Dynamic fragmentation causes low rock-on-rock friction

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ABSTRACT: The low frictional resistance to rock-on-rock sliding reported in large blockslides, in co-seismic fault rupture and in laboratory-scale rock friction tests has been attributed to a variety of causes. Herein we propose a mechanical explanation for the reduced friction, which seems likely to be universally relevant to complement other mechanisms. Rock-on-rock sliding of intact brittle rocks always generates a layer of comminuted debris. Rock must fail in order to form and further comminute debris; at local strain rates $\gg 100 \text{ s}^{-1}$, recycling of elastic strain energy stored in accomplishing fragmentation generates instantaneous, local, GPa-range isotropic pressures similar to the rock's Hugoniot elastic limit (Q). Under rapid strain, simultaneously fragmenting grains deliver large normal forces to the boundaries of the comminuting layer, reducing the confining stress on the debris (and hence its resistance to shear), thus lowering the frictional resistance to slip. This behaviour corresponds quantitatively to published laboratory data on granite friction; to the dynamics of low-angle blocksliding and faulting; and to our data on rapid shearing of fragmenting dry coal.

1 INTRODUCTION

Anomalously low frictional resistance to the motion of rock masses has been reported or inferred from time to time in a number of geological situations. Lower-than-normal resistance to rock-on-rock slip is widely inferred in fault motion (e.g. Scholz, 2002; Parsons, 2002; Townend and Zoback, 2004), which involves shear in comminuted rock debris. Rapid, low-angle motion of large intact blockslides generates comminuted material at the sliding surface (Anders *et al*, 2000) and unequivocally requires low basal friction (Davies *et al*, 2006). Reduced slip resistance is also reported in laboratory-scale rock-on-rock sliding (Kilgore *et al*, 1993; Byerlee, 1978; Di Toro *et al*, 2004; Goldsby and Tullis, 2002).

Explanations of these phenomena have focussed on particular environments. For example, Di Toro *et al* (2004) attribute their dramatically low friction data to formation of weak silica gel in the presence of moisture on the laboratory rock surface, which is difficult to apply to the much larger-scale situation of blocksliding. Reduced resistance to slip in faulting is often explained (Lachenbruch, 1980; Sibson, 1994) with high-pressure pore water, which likewise cannot apply to the laboratory result.

Herein we explore a mechanism capable of reducing slip resistance in these situations; the intense

local pressure generated by dynamic failure of brittle rock reduces frictional resistance to motion by reducing the effective stress between the sliding surfaces, under a normal coefficient of friction.

2 DYNAMIC ROCK FRAGMENTATION

2.1 General description

In brittle rock sliding, wear of the sliding surfaces always occurs (Scholz, 2002); rock fragments accumulate between the sliding surfaces and further comminute, forming a debris layer. Rapid breakage of intact (i.e. unjointed) rock – (“dynamic fragmentation”: Grady and Kipp (1987) - requires that the rock is stressed beyond its dynamic (or Hugoniot) elastic limit (Q – the temperature-, strain-rate- and confinement-stress-related value of the compressive strength). When a particle deforms rapidly in a shear field, its internal elastic-strain energy increases; this energy is extracted locally from the surrounding kinetic energy field. The stress state of the deforming particle rapidly approaches, and then exceeds, its strength. It suddenly fails, breaking on newly-generated fracture surfaces defining smaller sub-fragments (Sagy *et al*, 2001; Zhang *et al*, 2000, 2004).

2.2 Energy transformations in fragmentation

As an intact clast subjected to compressional and confining stresses initially deforms elastically, the internal stress approaches the elastic limit Q of the intact rock material; this is a function of strain rate and confining stress. Immediately before Q is reached (somewhere in the clast), the elastic strain energy stored by the clast is SE_1 . When Q is exceeded somewhere in the clast, a set of cracks appears defining a set of sub-clasts a_1, a_2, a_n .

Development of a crack requires inter-molecular bonds to be broken; by definition this takes place at the stress Q , which is the areally-averaged strength of the bonds. The genesis of the crack is the crack walls moving apart by a perpendicular distance δy , so the work done in generating the crack $= Q\delta y$. This is exactly the same as the work that would be done in further straining the clast by δy in the absence of crack development, because it is done at the same stress. The total work done to cause failure of the clast from a previously-defined state is exactly the work done in straining it from that previous stress state.

