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A note comparing parameters controlling low-angle normal and thrust movement

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Abstract—A simple model for the initiation of normal and thrust motion along pre-existing low-angle faults shows that the parameters necessary for fault movement are nearly the same in both cases. The parameters controlling fault movement (which are tectonic tension and tectonic compression as well as fault friction or fault strength) are considered in terms of available observations of crustal stresses, and possible crustal stresses predicted from laboratory rock friction experiments. Because tectonic compression can exceed tectonic tension it is concluded that thrust movement along low-angle zones of weakness is somewhat more likely than normal movement; the relative difficulty of the different motions depends upon the ratio of tectonic compression to tectonic tension. If available observational data of maximum and minimum horizontal stresses in the crust represent reasonable upper and lower limits to tectonic compression and tectonic tension, then low-angle movement is greatly facilitated by low coefficients of fault friction, similar to values obtained in a number of recent studies of weak faults. © 1997 Elsevier Science Ltd. All rights reserved.

INTRODUCTION

This paper compares the basic parameters controlling normal and thrust motion along pre-existing low-angle faults that penetrate the brittle crust. In essence, this involves a comparison of conditions for normal and thrust reactivation along low-angle faults. The reason for examining this problem is that considerable disagreement exists as to the origin of low-angle normal faults that penetrate the brittle crust (so-called detachment faults; e.g. Wernicke, 1981; Johnson and Loy, 1992; Burchfiel *et al.*, 1992). Johnson and Loy (1992) express the need to explain why low-angle (normal) faults should move at all. The present model, which is surely a simplification of natural processes, may be described as restrained gravity sliding for normal motion. The model straightforwardly demonstrates the possibilities of normal motion, as compared to thrust motion, along low-angle faults.

Although the analysis in the present study assumes that the low-angle fault already exists, it is instructive to review very briefly some of the geologic studies relating to low-angle normal faults. Hamilton and Myers (1966) suggest large amounts of Cenozoic extension in the Cordillera generally before high-angle Basin and Range faulting; an idea consistent with mid-crustal rocks opposite upper crustal rocks along low-angle normal faults (Crittenden *et al.*, 1980). Much work on low-angle normal faulting has since been done (e.g. Davis and Lister, 1988 and Lister and Davis, 1989; Burchfiel *et al.*, 1992; Wernicke, 1992). Anderson's theory of faulting suggests low-angle faults initiate in compressive environments (e.g. Turcott and Schubert, 1982), although some studies suggest principal stress reorientation or additional stress development, so that low-angle faulting might initiate in extensional environments (e.g. Reiter and Minier, 1985; Spencer and Chase, 1989; Yin, 1989; Melosh, 1990). Rotation of faults during extension has been suggested (e.g. Buck, 1988; Wernicke and Axen, 1988); other studies in western Arizona suggest detachment faults there originate at low angles (Scott and Lister, 1992). The lack of seismicity indicating active low-angle faulting has been noted (Jackson and White, 1989); more recent studies however suggest seismogenic low-angle faulting (Abers, 1991; Johnson and Loy, 1992). Low-angle normal faulting may suggest detachment faults are weak due to high pore pressure causing a change in crustal stresses within the fault zone (Byerlee, 1990; Axen, 1992; Rice, 1992). Axen and Selverstone (1994) do suggest elevated shear stress along the Wipple detachment. Similarities of major thrust and detachment faults in terms of regional extent and fault orientation, and the proximity of thrusts to detachment faults in some areas of the Cordillera, suggest some detachments are faults initiated in compression. Normal movement along previously initiated low-angle thrusts, or gently dipping joints and cracks, in regions of the North American Cordillera has been noted (Bruhn *et al.*, 1982; Lamerson, 1982; West, 1992).

MODEL

Model basics

In the present study a straightforward analysis of forces which operate along a low-angle fault in either extensional or compressional tectonic environments has been used. This analysis allows a comparison of the ease of thrust and normal movement along low-angle faults. Figure 1 illustrates the model and the force diagram used. A planar low-angle fault is assumed to exist and to penetrate the brittle (upper) crust, from the surface to the brittle–plastic transition (Rutter, 1986). The brittle upper crust rests on the plastic lower-crust, and the viscous drag experienced along the brittle–plastic transition during fault movement is assumed to be either negligible or the

