How faulting keeps the crust strong

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ABSTRACT

Deep drilling and induced seismicity experiments at several locations worldwide indicate that, in general, the brittle crust in intraplate regions is critically stressed, pore pressures are close to hydrostatic, and in situ bulk permeability is ~10^{-17} to 10^{-16} m^2. This high permeability, three or four orders of magnitude higher than that measured on core samples, appears to be maintained by critically stressed faults and greatly facilitates fluid movement through the brittle crust. We demonstrate that such high permeabilities can maintain approximately hydrostatic fluid pressures at depths comparable to the thickness of the seismogenic crust. This leads to the counterintuitive result that faulting keeps intraplate crust inherently strong by preventing pore pressures greater than hydrostatic from persisting at depth.

Keywords: fault mechanics, pore pressure, crustal strength, permeability.

INTRODUCTION

Three independent lines of evidence indicate that intraplate continental crust is in a state of failure equilibrium: (1) the widespread occurrence of seismicity induced by either reservoir impoundment (Simpson et al., 1988; Roeloffs, 1996) or fluid injection (Raleigh et al., 1972; Pine et al., 1983; Zoback and Harjes, 1997), (2) earthquakes triggered by other earthquakes (Stein et al., 1992, 1997), and (3) in situ stress measurements in deep wells and boreholes (Zoback and Healy, 1992; Brudy et al., 1997). The in situ stress measurements further show that Coulomb frictional-failure theory incorporating laboratory-derived frictional coefficients, μ, of 0.6–1.0 (Byerlee, 1978) gives predictions that are consistent with measured stress states in the upper crust. For instance, at virtually all locations where deep stress levels have been measured, the ratio of the maximum differential stress, ΔS, to the effective mean stress, $\overline{S} - P_f$ (where $\overline{S}$ is the mean stress and P_f is the pore pressure), agrees well with that predicted using Coulomb frictional-failure theory, namely

$$\Delta S \Big/ \left(\overline{S} - P_f\right) = 2\mu \Big/ \sqrt{\mu^2 + 1}. \qquad (1)$$

This is illustrated in Figure 1; it can be clearly seen that at each of the six locations illustrated, the effective stress data are consistent with values of μ between ~0.6 and 1.0. These data support the hypothesis that the crust contains critically stressed faults that limit its strength.

However, because the frictional strength of a faulted rock mass depends on pore pressure (Hubbert and Rubey, 1959), estimates of the frictional strength of the brittle crust depend on the pore pressure at depth (Sibson, 1973; Brace and Kohlstedt, 1980). In particular, utilization of Coulomb faulting theory with laboratory-derived coefficients of friction leads to the conclusion that the crust's brittle strength is quite high (hundreds of megapascals) under conditions of hydrostatic pore pressure.

In this paper we demonstrate that critically stressed faults maintain high crustal permeability, resulting in near-hydrostatic pore pressures, and high crustal strength.

CRUSTAL PERMEABILITY AND ITS SCALE DEPENDENCE

The high permeability of upper crustal crystalline rocks was first noted by Brace (1980), who observed that, even given the relatively limited number of permeability measurements available at the time, the crust was unlikely to be able to sustain pore pressures much greater than hydrostatic. Recently acquired in situ permeability data support this observation, and furthermore suggest a gross scale dependence in which permeability increases with increasing scale (Clauser, 1991). This relationship is particularly well illustrated by hydraulic tests made in the German Continental Deep Drilling Program (KTB; Kontinentales Tiefbohrprogramm der Bundesrepublik Deutschland) pilot and main holes at depths as great as 9.1 km (Huenges et al., 1997; Fig. 2). During these tests, an interval of the borehole was mechanically isolated, and fluid was pumped into it. The test intervals used for these experiments varied between a few tens of meters and almost 3.5 km, providing estimates of the gross vertical permeability of kilometer-scale sections of the upper crust. Data from the most reliable experiments—7 low-volume buildup drill stem tests and 12 open-hole buildup tests—indicated permeabilities of between 10^{-20} and $>10^{-16}$ m^2: the majority of the most reliable measurements consistently gave permeabilities of $>10^{-17}$ m^2. In comparison, laboratory measurements made under estimated in situ pressure and temperature conditions on centimeter-scale core samples obtained in the 0–7.5 km depth range indicated permeability of between 10^{-20} and 10^{-18} m^2 (Huenges et al., 1997). A three to four order of magnitude discrepancy existed therefore

