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Research note

Frictional heating on a fault zone with finite thickness

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Summary. The problem of frictional heating on a fault of finite thickness is considered. Temperature distributions during and after faulting are obtained for faults of various thicknesses. A thick fault approximation is shown to be valid if a fault has a thickness greater than about 1 cm. The thickness of the melted zone is predicted for various frictional stress levels and displacements. The predicted thicknesses are shown to be in reasonable agreement with field observations.

Introduction

Most earthquakes are the result of slip on pre-existing faults. Field studies show that faults range from relatively clean fractures to wide zones of brecciated rock (fault gouge). Theoretical studies suggest that displacements during faulting are unlikely to overshoot significantly the final static displacement (Richards 1976). This indicates that the frictional stress transmitted across the fault during slip must be a sizeable fraction of the stress transmitted across the fault prior to faulting. The frictional stress acting on the fault during slip results in frictional heating. This heating occurs across some fraction of the width of the fault zone and is dissipated by conduction. Because the thermal conductivity of rock is low and the time increment over which faulting occurs is small, high temperatures can be generated.

Field studies indicate that melting has occurred on some faults. Zones of glass-rich rock with evidence of flow (pseudotachylite or hyalomylonite) have been found in a number of cases (Scott & Drever 1953; Philpotts 1964; Ermanovics, Helmstead & Plant 1972; Masch 1973; Sibson 1975; Wallace 1976). However, most faults show no evidence of melting or high temperatures. Melting has also occurred during frictional sliding experiments on sandstone and limestone in the laboratory (Friedman, Logan & Regent 1974; Teufel & Logan 1976).

An analysis of the problem of frictional heating on a fault has been carried out by McKenzie & Brune (1972) and Richards (1976). These authors obtained solutions for planar faults of zero width. It is the principal purpose of this paper to provide solutions for fault

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zones of finite width. We will show that high temperatures will be generated by frictional heating only if the fault zone is relatively thin.

Analysis

Our formulation of the problem follows that given by McKenzie & Brune (1972). The equation for the one-dimensional, unsteady transport of heat is

$$\rho c_p \frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial x^2} + Q(x, t) \tag{1}$$

where ρ is density, c_p specific heat at constant pressure, k thermal conductivity, and Q the heat generation per unit volume and time.

We assume that $T = T_0$ and $Q = 0$ for $t < 0$ prior to faulting. The appropriate solution of (1) is (Morse & Feshbach 1953)

$$T(x, t) = T_0 + \frac{1}{2\rho c_p \sqrt{\pi\kappa}} \int_0^t \int_{-\infty}^{\infty} \exp\left[\frac{-(x - x_0)^2}{4\kappa(t - t_0)}\right] \frac{Q(x_0, t_0)}{(t - t_0)^{1/2}} \cdot dx_0 dt_0 \tag{2}$$

where $\kappa = k/\rho c_p$ is the thermal diffusivity. We further assume that the fractional heating is uniform across the width w of the fault and is independent of time during the faulting

$$\begin{aligned} Q(x_0, t_0) &= \frac{\sigma_f D}{w\tau}\left[H\left(x_0 + \frac{w}{2}\right) - H\left(x_0 - \frac{w}{2}\right)\right], \quad 0 < t < \tau \\ &= 0 \quad t_0 < 0, \quad t_0 > \tau \end{aligned} \tag{3}$$

where w is the width of the fault, D the displacement on the fault, τ the time of faulting, and H the Heavyside step function. Substitution of (3) into (2) and integration of the space coordinate yields

$$\begin{aligned} T &= T_0 + \frac{\sigma_f D}{2\rho c_p w\tau} \int_0^t \left\{\operatorname{erf}\left[\frac{x + (w/2)}{(4\kappa[t - t_0])^{1/2}}\right] - \operatorname{erf}\left[\frac{x - (w/2)}{(4\kappa[t - t_0])^{1/2}}\right]\right\} dt_0, \quad 0 < t < \tau \\ &= T_0 + \frac{\sigma_f D}{2\rho c_p w\tau} \int_0^\tau \left\{\operatorname{erf}\left[\frac{x + (w/2)}{(4\kappa[t - t_0])^{1/2}}\right] - \operatorname{erf}\left[\frac{x - (w/2)}{(4\kappa[t - t_0])^{1/2}}\right]\right\} dt_0, \quad t > \tau \end{aligned} \tag{4}$$

and the integrals must be evaluated numerically.

A characteristic length for heat conduction is given by $(\kappa\tau)^{1/2}$. This is the distance heat is conducted during faulting. If $w/(\kappa\tau)^{1/2} \ll 1$, the maximum temperature increase at $t = \tau$ and $x = 0$ from (4) is

$$T_{m1} - T_0 = \frac{\sigma_f D}{\rho c_p (\pi\kappa\tau)^{1/2}}. \tag{5}$$

This is the maximum temperature increase on a fault of zero width and is identical to the result obtained by McKenzie & Brune (1972).

