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Pressure-related feedback processes in the generation of pseudotachylytes

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Abstract

Pseudotachylyte formation is typically viewed as a self-limiting process. Quantitative models have demonstrated that frictional melting is possible at seismic slip rates, but once a thin film of melt forms on a seismic fault plane, the dramatic decrease in fault friction is thought to suppress further melting. Volumes of melt-generated pseudotachylyte observed in field studies, however, are in many cases substantially larger than the amounts of frictional melt predicted by theoretical models. This suggests that previously unrecognized physical processes may enhance melting during seismic slip. Localized decompression at dilational jogs may be one such phenomenon. Transient unloading could play two important roles in the dynamics of pseudotachylyte generation. First, by setting up significant fluid pressure gradients, it would lead to rapid migration of the melt to sites of low pressure, thereby reestablishing frictional contact across the fault surface and favoring further melt generation. Second, at significant depths, sudden depressurization might lead to in situ decompression melting.

Keywords: Pseudotachylyte formation; Pressure-related feedback processes; Localized decompression; Transient unloading; Deep earthquakes

1. Introduction

The nature and genesis of pseudotachylytes has been controversial in the past (e.g. Wenk, 1978), but the current consensus is that these distinctive rocks are records of seismic events and are formed by a combination of ultracataclasis and frictional melting, with the latter a requirement for 'true' pseudotachylyte (e.g. Spray, 1987, 1995; Magloughlin, 1992; Lin, 1999; Fabbri et al., 2000; O'Hara, 2001). Ultramylonites and ultracataclasites may be spatially associated with and superficially similar to pseudotachylyte, but they lack features indicative of melting (e.g. evidence of liquid flow, high-temperature microlites, dendritic and skeletal crystals, spherulitic textures, embayed clasts [Magloughlin and Spray, 1992]).

The physical plausibility of melt generation on seismic slip surfaces has been demonstrated theoretically (McKenzie and Brune, 1972; Sibson, 1975; Cardwell et al., 1978) and experimentally (Spray, 1987, 1995; Tsutsumi and Shimamoto, 1996). There remains, however, a significant gap between the small volumes of frictional melt that these studies predict on slip surfaces (millimeter- to centimeter-thick films), and the larger volumes commonly reported for fault veins that appear to record single slip events (centimeter- to meter-thick sheets). There are two possible explanations for this inconsistency. First, the volume of melt-generated pseudotachylyte might be routinely overestimated in field studies. But even where pseudotachylyte is carefully distinguished from ultramylonites and ultracataclasites, the observed thickness of pseudotachylyte in generation zones commonly exceeds 1 cm (Fig. 1; Table 1). Moreover, because a large fraction of melt commonly migrates out of the generation zone (Magloughlin, 1989) and forms injection veins (Grocott, 1981; Swanson, 1992), these reported thicknesses are minimum estimates for the total amount of melt formed.

In contrast, several theoretical approaches to estimating the magnitude of frictional heating and melting (if any) during seismic faulting all lead to the conclusion that, although a significant temperature increase is possible on a fault plane during a seismic event, only a trivial volume of rock 'feels' the thermal effects (see next section). This

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points to a second explanation for the incongruence between theoretical predictions and field observations: that transient conditions along a slipping fault contribute to the generation of melts. The purpose of this paper is to explore the possibility that dynamic changes in pressure within the fault zone may be such a factor.

2. Frictional heating on faults: summary of earlier work

The energy production Q per unit area on a slipping fault (Scholz, 1980), equal to the product of resolved shear stress τ times the slip distance d, can be divided into two parts: the work done dynamically against resistance to fracture propagation (Q_{dyn}) and the work done against friction (Q_{fric}). The frictional term is further partitioned into a non-thermal component (the work allocated to overcoming surface energy, Q_s) and a thermal component (frictional heat generation, Q_h), such that:

$$Q = \tau d = Q_{dyn} + Q_{fric} = Q_{dyn} + Q_s + Q_h \quad (1)$$

At non-seismic slip rates, conductive and advective heat transport can nearly keep pace with the rate of frictional heat generation, and the temperature increase in the fault zone is minimal. For seismic slip rates (typically 0.1–2 m/s; Spray, 1992), most of the frictionally generated heat Q_h is retained, at least momentarily, and its value is orders of magnitude greater than either Q_{dyn} or Q_s (Scholz, 1980; Kanamori et al., 1998). Thus in the seismic case, frictional heat is approximately equal to total energy production, and all of this energy goes into raising the temperature along the fault. This assumption has been the starting point for most attempts to calculate plausible values for temperature increases associated with seismic faulting.