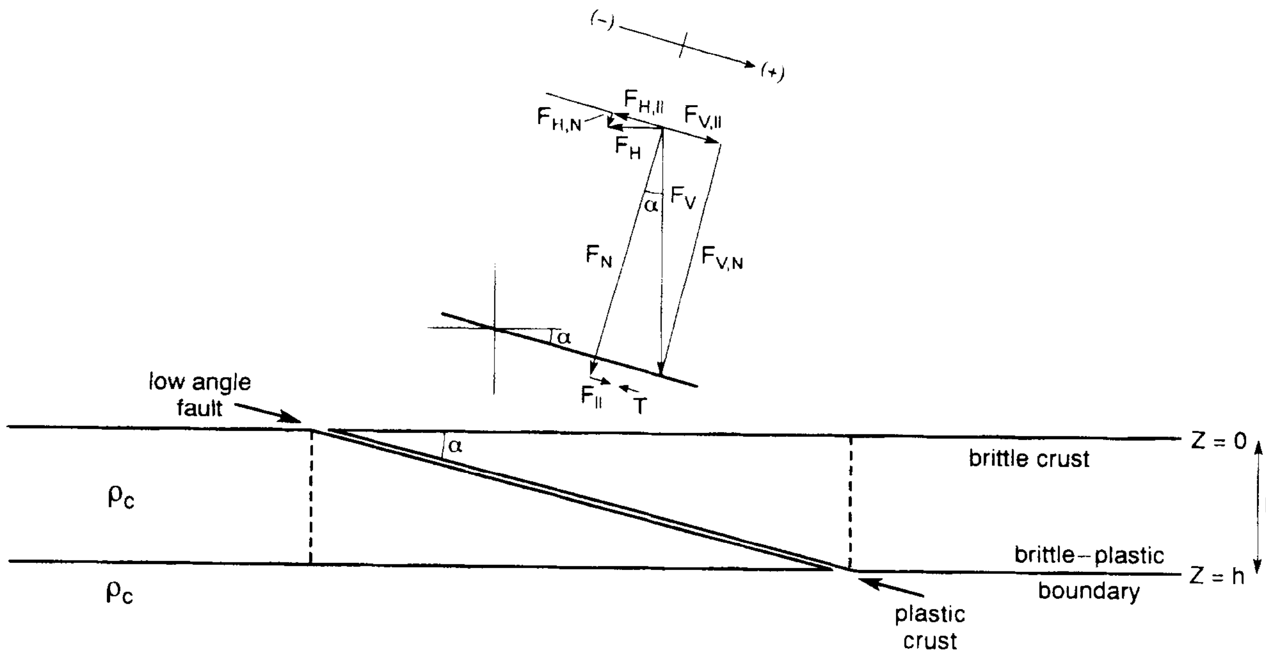

Fig. 1. Diagram of forces acting on low-angle detachment fault (above); illustration of low-angle fault structure (detachment fault) in crust (below). Note (+) and (−) coordinates, ρ_c is crustal density.

same for both thrust and normal movement. The upper crust may also move along a mid-crustal fluid layer like the one detected from *S*-wave screening studies in New Mexico (Sanford *et al.*, 1973).

Tectonic stresses in the brittle crust should control motion along upper crustal faults, whatever the cause of stress development, e.g. by such boundary conditions as plate interactions, hot spots, or other displacements. A simple reversal of the tectonic stresses will promote conditions for different fault movement. Vertical and horizontal forces can be derived from the vertical and horizontal components of stress in the brittle crust, and it is the balance between the components of these forces and the resistance to motion along the fault which determines if fault motion is possible. From Fig. 1 we see that to balance forces parallel to the fault plane, for possible normal movement,

$$F_\| + (-T) = 0 \tag{1a}$$

or

$$F_\| = F_{V,\|} - F_{H,\|} = T \tag{1b}$$

where $F_{V,\|}$ is the fault-parallel component of the vertical force, $F_{H,\|}$ is the fault-parallel component of the horizontal force, and T is the shear or friction force opposing motion along the fault plane. Note that T and the resultant force $F_\|$ are oppositely directed, T being in the negative direction for the normal motion case (see Fig. 1). If we let

$$T = \mu_f F_N \tag{2}$$

where F_N is the normal force across the fault, and μ_f is the coefficient of friction along the fault, then

$$F_{V,\|} - F_{H,\|} = \mu_f F_N \tag{3}$$

Byerlee (1968, 1975) showed that in laboratory experiments

$$\tau = \mu_R \bar{\sigma}_N = 0.85\bar{\sigma}_N, 3 < \bar{\sigma}_N < 200 \text{ MPa} \tag{4a}$$

$$\tau = 60 \pm 10 + 0.6\bar{\sigma}_N, \bar{\sigma}_N > 200 \text{ MPa} \tag{4b}$$