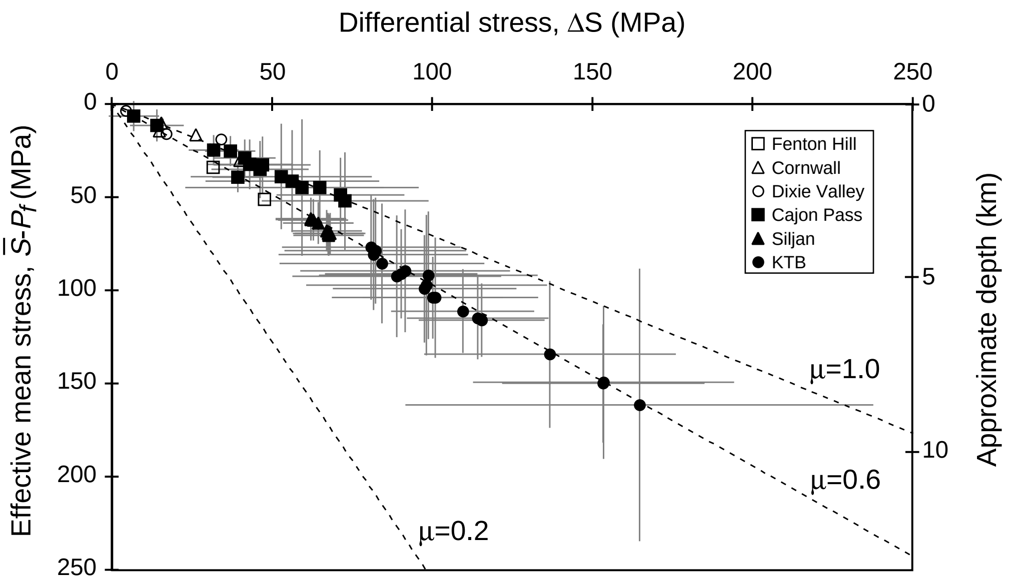

Figure 1. Dependence of differential stress, ΔS, on effective mean stress, $\overline{S} - P_f$, at six locations where deep stress measurements have been made. Dashed lines illustrate relationships predicted using Coulomb frictional-failure theory for various coefficients of friction, μ. References: Fenton Hill—Barton et al. (1988); Cornwall—Pine et al. (1983), Batchelor and Pine (1986); Dixie Valley—Hickman et al. (1997); Cajon Pass—Zoback and Healy (1992); Siljan—Lund and Zoback (1999); KTB (German Continental Deep Drilling Program)—Brudy et al. (1997).