If $w/(\kappa\tau)^{1/2} \gg 1$, the maximum temperature increase at $t = \tau$ and $x = 0$ from (4) is

$$T_{m2} - T_0 = \frac{\sigma_f D}{\rho c_p w}. \tag{6}$$

In this thick fault limit thermal conduction is not important during faulting and the frictional heat is uniformly distributed across the width of the fault.

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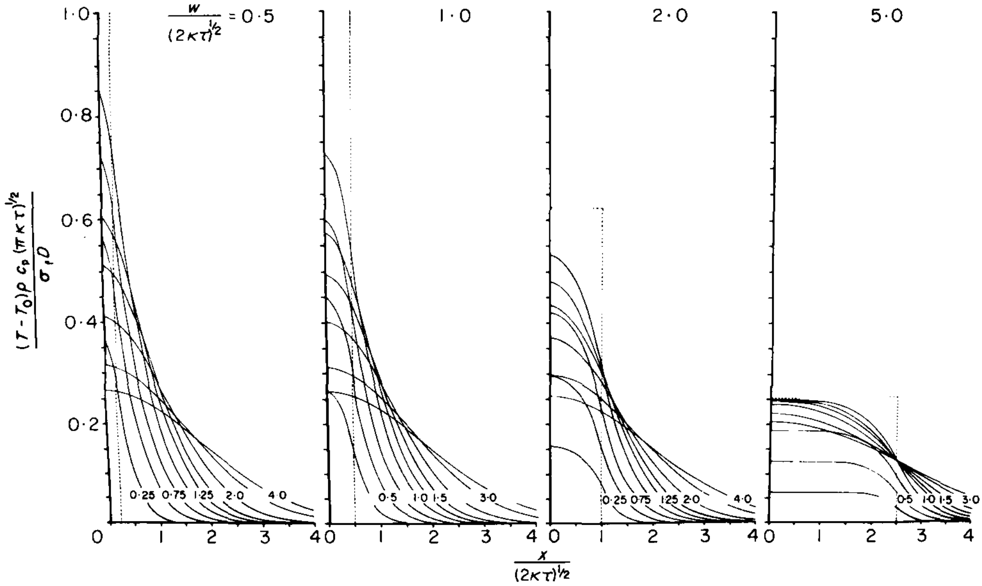

Figure 1. Dependence of the non-dimensional temperature on the distance from the centre of the fault at various values of the non-dimensional time t/τ; results are given for four values of the non-dimensional fault width. The vertical dashed line is the edge of the fault; the horizontal dashed line is the maximum temperature predicted by (6).

In order to illustrate this transition from thin to thick fault behaviour, a series of temperature profiles obtained by integrating (4) are given in Fig. 1. Non-dimensional temperatures are given as a function of the non-dimensional distance from the centre of the fault for several non-dimensional times; results are given for $w/(2\kappa\tau)^{1/2} = 0.5, 1, 2, 5$. The temperatures are given as a fraction of the maximum temperature increase for a fault of zero thickness, as given in (5). The finite thickness of the fault reduces the maximum temperature. The horizontal dashed lines show the maximum temperatures predicted for a thick fault by (6). Thermal conduction during faulting reduces the maximum temperature below the values given by (6). Therefore, (5) and (6) provide upper limits to the temperatures that can be expected on faults of finite width.

The distance from the fault is non-dimensionalized with the conduction distance $(2\kappa\tau)^{1/2}$. The half width of the fault is shown by the vertical dashed lines. The influence of the finite fault width is evident for the case $w/(2\kappa\tau)^{1/2} = 5$ where near uniform heating is occurring.

A reasonable upper limit for the time of faulting is 1 s. Taking $\kappa = 10^{-2}\,\mathrm{cm^2/s}$ we find that the conduction distance is $(2\kappa\tau)^{1/2} = 0.14$ cm. For fault widths greater than about 1 cm, (6) should give the maximum temperature on the fault. In Fig. 2, the fault widths for various temperature increases are given as a function of $\sigma_f D$. We have taken $\rho = 2.8\,\mathrm{gm/cm^3}$ and $c_p = 0.25$ cal/gm K. The value of $\sigma_f D = 10^{12}$ dyne/cm could correspond to a stress $\sigma_f = 1$ kbar and a displacement $D = 10$ m, that is a very large earthquake. The value of $\sigma_f D = 10^{10}$ dyne/cm could correspond to $\sigma_f = 100$ bar and $D = 1$ m. A conclusion from Fig. 2 is that even the largest earthquake would not produce a melt zone wider than about a metre.

