Frictional Melting During the Rupture of the 1994 Bolivian Earthquake

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The source parameters of the 1994 Bolivian earthquake (magnitude M_w = 8.3) suggest that the maximum seismic efficiency η was 0.036 and the minimum frictional stress was 550 bars. Thus, the source process was dissipative, which is consistent with the observed slow rupture speed, only 20% of the local *S*-wave velocity. The amount of nonradiated energy produced during the Bolivian rupture was comparable to, or larger than, the thermal energy of the 1980 Mount St. Helens eruption and was sufficient to have melted a layer as thick as 31 centimeters. Once rupture was initiated, melting could occur, which reduces friction and promotes fault slip.

The possibility of frictional melting during faulting has been suggested by several investigators (*1–4*). McKenzie and Brune (*2*) quantitatively investigated this problem as a one-dimensional heat conduction problem. They assumed that the fault surface is simultaneously heated during slippage (that is, infinite rupture speed) over a finite time. The temperature was determined mainly by generation of heat (the integrated product of slip and frictional stress) and diffusion of heat (controlled by duration of the heating event and thermal diffusivity). The duration of heating was determined by the driving stress on the fault. If the driving stress was lower, the heating process was slower, allowing heat to diffuse over a larger distance away from the fault; this case results in less temperature rise. They concluded that if the frictional and driving stresses are of the order of 1 kbar, melting could occur for fault slips as small as 1 mm. Richards (*4*) solved elastodynamic equations for a propagating elliptical crack and estimated the frictional heating rate behind a rupture front. He showed that if the driving stress is 100 bars and the fault particle velocity is 10 cm/s at the time of rupture nucleation, a temperature rise of about 1000°C can occur within a few seconds at a point halfway between the rupture front and the point of nucleation. These studies indicate that frictional melting can occur if the stresses involved in faulting are sufficiently high. Despite these studies, frictional melting is not generally regarded as an important process during earthquake faulting because of uncertainties in the stress levels—especially the magnitude of frictional stresses—associated with earthquakes and uncertainties in the detailed fault-zone structures. Sibson (*3*) noted that production of pseudotachylyte (glassy material presumably formed by frictional melting) should take place during faulting, but very few faults contain pseudo-tachylytes.

The 9 June 1994 M_w = 8.3 Bolivian event (13.86°S, 67.54°W; depth = 637 km) was a large deep-focus earthquake (source parameters shown in Table 1). Although the results obtained by different studies vary in detail, the values of most source parameters

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agree. The seismic moment was $M_0 = 3 \times 10^{21}$ N·m (the median of the values listed in Table 1). The main shock consisted of several discrete events, here called subevents. The rupture area S was determined from the spatial spreads of all of the subevents, and the estimates range from 1000 to 2500 km^2; the range in S is mainly due to the difference in how the outer boundary of the subevent distribution was delineated. Here we used the estimate obtained by Kikuchi and Kanamori (*5*), $S = 40$ km $\times$ 40 km, which is close to the middle of the range of S listed in Table 1. If we use only the area where the displacement was larger than 10 m, S can be smaller (15 km $\times$ 30 km) (*6*). The spatial extent of the aftershock area was about 50 km (*7*), the same order as the rupture dimension estimated from waveform analysis. The source duration was about $\tau = 40$ s. The variation of the rupture speed V reflects the different estimates of speed for different subevent pairs. Because the details of the rupture pattern could not be determined, the rupture speed was defined by the ratio of the distance between the subevents to the time difference between them. If we define the average rupture speed by the ratio of the total rupture length (about 40 km) to the rupture time (about 40 s), we obtain $V = 1$ km/s (Table 1). The estimated rupture speed of Chen (*8*) was high, 3 to 3.5 km/s, because he assumed that the subevents were distributed en echelon. Here we use $V = 1$ km/s, which corresponds to $V/\beta = 0.2$, where β is the local S-wave velocity. From these parameters, the average fault offset is estimated to be $D = 15.6$ m, and the static stress drop $\Delta\sigma = 1.14$ kbar (*9*).

Another important source parameter is the radiated wave energy E_R. For shallow earthquakes, the determination of E_R is uncertain because of complex wave propagation effects, but for deep-focus earthquakes, it can be determined relatively accurately. Winslow and Ruff (*10*) determined the radiated energy to be $E_R = 5.2 \times 10^{16}$ J by integrating the observed velocity records. This value is in good agreement with the previously reported values of 3.2×10^{16} J (*11*) and 5.2×10^{16} J (*12*).