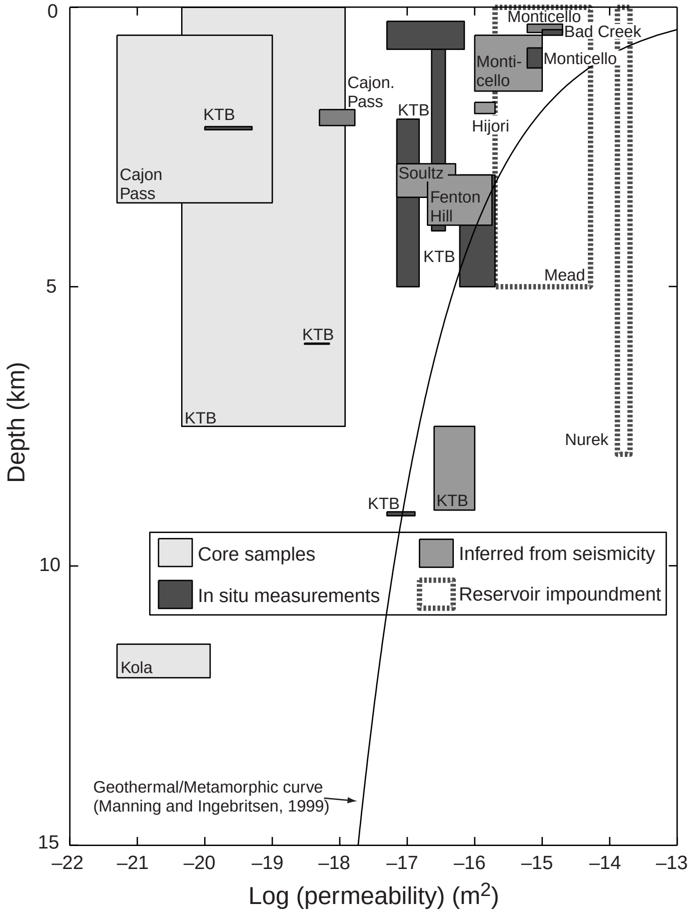

Figure 2. Deep crustal permeability data acquired from core samples, in situ hydraulic tests, and induced seismicity. References: core samples—Huenges et al. (1997), Lockner et al. (1991), Morrow and Byerlee (1988, 1992), Morrow et al. (1994); in situ measurements—Coyle and Zoback (1988), Huenges et al. (1997); seismicity data—Sasaki (1998), Shapiro et al. (1997, 1999), Zoback and Hickman (1982); reservoir impoundment data—Roeloffs (1988), Talwani et al. (1999); geothermal/metamorphic curve—Manning and Ingebritsen (1999). KTB—German Continental Deep Drilling Program.

between the large-scale and small-scale permeabilities of rocks tested under approximately the same effective confining pressures (Fig. 2).

At the same location, Shapiro et al. (1997) concluded that progressive hypocentral migration over distances of >1 km during an induced seismicity experiment performed at the bottom of the KTB main hole (Zoback and Harjes, 1997) indicated bulk permeability of ~10^{-17} to 10^{-16} m^2. Analogous experiments at the Fenton Hill (Nevada), Soultz (Alsace, France), and Hijori (Yamagata, Japan) hot dry rock sites gave similar permeabilities of 10^{-17} to 10^{-16} m^2 at depths of 3.0–3.9 km, 2.8–3.4 km, and 1.7–1.9 km, respectively (Shapiro et al., 1999; Sasaki, 1998). Slightly higher permeabilities (10^{-16} to 10^{-15} m^2) were estimated from hydraulic tests and induced seismicity diffusion at the Monticello Reservoir, South Carolina, by Zoback and Hickman (1982) at very shallow depths (<1 km).

A similar result was obtained in experiments made at the Cajon Pass borehole in southern California. Morrow and Byerlee (1988, 1992) obtained permeabilities of 10^{-22} to 10^{-19} m^2 from core samples retrieved from 0.5 to 2.1 km depth (which also exhibited a systematic one order of magnitude decrease per kilometer), whereas Coyle and Zoback (1988) measured a permeability of ~10^{-18} m^2 over 100 m and 300 m hydraulic test intervals.

Core measurements at in situ confining pressures (for hydrostatic fluid pressures) on samples from the 12-km-deep Kola Peninsula superdeep well consistently show extremely low permeability values of <10^{-20} m^2 (Lockner et al., 1991; Morrow et al., 1994). Unfortunately, hydraulic testing was not performed on the Kola borehole, so no large-scale permeability data were obtained directly. However, thermal models of borehole temperature data constrain the kilometer-scale permeability of the 0–2 km and 6–8 km intervals to 10^{-14} m^2 and 10^{-17} m^2, respectively (Kukkonen and Clauser, 1994).

