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Fluid overpressures and strength of the sedimentary upper crust

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ABSTRACT

The classic crustal strength-depth profile based on rock mechanics predicts a brittle strength $\sigma_1 - \sigma_3 = \kappa(\overline{\rho}gz - P_f)$ that increases linearly with depth as a consequence of [1] the intrinsic brittle pressure dependence κ plus [2] an assumption of hydrostatic pore-fluid pressure, $P_f = \rho_w gz$. Many deep borehole stress data agree with a critical state of failure of this form. In contrast, fluid pressures greater than hydrostatic $\overline{\rho}gz > P_f > \rho_w gz$ are normally observed in clastic continental margins and shale-rich mountain belts. Therefore we explore the predicted shapes of strength-depth profiles using data from overpressured regions, especially those dominated by the widespread disequilibrium-compaction mechanism, in which fluid pressures are hydrostatic above the fluid-retention depth z_{FRD} and overpressured below, increasing parallel to the lithostatic gradient $\overline{\rho}gz$. Both brittle crustal strength and frictional fault strength below the z_{FRD} must be constant with depth because effective stress $(\overline{\rho}gz - P_f)$ is constant, in contrast with the classic linearly increasing profile. Borehole stress and fluid-pressure measurements in several overpressured deforming continental margins agree with this constant-strength prediction, with the same pressure-dependence κ as the overlying hydrostatic strata. The role of z_{FRD} in critical-taper wedge mechanics and jointing is illustrated. The constant-strength approximation is more appropriate for overpressured crust than classic linearly increasing models.

1. Introduction

The classic Brace and Kohlstedt (1980) strength-depth graph, based on experimental rock mechanics, predicts an initial linear increase in brittle frictional strength with increasing pressure, followed by an exponential decay in ductile strength with increasing temperature (Fig. 1A). This first-order prediction has been remarkably successful, for example in predicting the temperature of the brittle-ductile transition in common rock types. The intricacies of this graph in the lower crust and upper mantle continue to be widely discussed and are even controversial in the case of the proposed “jelly sandwich” of weak lower continental crust above a presumed stronger mantle (*e.g.* Chen and Molnar, 1983; Suppe, 1985, p. 188–189; Jackson, 2002; Burov and Watts, 2006; Bürgmann and Dresen, 2008). In contrast, the predicted linear increase in brittle strength is generally assumed without discussion and agrees with deep borehole stress data in crystalline rock, showing that the crust is commonly close to a critical state of brittle failure (*e.g.* Townend and Zoback, 2000; Zoback, 2007), for example the deep KTB borehole in Germany (Fig. 1B).

1.1. Classic brittle strength-depth relationship

The predicted linear increase in brittle strength $\sigma_1 - \sigma_3$ with depth is a consequence of [1] the intrinsic pressure dependence κ of brittle strength, plus [2] an assumption that pore-fluid pressure is hydrostatic and therefore linearly increasing $P_f = \rho_w gz$. Ignoring cohesion we approximate brittle crustal strength very simply as the vertical effective stress $\overline{\rho}gz - P_f$, times the pressure dependence κ.

$$\sigma_1 - \sigma_3 = \kappa\left(\overline{\rho}gz - P_f\right), \quad (1a)$$

(Suppe, 1985, p. 185), where $\overline{\rho}$ is the mean bulk density of the overlying rock. Eq. (1) also can be written in terms of the classic Hubbert-Rubey (1959) fluid-pressure ratio $\lambda = P_f/\overline{\rho}gz$

$$\sigma_1 - \sigma_3 = \kappa(1 - \lambda)\overline{\rho}gz, \quad (1b)$$

where $(1 - \lambda)$ may be considered the fractional fluid-pressure weakening. In submarine cases λ is measured with respect to the sea bottom (Davis et al., 1983). For the hydrostatic case (1a) becomes.

