Strength of chrysotile-serpentinite gouge under hydrothermal conditions: Can it explain a weak San Andreas fault?

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ABSTRACT

Chrysotile-bearing serpentinite is a constituent of the San Andreas fault zone in central and northern California. At room temperature, chrysotile gouge has a very low coefficient of friction ($\mu \approx 0.2$), raising the possibility that under hydrothermal conditions μ might be reduced sufficiently (to $\leq$0.1) to explain the apparent weakness of the fault. To test this hypothesis, we measured the frictional strength of a pure chrysotile gouge at temperatures to 290 °C and axial-shortening velocities as low as 0.001 μm/s. As temperature increases to $\approx$100 °C, the strength of the chrysotile gouge decreases slightly at low velocities, but at temperatures $\geq$200 °C, it is substantially stronger and essentially independent of velocity at the lowest velocities tested. We estimate that pure chrysotile gouge at hydrostatic fluid pressure and appropriate temperatures would have shear strength averaged over a depth of 14 km of 50 MPa. Thus, on the sole basis of its strength, chrysotile cannot be the cause of a weak San Andreas fault. However, chrysotile may also contribute to low fault strength by forming mineral seals that promote the development of high fluid pressures.

INTRODUCTION

The serpentine mineral chrysotile is commonly associated with faults and shear zones (e.g., O'Hanley, 1991). At several localities in California, for example, the chrysotile content of sheared serpentinite is higher than that of nearby unsheared serpentinite (Page, 1968; Coleman and Keith, 1971; Mumpton and Thompson, 1975). Knowledge of the frictional properties and mode of occurrence of chrysotile may therefore be important to understanding the behavior of serpentinite-bearing sections of the San Andreas and related faults.

The absence of a significant conductive heat-flow anomaly directly over the San Andreas fault suggests, on the basis of models of conductive heat transport, that the average frictional resistance in the seismogenic zone during fault motion is $\leq$20 MPa (Brune et al., 1969; Lachenbruch and Sass, 1980). In addition, Mount and Suppe (1987) and Zoback et al. (1987) determined that the maximum principal stress generally is almost perpendicular to the San Andreas fault. Fault-normal compression imposes additional limitations on average fault strength, to the extent that if a hydrostatic fluid-pressure gradient is assumed, the coefficient of friction, μ (= shear stress/[normal stress − pore pressure]), of fault-zone materials may need to be $\leq$0.1 to account for this weakness (Lachenbruch and Sass, 1992). Previous strength investigations on serpentinite have concentrated on the antigorite- and lizardite-rich types, both of which are relatively strong (Raleigh and Paterson, 1965; Dengo and Logan, 1981; Reinen et al., 1991). The strength of chrysotile has only recently been considered, but preliminary reports indicate that it may be less than half as strong as the other two varieties, averaging $\mu \approx 0.20–0.25$ at room temperature (Reinen et al., 1993; note: Reinen et al., 1993, 1994, reported a low strength for a serpentinite gouge that they originally identified as lizardite but that Reinen and Tullis, 1995, subsequently determined to be chrysotile). Furthermore, it has been suggested that at higher temperatures (Reinen et al., 1993) and/or lower strain rates (Reinen and Tullis, 1995) the frictional strength of chrysotile may be reduced to $\mu \leq 0.1$. As a test of this hypothesis, we report here on the strength of a pure chrysotile gouge at elevated temperatures and a wide range of velocities, intended to more closely represent conditions at depth in the fault. Our results suggest that chrysotile alone is not a likely explanation for a weak San Andreas fault.

EXPERIMENTAL PROCEDURES

The chrysotile gouge used in this study was prepared from a soft, layered, light to medium grayish-green rock from the New Idria district, California. The rock is essentially all serpentine, and X-ray diffraction analysis combined with petrographic observations suggests the presence of clinochrysotile alone. The chrysotile has a uniform composition of approximately $Mg_{5.62}Fe^{2+}_{0.30}Al_{0.05}Si_4O_{10}(OH)_8$, calculated from electron microprobe analyses. The rock was hand ground and passed through an 88-μm-diameter sieve to produce a simulated gouge.