Philpotts (1990) used a 'rate of working' approach to determine the local temperature rise ΔT along a fault with no heat loss:

$$\Delta T = \frac{\tau (\mathrm{d}\varepsilon/\mathrm{d}t) t}{C_p \rho} \quad (2)$$

where τ is shear stress, $\mathrm{d}\varepsilon/\mathrm{d}t$ is strain rate, t is the duration of slip, C_p is heat capacity and ρ rock density. The strain rate, however, is more easily quantified if expressed as the ratio of the particle or slip velocity, v to fault zone thickness w, so that:

$$\Delta T = \frac{\tau v t}{w C_p \rho} \quad (3)$$

Assigning plausible values to these variables for a seismic slip event in the middle crust ($\tau = 0.1$ GPa, $v = 0.5$ m/s; $t = 1$ s; $C_p = 1$ kJ/kg/K; $\rho = 2800$ kg/m^3) on a fault 1 cm in width (w) yields a temperature rise of ca. 1800 °C, sufficient to cause localized melting of most minerals in crustal rocks. However, Spray (1995) and Kanamori et al. (1998) emphasized that the total heat generated and the

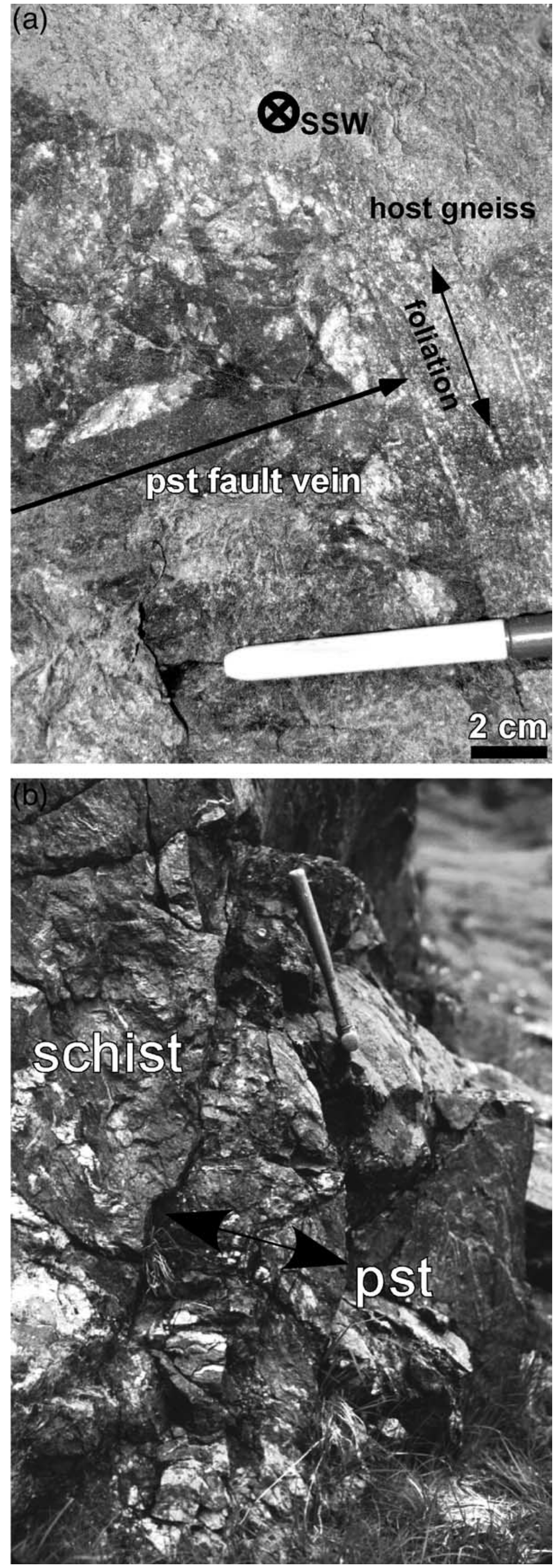

Fig. 1. Examples of pseudotachylyte fault veins whose thickness exceeds theoretical predictions. (a) Edge of 30–40-cm-thick pseudotachylyte fault vein, concordant to gneissic foliation, Outer Hebrides Fault Zone, Isle of

