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A naturally constrained stress profile through the middle crust in an extensional terrane

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ABSTRACT

We present a method in which paleopiezometry, Ti-in-quartz thermobarometry (TitaniQ), and 2-D thermal modeling are used to construct a naturally constrained stress profile through the middle crust in an area of exhumed mid-crustal rocks. As an example, we examine the footwall of the Whipple Mountains metamorphic core complex (WMCC). Rocks in the WMCC were initially deformed at ~20 km depth by distributed ductile shear, and were then progressively overprinted by localized ductile shear zones and eventually by discrete brittle fracture as the footwall was cooled and exhumed toward the brittle–ductile transition (BDT). Increasing strain localization and cooling during exhumation allowed earlier microstructures to be preserved, and rocks in the WMCC therefore represent several points in temperature–stress space (and by inference depth–stress space). We identify enough of these stress–depth points to construct a complete profile of the flow stress through the middle crust to a depth of ~20 km, from which we derive regional estimates of the ambient stresses in the brittle upper crust, and the peak strength at the brittle–ductile transition in this region during Miocene extension.

Maximum differential stress reached ~136 MPa just below the brittle–ductile transition at a depth of ~9 km. Stress levels are consistent with Byerlee's law in the upper crust assuming a vertical maximum principal stress and near-hydrostatic pore fluid pressures, and suggest a coefficient of friction on the 25°-dipping Whipple fault of ~0.4. Differential stress decreases to 10–20 MPa at 20 km depths and ~500 °C. For strain rates typical of actively deforming regions (10^{-12} to 10^{-15}/s), our stress profile is bracketed by the Hirth et al. (2001) flow law for wet quartzite, whereas the flow law of Rutter and Brodie (2004) overestimates the strength of this particular region.

1. Introduction

The magnitude and spatial distribution of deviatoric stresses in the earth's crust has remained a fundamental question in geodynamics for over four decades (Brune et al., 1969; Burov and Watts, 2006; Hanks, 1977; Hanks and Raleigh, 1980; Jackson, 2002a; Lachenbruch and Sass, 1992; McGarr and Gay, 1978; Scholz, 2000; Thatcher and Pollitz, 2008; Zoback and Healy, 1992). Much of our understanding of crustal strength is framed in the context of laboratory experiments, which predict that rocks in the upper crust follow a Coulomb frictional failure criterion in which the differential stress is linearly related to the effective normal stress via a coefficient of friction (Brace and Kohlstedt, 1980; Sibson, 1983). Rocks in the lower crust are predicted to deform plastically, so that the stress depends on strain rate, temperature and grainsize, as a function of deformation mechanism (i.e. dislocation or diffusion creep) (Brace and Kohlstedt, 1980). These laboratory constraints, extrapolated over several orders of magnitude of strain-rate and temperature to natural conditions, predict that peak crustal strength resides at the brittle–ductile transition (BDT), such that rocks around the BDT may act as a 'stress guide' during continental deformation (Sibson, 1983). This high strength crustal beam is especially influential in continental deformation if its integrated strength exceeds the strength of the upper mantle (Jackson, 2002b).

Direct observations of deviatoric stress levels are limited to the brittle upper crust, for which deep boreholes in several locations worldwide have confirmed that ambient stresses are consistent with cohesionless friction with a coefficient of friction on favorably oriented faults of 0.6 to 1.0 (i.e. Anderson–Byerlee mechanics) (Brudy et al., 1997; Byerlee, 1978; Fuchs et al., 1991; McGarr, 1980; Zoback and Harjes, 1997). Several observations, however, such as the absence of a heat flow anomaly along major transform faults (Brune et al., 1969; Lachenbruch and Sass, 1992), and the relatively low magnitude of earthquake stress drops, suggest that faults themselves may be significantly weaker, perhaps due to high pore fluid pressures, or the presence of low friction materials along the fault such as clay gouge (Boulton et al., 2009). This has led to several conflicting hypotheses: for example, all faults in the brittle upper crust are weak

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(e.g. Hardebeck and Michael, 2004); only large-displacement faults are weak in an otherwise strong crust (e.g. Townend and Zoback, 2000), or all faults are strong and the heat flow argument is flawed (Scholz, 2000; Scholz and Hanks, 2004).

For the ductile field, the applicability of laboratory-derived flow laws can only be assessed indirectly, using, for example, inferences from field observations of exhumed middle and lower crustal rocks (Handy and Zingg, 1991; Mehl and Hirth, 2008; Stipp et al., 2002a), modeling of GPS velocity fields in actively deforming regions (Fialko, 2004; Pollitz, 2003; Thatcher and Pollitz, 2008), measurements of elastic thickness (Burov et al., 1998; Jackson, 2002a), distribution of seismicity (Maggi et al., 2000), and thermomechanical modeling (Kusznir and Park, 1987). Geophysical estimates based on the time-dependent response of the lithosphere to tectonic loads are limited by the non-uniqueness of model rheology structures – Maxwell viscoelastic vs. nonlinear-viscous, for example – and by trade-offs between the viscosity of the upper mantle and the lower crust. Field observations from exhumed rocks place useful constraints on the dominant rheology when the deformation was occurring, but often represent deformation occurring at rates well outside the reach of geophysical measurements.

Thus, large uncertainties contribute to a general lack of agreement as to how continental lithosphere responds to plate boundary and internal forces. It is clear, however, that crustal and lithospheric strength are likely to vary from region to region, and cannot be represented by a single universal strength profile. Methods of quantifying lithospheric strength in specific regions are thus extremely valuable.