The strength tests were run in a triaxial machine fitted with an internal furnace. For each experiment, a 1-mm-thick layer of gouge was placed along a 30° sawcut in a 19.1-mm-diameter cylinder of antigoritic serpentinite. Serpentinite cylinders were used to avoid the possibility of reaction with the gouge during tests at elevated temperatures. A borehole for pore-fluid entry was drilled into the upper half of each cylinder. The sample was contained in a copper jacket during the experiment, to separate it from the confining-pressure medium; reported strengths have been corrected for the strength of the jacket. Selected runs, using polyurethane and viton jackets, were conducted to verify the corrections for copper-jacket strength. Confining pressure was applied first to the sample, followed by the pore pressure. After the pressures had equilibrated, the temperature was raised to the desired value. Temperature was monitored by a thermocouple inserted along the pore-pressure inlet. The experiments were all run at a constant normal stress, which was maintained by means of computer-controlled adjustments to the confining pressure. As a test of pore-fluid communication, at the end of two 200 °C experiments, pore pressure.and normal stress were increased together by 5 MPa, at a constant velocity of 0.2 μm/s. In both cases, essentially no change in shear strength was measured, as expected for a sample in good hydraulic communication with the external pore-pressure system. Had the fault been hydraulically isolated from the pore-pressure system, the increase in normal stress would have resulted in a 5% increase in shear strength.

We conducted two groups of experiments for this study. (1) A series was run at 10 MPa pore pressure of deionized water and a constant normal stress of 110 MPa, to yield a constant effective normal stress of 100 MPa. The experiments were run at temperatures of 25, 100, and 200 °C, with axial shortening rates stepped in the range 0.001 to 10 μm/s. The slowest rate is equivalent to a sliding velocity along the inclined gouge layer of 36.4 mm/yr, which is close to the average slip rate of the San Andreas fault in central California. (2) A series was run at the temperature-pressure conditions corresponding to depths of 3, 6, and 9 km in the fault, assuming a hydrostatic pore-pressure gradient and a geothermal gradient of 30°/km (Lachenbruch and Sass, 1973). The 9 km simulation, conducted at 290 °C, is close to the upper temperature limit of stability of chrysotile (Evans et al., 1976). At greater depths, chrysotile should be replaced by antigorite. Combined petrographic, X-ray diffraction, and scanning electron microscope analyses of the samples yielded no evidence of chrysotile breakdown, to ≈1 μm resolution, during any of the experiments.

RESULTS

The low room-temperature strength of the chrysotile gouge ($\mu \approx 0.20$) measured at 100 MPa effective normal stress (Fig. 1) is consistent with the values reported by Reinen et al. (1993). At the faster rates of 0.1 to 10 μm/s (Fig. 1A), the two heated samples are stronger overall than the one run at room temperature, although the differences are slight. At lower velocities (Fig. 1B), however, the 200 °C sample is significantly stronger, with μ approaching 0.35 when the jacket failed. Much of this increase is attributable to a time-dependent hardening of the gouge during the interval at 0.001 μm/s; 0.5 mm axial displacement at that rate required nearly 5.8 days to complete.

Our room-temperature results (Fig. 1A) are consistent with the findings of Reinen et al. (1993) that chrysotile exhibits velocity-weakening behavior (a decrease in strength with a step increase in velocity) at fast velocities and velocity-strengthening behavior (an increase in strength with velocity increase) at slow velocities. At elevated temperatures, however, the velocity-strengthening behavior of the chrysotile gouge reaches a maximum at intermediate to high velocities, in the range tested, and diminishes at both slower and faster rates. The low-velocity behavior is illustrated in Figure 1B, in which decreasing the velocity by steps from 1.0 to 0.001 μm/s leads to successively smaller increments of strength decrease. At the slowest rates, chrysotile strength is almost independent of velocity change. Quantitative determinations of the velocity dependence of chrysotile were presented by Lockner et al. (1996).

