Stress release in exhumed intermediate and deep earthquakes determined from ultramafic pseudotachylyte

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ABSTRACT

Stresses released by coseismic faults during subduction toward lawsonite-eclogite facies conditions in the Alpine subduction complex of Corsica can be estimated based on the energy required to form pseudotachylyte fault veins where shear strain can be measured. Congruent peridotite melting at ambient conditions of 1.5 GPa and 470 °C requires a temperature increase of 1280 °C to 1750 °C. We assume that more than 95% of the work is converted to heat during faulting, hence that the stress drop is nearly proportional to the amount of melting and inversely proportional to shear strain. Minimum estimates of released stress are typically greater than 220 MPa and as high as 580 MPa. The abundance of pseudotachylyte on small faults in the studied peridotite suggests that melting is very common on intermediate and deep earthquakes and that shear heating is important for seismic faulting at depth.

INTRODUCTION

The main strength of the lithosphere is carried by upper mantle peridotite and possibly also by dry granulites in the lowermost continental crust (e.g., Jackson et al., 2004). Rock-mechanical experiments and numerical models based on results from such experiments, extrapolated to subduction zone conditions, suggest that old (10^8 yr) and cold (500 ± 50 °C) upper mantle rocks with olivine rheology are strong (>1.5 GPa) at geological strain rates (10^{-14} s^{-1}). Strength is also grain-size-dependent, and coarse-grained peridotite, as studied here, is particularly strong (e.g., Kelemen and Hirth, 2007; Stüwe, 2007). Byerlee's (1978) friction data also indicate that faults are strong (~10^2 to 10^3 MPa at confining pressures discussed here). The strong versus weak fault disparity may be a result of most earthquakes being associated with variably efficient weakening processes (Rice, 2006), such as high pore fluid pressure and/or prefractured, noncohesive, or extremely fine-grained rocks, implying reactivation (e.g., Faulkner et al., 2006). The shortcomings of using seismics and field studies to quantify heating may also be important (e.g., d'Alessio et al., 2003; Scholz, 2006).

The study of pseudotachylyte (PST) where temperature change (ΔT) and shear strain (γ) can be determined provides an independent avenue to explore the strength of exhumed deep-seated rocks. This method has been used previously for faults at middle and high crustal level (e.g., Di Toro et al., 2005; Sibson, 1977; Wenk et al., 2000). Here, we apply this technique to determine minimum stresses released by coseismic faulting near the crust-mantle boundary in an ophiolite subjected to high-pressure and low-temperature metamorphism in a paleo-subduction environment. Shear-induced melting down to the base of the seismogenic zone (~15 km) is to be expected (e.g., Sibson, 1977). In the subduction environment, both the order(s) of magnitude of stresses and the mechanism of deep earthquakes are the subject of debate mostly based on seismological constraints or numerical models. We present stress estimates from recently discovered PST occurrences within the Alpine blueschist to eclogite facies complex in Corsica (Andersen and Austrheim, 2006; Austrheim and Andersen, 2004). This complex is considered a type example of fossil subduction (e.g., Jolivet et al., 2003). Our stress estimate is based on the energy balance combined with direct field and microstructural observations, and is independent of much-debated mechanisms of strain localization (e.g., Braeck and Podladchikov, 2007; Green and Houston, 1995; Kelemen and Hirth, 2007).