Another notable feature of the Bolivian earthquake is the lack of any isotropic component (net volume change) over long periods. The isotropic component was insignificant, probably less than 2 to 4% of the double couple component (*5, 13, 14*). This value corresponds to a maximum volume change of 7.5×10^7 to 1.5×10^8 m^3, or a maximum fault-normal displacement of 12 to 24 cm.

The mechanism of deep-focus earthquake has been a matter of significant interest for many years. Because of the high pressure and temperature in the source region, the ordinary brittle failure is not likely to occur, and other mechanisms need to be invoked. Recently, Green and Burnley (*15*) and Kirby *et al.* (*16*) proposed transformational faulting as a plausible mechanism of deep-focus earthquakes. In this case, a phase transition triggers a rupture, but the main rupture occurs under the ambient shear stress. Also, rupture may nucleate on a weak zone established in a slab at the surface (*17*). Here we assume that some triggering mechanism initiated the Bolivia earthquake and address the question of what happened after the rupture began.

We first investigate the energy budget of the earthquake using only the quantities determined from seismic observations. The total potential-energy (strain energy plus gravitational energy) change W can be written as (*18*)

$$W = (\sigma_0 + \sigma_1)DS/2 = \Delta\sigma DS/2 + \sigma_1 DS \quad (1)$$

where σ_0 and σ_1 are the initial and the final stresses on the fault plane, respectively, and $\Delta\sigma = \sigma_0 - \sigma_1$ is the static stress drop.

We define seismic efficiency η by

$$\eta = (W - H_f)/W = E_R/W \quad (2)$$

where $H_f = \sigma_f SD$ is the frictional energy loss during faulting and σ_f is the average frictional stress (*19*). Using the seismic moment $M_0 = \mu DS$, where μ is the rigidity, we can rewrite Eqs. 1 and 2 as $\eta = \mu(E_R/M_0)/(\sigma_1 + \Delta\sigma/2)$ and $\sigma_f = \mu(1 - \eta)(E_R/M_0)/\eta$.

For any physically reasonable stress release mechanism, $\sigma_1 \geq 0$. With this constraint, an upper bound for η is given by

$$\eta_{max} = 2(E_R/M_0)\mu/\Delta\sigma \quad (3)$$

and a lower bound for σ_f is given by

$$\sigma_{fmin} = (1-\eta_{max})\Delta\sigma/2 \quad (4)$$

Using $E_R = 5.2 \times 10^{16}$ J (*10–12*), $\mu = 1.2$ Mbar, and $\Delta\sigma = 1.14$ kbar for the Bolivian earthquake, we obtain $\eta_{max} = 0.036$ and $\sigma_{fmin} = 550$ bars. From η_{max}, the lower bound of the nonradiated energy can be estimated as $H_{fmin} = W - E_R = E_R(1 - \eta_{max})/\eta_{max} = 1.4 \times 10^{18}$ J. For comparison, the total thermal energy released during the 1980 eruption of Mount St. Helens was about 10^{17} to 10^{18} J (*20*). Although this amount of heat does not significantly contribute to the global heat flow, it can influence the local thermal budget in subduction zones.

To estimate the overall thermal state in the focal region of the Bolivian earthquake, we consider the gross energy budget. If $H_f = \sigma_f SD$ is used to raise the temperature of a thin layer with thickness Δd, then the temperature increase ΔT over the volume $S\Delta d$ is given by

Table 1. Seismic moment M_0, rupture area S, source duration τ, and rupture speed V of the 1994 Bolivian earthquake as determined in various studies. The first moment is from the Harvard CMT solution.