In this paper, we illustrate a new method in which recently developed microstructural and thermobarometric techniques can be combined to produce naturally constrained depth profiles of the stress associated with lithospheric deformation in specific regions. The technique we outline has the potential to be applied to both crustal and mantle rocks that have been exhumed to the surface from depth and that preserve various stages of their exhumation histories. As an example, we present results from the middle crust of the Basin and Range province of the North American Cordillera, by focusing on the well-described Whipple Mountains metamorphic core complex (WMCC) in eastern California. Rocks in the WMCC were initially deformed at mid-crustal depths by distributed ductile shear, and were then progressively overprinted by localized ductile shear zones, and eventually by discrete brittle faulting as the footwall rocks were captured and exhumed toward the brittle–ductile transition (Davis, 1988; Davis et al., 1986). Increasing localization and cooling during exhumation allowed earlier microstructures produced by distributed deformation to be preserved, and we demonstrate that rocks in the WMCC footwall represent several points in stress–temperature space, and by inference, stress–depth space (Fig. 1). We estimate the magnitude of differential stress and the temperature at each point using recrystallized grainsize paleopiezometry and thermobarometry, respectively. We identify enough stress–depth points to construct a complete profile of the flow stress through the middle crust to a depth of ~20 km, from which we derive regional estimates of the ambient stresses in the brittle upper crust, the peak strength at the brittle–ductile transition (BDT), and the integrated strength of the continental crust in this region during Miocene extension.

2. Whipple Mountains Core Complex

The WMCC is one of several Miocene metamorphic core complexes within the Colorado River extensional corridor of eastern California and Arizona. It forms a NE–SW-trending elongate dome with lower plate rocks exposed in the core beneath the Whipple detachment (Fig. 2) (Davis et al., 1986). The hanging wall rocks include Tertiary volcanic and sedimentary strata cut by moderately to steeply dipping normal faults that sole into the Whipple detachment (Davis, 1988; Davis et al., 1986; Yin and Dunn, 1992). Rocks in the lower plate consist of Proterozoic gneisses and Cretaceous granitoids intruded by several suites of Tertiary dikes (Anderson and Rowley, 1981; Anderson et al., 1988). In the eastern half of the WMCC, lower plate rocks are mylonitized, and early high-angle gneissic fabrics are transposed into a gently SW-dipping mylonitic foliation with a top-NE sense of shear (Davis, 1988; Davis et al., 1986). Toward the west, however, the mylonitic foliation swings through the horizontal and then dips west beneath undeformed footwall granitoids

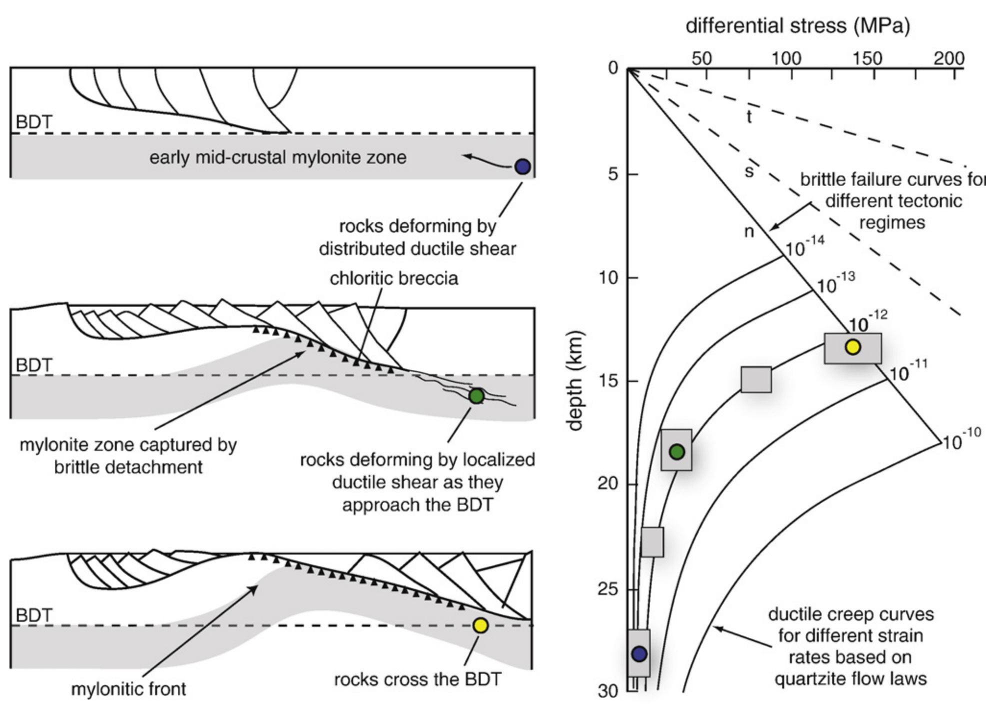

Fig. 1. Concept of constructing a crustal strength profile for the lower crust by examining rocks within a metamorphic core complex. Rocks initially deformed at mid-crustal depths via distributed ductile shear under low stress conditions are progressively overprinted by localized ductile shear zones under higher stress conditions and eventually by discrete brittle fracture as the footwall rocks are captured and exhumed toward the BDT. Each stage of deformation represents a point on a stress–depth profile as shown on the right-hand diagram. Abbreviations *t*, *s*, and *n* stand for thrust, strike-slip, and normal faulting regimes, respectively.