Clearly we have made a number of serious approximations. The first is that the heating is uniform across the fault zone. However, the limits of heating on a plane and uniform heating across a finite width provide the two limits between which other heating models would lie. We have not specified the width of the heating zone. In general it is likely to be

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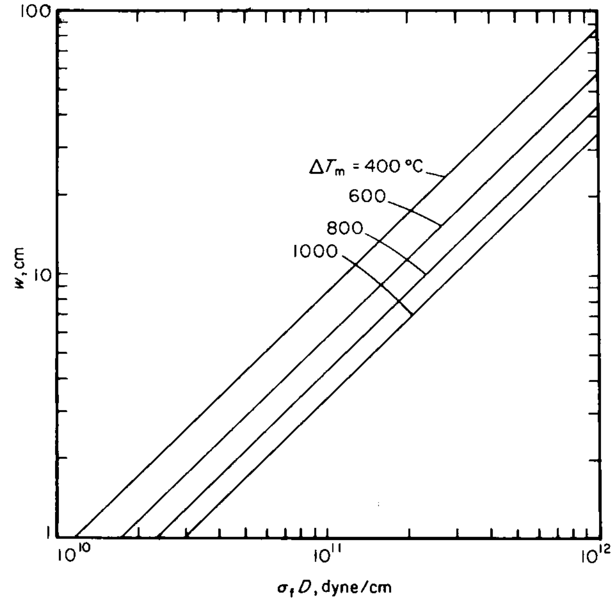

Figure 2. Fault widths as a function of $\sigma_f D$ for various values of the maximum temperature increase.

some fraction of the relatively wide observed mylonite zones. Laboratory experiments (Engelder, Logan & Handin 1975) indicate that slip zones are thin even in the presence of fault gauge. However, it is not clear whether these results are directly applicable to actual faults. Also, the heat of fusion has not been considered so that the degree of partial melting cannot be predicted. Our assumption that the heating is independent of time can also be questioned. However, the uncertainties in specifying the time of faulting τ are such that it is of doubtful value to utilize more sophisticated models of the dynamic behaviour of the fault such as that given by Richards (1976).

Field observations

Zones of fused rock associated with fault heating are generally observed within relatively wide mylonite zones in metamorphic terranes. Usually there is a gradation from glassy rock through mylonite into undeformed rock (Scott & Drever 1953). The glassy (melted) rock occurs either as the matrix of breccias or as thin, black veinlets. The veinlets are usually 1–3 cm in width but have been observed to be as wide as 10–20 cm (Scott & Drever 1953; Sibson 1975, 1977). These maximum widths are in good agreement with the widths which will yield melting as obtained above.

The intrusive veinlets may be parallel or may cross-cut the metamorphic foliation of the host rock and show no trace of the fabric of the host rock. Veins may be scattered through the fault zone, and in one case are observed several metres from the fault zone (Philpotts 1964). When observed as a matrix the glass may contain microlites of feldspar, minute bubbles and vesicles, spherulites, amygdules, or partially melted rock and mineral fragments (Higgins 1971; Sibson 1975). The rock fragments may be cataclastic or undeformed. The rock fragments in the Himalayas observed by Scott & Drever (1953) consisted of opaque ultramylonite containing quartz and feldspar xenoclasts and subparallel veins of light and medium brown glass 1–2 mm wide. The clearer glass commonly occurred as fine stringers drawn out into the darker glass.

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Most of the occurrences of fused rock associated with fault zones are found in metamorphic terranes. The metamorphic grade ranges from greenschist facies, e.g. the Moine and outer Hebrides thrust zones (Christie 1963; Sibson 1975), to much higher grades, e.g. the Colorado Front Range (Abbot 1972) and the Lewisian basement of the Scottish Highlands (Park 1961). Greenschist metamorphism requires a minimum pressure of 1.5 kbar and minimum temperature of 250°C (Turner 1968). Theodore (1970) estimated pressures of 3.4–7 kb (11–23 km depth) and initial temperatures greater than 580°C during mylonitization of rocks in the Peninsular Ranges of southern California. Philpotts (1964) suggested that the mylonite rocks on Quebec were at a temperature of 400°C prior to faulting and the generation of glassy veins. Field evidence and laboratory studies indicate that these high temperatures were synchronous with faulting.

The temperature at which rocks begin to melt during frictional sliding is difficult to determine. The effects of pressure and water on this type of melting are not well known, and the data are still not conclusive as to whether the rocks undergo total or partial melting. Scott & Drever (1953) concluded that hylomylonite in the Himalayas is produced by differential melting with fluxing by iron. They suggested that the temperature of melting is determined by the constituents that have low melting points. Wallace (1976) believes that this is not necessarily true, and that glass is enriched in a granitic fraction and probably formed by fusion of feldspar xenocrysts. The data of Philpotts (1964) indicate partial melting, which Masch (1973) and Sibson (1975) showed that total melting has occurred. Wallace (1976) suggests that unless the initial melt is able to lubricate the fault plane, frictional heating will continue to increase the temperature above the melting point of most of the mineral constituents. This may explain why in some localities partial melting is observed, while total melting is observed in other localities. Ermanovics *et al.* (1972) suggested that ultramylonites result from the mechanical granulation of the host rock, and that fusion will occur if sufficient water is present, while Sibson (1975) believes that melting only occurs in the absence of interstitial water. Wallace (1976) suggests on the basis of chemical properties of the glass associated with the Alpine fault zone in New Zealand and comparisons with experimental work that the melting of these rocks should have taken place at about 750°C.

It is clear from the above observations that relatively low temperature increases may cause melting on faults. With relatively high ambient temperatures (approximately 400°C) and relatively low melt temperatures (approximately 800°C), a temperature increase of 400°C may cause melting. Our calculations (see Fig. 2) show that partial melting should be expected on faults zones 1–10 cm wide during moderately large earthquakes.

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