where μ_R is the coefficient of rock friction in lab experiments, $\bar{\sigma}_N$ is the effective normal stress ($\bar{\sigma}_N = \sigma_N - P$, P is pore pressure) and τ is the shear stress at which friction across a cut or fracture is overcome. Equations (2) and (3) assume that all resistance to slip is encompassed by a single coefficient, μ_f. On a natural fault zone extending across the brittle crust, movement on some segments of the fault may occur by frictional slip whereas movement on other segments may occur by other processes, such as ductile deformation within a narrow zone (Mitra, 1984; Wojtal and Mitra, 1986). In this simple analysis I presume that the coefficient of friction μ_f is representative of the coefficients of resistance to slip across faults in nature. If shear stresses transmitted in laboratory experiments are representative of shear stresses transmitted along faults, then the effects of pore pressure will be included in μ_f; i.e. $\mu_f = \mu_R(\sigma_N - \text{P})/\sigma_N$. If low values of μ_f are necessary for fault movement, then nonfrictional processes may well operate along fault surfaces, e.g. ductile behavior. With these concerns we proceed from equation (3) because it is constructive to consider the potential strength of a fault in terms of both a frictional model and the observational stress data available for the brittle crust.

The terms in equation (3) may be written as

$$F_{V,\parallel} = F_V * \sin\alpha \qquad F_{H,\parallel} = F_H * \cos\alpha \qquad \text{(5a), (5b)}$$

and

$$F_N = F_{V,N} + F_{H,N} = F_V * \cos\alpha + F_H * \sin\alpha \qquad \text{(5c)}$$

Equation (5c) is equivalent to equation (8), page 55, in Jaeger and Cook (1969).

In order to write the force balance equation for potential thrust motion we must recognize that $F_{H,\parallel}$ becomes greater than $F_{V,\parallel}$. This being the case, the resultant force in equation 1a, $F_\parallel$ (which will now be $F_{H,\parallel} - F_{V,\parallel}$), is negatively directed (see Fig. 1). The frictional resistance force in this case is positively directed. Therefore

$$-F_\parallel + T = 0, \therefore, F_{H,\parallel} - F_{V,\parallel} = T = \mu_f F_N \qquad (6)$$

Consequently, the equations describing force balance are somewhat different for the cases of possible normal and thrust movement.

Estimates of crustal stress

At this point it is important to consider possible stress distributions in the brittle crust in order to calculate the forces acting along the fault. An attempt is made to estimate crustal stresses primarily from *in-situ* observational data, although laboratory data are also considered. Observational field data show that the vertical stress at depth may be well approximated as the weight of the overburden (McGarr and Gay, 1978). McGarr (1988) also points out that the magnitude of the overburden stress is the most reasonable stress condition for all three coordinates in the absence of applied tectonic stress (i.e. lithostatic stress where $\sigma_x = \sigma_y = \sigma_z = \rho g z$).

Although the horizontal stress field in the crust is not well known (Hanks and Rayleigh, 1980), McGarr and Gay (1978) show that the horizontal stress field in the upper-brittle crust may be reasonably approximated as a fraction of the overburden; i.e. $f = \sigma_H / \sigma_V$, where $f < 1.0$ implies an extensional tectonic environment, $f > 1.0$ a compressional tectonic environment, and $f = 1.0$ a state of lithospheric stress without tectonic forces.

Brace and Kohlstedt (1980) apply Byerlee's Law (with and without pore pressure) to obtain estimates of horizontal stress as a function of depth; they compare these estimates to the data presented in McGarr and Gay (1978). The data for minimum horizontal stress from South Africa generally lie between two appropriate linear stress profiles generated from Byerlee's Law; one profile with hydrostatic pore pressure and one without pore pressure. In United States basins, a linear fit to all of the minimum horizontal stress observational data is in close agreement with the minimum horizontal stress distribution predicted using Byerlee's Law for hydrostatic pore pressure conditions (15 MPa km^{-1} vs 13 MPa km^{-1}, respectively). The estimates of maximum horizontal stress gradients calculated using Byerlee's Law for hydrostatic pore pressure are somewhat greater than the least mean squares fit to the maximum horizontal stress data taken near the margin of the Canadian Shield (96 MPa km^{-1} vs 44 MPa km^{-1}, data number = 17, or 64 MPa km^{-1}, data number = 16; see also Table 1).

Byerlee's Law suggests a linear increase of horizontal stress vs depth, with a slope and intercept change at about 4 km. From Table 1, one can estimate the appropriateness of a least squares linear fit to observations of *in-situ* horizontal stress. A linear fit to the data is as reasonable as any other possibility and makes sense in terms of Byerlee's Law (although a different coefficient of friction may apply). Therefore, I suggest that the horizontal stress (minimum or maximum) is a fraction (less or greater than 1.0) of the overburden.

