma_{xx}(x,z)\, z\, dz \tag{A2}$$

The elastic core of the lithosphere is the portion in which bending stresses are less than the yield stress. In the elastic core of the plate the horizontal bending stresses vary linearly through the plate. The parts of the plate at its top and bottom, which deform plastically, support stresses equal to the yield stress (see Figure 9a). For equilibrium within the plate:

$$w''(x) = \frac{M_T(x)}{D_T(x)} = \frac{M_{ec}(x)}{D_{ec}(x)} \tag{A3}$$

where the subscript "T" denotes the total integral through the yield stress envelope, while "ec" represents the integral through the elastic core alone.

Combining equation (A3) and the relation between rigidity and effective elastic thickness (equation (3)) gives:

$$T_e(x) = T_{ec}(x) \left[\frac{M_T(x)}{M_{ec}(x)} \right]^{1/3} \tag{A4}$$

Integrating through the elastic core of the yield stress envelope gives M_{ec} while integrating through the entire envelope gives M_T. Using the simple geometry of a diamond-shaped yield stress envelope gives:

$$T_{ec}(x) = 2 Z_{bd} \left[1 + \frac{Z_{bd} E\, w''(x)}{\sigma_{bd}(1 - \nu^2)} \right]^{-1} \tag{A5}$$

and

$$T_e(x) = T_{ec}(x) \left\{ \frac{[2 Z_{bd}^2 + T_{ec}(x) Z_{bd}]}{[T_{ec}(x)]^2} \right\}^{1/3}$$

The relation between T_e and w" is shown in Figure 9c for $\sigma_{bd} / Z_{bd} = 0.33$ kbar/km.

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