$$\sigma_1 - \sigma_3 = \kappa(\overline{\rho} - \rho_w)gz. \quad (2)$$

These crustal-strength equations express the fact that brittle pressure-dependent strength is a function of effective stress

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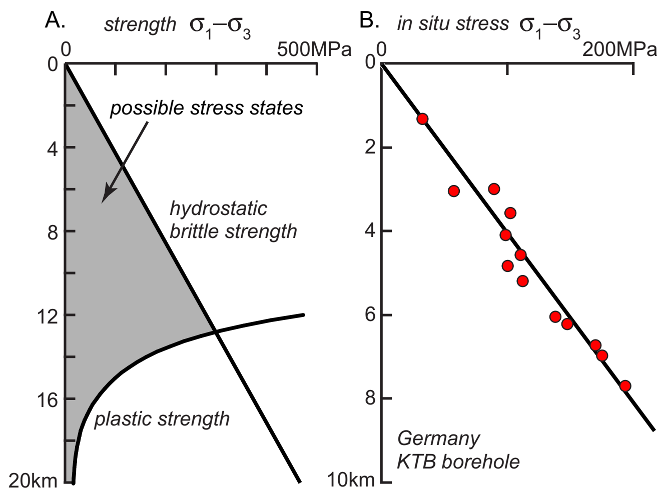

Fig. 1. (A) The classic strength-depth graph with linearly increasing brittle strength, which is based on the assumption of hydrostatic pore-fluid pressures (Eq. (2)). (B) This brittle-hydrostatic model agrees well with stress data from the deep KTB borehole in Germany where fluid pressures are hydrostatic (Brudy et al., 1997; Zoback and Harjes, 1997; Grawinkel and Stöckhert, 1997), as well as data elsewhere (*see* Townend and Zoback, 2000; Zoback, 2007).

$\overline{\rho}gz - P_f$ and not the solid pressure $\overline{\rho}gz$ or even depth, which is a distinction that becomes more important in the case of fluid overpressures.

The pressure dependence κ can be expressed in terms of the familiar Coulomb internal friction ($\mu = \tan\varphi$) and varies between $\kappa_c = 2\sin\varphi\ (1 - \sin\varphi)$ in pure-thrust compression and κ_e=2sinφ $(1 + \sin\varphi)$ in pure normal-fault extension, with strike-slip stress states (σ_2 vertical) lying between these limits, as can be shown from the Mohr diagram.

1.2. Success of the classic brittle-hydrostatic assumption

Brace and Kohlstedt (1980) assumed that brittle crustal strength is controlled by an optimally oriented static fault friction of $\mu = 0.6{-}0.85$ based on lab static friction measurements from a wide variety of rock types (Byerlee, 1978) ($\kappa_c = 2.1{-}3.7$, $\kappa_e = 0.68{-}0.79$). Townend and Zoback (2000) showed that deep borehole stresses are typically close to a critical state of failure consistent with $\mu = 0.55{-}1$ ($\kappa_c = 1.9{-}4.8$, $\kappa_e = 0.65{-}0.83$) and an assumed hydrostatic fluid pressure, implying that stress ≈ regional strength in many regions (Zoback, 2007). Therefore the Brace and Kohlstedt (1980) brittle-hydrostatic proposal (Eq. (2)) is in good agreement with available observations, as is illustrated by stress data from the deep KTB borehole in Germany (Fig. 1B).

We have expressed brittle crustal strength (Eq. (1a)) in terms of the vertical effective stress because this quantity can be directly calculated from commonly available borehole data or estimated from seismic velocities in clastic sedimentary sections that have not undergone uplift and erosion (*e.g.* Fertl, 1976; Magara, 1978; Swarbrick et al., 2002; Zoback, 2007). We will use Eq. (1a) to compare the observed vertical effective stress $\overline{\rho}gz - P_f$ with *in situ* borehole stress measurements of $\sigma_1 - \sigma_3$ to determine if they are consistent with a constant value of the pressure dependence κ over the entire depth range of the data from actively deforming sedimentary basins. This strategy provides a more explicit test of the role of pore-fluid pressures on crustal strength and avoids assuming what processes may control crustal brittle strength, which is an open question. For example, the regional-scale pressure dependence of the crust κ may be controlled by a combination of faulting, off-fault fracturing, and folding processes, some of which may be coseismic and may be controlled by dynamic frictional processes in earthquakes (*e.g.* Di Toro et al., 2011).