The conditions of the 200 °C, 6 km experiment in Figure 2 are similar to those of the 200 °C experiments in Figure 1, and the coefficients of friction are also similar. However, the 110 and 290 °C samples yield the lowest and highest coefficients of friction, respectively, obtained for chrysotile in this study. At 290 °C, the coefficient of friction exceeds 0.5 and is comparable to estimates of μ at 290 °C for lizardite- and antigorite-rich serpentinite gouges at 290 °C, extrapolated from our data at 200 °C (Moore et al., 1995). The 200 and 290 °C samples show pronounced increases in μ during the 0.001 μm/s step, which occurs early in these experiments, and μ remains high for the rest of the experiments. Changing velocity has little effect on chrysotile strength at 200 and 290 °C. At 110 °C, μ is as low as 0.1 at the slowest velocity steps, compared to a minimum of $\mu = 0.15$ for comparable velocities in the 100 °C experiment (Fig. 1B). The 110 °C experiment was conducted at about one-half the effective normal stress of the 100 °C experiments, which suggests that μ for chrysotile may increase with increasing effective stress. In contrast, the coefficient of friction of both lizardite and antigorite serpentinite decreases with increasing effective pressure (e.g., Raleigh and Paterson, 1965). We have planned a series of experiments to better separate the controls of temperature, effective stress, and fluid pressure on chrysotile strength.

DISCUSSION

Our results corroborate the findings of Reinen et al. (1993) that at low temperatures chrysotile is only about half as strong as the more abundant serpentine minerals lizardite and antigorite. We find, however, that chrysotile is essentially as strong as other serpentine varieties near its high-temperature stability limit. The strength of the heated gouge is nearly independent of velocity at the lowest tested rates, suggesting that further velocity reductions should have little effect on chrysotile strength.

Figure 3 compares the shear strength of chrysotile gouge subjected to a hydrostatic fluid-pressure gradient (Fig. 2) to the constraints on fault strength imposed by heat-flow measurements (Brune et al., 1969; Lachenbruch and Sass, 1980). At depths shallower than about 4 km, the strength of pure chrysotile gouge falls within the strength envelope. However, between 5 and 9 km depth, the shear strength of chrysotile is well above the limits imposed from modeling the heat-flow data. At still greater depths, antigorite is the stable serpentine mineral (Evans et al., 1976); its strength is approximated in Figure 3 by extrapolating our experimental data to 200 °C on antigorite (Moore et al., 1995) to higher temperatures.