M_0 (10^{21} N·m)	S (km^2)	τ (s)	V (km/s)	Ref.
3.0				CMT
3.0	40 × 40*	40	1	(*5*)
	30 × 50	40	1 to 2.6	(*17*)
3.0		46		(*14*)
2.6	50 × 20†	40	1	(*34*)
3.0	30 × 40†	40	2 to 3‡	(*35*)
2.7		50		(*36*)
	30 × 50	40	1 to 2	(*37*)
		40	3 to 3.5§	(*8*)
3.1				(*13*)
	50 × 50	40		(*38*)
	60 × 40‖		<2	(*6*)

*The area of main energy release is even smaller. †Area of main energy release. ‡Maximum rupture speed. §Rupture speed within en echelon subevents. The apparent rupture speed is 1 to 2 km/s. ‖Total area of rupture.

$$\Delta T = \sigma_f SD/C\Delta dS\rho \quad (5)$$

if no melting occurs. Here, ρ is the density and C is the heat capacity of the bulk rock. If melting occurs at temperature T_m, then $C\Delta T$ should be replaced by $C\Delta T_m + H + C_L(T - T_m)$, where ΔT_m (estimated as 600°C) is the temperature difference between the melting temperature and the ambient temperature in the slab (*21*), C_L is the heat capacity for the liquid phase, and H is the latent heat.

The thickness Δd depends on how the heat is generated. If faulting occurs on an infinitesimally thin layer and heat diffuses out from there, Δd will be of the order of the thermal penetration depth given by $\Delta d = (k\tau_d)^{1/2}$, where τ_d is the time scale of faulting and k is the thermal diffusivity. If we choose $\tau_d = 10$ s as a characteristic local time scale of faulting for the Bolivian earthquake and $k = 1.35 \times 10^{-2}$ cm^2/s, then $\Delta d = 3.7$ mm (*22*). If faulting occurs on a distributed fault zone or in shear bands, Δd can be larger. The resulting ΔT from Eq. 5 is shown in Fig. 1. In this computation, H and the difference between C and C_L are ignored. We find that melting is likely to occur in some region near the fault zone if the slip zone is over a thickness of less than 31 cm (indicated by a dash-dot line at $\Delta T = 600$°C in Fig. 1). If the thermal penetration depth, $\Delta d = 3.7$ mm, is used, the local temperature rise is of the order of 52,000°C (indicated by a dash-dot line in Fig. 1). Figure 1 also compares the thermal energy density $H_f/\Delta dS\rho$ in the slip zone with the latent heat for melting, enthalpy for phase changes (because both are about the same on this scale, the range of ΔH for both is indicated by a dash-dot line at about

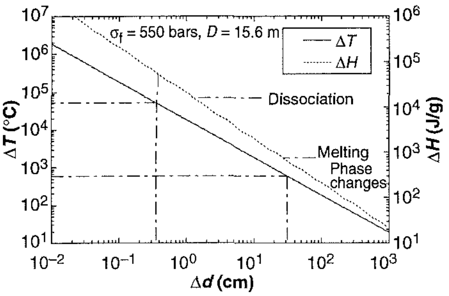

Fig. 1. Changes in temperature ΔT and enthalpy ΔH due to frictional heating as a function of the thickness of the slip zone Δd (σ_f = 550 bars and D = 15.6 m). The density ρ = 4 g/cm^3 for the depth of 637 km was used. The value of C (1.14 J/g°C) used to calculate ΔT, and the values of ΔH of melting, phase changes, and dissociation for Mg_2SiO_4 are taken from (*33*).

$\Delta H = 10^3$ J/g in Fig. 1), and dissociation (to elements) energy (indicated by a dash-dot line at about $\Delta H = 1.5 \times 10^4$ J/g in Fig. 1) of Mg_2SiO_4. If Δd is less than a few millimeters, the available energy is sufficient to dissociate olivine to elements, and the material will be in a plasma-like state.

The thickness Δd cannot be determined directly from seismological data, but weakening as a result of melting is likely to localize deformation on a thin zone, as is seen in pseudo-tachylytes. The small upper bound for the fault-normal displacement (*5*, *13*, *14*) also suggests a fairly simple dislocation source, and a large complex volumetric source is probably ruled out. Thus, Δd as small as a few millimeters is plausible.

The slow rupture speed $V = 0.2\beta$ observed for the Bolivian earthquake is an independent observation that also suggests a dissipative rupture process. We discuss the implication of slow rupture speed using a simple crack theory. For Mode III cracks (antiplane shear crack), Kostrov (*23*) and Eshelby (*24*) showed that the energy flux to the crack tip (energy release rate) G is given by

$$G = g(V)G^* \tag{6}$$

where G^* is the value of G computed for a quasi-static situation, and $g(V)$ is a universal function of V. Eshelby gives

$$g(V) = [(1 - V/\beta)/(1 + V/\beta)]^{1/2} \tag{7}$$

The equation of motion for rupture propagation can be obtained by equating G to 2γ, where γ is the surface energy. If γ is independent of V, V increases from 0 to β as the crack length increases. Thus, the limiting rupture speed is β, which does not fit the characteristics of the Bolivian earthquake. However, if γ increases with V as a result of extensive plastic deformation near the crack tip, as experimentally demonstrated by Rosakis and Zehnder (*25*), V can be significantly lower than the S-wave velocity.