Table 1
Thicknesses of pseudotachylyte fault and injection veins

Reference	Location	Maximum fault vein thickness and [traceable length]	Maximum injection vein thickness and [traceable length]
Austrheim and Boundy, 1994; Bjørnerud et al., 2002	Holsnøy, W Norway	5 cm [100 m]	5 cm [0.5 m]
Camacho et al., 1995	Musgrave Ranges, central Australia	8 cm	4 m [10 m]
Wenk et al., 2000	Eastern Peninsular Ranges, California, USA	10 cm [0.5 m]	
Magloughlin, 1989	Northern Cascades, Washington USA	20 cm	
McNulty, 1995	Sierra Nevada, California, USA	ca. 30 cm [10 m]	
Sibson, 1975; Magloughlin (unpublished data)	Outer Hebrides, Scotland	40 cm	
Curewitz and Karson, 1999	SE Greenland	10 m (thickness of 'reservoir zones' containing meter-size clasts)	2 cm [1 m]

potential for melting on a fault is an extremely sensitive function of the slip rate v relative to the width of the process zone, w (since v/w is equivalent to strain rate). For example, if slip rate v in the calculation above is decreased by just a factor of 5, to 0.1 m/s (still considered to be 'seismic'), then the width of the heated zone, for the same temperature rise, is reduced to 2 mm.

In an earlier analysis of frictional heating on seismic faults, McKenzie and Brune (1972) developed a numerical model that took into account temporal effects including heat conduction away from the fault and variations in slip rate, based on the heat flow equation:

$$-H(x,t) = \rho C_{\mathrm{p}}\left(\frac{K\delta^2 T}{\delta x^2} - \frac{\mathrm{d}T}{\mathrm{d}t}\right) \quad (4)$$

where x is the direction perpendicular to the fault, $H(x,t)$ is the rate of heat generation per unit volume of rock, a function of slip rate and frictional stress; and K is the thermal diffusivity (equal to $k/\rho C_{\mathrm{p}}$, where k is thermal conductivity). Note that the second term is essentially identical to Eq. (1) above. McKenzie and Brune (1972) presented temperature calculations for two slip scenarios (constant vs. exponentially decaying slip rate) and two different boundary conditions (formation of continuous melt films vs. formation of isolated melt pockets). In all cases, the melt zones predicted to form along the fault zone are on the order of 1–2 mm thick, except for the very largest seismic events (largest stress drops), which may form melt layers up to 1 cm thick. The model simulations were restricted, however, to seismicity in the shallow crust.

Although these studies indicate that melting is possible during seismic faulting, they do not take into account several factors that may conspire against the production of significant amounts of melt. First, the heat of fusion must be supplied to the host rock before melting can occur. The presence of unmelted clasts within many pseudotachylytes (Lin, 1999; O'Hara, 2001) indicates that this is an important sink for frictional heat. Second, the melting recorded by most mid- and lower-crustal pseudotachylytes appears to have occurred in the absence of intergranular aqueous fluids that could have acted as fluxes (e.g. Sibson, 1975; Camacho et al., 1995; Bjørnerud et al., 2002). Indeed, anhydrous conditions may be a prerequisite for pseudotachylyte generation at these depths because free water (and the vapor phase it would form when heated) would substantially reduce the area of frictional contact within the fault zone and also facilitate advective heat transport away from the zone (Sibson, 1977). Thus the instantaneous coseismic temperatures must in many cases exceed the dry melting temperatures of the predominant minerals for significant amounts of pseudotachylyte to form. A third factor that makes large volumes of pseudotachylyte difficult to explain is that as soon as any melt is formed, the frictional properties of the system change rapidly, and the shear stress value (the product of the effective normal stress and frictional sliding coefficient) may fall by orders of magnitude, inhibiting additional heat generation (McKenzie and Brune, 1972; Sibson, 1975). Although some experimental studies (e.g. Tsutsumi and Shimamoto, 1996) have shown that friction may remain constant or even increase slightly at the onset of melting (apparently owing to welding of asperity junctions), this transient effect gives way immediately to velocity weakening behavior (a prerequisite for any seismic instability). The volumetric increase of ca. 10% associated with melting leads to an enhanced pore pressure effect along the fault surface that drastically reduces frictional contact. Scholz (1990, p. 145) suggested that fault weakening from transient heating effects is the only circumstance in which coseismic stress drops can locally approach 100%.