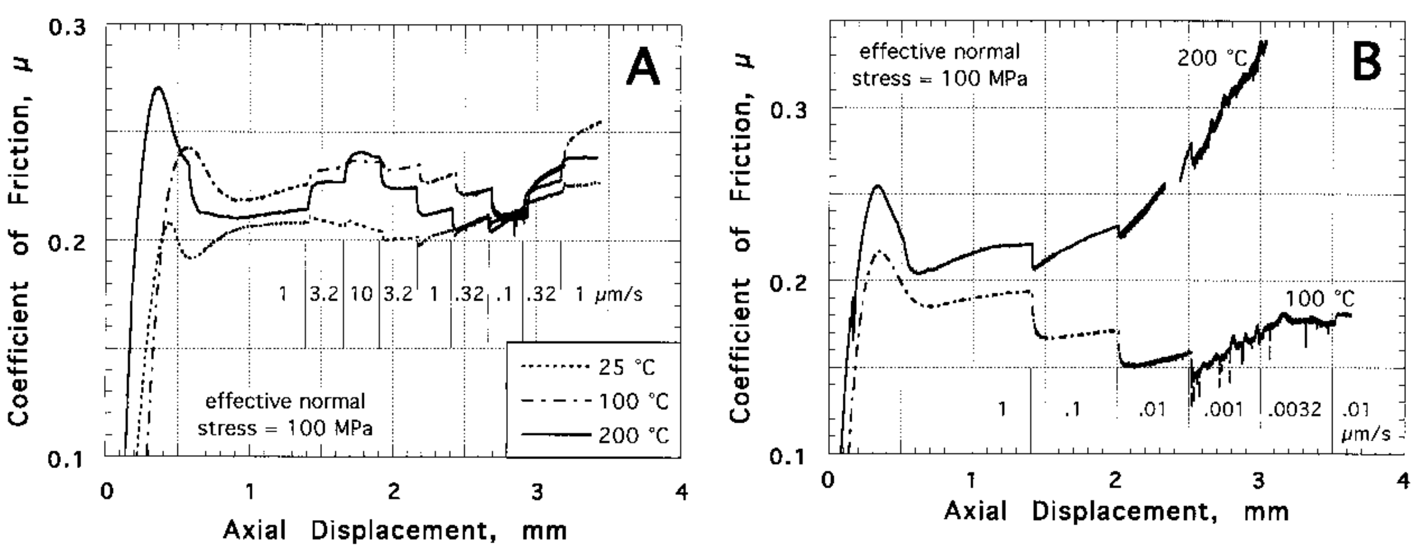

Figure 1. Variations in coefficient of friction of chrysotile gouge with increasing temperature, at axial-shortening velocities in range (A) 0.1–10.0 μm/s and (B) 0.001–1.0 μm/s. Other conditions: 10 MPa pore pressure of deionized water; 100 MPa effective normal stress. In B and in Figure 2, irregularities in data during 0.001 and 0.0032 μm/s steps are caused by daily variations in room temperature, which affect fluid volumes in pore-pressure lines. Previous studies conducted on chrysotile-bearing samples under range of normal stresses indicate that cohesion term in general friction relationship is essentially zero to 200 °C (Reinen et al., 1994; Moore et al., 1983).

Some corroboration for continuously increasing strength to the base of the seismogenic zone was provided by Moore et al. (1986), who found that a serpentinite gouge consisting of roughly equal amounts of chrysotile and lizardite was considerably stronger at 400 °C than at 200 °C. If correct, the estimated high-temperature strength will further raise the average shear strength of the gouge in the seismogenic zone to values well above the limits imposed by the heat-flow data. If we integrate the curve for pure chrysotile shown in Figure 3, we find that shear strength averaged over the upper 14 km is $\approx$50 MPa. This clearly exceeds the heat-flow constraints of 10–20 MPa. It should also be noted that the chrysotile curve in Figure 3 is a lower limit for the strength of serpentinite-bearing parts of the San Andreas fault, because natural serpentinite gouges in the upper seismogenic zone will typically consist of stronger mixtures of lizardite + chrysotile (e.g., Morrow et al., 1982; Moore et al., 1986).

Thus, although chrysotile is one of the weakest rock-forming minerals under certain conditions, it is too strong to explain a weak San Andreas fault (Fig. 3). Similar arguments have been advanced for montmorillonite, the room-temperature strength of which is close to that of chrysotile (Summers and Byerlee, 1977; Morrow et al., 1982, 1992). Morrow et al. (1992) conducted room-temperature tests on montmorillonite at confining and fluid pressures consistent with depths of burial to 15 km, assuming a hydrostatic fluid-pressure gradient. The strength of the montmorillonite gouge increased with increasing effective stress, and extrapolating the data to lower slip rates did not substantially decrease strength. They concluded that even a pure montmorillonite gouge would not produce the postulated low shear stresses across the San Andreas fault. Vermiculite is the only other low-strength mineral of geologic significance that has been identified (Summers and Byerlee, 1977). However, vermiculite has not been found in abundance in near-surface exposures of San Andreas fault gouge, and its probable restriction to supergene environments (de la Calle and Suquet, 1988) indicates that vermiculite will not be more abundant at greater depths in the fault. The available data therefore strongly suggest that the weakness of the San Andreas fault cannot be a consequence of the inherently low strength of any of the materials within it.

