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Lithosphere–asthenosphere interactions near the San Andreas fault

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ABSTRACT

We decipher the strain history of the upper mantle in California through the comparison of the long-term finite strain field in the mantle and the surface strain-rate field, respectively inferred from fast polarization directions of seismic phases (SKS and SKKS), and Global Positioning System (GPS) surface velocity fields. We show that mantle strain and surface strain-rate fields are consistent in the vicinity of San Andreas Fault (SAF) in California. Such an agreement suggests that the lithosphere and strong asthenosphere have been deformed coherently and steadily since >1 Ma. We find that the crustal stress field rotates (up to 40° of rotation across a 50 km distance from 50° relative to the strike of the SAF, in the near-field of SAF) from San Francisco to the Central Valley. Both observations suggest that the SAF extends to depth, likely through the entire lithosphere. From Central Valley towards the Basin and Range, the orientations of GPS strain-rates, shear wave splitting measurements and seismic stress fields diverge indicating reduced coupling or/and shallow crustal extension and/or presence of frozen anisotropy.

1. Introduction

Despite several decades of study there remains a wide range of views on how the upper mantle deforms and interacts with the overlying crust. For instance, Bourne et al. (1998) proposed that the deformation of the upper mantle completely drives the stress field in the shallow crust. In this configuration the mantle imposes forces at the base of the lithosphere potentially creating with time large strike slip faults systems located near plate boundaries (Alpine, San Andreas, Anatolian). Others such as Jackson (2002) argue that the lithospheric stress is mostly concentrated in a strong seismogenic layer, and the contribution of upper mantle strength to crustal stress is negligible when considering faulting on the 100 km scale.

In this paper we assess the suitability of two possible end members of mantle deformation (Fig. 1) beneath continental San Andreas Fault (SAF) zone in California, utilizing the technique applied by Houlié and Stern (2012) in New Zealand. We use a GPS velocity field to compute the strain-rate field of the western USA (Fig. 2a) to compare the perpendicular to the maximal compressional strain-rate component ($\dot{\varepsilon}_1$) field with the directions of the

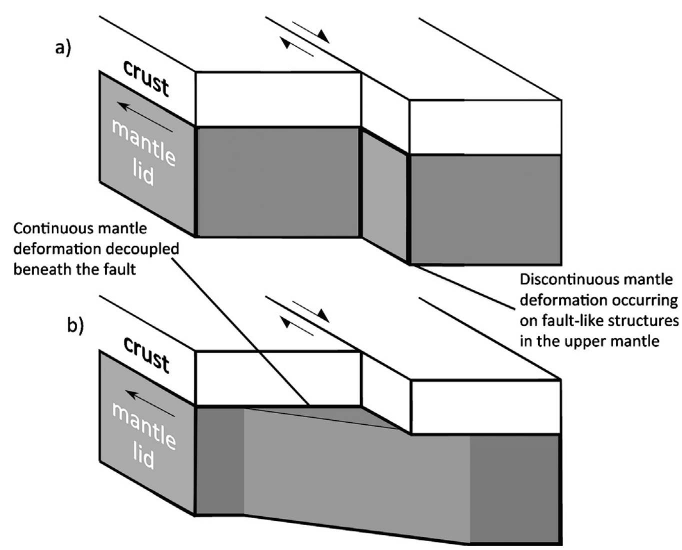

Fig. 1. Cartoon illustration of two possible end members of faulting related deformation in the mantle, after (Bourne et al., 1998; Little et al., 2002; Molnar et al., 1999). (a) Upper mantle displaying discontinuous deformation beneath the crustal fault structure. Displacement occurs in the upper mantle on fault-like structures at depth. (b) Upper mantle decoupled from the fault zone and undergoing simple shear as distant background forces in the mantle drive plate motion.

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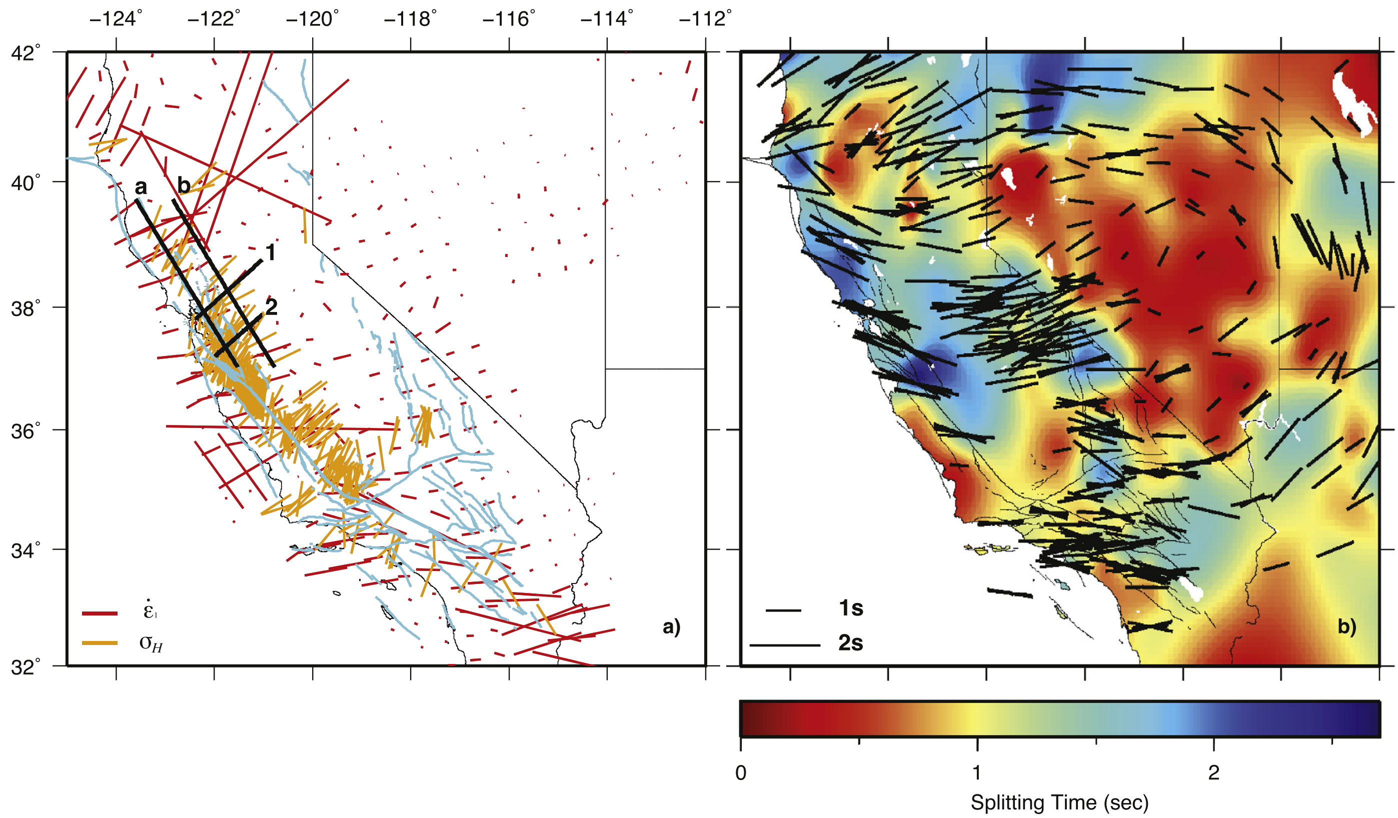