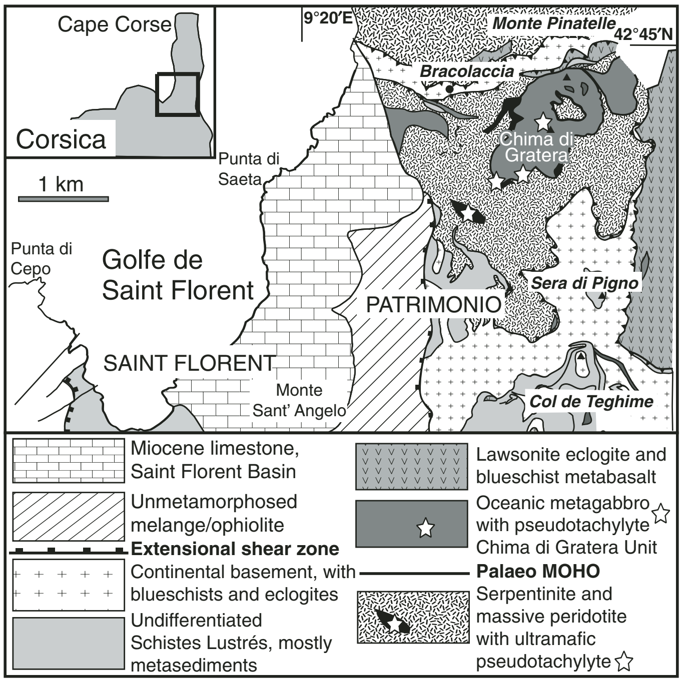

Figure 1. Simplified map of the Chima di Gratera area, Cape Corse, showing main geological units and the locations (stars) of ultramafic and mafic pseudotachylytes (in gabbro). The paleo-Moho is shown as a thick solid line.

FAULTING IN THE MANTLE PERIDOTITE

The *P-T* conditions of subduction-related faulting in Cape Corse (Fig. 1) are given by the regional metamorphism (e.g., Lahondère [1988] and unpublished lawsonite eclogite data giving pressures of 2.4 GPa at 470 °C [E.K. Ravna, 2008, personal commun.]). The eclogites were exhumed at blueschist and greenschist facies conditions (e.g., Fournier et al., 1991). The PSTs are variably pristine to completely overprinted by deformation and metamorphism associated with the exhumation (Andersen and Austrheim, 2006). Faulting took place at earlier stages of subduction, but it is difficult to precisely constrain at which stage in the prograde loop toward maximum *P-T* conditions individual faults were active. We use 470 °C and 1.5 GPa as the *P-T* conditions during faulting. Crosscutting PST veins are common in some of the larger fault zones (see Andersen and Austrheim, 2006) and show that faulting occurred repeatedly in this zone as predicted by the Kelemen and Hirth (2007) model. The faults analyzed in the present study, however, formed by single displacement events.

Omphacite, fassaitic (high-Al_2O_3) pyroxene, and glaucophane are present in quenched mafic PST (Austrheim and Andersen, 2004), whereas ultramafic PST has zoned olivine (Fig. 2D), high-*T* pyroxenes, spinel, and in some cases preserved glassy or hydrated glassy material (Andersen and Austrheim, 2006). Exhumation-related fabrics mostly obliterate evidence for the subduction paleoearthquakes, but quench textures in PST (Fig. 2) are well preserved within peridotite least affected by serpentinization.

The PST occurs near the contact previously interpreted as paleo-Moho within a Ligurian ophiolite of the "Schistes Lustés" nappe complex at Chima di Gratera (Fig. 1). The Jurassic Ligurian ophiolites (155–160 Ma) were cold and strong during the early Alpine subduction event (e.g., Jolivet et al., 2003). Because the well-preserved peridotite lenses (max 0.2 km^2) are mostly without markers, it is difficult to determine fault displacements. It is therefore difficult to ascertain how vast amounts of energy required for melting large volumes of peridotite along 1–15-cm-thick major fault veins (~30–450 kg m^{-2}) partitioned between stress and displacement. To quantify stresses in the fault energy budget we have instead studied a number of small faults (Fig. 2) where apparent displacement d_a (2–990 mm, no piercing points) and melt thickness *h* can be determined (Fig. 3). Melt thickness, taken to represent the width across which the displacement occurred, is measured in the microscope to be from 0.15 to 12.9 mm. Outcrop conditions make sampling difficult, therefore only 14 of 51 faults were sampled successfully (mostly drill cores). Optical and electron backscatter microscopy and probe analyses document near-complete congruous melting, common injection veins, and quench textures in 5%–25% of the damage zone of the faults measured in the field (Figs. 2C and 2D). Based on the microtextures of the 14 studied fault rocks, we assume that an average of 10% of the field-measured fault rock thicknesses of the 37 faults not studied microscopically are constituted by melted peridotite. Hence, we use *h* measured directly from thin sections, or alternatively, 10% of the thicknesses measured in the field as *h* in stress calculations.