In the above, frictional energy loss is not explicitly included, but during complex rupture propagation that involves large plastic deformations near the crack tip, distinction between surface energy and frictional energy is not obvious, especially if rupture propagation is slip-like (*26*). If we include frictional energy loss in the surface energy (*27*), then the efficiency η is given by $(G^* - G)/G^*$, and using Eqs. 6 and 7, we obtain

$$\eta = 1 - g(V) = 1 - [(1 - V/\beta)/(1 + V/\beta)]^{1/2} \tag{8}$$

The exact relation between η and V/β depends on the specific crack model. If we use the energy-based crack model of Mott (*28*), the kinetic energy is proportional to V^2, and a relation

$$\eta = (V/\beta)^2 \tag{9}$$

is inferred instead of Eq. 8. Because V/β = 0.2 for the Bolivian earthquake, we obtain η = 0.18 and 0.04 from Eqs. 8 and 9, respectively. These values are consistent with the low η_{max} estimated from seismic data for the Bolivian earthquake.

The molten layer can be very thin compared with the dimension of the fault plane. Then a question may be raised as to whether such a thin layer can have significant effects on sliding friction or not. In this connection, the effect of a thin molten layer on ski sliding, studied by Bowden and Hughes (*29*), provides an interesting analog of frictional melting during earthquake faulting. Bowden and Hughes experimentally demonstrated that a sliding ski can produce a thin film of water 10^{-2} cm thick or less and promote sliding with a low coefficient of friction μ_f = 0.03.

It is unclear whether the Bolivian earthquake is fundamentally different from other deep-focus earthquakes. The ratio of E_R to M_0 is generally low for most deep-focus earthquakes (*10*, *30*), suggesting a fairly dissipative rupture mechanism. However, because of the limited resolution of determination of source dimensions, the static stress drop and rupture speed for smaller deep-focus earthquakes are not well-determined. No evidence for slow rupture speed has been found for other deep-focus earthquakes, with the possible exception of the equally large 1970 Colombia earthquake (M_w = 8.2), for which a rupture speed of 1 to 3.2 km/s was suggested (*31*). Because of these uncertainties, it is unclear whether melting plays a major role in other deep-focus earthquakes. Deep-focus earthquakes may be different from event to event (*32*). It is possible, however, that when the slip caused by some triggering mechanisms exceeds a threshold, melting occurs and promotes extensive sliding, which results in an exceptionally large deep-focus earthquake, such as the 1994 Bolivian event. As the quality and quantity of seismic data improve, the accuracy of source-parameter determinations will improve, so that we may eventually be able to resolve this question.

For most shallow earthquakes, the ratio of the rupture speed to the average crustal S-wave velocity is about 0.7 to 0.8. Considering the relatively low S-wave velocity near the fault zone and the various rupture modes in faulting, this ratio may represent an even higher ratio of rupture speed to the limiting velocity (for example, Rayleigh wave velocity) appropriate for a given rupture mode. If the mechanism of friction for shallow earthquakes is similar to that of the Bolivian earthquake, then the high ratio for shallow earthquakes may suggest a relatively nondissipative brittle rupture process, that is, faulting with low frictional stress. Resolution of this problem, however, requires further studies on the mechanism of friction during seismic rupture.