Momentarily, each subclast has almost the same shape as their corresponding mass had prior to cracking; some energy is transformed to noise and heat during the creation of the cracks, so the elastic strain energy is reduced somewhat. Hence $SE_1 \sim \Sigma SE_a$; the elastic strain energy stored in the mass is only slightly reduced by crack appearance. Once through-going cracks exist, subclasts can move relative to each other under the applied stress field. Instantly, much remaining elastic strain energy converts to kinetic energy. The total released elastic strain energy is close to $Q^2/2E$ per unit volume (Herget, 1960). Displacement of clasts generates frictional resistance to relative subclast motion,

Fig. 1 Stress-strain curve for failing solid

dissipating some of the kinetic energy. The above is still valid if the cracks develop progressively through the clast from the point of initial failure; in this case kinetic energy is generated locally as failure progresses through the clast. The strain rate needed to cause dynamic fragmentation is $\sim 100 \text{ s}^{-1}$ (Melosh *et al*, 1992). At strain rates ($> 100 \text{ s}^{-1}$) of interest, the kinetic energy is effectively available throughout the clast immediately; Li *et al* (2005) show that duration of fragmentation of a 70 mm granite clast is of the order of $100\mu\text{s}$ at strain rates of $20 - 60 \text{ s}^{-1}$.

This process can be represented by Fig. 1. The initial strain of the clast is linear (O-a); prior to failure, in general, some plastic strain occurs in the curved path between b and c, the extent of which is determined by the material properties and strain rate. In the case of a brittle solid strained rapidly, the difference between the elastic limit Q_e and the ultimate or plastic limit Q_p will be very small. The total work done in stressing the clast to failure is the sum of the dotted and shaded areas in Fig. A. The release of elastic strain energy on failure causes the curve to fall along b-c, and the energy thus released is that of the dotted area. In the case of brittle failure under rapid strain, the plastic strain δs_p is very small, and the energy dissipated during failure – the shaded area – is also small. Hence most of the work done in brittle failure of a clast is transformed into kinetic energy of the sub-clasts. This is the “explosive fragmentation” reported by Wawersik and Fairhurst (1971) and Li *et al* (2005).

Note that the energy expended is the work done in deforming the rock mass; it is *not* expended in breaking the rock and is *not* associated with the newly created fragment surfaces. Note also that in the above the only elasticity involved is that of the clast. The elasticity of the environment applying the stress is unspecified; in effect it is subsumed into Q .

It is often supposed that explosive fragmentation in a laboratory situation is caused entirely by the elasticity of the testing machine, but this is clearly untrue; machine elasticity has some influence, but in principle the machine could have the same elasticity as the clast – deep-mine rockbursts are exactly this case, as are dynamic fragmentations in shearing granular rock masses. The only situation in which brittle fragmentation is not explosive is in a servo-controlled testing machine that reduces stress as soon as strain occurs – which is in effect operating the machine with *negative* elasticity numerically equal to that of the clast so that the total elasticity is always zero (Hazzard *et al*, 2000, Fig 12). This situation does not occur in nature.

The elastic-strain energy released at failure thus converts mostly to kinetic energy of sub-fragments moving away from the original centre of mass at ve-

locities $> 10 \text{ ms}^{-1}$ for common rock types (Herget, 1988; Bergstrom, 1963), with some loss to heat in the imperfectly elastic strain and release processes. The sum of fragment momenta with respect to the original particle centre of mass is zero. In tightly-packed debris, the travel distance of the high-velocity sub-fragments is very small, and their effect on the surroundings is that of a short-lived local dispersive pressure.

2.2 Effects of dynamic fragmentation

Crack propagation in rapid failure of brittle rock takes place as high-speed (typically $\sim \text{km s}^{-1}$) expansion of a “failure front” through the particle (Espinoza and Xu, 1997). The shear strength of the fragmenting material is very low (e.g. Benz and Ausphaug, 1995), hence the resulting dispersive pressure is isotropic. Since the pressure in the vicinity of the failure front is slightly above Q (Espinoza and Xu, 1997; Reches and Dewers, 2005), the magnitude of the dispersive pressure on the surrounding grains is similar to Q . Q for crustal rocks at high shear rates is $\sim 2 - 4 \text{ GPa}$ (Sharpton *et al*, 1996), so instantaneous local isotropic dispersive pressures of this order result from rapid compressive grain failure (Reches and Dewers, 2005).