2. Implications of fluid overpressures for regional strength

In contrast with the classic view of linearly increasing crustal strength dominated by hydrostatic pore-fluid pressures, it is well established from petroleum boreholes and seismic velocity analysis that fluids are overpressured in deeper parts of thick, fine-grained clastic sedimentary basins and in deforming shale-rich plate-boundary mountain belts, largely due to disequilibrium compaction from stratigraphic and tectonic loading, but with additional effects including vertical and lateral pressure redistribution and gas generation (Bredehoeft and Hanshaw, 1968; Fertl, 1976; Magara, 1978; Hart et al., 1995; Swarbrick and Osborne, 1996; Yardley and Swarbrick, 2000; Tingay et al., 2009a). The basins of most interest for crustal strength and large-scale tectonics are continental margins built on oceanic or highly-thinned continental crust that have vast deforming clastic-rich volumes that approach a significant fraction of crustal thickness, even spanning the brittle-ductile transition (*e.g.* Gulf of Mexico, Niger delta, Bangladesh/Myanmar, Sumatra, Makran, Gulf of Alaska, New Zealand, Nankai trough, Barbados/Trinidad, and offshore southwestern Taiwan). Here we illustrate typical observed fluid-pressure/depth relationships from boreholes. We then compute the expected strength-depth profiles (Eq. (1a)) and compare them with borehole stress data in actively deforming regions, where stress is expected to be close to regional strength, or at least limited by Eq. (1a) as an upper bound.

2.1. Disequilibrium-compaction mechanism and crustal strength

The fluid-pressure depth relationship for the Yinggehai basin offshore south China (Fig. 2A) (Luo et al., 2003) is typical of the disequilibrium-compaction mechanism (*e.g.* Swarbrick and Osborne, 1996; Magara, 1978), which has also been called "leaky overpressure" (Crans and Mandl, 1980; Mandl and Crans, 1981). Pressures are hydrostatic down to a critical depth of ~1.8 km, the fluid-retention depth z_{FRD}, below which fluid pressures increase parallel to the lithostatic gradient (Fig. 2A), because permeability becomes sufficiently low that the incremental increases in sedimentary or tectonic load are supported by the pore fluid, rather than by incremental compaction of the solid-grain framework. Fluid pressures under the disequilibrium-compaction mechanism are a simple function of the fluid-retention depth, representing the sum of the hydrostatic and lithostatic contributions (Fig. 2A).

$$P_f = \rho_w g z \quad \text{for } z \le z_{FRD} \tag{3a}$$

$$P_f = \rho_w g z_{FRD} + \overline{\rho} g (z - z_{FRD}) \quad \text{for } z \ge z_{FRD} \tag{3b}$$

It follows that the Hubbert-Rubey fractional fluid-pressure weakening $(1 - \lambda)$ (see Eq. (1b)) is a simple function of the fluid-retention depth z_{FRD}

$$(1 - \lambda) = [1 - (\rho_w/\overline{\rho})] z_{FRD}/z \approx 0.6 z_{FRD}/z \quad \text{for } z \ge z_{FRD} \tag{4}$$

We make use of Eq. (4) in the discussion section of this paper.

Because of the observed parallelism of fluid pressures to the lithostatic gradient, the vertical effective stress $\overline{\rho}gz - P_f$ is approximately constant below the fluid-retention depth (Fig. 2A) and equal to the effective stress at the fluid-retention depth $(\overline{\rho}gz - P_f) = (\overline{\rho} - \rho_w) g z_{FRD}$. Therefore pressure-dependent strength