STRESS ESTIMATES

In calculating the stress release by a fault breaking unfractured rock, we assume (Scholz, 2004) that the work (*W*) is partitioned into thermal energy (*Q*), seismic energy (*S*), and surface energy (*U*). Laboratory studies (Lockner and Okubo, 1983) and seismological studies (McGarr, 1999) suggest that both *S* and *U* are very small (<5%) in unfractured rock compared to the total energy in an earthquake; hence, we consider work per unit area, $W \approx Q$, which can be written

$$W = d\sigma;\ \text{rearranged as}\ \sigma = W/d, \qquad (1)$$

where *d* is true displacement and σ is the shear stress at the time of deformation. Lacking piercing points, the field-measured d_a given by offset markers is the best estimate of true displacement. Not considering potential superheating, the thermal energy required to heat and melt a unit mass is

$$Q = C_p \Delta T + H, \qquad (2)$$

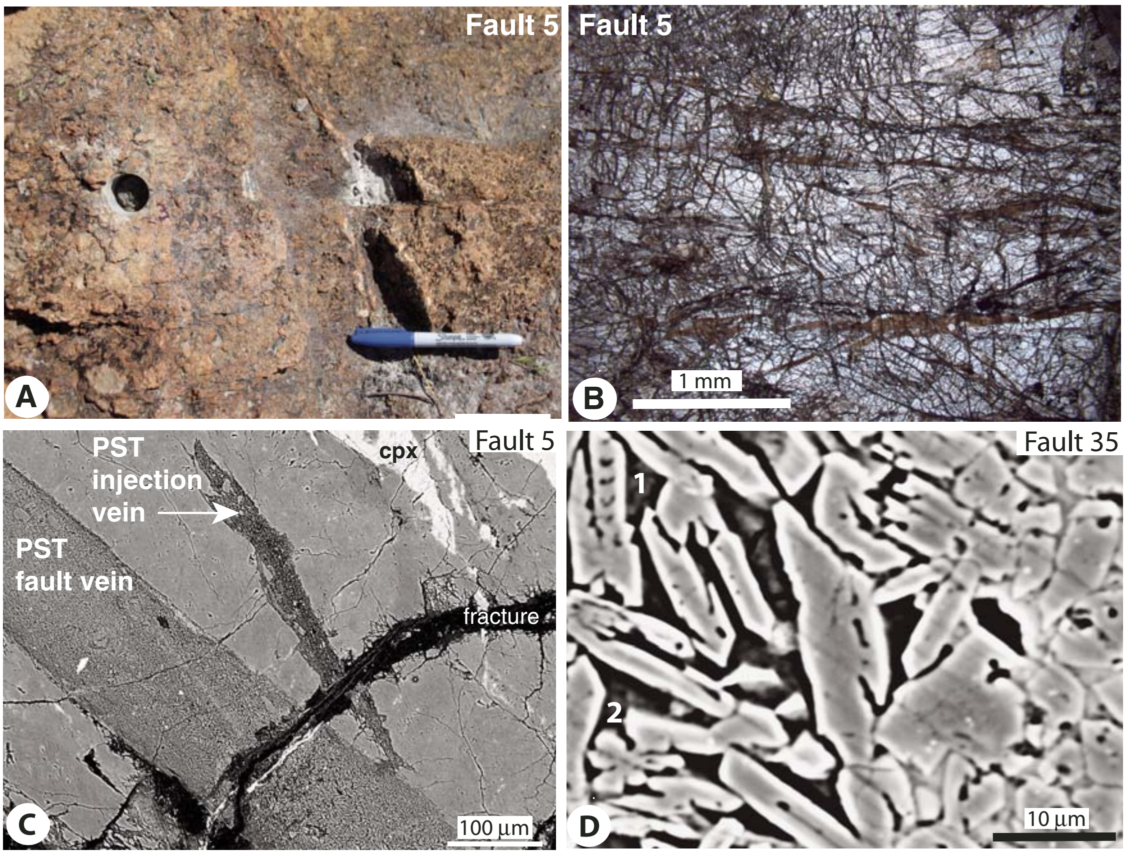

Figure 2. A: Small fault (fault 5 in Fig. 3A) offsetting an ~5-cm-thick gabbro vein by 40 mm. Notice that the drill core sample is taken entirely within the peridotite. B: Micrograph (plane-polarized) of fault 5 drill core (Fig. 2A) showing micro fault strands, fractured peridotite, and numerous injection veins in the damage zone of the fault. The pseudotachylyte thickness *h* is 0.3 mm. C: Electron backscatter image of fault strand and injection vein from fault 5. Notice near-complete melting and well-preserved quench texture of the pseudotachylyte (PST), which truncates the coarse-grained but fractured peridotite. Ol—olivine; Cpx—clinopyroxene. Pseudotachylyte is truncated by fractures associated with serpentinization. D: Backscatter image of "spinifex-like" olivine