Reservoir impoundment provides another method of inducing seismicity and has been used by several authors to estimate kilometer-scale permeability. Roeloffs (1988) used seismicity occurring beneath the Mead (Arizona-Nevada) and Nurek (Tadjikistan) Reservoirs following peak impoundment to estimate hydraulic diffusivity. When transformed into equivalent permeabilities, Roeloff's data suggest permeabilities of 10^{-16} to 10^{-15} m^2 between 0 and 5 km beneath the Mead Reservoir and ~10^{-14} m^2 between 0 and 8 km beneath the Nurek Reservoir. Both these results are an order of magnitude higher than those obtained from direct fluid-injection results. The hydraulic response of a 250-m-long shear zone to fluctuations in reservoir level at the Bad Creek Reservoir (South Carolina) was used by Talwani et al. (1999) to calculate permeability, giving a result of 10^{-15} m^2.

Manning and Ingebritsen (1999) compiled and interpreted geothermal and metamorphic data to provide crustal permeability estimates at depths greater than observable in boreholes (>10 km). The permeability threshold above which fluid advection transports heat more effectively than conduction appears to be ~10^{-16} m^2, whereas the corresponding threshold for advective solute transport is only 10^{-20} m^2. Similarly, time-integrated fluid fluxes during metamorphism—manifested geochemically, petrologically, and isotopically in both metamorphic protoliths and the associated fluids—indicate permeabilities of 10^{-19} to 10^{-18} m^2 during regional metamorphism. Manning and Ingebritsen (1999) demonstrated that crustal permeability obeys a power-law decrease with depth according to the relationship (Fig. 2)

$$\log k = -3.2 \log z - 14. \tag{2}$$

Figure 2 clearly illustrates that with the exception of laboratory measurements on cores, different methods of estimating in situ permeability give relatively consistent results. Core measurements, even when made under in situ pressure and temperature conditions, give very low permeabilities that vary substantially owing to local heterogeneities. Consequently, although the core measurements determine the intrinsic permeability of the rock mass, they are not indicative of the effective permeability controlling large-scale upper crust hydraulics. In contrast, borehole measurements and experimentally and reservoir-induced seismicity at several locations give almost uniformly high permeabilities of >10^{-17} m^2. Furthermore, these permeabilities agree extremely well with independent estimates based on geochemical and geothermal considerations. We conclude that the permeability of the upper crust is ~10^{-17} to 10^{-16} m^2 over 1 to 10 km scales.

HYDROSTATIC PORE PRESSURE AND CRITICALLY STRESSED FAULTS

Fluid pressures at depths of several kilometers have been measured using several independent techniques in deep boreholes drilled into crystalline basement and have been consistently found to be approximately hydrostatic. Table 1 lists the deepest of these boreholes, in each of which fluid pressures are unequivocally near hydrostatic. Stress magnitudes at each of these locations (except the Kola borehole, where stress measurements were not performed) are consistent with Coulomb frictional-failure theory for coefficients of friction of 0.6–1.0 such as are measured experimentally in laboratory settings (Byerlee, 1978; Brace and Kohlstedt, 1980; Fig. 1), and seismicity was induced by fluid injection at a number of these sites.

By using borehole televiewer images and high-resolution temperature logs from the Cajon Pass, Long Valley (California), and Yucca Mountain USW-G1 (Nevada) boreholes, Barton et al. (1995) showed that critically stressed faults—i.e., faults with ratios of resolved shear to normal tractions of 0.6–1.0—are hydraulically conductive, whereas those that are not critically stressed are not hydraulically conductive. Figure 3 presents these data in a somewhat different form from that shown by Barton et al. (1995), and it is clear that the hydraulically conductive fractures are critically stressed