A single layer of fragments sheared and comminuted between two rock surfaces delivers this dispersive pressure directly to the surfaces, opposing the external confining stress. This is likely to be the situation early in a rock-on-rock friction experiment, or a rupture in re-agglomerated fault gouge, or a first-time blockslide; very large local forces are applied to the intact rock surfaces by fragmenting grains, reducing the effective stress on the fragment layer and thus reducing the frictional resistance to motion that the layer can develop with a friction coefficient in the usual range 0.5-0.7. As shearing and fragmentation continue, the single layer of fragments develops into a granular mass comprising many layers of smaller fragments (Fig. 2); while new large fragments are introduced by continued failure of the intact boundaries.

3 GRAIN FLOW

3.1 Shear localisation

Shear bands, indicating localisation of shear strain and comprised of fine, highly-communited debris, are frequently reported from fault motion (Sibson, 2003) and blockslides (Anders *et al*, 2000). Laboratory experiments show that shear localisation occurs preferentially under high confining stresses

in association with comminution (Aharonov and Sparks, 2006; Mair *et al*, 2002), and that the inclination of the shear band to the direction of shear reduces from $\sim 12^\circ - 20^\circ$ to a very low value as the rate of strain increases (Morrow and Byerlee, 1989).

Shear bands reported in non-fragmenting granular materials are ~ 10 grain diameters deep (François *et al*, 2002); frictional resistance within such bands is somewhat lower than in deeper layers Siavoshi *et al*, 2006).

If confining pressures are significant, the shear rate in the shear band may become sufficient to cause grain fragmentation to occur as a result of compressive stress in a grain bridge (see below). Where a new failure surface is generated, as in initiation of a fault rupture in intact or rewelded rock, or of a first-time blockslide, the debris at the failure surface will initially be about the same size as the gap between the bounding intact rock surfaces; this will however rapidly comminute, and when the ratio of debris thickness to grain diameter exceeds ~ 10 , shear localisation will begin to occur (Fig. 2). Shear banding has the effect of substantially increasing the local strain rate.

3.2 Grain bridges and fragmentation

Shearing of a non-fragmenting grain mass is dominated by stresses in short-lived, quasi-linear grain columns brought into compressive contact by shear strain. These have been called force chains or grain bridges (Anthony and Marone, 2005). Forming in the direction of maximum local compressive strain, grain bridges transmit normal and resisting stresses between the boundaries; the many grains not instantaneously part of a bridge carry lower stresses than those in bridges. Continued shearing of the grain mass steepens and thus shortens the bridges, increasing their intergranular compressive stress. In three dimensional, widely graded debris

Fig. 2 Development of shear localization in comminution on a new failure surface

with irregularly-shaped particles (Hazzard and Mair, 2003), the bridges are possibly somewhat diffuse and irregular, nevertheless they retain their function. Grain bridges can fail and become shorter by buckling, or by grain fragmentation under compressive stress; when the latter happens the failed bridge shortens immediately, transferring stress to adjacent grains which immediately form other bridges.

Once assembled, a grain bridge in a shear field is advected so that its inclination to the shear direction increases; this causes it to shorten, and the compressive stresses in it increase rapidly (at a rate dependent on the end conditions). At the same time the shear stress tending to cause grain slip increases as the inclination of the bridge increases, but much more slowly than the increase in compressive stress, for rock of normal elasticity. Advection therefore rapidly increases the available frictional resistance to intergranular sliding or slip. Hence, once a bridge is assembled with compressive stresses between all grains, failure by slipping is unlikely. This conclusion corresponds to the observation in two-dimensional computer simulations of shearing, that bridge failure by slip is rare (Kuhn, 1999). In a natural shear zone with highly non-spherical comminuted grains, the possibility of failure by grain rolling is also slight.

The most likely grain-bridge failure mode is therefore that of *buckling* (King & Sammis, 1992). Confining pressure of surrounding material significantly increases the ability of the column to resist buckling failure (Toakley, 1965; Battacharya *et al*, 2004). A grain bridge in a low confining pressure is thus able to fail by buckling before the compressive strength of a constituent grain is reached. With higher confining pressure, it is more likely that bridge failure will occur by grain crushing. This corresponds to the common observation of a surface “carapace” of less-fragmented rock above the pervasively-fragmented interior of many large rock avalanche deposits (e.g. Dunning, 2004), and indicates that the confining pressure needed to cause fragmentation is that at the base of the carapace.