In summary, the thermal budget along a seismically active fault would appear to be only marginally sufficient to

Lewis, Scotland. (b) Thick pseudotachylyte fault vein/breccia zone, 15–23 cm wide, marked by black double arrow (pst). Vein is approximately concordant with foliation in host schist; view is to the northwest. Near Lake Ethel, Chelan County, Washington.

melt rock within the fault zone. Extreme temperatures can be achieved, but only if extremely small amounts of rock are processed along the fault plane. Also, pseudotachylyte production seems to be a self-limiting process owing to the abrupt decline in frictional heat generation expected once the first melt forms. However, the sizeable volumes of pseudotachylyte observed in field studies suggest that some sort of positive-feedback mechanism contributes in some cases to melting on slipping fault planes.

3. Evidence for dilational strains during pseudotachylyte generation

Many reports on natural pseudotachylytes include descriptions of chaotically brecciated zones in which pseudotachylyte forms a web-like network around angular clasts of the host rock (Fig. 2). In his study of pseudotachylytes in Lewisian gneisses in the Outer Hebrides of Scotland, Sibson (1975) noted that such bodies tend to occur at releasing bends in the 'pseudotachylyte source zones' and interpreted them as implosion breccias formed by instantaneous dilation followed by collapse. Grocott (1981), Swanson (1992), and Curewitz and Karson (1999) similarly observed that many pseudotachylyte generation zones consist of paired shear surfaces and that pseudotachylyte breccias commonly occur in pull-apart structures in the space between these surfaces.

Pseudotachylyte breccias on the island of Holsnøy, Bergen arcs region, western Norway provide additional evidence for local dilatancy during pseudotachylyte formation. Pseudotachylytes on Holsnøy are exceptional in that they appear to have formed at depths of >50 km (15–17 kbar), based on the occurrence of eclogite-facies minerals (jadeite, kyanite, garnet) with skeletal and dendritic habits within the pseudotachylytes (Austrheim and Boundy, 1994). This is well below the typical depths for crustal seismicity (Sibson, 1989), but in this case ductile behavior was apparently suppressed by the extremely anhydrous nature of the host rocks. The pseudotachylytes were formed during Caledonian (ca. 420 Ma) reactivation of gabbroic to anorthositic banded gneisses that had been previously metamorphosed, and completely dehydrated, under granulite facies conditions (ca. 800 °C and 10 kbar) at about 945 Ma (Austrheim, 1987). The gneissic banding seems to have played an important mechanical role in the formation of the pseudotachylyte; many of the largest bodies of pseudotachylyte are laterally extensive sheets (surface areas >800 m^2) that parallel mafic layers in the gneisses (Bjørnerud et al., 2002). The planar and continuous nature of these bodies suggests that they are fault veins generated in situ, with local migration of the melt into apophyses, or injection veins, that branched off from the source. In some cases fault veins can be traced into complex 'crush zones' in which millimeter–centimeter thick selvages of pseudotachylyte surround angular blocks of chaotically rotated granulite. At many of these localities, the amount of rotation of the granulite banding between adjacent blocks seems geometrically impossible to achieve in the space filled by the pseudotachylyte (Fig. 3; cf. Swanson, 1992, his fig. 3b). This geometric paradox can be explained if there had been violent fragmentation due to instantaneous dilation during the formation of these breccias, followed by, or simultaneous with, the influx of pseudotachylyte melt. In other words, the breccias may represent compressed, pseudotachylyte-cemented 'rock bursts'.

Without three-dimensional exposure, it is impossible to assign a precise value to the dilational strains that are suggested by the rotation of the blocks in Fig. 3. A two-dimensional estimate of dilation was made from a digital image of the photograph, treating the blocks as puzzle pieces that cannot intersect during rotation. Although there is no unique solution to this geometric puzzle owing to the possibility of movement in the third dimension, areal dilations of at least several percent are necessary to explain the observed configurations. Such dilations imply significant, albeit momentary, pressure drops. If dilational jogs (Fig. 2) and crush zones (Fig. 3) represent actual voids that opened during coseismic slip, the pressure drop could be the lithostatic value (in the Holsnøy case, at least 1.5 GPa). While this seems exceptionally high, it is of the same order of magnitude as the pressure drop that can be calculated