Fig. 2. (a) Principal shortening orientations derived from GPS with length of bar indicating strain-rate; the legend bar is of length 1.25×10^{-5} yr^{-1}. The strain-rate field has been inverted using the SSPX algorithm (Allmendinger et al., 2007) from the BARD and PBO GPS networks with nodes every 50 km. The shallow stress orientations depict σ_H^{max} as derived from borehole breakout orientations and cluster focal mechanism inversions. The transects marked, with profiles a and b (fault parallel) and profiles 1 and 2 (fault perpendicular) are used for analysis of shallow stress field rotation (Fig. 5). Fault parallel profiles are plotted at the angle used for calculation of alpha (Fig. 3). (b) Map of the interpolated splitting time delays (δt) overlayed by the SKS orientations with length proportional to δt.

fast axis of shear wave splitting (Φ) based on the measurements of the SKS and SKKS seismic phases (Fig. 2b). We term the angle between these two orientations, θ. If θ is close to zero then the $\dot{\varepsilon}_1$ and ϕ values are perpendicular to one another, (i.e. the maximum extensional direction and ϕ are parallel), and there is an agreement between strain orientations. Also, because the periods during which the anisotropy builds up in the mantle and the period of the crustal deformation are likely different, a small value of θ implies that (1) the stress regime has been steady since the strain accumulated in the mantle and (2) by consequence, the asthenosphere is strong and coupled to the lithosphere.

If θ is not small a wide range of possibilities is offered (lack of coupling resulting from fluid presence, crustal anisotropy, frozen anisotropy in the mantle, maximal component of anisotropy is vertical, etc.) and must be considered in the light of other geophysical and geological observables.

We assume that the GPS derived strain-rate field measured inter-seismically and computed over long-wavelength grids (>30 km spacing) is representative of the long-term strain-rate averaged over many seismic cycles and therefore corresponds to the deformation at timescales of >1 Ma; also thought to be the minimum period necessary to accumulate strain in the mantle. In the case of a steady uniaxial stress applied from the upper mantle to the surface over the period necessary to accumulate strain in the mantle, the surface strain-rate $\dot{\varepsilon}$ and, ε_{mantle}, the deformation accumulated in the mantle, are presumed to be aligned. If short- and long-term deformation are consistently oriented and the medium is isotropic, the principal axis of stress (σ_1) and principal axis of the strain-rate ($\dot{\varepsilon}_1$) must be perpendicular to the fast axis of the shear wave splitting, which we take to be a proxy for ε_{mantle}.

Taking the GPS strain-rate field as a proxy for shear strain direction in the upper mantle is in contrast to the more commonly used method of using fault orientation as a proxy for strain direction in the upper mantle (Bourne et al., 1998; Moore et al., 2002; Savage et al., 2004) and provides a new way of assessing deformation across the lithosphere–asthenosphere boundary for various time scales. This method (Houlié and Stern, 2012) is especially useful to test whether large strike-slip fault systems overlay areas where the upper-mantle is strong. For the Alpine fault in New Zealand, the relative orientation of strain-rate axes (defined here as the angle α) with the fault strikes is ~60°. That angle is similar to those estimated by using inversions of the principal stress axis projected onto the horizontal (σ_1) in the near-surrounding (<10 km) of the San Andreas Fault in Northern California (Hardebeck and Michael, 2004) or in the vicinity of North Anatolian fault (Biryol et al., 2010).

Finally, we compare our result with crustal and shallow stress field (Hardebeck and Michael, 2004; Hardebeck and Hauksson, 2001; Provost and Houston, 2001; Townend and Zoback, 2001; Zoback et al., 1987) in the light of GPS strain-rate and SKS fast orientation fields.

2. Data

In this study we use four datasets (GPS velocity field, shear wave splitting, principal stress directions and borehole breakouts), which highlight strain, strain-rate or stress in the lithosphere or/and asthenosphere for various time-scales (from decades to millions of years).

The GPS velocity field (Nikolaidis, 2002) has been measured at 395 locations including at sites of the Plate Boundary Observatory