Assessment of the confining pressure causing grain bridges to fail by crushing rather than buckling is presently beyond our capability: details of bridge geometry and end conditions affect any calculation, and appear unknowable; and the passive resistance able to be mobilised by the surrounding fractally-graded granular material, and the strain needed to mobilise it, are also imponderables.

Grain bridges must span the width of an active shear zone, and shear bands are usually ~ 10 grains wide (François *et al*, 2002), hence grain bridges are ~ 15 grains long. Both grain and bridge strengths increase with shear rate (Biegel *et al*, 1989; Marone and Kilgore, 1993). These considerations suggest

that stronger bridges more likely to fail by fragmentation occur:

1. in the shear zones associated with shear localisation at high shear rates, which are required for dynamic grain fragmentation, and are also typical of fault rupture and blocksliding; and
2. under high confining stresses.

3.3 Effects of fragmentation stresses

A fragmenting grain loses most of its shear strength (e.g. Benz and Ausphaug, 1995), so little frictional resistance to slip can be transferred between the boundaries through a failing bridge. The effect of this on frictional resistance can be demonstrated by considering a gouge comprising comminuting grains with $Q = 1$ GPa and applied normal stress of 100 MPa; a single grain fragmenting simultaneously in each of as few as 2% of grain bridges reduces the effective normal stress on the debris by 2% of ~ 1 GPa to ~ 80 MPa, and the frictional resistance to $\sim 80\%$ of its non-fragmenting value, assuming conventional friction in the shearing grain mass. The reduced frictional resistance to relative motion of the rock boundaries is partly offset by energy abstracted from the boundary motion to cause fragmentations (Bergstrom, 1963); however this is a small proportion of the total frictional energy dissipation, and so does not greatly affect friction (Tarantino and Hyde, 2005).

3.4 Fragmentation stress in a shear band

Consider a thin granular, shearing fragmenting stratum between a deep moving non-fragmenting layer and a stationary base (Fig. 3): in which P_{OB} is the overburden pressure at top of the fragmenting layer, P_F is the average fragmentation pressure in the fragmenting layer and P_E is the effective intergranular direct stress in the fragmenting layer. All stresses are assumed time-invariant, as is strain rate.

Applying a vertical stress balance at the top of the fragmenting layer:

$$P_E = P_{OB} - P_F \quad (1)$$

Fig. 3 Diagram for derivation of P_E

Each fragmenting grain generates a pressure = Q, so

$$P_F = k_1 F_f \quad (2)$$

where F_f = proportion of grains fragmenting simultaneously; $k_1 = Q \sim 10^9$ Pa.

It seems reasonable to expect that the proportion of grains fragmenting simultaneously (at a given shear rate) will depend on the amount by which the effective intergranular direct stress P_E exceeds the minimum value needed to cause fragmentation P_C :

$$F_f = k_2(P_E - P_C) \quad (3)$$

assuming the dependency is linear.

In the Falling Mountain simulation (Davies & McSaveney, 2002), fragmentation occupied > 80% of the (unconfined) flow depth and no shear banding was apparent. The mean fragmentation stress explaining runout distance was ~ 2.5 MPa, which required $F_f \sim 0.01$. The mean overburden pressure resulting in this stress was ~ 0.75 MPa. The minimum stress required to cause fragmentation (P_C) was that at the base of a relatively unfragmented carapace, ~ 0.3 MPa, or 10^{-3} times the ambient Q .

Applying these empirical data to (3), $0.01 = k_2 \times 0.45 \times 10^6$ and $k_2 = 2 \times 10^{-8}$.

From (1), (2) & (3),

$$P_F = k_1 k_2 (P_{OB} - P_C) / (1 + k_1 k_2) \quad (4)$$

$$k_1 k_2 = 10^9 \times 2 \times 10^{-8} = 20 \text{ so}$$

$$P_F = 20(P_{OB} - P_C) / 21 = 0.95(P_{OB} - P_C) \quad (5)$$

Hence

$$P_E = P_{OB} - P_F = 0.05P_{OB} + 0.95P_C \quad (6)$$

Thus any rock shear whose motion is dominated by a fragmenting layer will experience a frictional resistance to motion ($\mu \times P_E$) of the order of that given by (6).