crystals from fault vein 35 (Fig. 3A). Notice the compositional zoning (Fo_{93-89}) with forsterite-rich cores and the skeleton and dendritic crystals (near points marked 1 and 2, respectively). The glassy matrix is partly hydrated (dark) and partly nonhydrated (lighter-colored along right-hand side of the image) where it has a cpx-like, but nonstoichiometric, composition.

where C_p is specific heat, ΔT is the temperature difference between ambient and melting temperature, and H is the latent heat of fusion. Because nearly all the energy is preserved as heating and melting, Equation 1 can be written

$$\sigma = \rho\left(C_p \Delta T + H\right)\gamma^{-1}, \tag{3}$$

where ρ is density and γ is shear strain. In our case, $\gamma = d_a/h$, recorded by field and/or microstructural analysis. Estimates of Q for a given γ therefore give quantitative information regarding the dynamic conditions that existed during earthquake rupture. A more precise definition of the shear stress estimate from Equation 3 is the *strain-averaged stress.* The energy balance across the shear zone integrated over time of deformation yields

$$C_p\int_{-h/2}^{h/2} \Delta T dx + hH = \int_{-h/2}^{h/2}\left(\int_0^{t_d} \sigma(t)\frac{\partial}{\partial t}\left(\frac{\partial u}{\partial x}\right)dt\right)dx = \int_0^{d} \sigma(u)\,du, \tag{4}$$

where t is time, x is the spatial coordinate across the shear zone, t_d is the time to achieve maximum displacement d, and u is the shear zone displacement field. Rearranging this equation we obtain the definition of the strain-averaged stress:

$$\sigma = \frac{1}{d}\int_0^{d} \sigma(u)\,du = \frac{h}{d}\left(C_p\frac{\int_{-h/2}^{h/2}\Delta T dx}{h} + H\right). \tag{5}$$