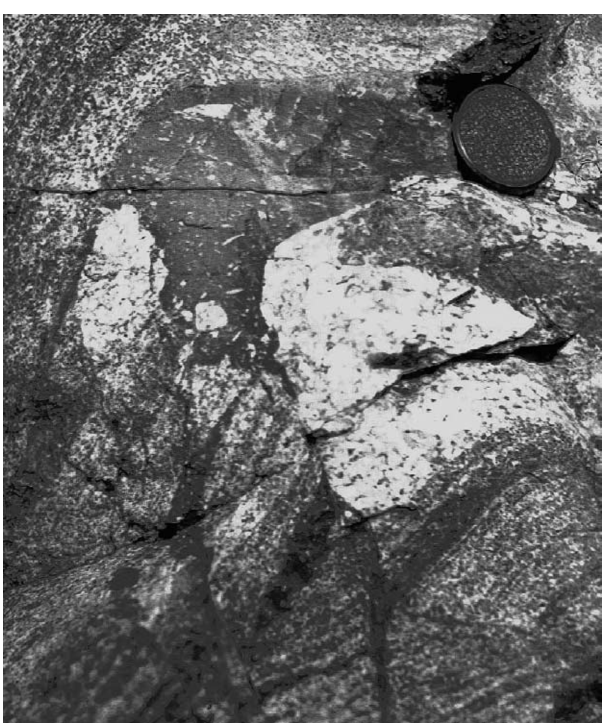

Fig. 2. Pseudotachylyte breccia probably representing a dilational jog along a fault surface in foliated granulite. Lens cap is 5 cm in diameter. Locality: Øst Ådnefjellet, Holsnøy, western Norway.

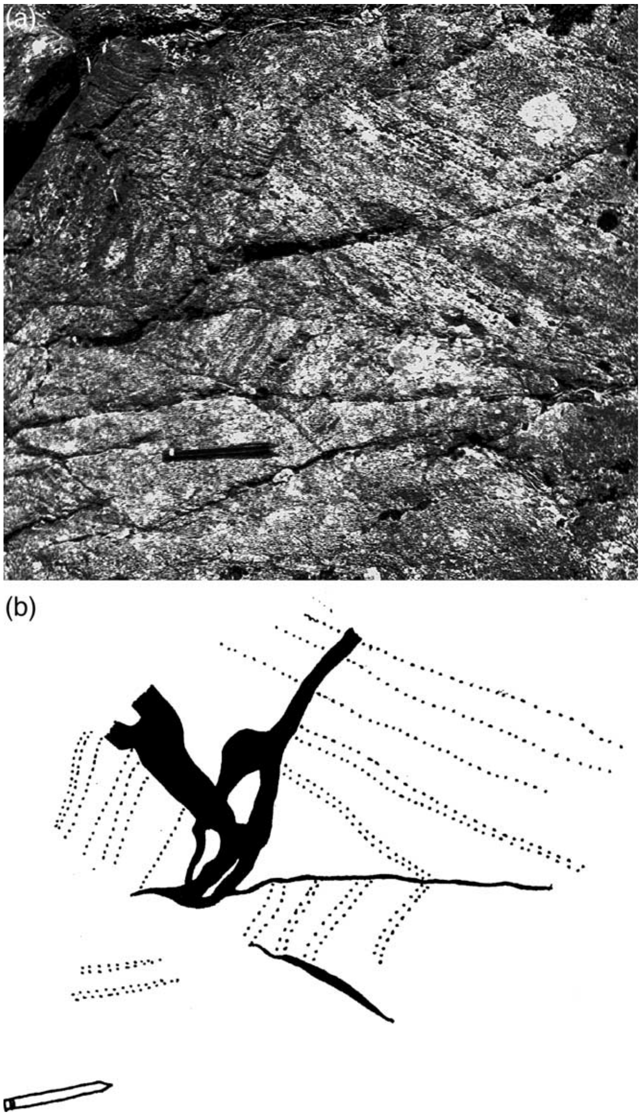

Fig. 3. 'Crush zone' in which rotated blocks of granulite are bounded by pseudotachylyte veins of varying thickness (black areas in sketch). Note abrupt changes in orientation of gneissic banding in the granulites (dotted lines); the apparent rotation seems geometrically impossible to achieve in the space filled by the pseudotachylyte. Pencil at lower left is 11 cm long. Locality: Vest Ådnefjellet, Holsnøy, western Norway.

from the equation for the bulk modulus:

$$\Delta P = \left(\frac{\Delta V}{V}\right) K \tag{5}$$

where ΔP is the transient pressure drop, $\Delta V/V$ is the dilation, and K is the bulk modulus of the rock, typically about 50 GPa (Turcotte and Schubert, 2002, pp. 112, 435). For a dilation value of 2%, ΔP would be on the order of 1 GPa.