4 APPLICATIONS

4.1 Blockslides – Waikaremoana and Heart Mountain

We have modelled the dynamics of the Waikaremoana blockslide in New Zealand (Davies *et al*, 2006). About 2200 years ago this 1.4 km^3 intact sandstone block slid ~ 2 km on a basal slope of the same rock inclined at $\sim 7^\circ$, accelerating from rest to about 40 ms^{-1} . A ~ 300 mm deep stratum of fine sand-sized rock debris has been found in the loca-

tion of the inferred sliding surface, suggesting an overall shear rate $\sim 130 \text{ s}^{-1}$ (but clearly much higher in narrow shear bands); such strata are common in large low-angle blockslides (Anders *et al*, 2000). Geostatic stress due to the weight of the 275-m deep block is about 6.6 MPa, and the value of Q at the ambient conditions is of the order of 0.2 GPa. Applying (6) to this situation, the frictional retarding stress in shear bands should be $\sim 0.05 \times 6.6 + .95 \times 0.2 = 0.52$ MPa. The net downslope stress (gravity component minus friction) is then $6.6 \sin 7^\circ - 0.52 = 0.28$ MPa. Since the mass of the block per unit base area is 0.69 ktm^{-2} , this stress would cause the intact block to accelerate at 0.4 ms^{-2} . Two kilometres of travel at this acceleration would develop a velocity of about 40 m/s, which is close to that suggested independently by Davies, *et al.* (2006) on the basis of physical modelling.

The $\sim 3500 \text{ km}^3$ Heart Mountain blockslide in Wyoming/Montana, USA had an initial thickness of several km, and travelled 45 km on a basal slope of $\sim 2^\circ$ (e.g. Beutner & Gerbi, 2006). The mechanics of its low-friction travel have remained a puzzle for over half a century. If we apply Eq. (6) to the situation, assuming normal rock density, $Q = 1$ GPa and $P_C = 1$ MPa, we find that the basal friction angle ($\tan^{-1} (\mu P_E / P_{OB})$) = 1.8° , indicating that basal layer fragmentation can allow the block to accelerate on a basal slope of 2° .

4.2 Fault rupture

Low apparent friction coefficients ($\mu_a = 0.1 - 0.3$) are often inferred in the motion of large strike-slip faults (e.g. Freed and Lin, 2002; Townend and Zoback, 2004). Townend & Zoback (2004) infer that the San Andreas fault ruptures locally with a friction coefficient as low as 0.06. Along the San Francisco peninsula, for example, the direct shear stress at rupture is ~ 180 MPa, while the shear stress is ~ 10 MPa. If we apply Eq. (6) to rupture at 15-km depth, with a confining stress of 400 MPa and $Q = 1$ GPa we obtain $P_E \sim 20$ MPa and the resisting stress is $\sim 0.6 \times 20 = 12$ MPa; close to the stress needed to rupture the San Andreas fault (Townend & Zoback, 2004).

4.3 Laboratory data

In a laboratory experiment measuring rock-on-rock friction of Arkansas novaculite (a very hard metasedimentary quartz rock), Di Toro *et al* (2004) found frictional resistance reduced to about 30% of normal at high shear rates ($\sim 100 \text{ s}^{-1}$) and normal stresses of 5 MPa, accompanied by extrusion of rock gouge. They tentatively attribute the reduced friction to development of a silica-gel layer. We ex

plain it differently — the gouge-forming rock fragmentation developed dispersive pressure, reducing the effective normal stress on the sliding surfaces. If we apply Eq (6), assuming $Q \sim 1$ GPa at the quoted shear rate gives $P_c = 1$ MPa, and $P_E = 5/20 + 1 = 1.25$ MPa. If $\mu = 0.6$, P_f is 0.75 MPa and the apparent friction coefficient is $0.75/5 = 0.15$, compared with their quoted 0.2.

That *dynamic* fragmentation is needed to significantly lower friction is indicated by the fact that only very minor friction variations from normal are reported at shear rates $\ll 1 \text{ s}^{-1}$, even at high normal stress (Kilgore *et al*, 1993; Byerlee, 1978; Mair *et al*, 2001; Beeler *et al*, 1996).