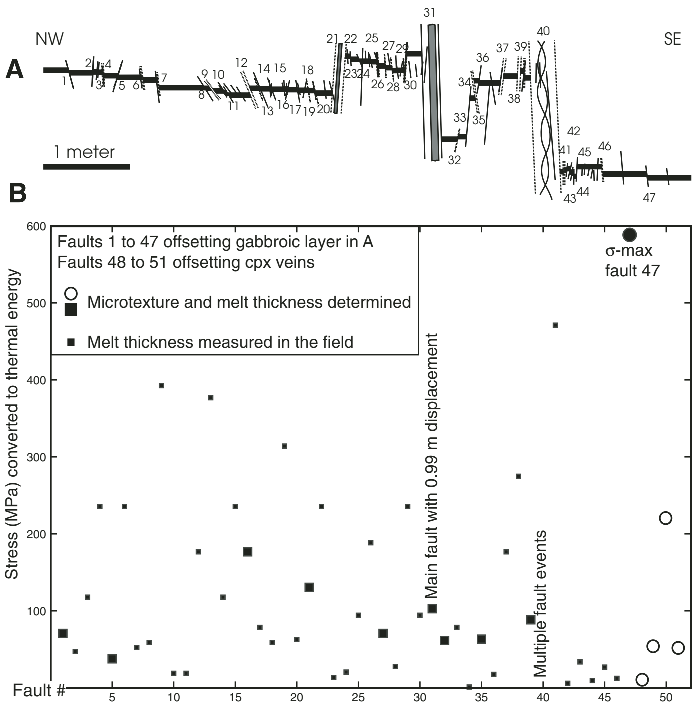

Figure 3. A: Field record of 47 faults offsetting a 5 cm gabbroic vein (see also Fig. 2A) cutting spinel peridotite. The main single-event fault is 31 with d_a = 0.99 m. Fault 40 records multiple events. Fault 47 gives the maximum strain-averaged stress of ~580 MPa. All samples studied by microtexture analyses were collected within the peridotite. B: Diagram showing strain-averaged stress from all measured faults (see text). Large circles and squares denote faults with microtexture information, i.e., melt volumes were measured in thin sections; small symbols denote faults where melt volumes are estimated as described in the text. Faults 1–47 (solid symbols) are shown in Figure 3A; faults 48–51 (open symbols) are small faults offsetting thin pyroxene veins.

Stresses obtained from Equation 3 are presented in Figure 3B. We assume that displacement took place across the melted zone h, which gives an overestimate of γ and an underestimate of σ. We have not corrected for injection (Figs. 2B and 2C), commonly assumed to be 50% or more in large fault veins (e.g., Di Toro et al., 2005; Wenk et al., 2000), since we have no good control on the scaling relations between the melt loss on the very small faults used here and more commonly studied larger faults. The field-measured d_a is always a minimum estimate due to lack of piercing points. Approximations of h and d_a therefore both have the effect of reducing estimates of σ. Improvement to these estimates relies on better determination of the melt volumes in a larger number of faults and wall rocks.

RESULTS AND DISCUSSION

Commonly accepted mechanisms for subduction earthquakes are embrittlement by devolatilization of minerals at progressive metamorphism during subduction (recent overview, Hacker et al., 2003; and case study, Rietbrock and Waldhauser, 2004). Shear heating may also lead to extreme localization of deformation at seismic strain rates in viscoelastic materials (e.g., Braeck and Podladchikov, 2007; T. John, 2007, personal commun.; Kelemen and Hirth, 2007). These models demonstrate that extreme temperatures similar to those suggested by Kanamori et al. (1998) may be viable for high-P faults in subduction and collision zones. Here we interpret σ as the strain-averaged stress associated with a single rupture event (Equation 5). We make no assumptions about the manner in which the stress drops during slip, but infer that both higher and lower stresses were available during the rupture process. We suggest that the rocks that hosted these events must have experienced stresses that were at least as high as the characteristic stress estimates from these small-scale faults, and most likely considerably higher. Stress estimates from large faults in the area cannot be carried out with this method because displacements are indeterminable (Andersen and Austrheim, 2006).