Calculation of forces along the fault

With the assumption that the vertical stress can be written as

$$\sigma_V = \rho_c g Z \qquad \text{(7a)}$$

where ρ_c, g, and Z are respectively the crustal density, gravity acceleration and depth, and the horizontal stress is a linear function of the vertical stress and can be written as

$$\sigma_H = f \rho_c g Z \qquad \text{(7b)}$$

we proceed to calculate the forces operating along the fault. To calculate the vertical force, one must integrate the vertical stress over the horizontal extent of the fault, therefore

Table 1. Compilation of horizontal stress data from McGarr and Gay (1978)

Location	Depth interval (km)	σ_H (min)/depth† (MPa km^{-1})	n (# data)	r (corr. coeff.)	Rk type (if given)
S. Africa	0–2.5	7.7	17	0.68	
	1.0–2.5	7.9	9	0.40	
U.S. basins	0–5.1	15.0	41 (est)	*	s.s., granite
	0.9–5.1	19.8	10	0.99	s.s.
	0.77–3.67	8.2	5	0.85	granite
Canadian Shield	0–1.7	64.3 [σ_H (max)]	16	0.88	
	0–2.1	43.9 [σ_H (max)]	17	0.80	

Note: * data bunched near surface make depth and stress difficult to read; correlation coefficient should be similar to depth interval 0.9–5.1 km.
† σ_H (max)/depth for Canadian Shield.

$$F_V = \rho_c g \int_0^h \cot\alpha * z \, dz = \rho_c g(\tfrac{1}{2})(\cot\alpha)h^2 \qquad (8a)$$

and

$$F_{V,\parallel} = F_V * \sin\alpha = \tfrac{1}{2}\rho_c g(\cot\alpha)\sin\alpha h^2 = \tfrac{1}{2}\rho_c g(\cos\alpha)h^2 \qquad (8b)$$

Note the analysis is two dimensional; therefore, the forces and weights calculated will be per unit strike length.

The horizontal force is calculated by integrating the horizontal stress over the vertical extent of the fault; therefore

$$F_H = f\rho_c g \int_0^h z \, dz = \tfrac{1}{2} f\rho_c g h^2 \qquad (9a)$$

and

$$F_{H,\parallel} = F_H * \cos\alpha = \tfrac{1}{2} f\rho_c g h^2 * \cos\alpha \qquad (9b)$$

From the expression for the normal force (equation 5(c)), we write

$$F_N = \tfrac{1}{2}\rho_c g h^2(\cot\alpha)(\cos\alpha) + \tfrac{1}{2}\rho_c g f h^2(\sin\alpha) \qquad (10)$$

By substituting equations (8b), (9b), and (10) into the basic force balance equations, either equation (3) or equation (6), we arrive at an expression relating the dip angle of the fault (α), the coefficient of fault friction (μ_f), and the ratio of horizontal to vertical stress (f), for normal and thrust motion, respectively. The relation for normal motion is