REFERENCES AND NOTES

1. H. Jeffreys, *Geol. Mag.* **79**, 291 (1942).
2. D. P. McKenzie and J. N. Brune, *Geophys. J. R. Astron. Soc.* **29**, 65 (1972).
3. R. H. Sibson, *Nature* **243**, 66 (1973).
4. P. G. Richards, *Bull. Seismol. Soc. Am.* **66**, 1 (1976).
5. M. Kikuchi and H. Kanamori, *Geophys. Res. Lett.* **21**, 2341 (1994).
6. M. Antolik, D. Dreger, B. Romanowicz, *ibid.* **23**, 1589 (1996).
7. S. C. Myers *et al.*, *ibid.* **22**, 2269 (1995).
8. W.-P. Chen, *ibid.*, p. 2261.
9. The value of D is computed from $M_0/\mu S$, where μ = 1.2 Mbar for the depth of 637 km; $\Delta\sigma$ is computed as $\Delta\sigma = 7\pi\mu D/16r$, for a circular fault with radius r estimated by $r = (S/\pi)^{1/2}$.
10. N. Winslow and L. Ruff, in preparation.
11. *U.S. Geol. Surv. Earthquake Data Report EDR No. 6-94* (1994), p. 145.
12. H. Houston, *Eos* **77**, 491 (1996).
13. T. Hara, K. Kuge, H. Kawakatsu, *Geophys. Res. Lett.* **22**, 2265 (1995).
14. J. Wu, T. Wallace, S. Beck, *ibid.*, p. 2237.
15. H. W. Green and P. C. Burnley, *Nature* **341**, 733 (1989).
16. S. H. Kirby, W. B. Durham, L. A. Stern, *Science* **252**, 216 (1991).
17. P. G. Silver *et al.*, *ibid.* **268**, 69 (1995).
18. E. Orowan, *Geol. Soc. Am. Mem.* **79**, 323 (1960).
19. Here we use "friction" to mean the resistance against fault rupture; it includes the conventional interface friction and the resisting force at the crack tip. In this sense, σ_f is the effective frictional stress. The actual energy loss is the sum of interface frictional energy and fracture energy. We assume that these energies will eventually be converted into heat.
20. J. D. Friedman, G. R. Olhoeft, G. R. Johnson, D. Frank, in *The 1980 Eruption of Mount St. Helens, Washington*, P. L. Lipman and D. R. Mullineaux, Eds. (U.S. Geological Survey, Washington, DC, 1981), pp. 557–567.
21. The value of ΔT_m depends on many parameters, such as the age of subducting slab and the convergence rate. Here we estimate it as 600°C from the category A and B thermal models of S. H. Kirby, S. Stein, A. Okal, and D. C. Rubie [*Rev. Geophys.* **34**, 261 (1996)].
22. The characteristic time scale of local dislocation τ_d is

estimated from the rise time of the far-field waveforms from the papers used in Table 1. The thermal diffusivity for olivine is taken from C. Clauser and E. Huenges, in *A Handbook of Physical Constants: Mineral Physics and Crystallography*, T. J. Ahrens, Ed. (American Geophysical Union, Washington, DC, 1995), pp. 105–126.
23. B. V. Kostrov, *J. Appl. Math. Mech.* **30**, 1241 (1966).
24. J. D. Eshelby, *J. Mech. Phys. Solids* **17**, 177 (1969).
25. A. J. Rosakis and A. T. Zehnder, *Int. J. Fract.* **27**, 169 (1985)..
26. T. Heaton, *Phys. Earth Planet. Inter.* **64**, 1 (1990).
27. Implicit in this assumption is that the frictional loss and surface energy have the same dependence on *V*. If friction does not depend on *V*, then η cannot be directly related to *V*.
28. N. F. Mott, *Engineering* **165**, 16 (1948).
29. F. P. Bowden and T. P. Hughes, *Proc. R. Soc. London Ser. A* **172**, 280 (1939).
30. H. Houston and Q. Williams, *Nature* **352**, 520 (1991).
31. M. Furumoto, *Phys. Earth Planet. Inter.* **15**, 1 (1977).
32. D. A. Wiens and J. J. McGuire, *Geophys. Res. Lett.* **22**, 2245 (1995).
33. A. Navrotsky, in *A Handbook of Physical Constants: Mineral Physics and Crystallography*, T. J. Ahrens, Ed. (American Geophysical Union, Washington, DC, 1995), pp. 18–28.
34. P. Lundgren and D. Giardini, *ibid.*, p. 2241.
35. S. Goes and J. Ritsema, *ibid.*, p. 2249.
36. P. F. Ihmle and T. H. Jordan, *ibid.*, p. 2253.
37. S. L. Beck, P. Silver, T. C. Wallace, D. James, *ibid.*, p. 2257.
38. C. H. Estabrook and G. Bock, *ibid.*, p. 2277.
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