TABLE 1. LOCATIONS EXHIBITING NEAR-HYDROSTATIC FLUID PRESSURE AT DEPTHS OF SEVERAL KILOMETERS

Well location	Regime	Depth (km)	Observation	Source	Evidence for critical stress
Cornwall HDR, England	SS	2.5	DST	Pine et al. (1983)	Stress magnitudes; induced seismicity
Fenton Hill HDR, New Mexico	N, SS	3.0	SWC	Barton et al. (1988)	Stress magnitudes
Dixie Valley, Nevada	N	2–3, 5–7	DST, SG	Hickman et al. (1997)	Stress magnitudes; prehistoric fault offsets
Cajon Pass, California	SS	3.5	DST	Coyle and Zoback (1988)	Stress magnitudes; breakout rotations
Soultz HDR, France	N, SS	5.0	DST	Baumgärtner et al. (1998)	Stress magnitudes; induced seismicity
Siljan, Sweden	SS	7.0	DST	Lund and Zoback (1999)	Stress magnitudes
KTB, Germany	SS	9.1	DST, SWC	Huenges et al. (1997) Zoback and Harjes (1997)	Stress magnitudes; induced seismicity
Kola, Russia	?R	12.2	SWC	Borevsky et al. (1987)	N.A.

Note: HDR—hot dry rock; KTB—Kontinentales Tiefbohrprogramm der Bundesrepublik Deutschland (German Continental Deep Drilling Program); SS—strike-slip faulting regime; N—normal faulting regime; R—reverse faulting regime; DST—drill stem test; SWC—static water column; SG—silica geothermometry; N.A.—not available.

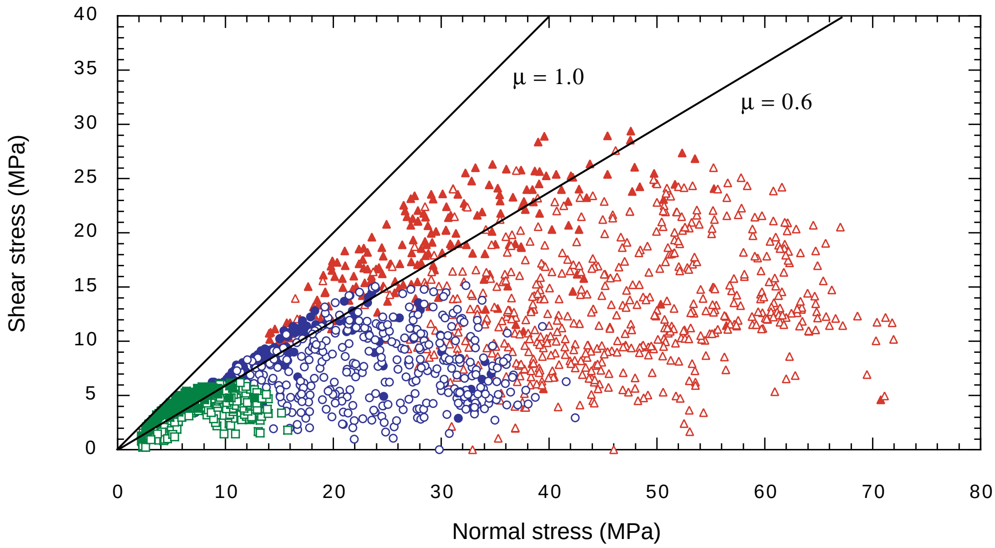

Figure 3. Shear and normal stresses on fractures identified with borehole imaging techniques in Cajon Pass (triangles), Long Valley (circles), and Nevada Test Site (squares) boreholes. Filled symbols represent hydraulically conductive fractures and faults, and open symbols represent nonconductive fractures. Original data are from Barton et al. (1995).

according to the Coulomb frictional-failure criterion. Hickman et al. (1997) and Barton et al. (1998) subsequently obtained similar results in the Dixie Valley geothermal field adjacent to the Stillwater fault, a range-bounding normal fault in the Basin and Range province, Nevada, on which M7.3 and M6.8 earthquakes occurred in 1915 and 1954, respectively. In this case too, the critically stressed fractures, including the Stillwater fault, were found to be hydraulically conductive, whereas the noncritically stressed faults and fractures were not. Ito and Zoback (2000) have reported similar results utilizing data from the KTB main borehole.

It seems clear from all of these in situ studies that, in general, the crust is in frictional failure equilibrium (even in relatively stable intraplate areas), near-hydrostatic pore pressures exist to great depth in crystalline intraplate crust, and the faults that are critically stressed maintain the crust's high permeability.