$$f = (1 - \mu_f \cot\alpha)/(1 + \mu_f \tan\alpha), \qquad (11a)$$

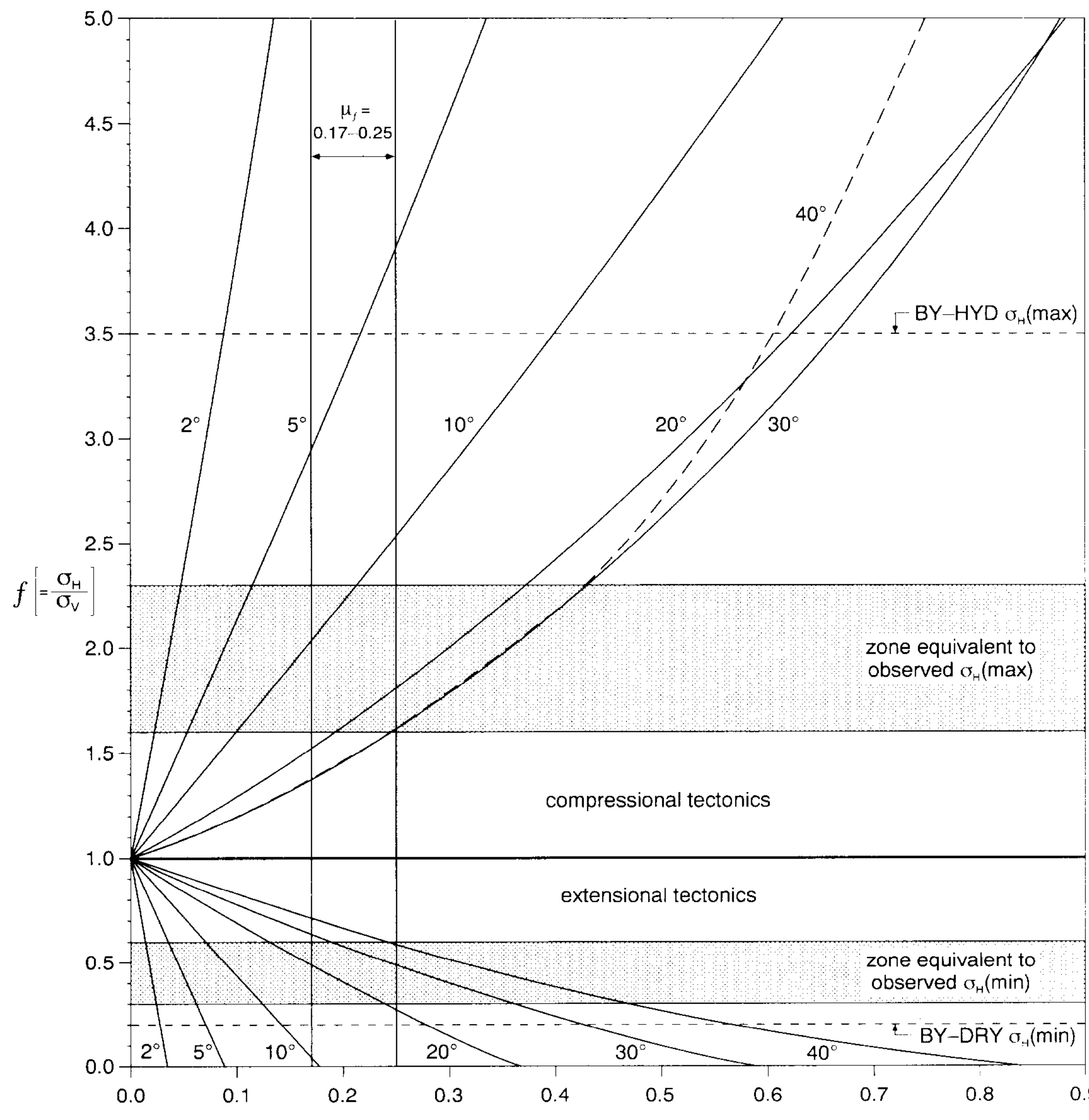

Fig. 2. Illustration of $f\,(=\sigma_H/\sigma_V)$ vs μ_f (fault friction) for fault angles between 2° and 40°. Zones of observed minimum f and maximum f are noted (from McGarr and Gay, 1978). σ_H (min) for Byerlee's Law (dry), and σ_H (max) for Byerlee's Law (with hydrostatic pore pressure) are also noted (after Brace and Kohlstedt, 1980; BY-DRY and BY-HYD, respectively). Thrust movement is appropriate for $f > 1.0$, normal movement for $f < 1.0$. Values of fault friction, μ_f, between 0.17 and 0.25 are indicated (after Bird & Kong 1994).