In summary, structural observations of pseudotachylyte bodies believed to have formed at middle to lower crustal depths indicate that non-trivial dilational strains may locally accompany seismic faulting. This, in turn, suggests that momentary depressurization may also occur locally, even at great depth. Microtextural features in some pseudotachylytes appear to record such pressure drops. Goodwin (1999) describes sieve-textured amphibole crystallites in a study of pseudotachylytes within granitic mylonites in an Arizona metamorphic core complex. These resorption textures, also observed in some volcanic rocks (Tsuchiyama, 1985; Nelson and Montana, 1992), indicate significant disequilibrium between crystals and melt. Such disequilibrium can occur in one of two ways: (1) by a sudden change in magma composition (i.e. by influx of melt from another source) or (2) by rapid decompression of the melt. Because there was no evidence for magma mixing in these pseudotachylytes, Goodwin concluded that the textures record decompression of the melt sometime during a crystallization period of less than 10 s (based on the sizes of the largest crystallites).

4. Effects of instantaneous pressure drops and temperature increases in a fault zone

How might such instantaneous depressurization affect the dynamics of pseudotachylyte generation? First, a rapid drop in pressure at one site along a fault zone, such as a dilational jog or pull-apart structure, would give rise to a tremendous fluid pressure gradient for any melt that had formed. The melt would be drawn into the low pressure areas, thereby reestablishing frictional contact between the rock surfaces adjacent to the fault. The importance of syndeformational melt migration in maintaining dynamic rock friction has been demonstrated recently by Rosenberg and Handy (2000), who used analogue materials to simulate the behavior of partially melted quartzofeldspathic rock undergoing simple shear. Berlenbach and Roering (1992) interpreted sheath-fold-like structures in impact-related pseudotachylyte from the Witwatersrand Basin as evidence of 'plug flow' driven by strong fluid pressure gradients within an irregular fault zone. If such flow happened fast enough, it could allow for at least localized melt generation over the full duration of the slip event. The pressure-driven flow rate might be approximated as a one-dimensional channel flow with a parabolic velocity profile given by:

$$u = \left(\frac{1}{2\mu}\right)\left(\frac{\mathrm{d}p}{\mathrm{d}x}\right)\left(\frac{y'2 - h^2}{4}\right) \tag{6}$$

where u is the flow velocity, μ is the melt viscosity, $\mathrm{d}p/\mathrm{d}x$ is the pressure gradient along the channel (fault zone), y' is the distance from the center of the channel, and h is the width of the channel (Turcotte and Schubert, 2002, p. 229). For a 1-cm-fault zone, a pressure difference of 1 GPa over a distance of 1 m, and viscosity value of 10^7 poise (10^6 kg/m/s), a comparatively high value that would be characteristic of a dry granitic melt (Scholz, 1990, p. 137), the flow rate at the center of the channel would be on the order of 1 cm/s. The flow profile in the melt would be modified by the viscous drag imposed by the moving fault wall, but the essential point is that pressure-induced flow rates in even the most viscous melts would be commensurate with seismic slip rates if the pressure gradient were high enough. In the Holsnøy case, the mafic host rock would likely have produced melts with significantly lower viscosity. Hence, it is very plausible that migration of the melt along the fault plane dynamically influenced coseismic slip.