The principal alternative group of models for low fault strength involves the generation of nearly lithostatic fluid pressures within the fault zone, which correspondingly lower the effective stress (e.g., Byerlee, 1990; Rice, 1992). Lachenbruch and Sass (1992) calculated that for the majority of fault-zone materials the ratio of fluid pressure to overburden pressure, λ, must exceed twice the hydrostatic level ($\lambda > 0.74$) to reduce the average fault strength to 20 MPa. By comparison, using values of μ from Figure 2, λ must be $\geq$0.65 at 6 km and $\geq$0.78 at 9 km (Fig. 3) to reduce the shear strength of a chrysotile-filled fault to 20 MPa. The recent focus of the models invoking high fluid pressures has been on mineral-sealing processes (Chester et al., 1993; Sleep and Blanpied, 1994; Lockner and Byerlee, 1995), which lead to the formation of permeability barriers that trap fluids within a fault zone. Chrysotile may help to generate such high fluid pressures through its involvement in reactions that promote fault-zone sealing. Chrysotile most commonly forms as a replacement of preexisting serpentine minerals (e.g., O'Hanley et al., 1989; O'Hanley, 1991). Mumpton and Thompson (1975) proposed that the extensive chrysotile asbestos deposits at New Idria, California, formed through solution-precipitation reactions involving antigorite and lizardite. Their model may be generally applicable to chrysotile occurrences worldwide (e.g., Craw et al., 1987; O'Hanley, 1991). Page (1968) suggested that brucite—$Mg(OH)_2$—a common accessory mineral in serpentinite, may also be converted to chrysotile through the passage of large volumes of silica-bearing ground waters.

The close association of chrysotile with faults and shear zones is thus a probable consequence of the channeling of fluids through faults. Correspondingly, the abundance of chrysotile in serpentinite

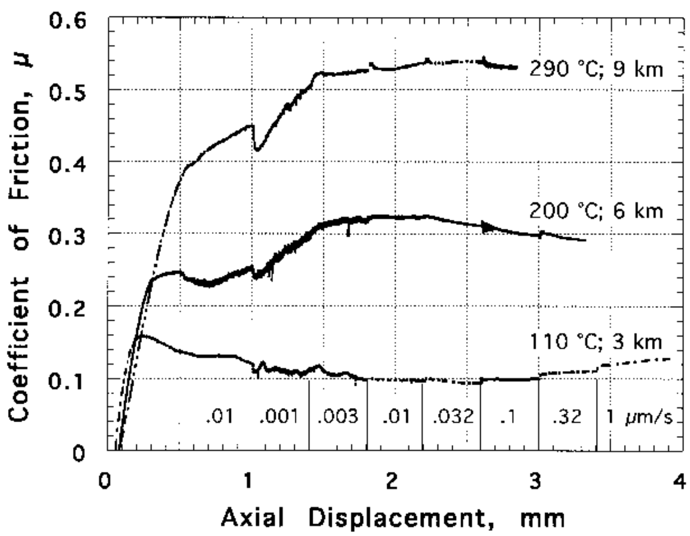

Figure 2. Strength of chrysotile-serpentinite gouge at conditions representative of 3, 6, and 9 km depth in fault zone: 3 km—110 °C, 30 MPa pore pressure, 76.5 MPa normal stress (46.5 MPa effective normal stress); 6 km—200 °C, 60 MPa pore pressure, 153 MPa normal stress (93 MPa effective normal stress); 9 km—290 °C, 90 MPa pore pressure, 229.5 MPa normal stress (139.5 MPa effective normal stress). Assumptions: hydrostatic fluid-pressure gradient of 10.0 MPa/km, serpentinite bulk density of 2.5–2.6 g/cm^3, lithostatic normal stress gradient of 25.5 MPa/km, surface temperature of 20 °C, and geothermal gradient of 30°/km. Correlation of normal stress with overburden pressure will be discussed by Moore et al. in the future.