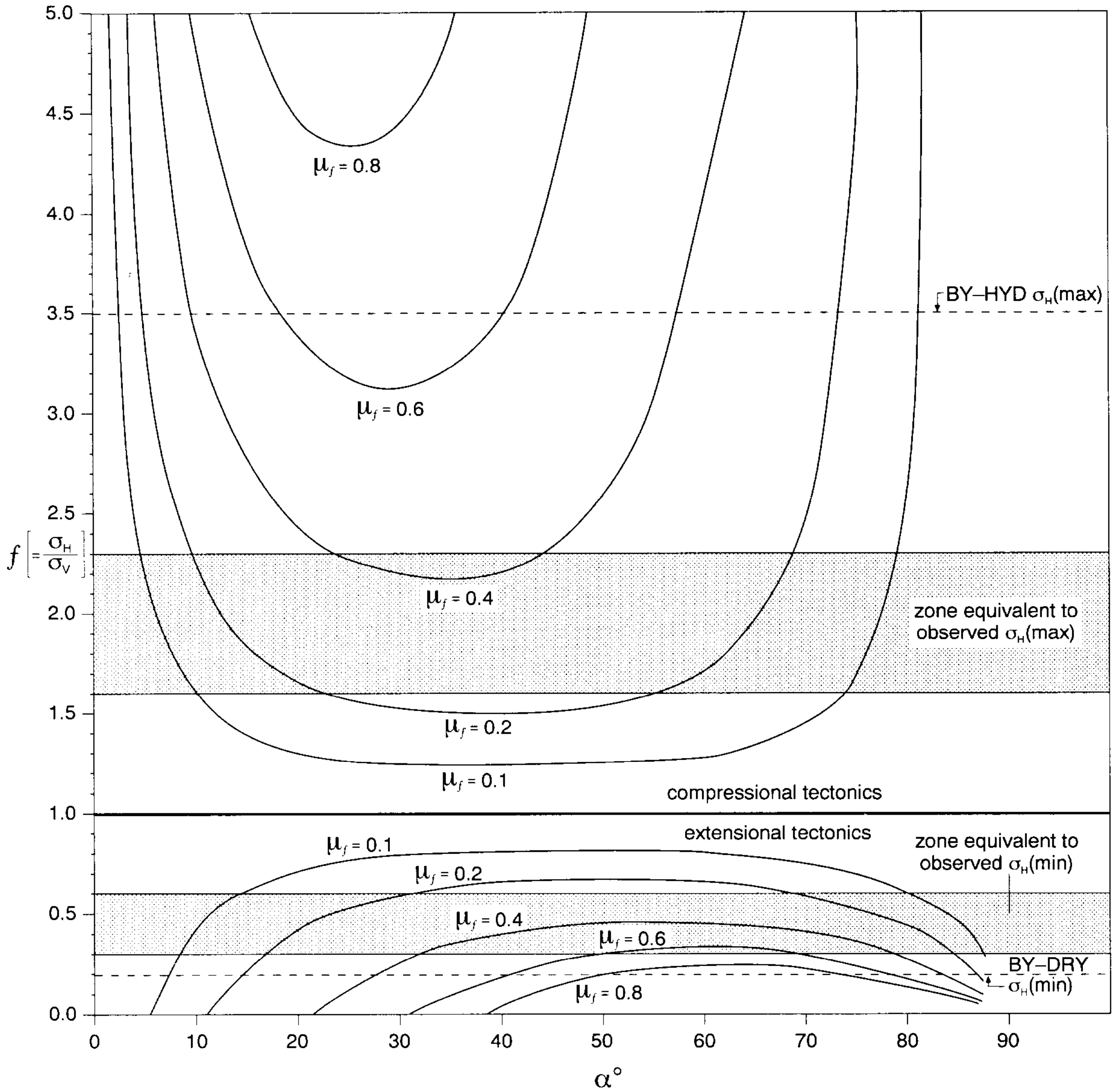

Fig. 3. Illustration of f ($=\sigma_H/\sigma_V$) vs α (fault angle) for various values of fault friction (μ_f). Zones of observed minimum f and maximum f are noted (from McGarr and Gay, 1978). σ_H (min) for Byerlee's Law (dry), and σ_H (max) for Byerlee's Law (with hydrostatic pore pressure) are also noted (after Brace and Kohlstedt, 1980; BY-DRY and BY-HYD, respectively). Thrust movement appropriate for $f > 1.0$, normal movement for $f < 1.0$.

and the relation for thrust motion is

$$f = (1 + \mu_f \cot\alpha)/(1 - \mu_f \tan\alpha). \qquad (11b)$$

Equation (11b) is similar to the expression presented by Sibson (1985). Note that the expressions relating α, μ_f, and f for normal and thrust faulting are somewhat different (i.e. equations (11a) and (11b)). This study compares parameters for normal and thrust faulting using the two different equations.

Comparison of parameters controlling normal and thrust movement

Equations (11a) and (11b) can be illustrated as f ($=\sigma_H/\sigma_V$) vs μ_f (fault friction) for various fault dip angles (α), or as f vs α for various fault friction values (μ_f), by iterating one of the independent variables as the other is changed sequentially. Figure 2 illustrates f as a function of μ_f for dip angles 2° to 40°, in both compressional and extensional environments. Figure 3 illustrates f as a function of dip angles (~2° to 88°) for various fault friction μ_f. A figure similar to the region of compressional tectonics shown in Fig. 3 appears in Lachenbruch and McGarr (1990); in Figs 2 and 3 the analysis is extended to incorporate regions of extensional tectonics as described by equation (11a). The first order observation from Figs 2 and 3 is that over similar departures from the lithostatic state (f= 1.0), the conditions controlling normal and thrust motion are nearly the same, with normal movement actually being somewhat easier than thrust movement for increasing dip angles. For example, equilibrium at f= 0.5 and α = 30° occurs in the extension regime at $\mu_f \simeq 0.24$, in the compressional regime equilibrium at f= 1.5 and α = 30° occurs at $\mu_f \simeq 0.22$. This seems reasonable because it is about as easy to pull something down an incline as to push it up when the incline is at a small angle;

and it becomes increasingly easier to pull an object down the incline as it becomes steeper. Note that the analysis breaks down at $\alpha = 0°$ and $\alpha = 90°$ because equations (11a) and (11b) become undefined. Also note that by choosing a specific brittle crustal thickness (h), and varying the dip angle (α), lower angle faults require greater deviation from the lithostatic case ($f = 1.0$) in order to move because the length of the fault increases trigometrically as the dip angle decreases.