A second possible effect of instantaneous dilation along a fault at significant depth could be decompression melting. This is the mechanism invoked for Type 'S' (or 'A') pseudotachylytes formed in impact settings during the excavation stage of crater formation (e.g. Martini, 1978; Spray and Thompson, 1995; Gibson et al., 1997). The nearly instantaneous decompression associated with passage of the rarefactive part of the shock wave through the heated rock is thought to lead to significant melting. Similarly, the local coseismic pressure drops that occur on faults in the middle to lower crust may be sufficient to cause decompression melting even if the increase in temperature is rather modest. Phase diagrams for rocks at typical crustal pressures and temperatures indicate that this is plausible (e.g. Thompson, 1972, p. 407). For the Holsnøy pseudotachylyte, as discussed above, the instantaneous pressure drop may have exceeded 1 GPa. At equilibrium this would be more than adequate to cause melting of a gabbroic rock at temperatures only 300–400 °C above the ambient eclogite facies conditions. Whether decompression melting can occur in the brief moment that pressures drop along a fault surface is a question of kinetics. Again, the analogy with impact-related S-type pseudotachylytes may be instructive; in these settings, decompression melting occurs in microseconds (Melosh, 1989; Spray and Thompson, 1995). In any case, the very large fluctuations in physical conditions that likely occur at sites of pseudotachylyte formation suggest that we should be alert to the possible role of second-order, dynamic feedback processes in seismic faulting and pseudotachylyte genesis.

5. Summary and conclusions

Existing quantitative models for frictional heat generation in fault zones (McKenzie and Brune, 1972; Sibson, 1975; Philpotts, 1990; Spray, 1995) strictly apply only up to the moment of onset of melting and predict volumes of frictional melt that are significantly smaller than those observed in many natural fault zones. This suggests that previously unrecognized dynamic feedback processes may facilitate melting during seismic slip and that more consideration should be given to transient physical effects that may occur throughout the process of pseudotachylyte formation. Rapid melt migration and flash decompression melting at dilational jogs are effects that should be explored further, both theoretically and observationally. Careful attention to cross-cutting relationships, at both the meso- and microscopic scales, may reveal information about the relative sequence and rates of processes that occur in the brief duration of a seismic event (Bjørnerud, 1997). Such considerations may, moreover, reveal unsuspected similarities between the mechanisms of pseudotachylyte formation in tectonic and impact settings.

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References

Austrheim, H., 1987. Eclogitisation of lower crustal granulites by fluid migration through shear zones. Earth and Planetary Science Letters 81, 221–232.

Austrheim, H., Boundy, T., 1994. Pseudotachylytes generated during seismic faulting and eclogitization of the deep crust. Science 265, 82–83.

Berlenbach, J., Roering, C., 1992. Sheath-fold-like structures in pseudotachylytes. Journal of Structural Geology 14, 847–856.

Bjørnerud, M., 1997. Superimposed deformation in seconds: breccias from the impact structure at Kentland, Indiana (USA). Tectonophysics 290, 259–269.

Bjørnerud, M., Austrheim, H., Lund, M., 2002. Processes leading to densification (eclogitization) in the deep crust. Journal of Geophysical Research, 107 (Bio), 2252. Doi:10.1029/2001JB000527.

Camacho, A., Vernon, R., FitzGerald, J., 1995. Large volumes of anhydrous pseudo-tachylyte in the Woodroffe Thrust, eastern Musgrave Ranges, Australia. Journal of Structural Geology 17, 371–383.

Cardwell, R.K., Chinn, D.S., Moore, G.F., Turcotte, D.L., 1978. Frictional heating on a fault zone with finite thickness. Geophysical Journal of the Royal Astronomical Society 52, 525–530.

Curewitz, D., Karson, J., 1999. Ultracataclasis, sintering and frictional melting in pseudo-tachylytes from East Greenland. Journal of Structural Geology 21, 1693–1713.

Fabbri, O., Lin, A., Tokushige, H., 2000. Coeval formation of cataclasite and pseudo-tachylyte in a Miocene forearc granodiorite, southern Kyushu, Japan. Journal of Structural Geology 22, 1015–1025.

Gibson, R., Reimold, W.U., Wallmach, T., 1997. Origin of pseudotachylyte in the lower Witwatersrand Supergroup, Vredefort Dome (South Africa): constraints from metamorphic studies. Tectonophysics 283, 241–262.

Goodwin, L., 1999. Controls on pseudotachylyte formation during tectonic exhumation in the South Mountains metamorphic core complex, Arizona, in: Ring, U., Brandon, M., Lister, G., Willett, S. (Eds.), Exhumation Processes; Normal Faulting, Ductile Flow and Erosion Geological Society of London Special Publications, 154, pp. 325–342.

Grocott, J., 1981. Fracture geometry of pseudotachylyte generation zones. Journal of Structural Geology 3, 169–178.