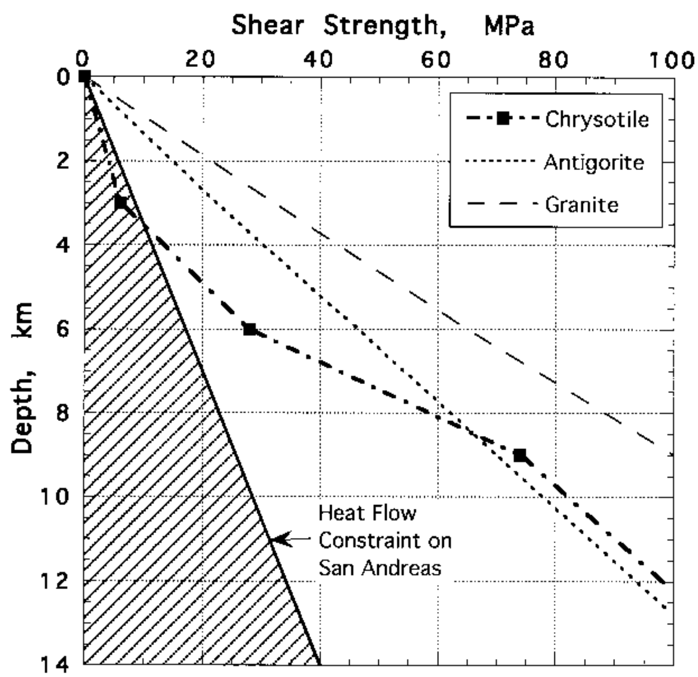

Figure 3. Shear strength of chrysotile relative to heat flow constraint on strength of San Andreas fault, in presence of hydrostatic fluid-pressure gradient. Cross-hatched field represents range of shear strengths that are consistent with heat-flow data of Brune et al. (1969) and Lachenbruch and Sass (1980). Upper limit of 20 MPa average shear strength is plotted assuming 14-km-deep seismogenic zone (Lachenbruch and Sass, 1980) and that μ remains constant at depth. Strength profiles for granite (Blanpied et al., 1995) and antigorite (Moore et al., 1995) gouges are plotted for comparison.

gouge may be roughly correlated with the volume of water that has moved through a fault zone. The newly crystallized chrysotile is likely to fill fractures and pore spaces instead of directly replacing a dissolving lizardite or antigorite grain (Mumpton and Thompson, 1975). In this way, the chrysotile-forming reaction may provide an important fault-sealing mechanism. Future laboratory research might be directed toward reproducing the chrysotile-forming reaction(s) under hydrothermal conditions, to determine its effect on the permeability and strength of fractured serpentinite and serpentinite fault gouge.

CONCLUSIONS

Possible explanations for the extreme weakness of the San Andreas fault include the presence of very weak materials within the fault zone and very high fluid pressures that lower the effective stress. Chrysotile-bearing serpentinite is considered to be a common fault-zone constituent in central and northern California. Because of its low room-temperature strength, chrysotile has been suggested as a possible cause of a weak San Andreas fault. However, our strength measurements of chrysotile under hydrothermal conditions indicate that chrysotile will become stronger at depth in the fault, with strength increasing markedly at temperatures ≥200 °C. Consequently, in the presence of a hydrostatic fluid-pressure gradient, chrysotile cannot generate the low shear stresses estimated for the San Andreas fault based on conductive heat-flow data. The available strength and mineralogical data suggest that no other gouge material of potential importance to the San Andreas fault will be any weaker than chrysotile. Fluid pressures approaching lithostatic levels must therefore be invoked to reduce the strength of all types of fault gouge to the required low levels. Chrysotile may help to generate high fluid pressures in serpentine-bearing parts of the fault by creating permeability barriers through its crystallization in fractures and pore spaces.

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