The remaining question to answer regarding the relative difficulty of reactivating normal and thrust faults is how much deviation from the lithostatic stress state is possible in compressional and extensional tectonic environments. I here define the deviation from lithostatic stress as the tectonic stress; i.e. tectonic compressive stress when $f > 1.0$ and tectonic tensional stress when $f < 1.0$, where the tectonic compression is $f - 1.0$ for $f > 1.0$, and the tectonic tension is $1.0 - f$ for $f < 1.0$.

Starting with extensional tectonic regimes, observational data presented in McGarr and Gay (1978) show that the minimum values for f are about 0.3 in U.S. basins and in South Africa; the value of f for all data in U.S. basins is about 0.6. Using Byerlee's Law, Brace and Kohlstedt (1980) calculate a stress profile equivalent to $f = 0.2$ for South Africa if no pore pressure is present; they calculate a stress profile equivalent to $f \simeq 0.5$ for both South Africa and U.S. basins given hydrostatic pore pressure. With these analyses it would appear that the minimum value one might expect for f is between 0.2 and 0.3. In Figs 2 and 3, a zone equivalent to observed σ_H (min), and a predicted σ_H (min) from Brace and Kohlstedt (1980) is indicated.

In compressive tectonic regimes, observations around the Canadian Shield yield values of σ_H (max) equivalent to $f = 1.6$ or $f = 2.3$ (depending on the data used, see Table 1). Using Byerlee's Law with hydrostatic pore pressure, Brace and Kohlstedt (1980) calculate a σ_H (max) profile equivalent to $f \simeq 3.5$. Byerlee's Law without pore pressure will yield a σ_H (max) profile equivalent to $f \simeq 4.7$. It therefore appears that much more tectonic compression than tectonic tension is possible in the brittle crust.

From Figs 2 and 3 we can estimate the various parameters necessary for normal or thrust movement along low-angle faults given conditions of greater tectonic compression than tectonic tension. For example, consider the conditions for σ_H (min) and σ_H (max) derived from observational data, i.e. $f = 0.3$ and $f = 2.3$, respectively (Figs 2 & 3). In this case normal movement is possible for dip angles of 5°, 10°, 20°, and 30° if $\mu_f \leq 0.06$, 0.12, 0.24, and 0.36, respectively; thrust movement is possible for the same dip angles if $\mu_f \leq 0.12$, 0.21, 0.36 and 0.42, respectively. For these cases where the tectonic compression is about twice the tectonic tension, the coefficient of fault friction necessary for normal movement is only slightly less than that necessary for thrust movement ($\Delta\mu_f$ being ~ 0.06 at $\alpha = 5°$, 0.09 at $\alpha = 10°$, 0.12 at $\alpha = 20°$, and 0.06 at $\alpha = 30°$). The difference in fault friction necessary for normal or thrust motion is somewhat greater if the ratio of tectonic compression to tectonic tension is greater. For example, consider fault friction values necessary for movement if $f = 3.5$ and $f = 0.2$ (where the ratio of tectonic stresses is 3.1; Fig. 2). In this case where $\alpha = 5°$, the fault friction values necessary for normal or thrust movement are less than about 0.07 and 0.22, respectively; when α is 30° the necessary μ_f value are less than about 0.42 and 0.66 for normal or thrust motion. Therefore, the similarity of parameters governing normal and thrust movement will depend upon the possible ratio of the tectonic stresses. Observational data are most important in this regard. At present, the observational data indicate a σ_H (max) equivalent to $f \simeq 2.3$; and although these data are very important, they are limited in extent and depth.

Another approach to the analysis is to consider the 'time averaged frictional coefficients of major faults', which are predicted to be between 0.17 and 0.25 by Bird and Kong (1994). In this case Fig. 2 indicates that normal faults of $\sim 13°$ are possible ($\mu_f = 0.17, f = 0.3$) and thrust faults of $\sim 8°$ are possible ($\mu_f = 0.17$, $f = 2.3$). Slightly lower angle faults are possible when $f = 0.2$ and $f = 3.5$ (Fig. 2).

Figure 3 indicates that as the coefficient of fault friction becomes smaller, a greater range of fault angles becomes possible for both normal and thrust motion. For example, if fault friction μ_f is $\simeq$ 0.2, normal movement appears possible for fault angles between 17° and 85° ($f = 0.3$); and thrust movement appears possible for fault angles between 10° and 69° ($f = 2.3$). Alternatively, if μ_f $\simeq$ 0.4, then normal movement is possible only between 33° and 79° ($f = 0.3$); and thrust movement is possible only between 25° and 44° ($f = 2.3$). From Fig. 3 estimates of fault parameters for thrust movement along high-angle faults are also possible.