Kanamori, H., Anderson, D., Heaton, T., 1998. Frictional melting during the 1994 Bolivian earthquake. Science 279, 839–842.

Lin, A., 1999. Roundness of clasts in pseudotachylytes and cataclastic rocks as an indicator of frictional melting. Journal of Structural Geology 21, 473–478.

Magloughlin, J., 1989. The nature and significance of pseudotachylyte from the Nason terrane, North Cascade Mountains, Washington. Journal of Structural Geology 11, 907–917.

Magloughlin, J., 1992. Microstructural and chemical changes associated with cataclasis and friction melting at shallow crustal levels: the cataclasite–pseudotachylyte connection. Tectonophysics 204, 243–260.

Magloughlin, J., Spray, J., 1992. Frictional melting processes and products in geological materials; introduction and discussion. Tectonophysics 204, 197–204.

Martini, J., 1978. Coesite and stishovite in the Vredefort Dome, South Africa. Nature 272, 1017–1049.

McKenzie, D., Brune, J., 1972. Melting on fault planes during large earthquakes. Geophysical Journal of the Royal Astronomical Society 29, 65–78.

McNulty, B, 1995. Pseudotachylyte generated in the semi-brittle and brittle regimes, Bench Canyon shear zone, central Sierra Nevada. Journal of Structural Geology 17, 1507–1521.

Melosh, H., 1989. Impact Cratering: A Geologic Process. Oxford University Press.

Nelson, S., Montana, A., 1992. Sieve-textured plagioclase in volcanic rocks produced by rapid decompression. American Mineralogist 77, 1242–1249.

O'Hara, K., 2001. A pseudotachylyte geothermometer. Journal of Structural Geology 23, 1345–1358.

Philpotts, A., 1990. Principles of Igneous and Metamorphic Petrology. Prentice Hall, Englewood Cliff, New Jersey, pp. 405–406.

Rosenberg, C., Handy, M., 2000. Syntectonic melt pathways during simple shearing of a partially molten rock analogue (norcamphor–benzemide). Journal of Geophysical Research 105, 3135–3149.

Scholz, C., 1980. Shear heating and the state of stress on faults. Journal of Geophysical Research 85, 6174–6184.

Scholz, C, 1990. The mechanics of earthquakes and faulting. Cambridge University Press.

Sibson, R., 1975. Generation of pseudotachylyte by ancient seismic faulting. Geophysical Journal of the Royal Astronomical Society 43, 775–794.

Sibson, R., 1977. Kinetic shear resistance, fluid pressures and radiation efficiency during seismic faulting. Pure and Applied Geophysics 115, 387–400.

Sibson, R, 1989. Earthquake faulting as a structural process. Journal of Structural Geology 11, 1–14.

Spray, J., 1987. Artificial generation of pseudotachylyte using a friction welding apparatus: simulation of melting on a fault plane. Journal of Structural Geology 9, 49–60.

Spray, J., 1992. A physical basis for the frictional melting of some rock-forming minerals. Tectonophysics 204, 205–221.

Spray, J., 1995. Pseudotachylyte controversy: fact or friction? Geology 23, 1119–1122.

Spray, J., Thompson, L., 1995. Friction melt distribution in a multi-ring impact basin. Nature 373, 130–132.

Swanson, M., 1992. Fault structure, wear mechanisms and rupture processes in pseutachylyte generation. Tectonophysics 204, 223–242.

Thompson, R., 1972. Melting behavior of two Snake River lavas at pressures up to 35 kb. Carnegie Institute of Washington Geophysical Laboratory Yearbook 71, 406–410.

Tsuchiyama, A., 1985. Dissolution kinetics of plagioclase in the melt system diopside–albite–anorthite, and origin of dusty plagioclase in andesites. Contributions to Mineralogy and Petrology 89, 1–16.

Tsutsumi, A., Shimamoto, T., 1996. Frictional properties of monzodiorite and gabbro during seismogenic fault motion. Journal of Geological Society of Japan 102, 240–248.

Turcotte, D., Schubert, G., 2002. Geodynamics, 2nd ed Cambridge University Press.

Wenk, H., 1978. Are pseudotachylites products of fracture or fusion? Geology 6, 507–511.

Wenk, H.-R., Johnson, L.R., Ratschbacher, L., 2000. A large pseudotachylite zone in the Eastern Peninsular Ranges of California. Tectonophysics 321, 253–277.