The temperature rise required for near-complete melting of mostly dry spinel peridotite at 1.5 GPa and 470 °C is $\Delta T \approx 1280$ °C (Katz et al., 2003). Because it is difficult to accurately determine pressures for individual events, we used 1.5 GPa, intermediate between 1 and 2.4 GPa in agreement with commonly referenced metamorphic conditions in the area (see summary by Jolivet et al., 2003). In Equation 3 we use $\Delta T = 1280$ °C, $C_p = 1150$ J kg^{-1} °C^{-1}, $H = 8.6 \times 10^5$ J kg^{-1}, and $\rho = 3200$ kg m^{-3}. The highest σ calculated from a fault where h is determined directly by microscopy is 220 MPa (fault 50), whereas the maximum σ is 580 MPa (fault 47) where h was estimated without microstructural inspection (Fig. 3B). The reactivated fault 40 is not used. The main fault 31 (d_a = 0.99 m) gives $\sigma \approx 110$ MPa. We suggest this is underestimated, as the fault rock thickness varies considerably in the field. The fault 31 sample was from a 43-mm-thick fault rock with only h = 2.9 mm measured in thin section. Field measurement of h varies from 35 to 57 mm along the studied 1.42 m segment of fault 31. The uncertainty in γ (based on h) is therefore close to a factor of 3, and the stress may be underestimated accordingly.

The conservative melt estimates, notwithstanding the problems of quantifying loss to injection veins, grain boundaries, and dilation bends, give a clear indication that the stresses released to thermal energy in most of the faults observed were commonly higher than 200 MPa, and higher than 580 MPa on fault 47 (Fig. 3B). It is interesting to observe that this is very similar to the estimates of 300–600, MPa by Obata and Karato (1995) based on dislocation density and grain size from olivine in PST from the Balmuccia peridotite. Kelemen and Hirth (2007) modeled coseismic failure at high stress (10^2 to 10^3 MPa) in subducting mantle peridotite and demonstrated that repeated seismic events could be related to shear heating and grain-size-sensitive creep laws. The modeling also indicates stress fluctuations, and it may be that our stress variations, approaching an order of magnitude, may be an effect of grain size (smaller grain size lowers strength and volume of melting). Melting along faults, however, may destroy evidence for preexisting grain size variation and render precise observational interpretation uncertain.

Byerlee's (1978) classic study suggested that frictional faulting is independent of composition. With shear heating as a possible failure mechanism, however, rheology and grain size also become important elements (Braeck and Podladchikov, 2007; John et al., 2007; Kelemen and Hirth, 2007). The absence of a fault-zone weakening mechanism such as pressurization of pore fluids (Rice, 2006; Sibson, 1977) has commonly been taken as a prerequisite for generation of PST and used to explain their relative scarcity compared to other exposed fault rocks (Sibson and Troy, 2006). Coseismic faulting occurs because faults weaken with increasing slip, observed in our data and by Di Toro et al. (2005). In cases where PST is produced, shear heating may in fact be the dominant weakening mechanism, since melting both lubricates and dries wall rocks, particularly at elevated pressures (e.g., Di Toro et al., 2004; Mysen and Wheeler, 2000). Subduction earthquakes are obvious candidates for high-stress failure. Fault rocks produced from subduction are, however, very rarely exposed, because they generally are lost by subduction, destroyed during exhumation, or not recognized in the field. These unique Corsican occurrences are therefore generally important for understanding failure mechanisms and for estimating the strength of rocks in subduction complexes. We document that peridotite can sustain stresses of several hundred MPa over geologic time, and that even very small faults at such high stress generate melt in refractory mantle. It is suggested that most intermediate and deeper earthquakes produce PST and that the apparent relative scarcity of PST is simply a function of the global sampling depth.

ACKNOWLEDGMENTS

We thank Q. Gautier and University of Oslo Physics of Geological Processes (PGP) colleagues for discussions and support in the field, and M. Erambert for help with the microprobe. Reviews by Peter Kelemen, Kurt Stüwe, and Chris Scholz improved the manuscript. A Centre of Excellence grant from the Norwegian Research Council to the PGP department financed the study.

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Manuscript received 5 June 2008
Revised manuscript received 1 September 2008
Manuscript accepted 3 September 2008

Printed in USA