DISCUSSION

From the above presentation it is concluded that for similar tectonic stresses and fault friction, the likelihood of normal and thrust movement along low-angle faults is about the same. The fact that more tectonic compression than tectonic tension is possible in the brittle crust makes thrust movement somewhat more probable than normal movement along low-angle faults. How much tectonic compression exists in the brittle crust is an important parameter and more observational data are needed. With present field data concerning maximum horizontal stresses in the crust it appears that both normal and thrust movement along low-angle faults are similarly facilitated by low coefficients of fault friction.

A number of other studies investigate the potential for low coefficients of friction along faults. In the geothermal areas of southwestern Utah, Bruhn *et al.* (1982) described low-angle normal faults associated with the movement of cover rocks from uplifting areas toward adjacent grabens. The faults dip between 5° and 35° to the west and seem to have developed simultaneously with steeply-dipping faults in the area. The authors presented an analysis derived from Jaeger and Cook (1969), for the ratio of strength along a plane of weakness, to the strength of pristine rock. Assuming a pore pressure of 125 % hydrostatic pressure (which may be possible due to the redistribution of stress in fault zones; Rice, 1992) they show that low-angle faulting is possible and predict an average coefficient of sliding friction probably between 0.15 and 0.40. Hubbert and Rubey (1959) first suggested that the effective normal stress across a fault zone can be reduced by high pore pressure, resulting in a fault that weakly transfers shear stress. Axen (1992) relates that available data suggest detachment faults are weak, and this weakness may be caused by high pore pressure resulting from a modification of the stress tensor (Rice, 1992). Chester *et al.* (1993) present observations from the internal structure of the San Gabriel and Punchbowl faults and suggest that inhomogeneous stress and elevated pore fluid pressure in a very narrow anisotropic core of the fault zone may explain a weak San Andreas Fault with no heat-flow anomaly. The fact that a dynamic weakening mechanism may be expected in mature fault zones (Chester *et al.*, 1993) could support the notion of some low-angle normal faults reactivating along previous thrust faults.

Other geophysical models also support the concept of weak active faults. Heat-flow studies along the San Andreas Fault show the lack of a heat-flow anomaly, which suggests a coefficient of fault friction less than about 0.2 (Lachenbruch and McGarr, 1990). Rotation of half grabens (Reiter *et al.*, 1992) and the deep subsidence of symmetrical grabens (Bott and Mithen, 1983; Reiter, 1995) are enhanced by low coefficients of fault friction. As mentioned above, Bird and Kong (1994) present a geodynamic model of active faults in California and suggest a time averaged coefficient of fault friction between 0.17 and 0.25.

Processes operating along fault zones that are not strictly frictional processes may contribute to fault weakness. Mitra (1984) describes high-angle reverse faults in the foreland of the Rocky Mountain Cordilleras, where a transformation from brittle fracturing to ductile deformation accompanies increasing strain and decreasing grain-size along the fault zone. In late stages of deformation, diffusional creep and strain softening may allow large strains to occur at very low deviatoric stress. Wojtal and Mitra (1986) indicate that observations of thrust faults in the southern Appalachian Mountains suggest non-frictional movement, and that extensive ductile deformation in a thin, fine-grained fault layer can accommodate sheet movement at low shear stress levels. Non-frictional processes which allow fault movement under relatively low shear stress would be masked in the present model as low coefficients of fault friction, not describing the actual phenomena along the fault zone.

Reorientation of principal stress axes is suggested as a potential mechanism for normal movement reactivation along low-angle faults (Sibson, 1985). From Jaeger and Cook (1969) (p. 14) and the analysis in the present study, one can show that equation (11a) can be applied to consider principal stress reorientation (rotation) if α is taken as the angle between the maximum principal stress and the normal to the fault plane (i.e. α is the sum of the fault dip angle and the off-vertical angle for the maximum principal stress). McGarr and Gay (1978) show that for regions of tectonic tension in southern Africa, most of the maximum principal stress measurements are within 30° of the vertical. If we allow the fault angle to be 25°, and a rotation of the principal stress axes of 30°, then from Fig. 3 we may notice that for an α of 55°, normal movement along a low-angle fault is possible if $\mu_f \leq 0.6$ ($f = 0.3$). If the maximum principal stress is vertical, $\alpha = 25°$, normal movement requires $\mu_f \leq 0.3$ ($f = 0.3$). Therefore a significant and appropriate reorientation of the principal stress axes could make normal movement much easier. However, weak detachment zones may reorient crustal stresses so that the angle between σ (max) and the fault plane actually increases rather than decreases (Byerlee, 1990). In this case, for normal movement to occur, fault friction should be even less than if the maximum principal stress was vertical.

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