FAULTS, FLUIDS, AND FLOW

Given that the upper crust's permeability, k, is ~10^{-17} to 10^{-16} m^2, we may ask over what lengths of time appreciable hydraulic diffusion occurs. The characteristic time, τ, for a diffusive process is given by

$$\tau = l^2/\kappa = \left(\phi\beta_f + \beta_r\right)\eta\, l^2/k, \tag{3}$$

where l is a characteristic length scale of the process, $\kappa \sim k/(\phi\beta_f + \beta_r)$ is the hydraulic diffusivity, β_f and β_r are the fluid and rock compressibilities, respectively, ϕ is the rock porosity, and η is the fluid viscosity. For low-porosity rocks ($\phi < 0.02$) at 150 °C, where $\beta_f = 5 \times 10^{-10}$ Pa^{-1}, $\beta_r = 2 \times 10^{-11}$ Pa^{-1}, and $\eta = 1.9 \times 10^{-4}$ Pa·s, the previous equation gives

$$\log\tau = 2\log l - \log k - 16, \tag{4}$$

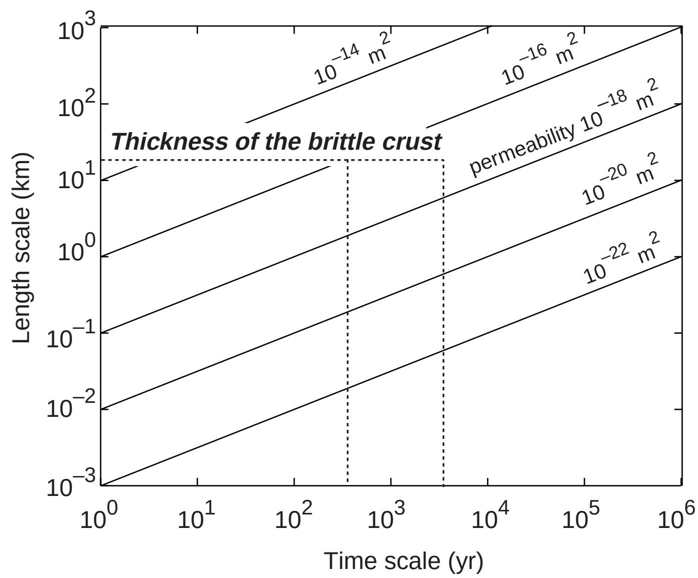

Figure 4. Length and time scales for diffusive fluid flow in rock masses of different permeabilities.

where τ and l are in years and kilometers, respectively. This relationship is illustrated for various values of permeability in Figure 4. For crustal permeabilities of 10^{-17} to 10^{-16} m^2, the characteristic times for fluid transport over length scales of 1–10 km are only 10–1000 yr. Thus fluid pressures in the crust are expected to equilibrate over relatively short time scales, which enables hydrostatic fluid pressure regimes to be maintained to depths of 10 km or more. We envisage brecciation associated with slip on critically stressed faults as countering fault-sealing mechanisms by incremental failure and thereby maintaining high permeability.

Continual faulting at a small scale appears necessary to maintain high permeability and low fluid pressures. Hence it appears that "stable crust" is only a relative term: with respect to stress and faulting we argue that stable intraplate crust is subject to continual small-scale failure. With respect to deformation, however, it is clear that long-term intraplate strain rates must be extremely low.

CONCLUSIONS

The bulk permeability of the upper crust in intraplate regions is ~10^{-17} to 10^{-16} m^2 over length scales of 1–10 km. Hence, the brittle crust is effectively permeable over time scales of 10–1000 yr and pore pressures can be maintained at hydrostatic values. We argue that this high permeability results from hydraulically conductive, critically stressed faults, presumably because brecciation associated with slip on active faults offsets permeability reductions associated with fault-zone sealing. Thus, intraplate crust is able to sustain higher differential stresses than would be possible if bulk permeability were sufficiently low to sustain fluid pressures higher than hydrostatic.

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