4.4 Rheometer tests

We have used a 300 mm diameter cone-and-plate rheometer (Davies *et al*, 2005) to measure the stresses generated by shearing ~ 1 litre of dry granular coal at strain rates $\sim 50 \text{ s}^{-1}$ and confining pressures up to 0.3 MPa, sufficient to cause pervasive grain fragmentation. Coal is a weak, brittle material with $Q \sim 10$ MPa and crushing behaviour characteristics similar to those of quartz (Charles, 1957). With grains of chosen size in the sample space, the sample bowl was rotated at a chosen rate and a chosen direct stress applied. Typical output data are shown in Fig. 5; using 4 – 8 mm grains, friction was initially very low and increased with strain as grains comminuted by fragmentation. At the conclusion of the test (total strain ~ 400) the sample was intensely comminuted, particularly against the roof, and because of the lack of new material being supplied the voids ratio of the sample became very small and comminution ceased. The behaviour shown in Fig. 5 requires a reduction of about 15% (or 0.04 MPa) in the normal stress on the sample; this would be provided by coal grains fragmenting simultaneously in about 1/500 grain bridges, or 1/5000 grains. If we apply Eq (6), assuming $P_c = 10^{-3} \times Q = 10$ kPa, then $P_E = 0.05 \times 125 + 0.95 \times 10 = 15.75$ kPa, compared with the measured 12.5 kPa.

Fig. 5 Rheometer test (raw data) from fragmenting coal granules (4–8 mm sieve fraction) shearing under 250 kPa normal load. Shear rate $\sim 50 \text{ s}^{-1}$.

5 RATE OF COMMINTION

Continuing re-fragmentation of grains causes an increasingly rapid reduction in grain-size with time; but it cannot cause grain-size to reduce to zero. In the situations considered above, new fragments are recruited from the rock boundaries of the shearing debris when shear bands are adjacent to them, offsetting to some extent the net reduction in average grain-size. Initial fragment sizes (Grady and Kipp, 1987; Zhang and Hao, 2003) of $\sim 10 - 100 \text{ mm}$ have been noted at strain rates $\sim 100 \text{ s}^{-1}$. Further complications arise from the wide range of fragment sizes and fragment numbers resulting from individual fragmentations. However, order-of-magnitude spreadsheet simulations using average conditions and uniform time-varying grain size (for example, that for comminution in a deep fault rupture with confining stress ~ 200 GPa is illustrated in Fig. 6) suggest that with realistically-sized initial fragments, fragmentation pressures in the GPa range, shear localisation with ~ 10 grains per bridge, and fresh fragments being recruited, a fragmenting shear band can persist long enough for fault motion to be realistic without creating zero-sized gouge. A similar exercise for block sliding reaches the same conclusion.

6 COMMENTS

The mechanism we propose is based on the stresses known to be associated with and generated by dynamic rock fracture, and requires only conventional friction *coefficients*; it also requires that a significant proportion of the strain energy applied to fragmentation is recovered as kinetic energy of the fragments. The latter requirement is met in reality.

While the kinematics and dynamics of three-dimensional fragmenting grain flows comprising widely-graded, angular grains are as yet poorly

Fig. 6 Comminution of 10 mm grains in a shear band under fault rupture conditions; duration of shear band $\sim 1/250$ sec. Low-friction fault motion of 0.1 sec requires 25 shear bands to act sequentially.

known (and difficult to study with real materials), the behaviour of simpler granular flows is compatible with the proposed mechanism by allowing transfer of stresses from fragmenting grains through grain bridges to the boundaries of the shear bands in which most of the motion is taking place.

6 CONCLUSIONS

Rapid fragmentation of intact rock under high confining stress generates dispersive pressures of the order of GPa on the surrounding rocks. These pressures result from recovery of much of the strain energy developed in rock failure, in the form of kinetic energy. Fragmentation in a thin band of localised shear opposes the confining stress and reduces the effective stress in the shear band, correspondingly reducing the available frictional resistance to shear. The residual frictional resistance is approximated by Eq. (6); it quantitatively explains the motion of the Waikaremoana and Heart Mountain blockslides. This mechanism also successfully explains the reduced friction inferred in coseismic fault rupture and measured in laboratory rock-on-rock friction tests. The ability of fragmentation-derived pressure to explain reduced frictional resistance to rock sliding hitherto seems likely to be significant either as a complement to, or as a replacement for, other mechanisms involving high pore water pressures, rock melting or high-pressure metamorphism.

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