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Electric currents along earthquake faults and the magnetization of pseudotachylite veins

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Abstract

Pseudotachylites occur in the form of thin glassy veins quenched from frictional melts along the fault planes of major earthquakes. They contain finely grained magnetite and often exhibit a high natural remanent magnetization (NRM). High NRM values imply strong local electric currents. These currents must persist for some time, while the pseudotachylite veins cool through the Curie temperature of magnetite around 580 °C. There is no generally accepted theory explaining how such powerful, persistent currents may be generated along the fault plane. Data presented here suggest the activation of electronic charge carriers, which are present in igneous rocks in a dormant, inactive form. These charge carriers can be "awakened" by the application of stress. They are electrons and defect electrons, also known as positive holes or p-holes for short. While p-holes are capable of spreading out of the stressed rock volume into adjacent p-type conductive unstressed rocks, electrons require a connection to the hot, n-type conductive lower crust. However, as long as the (downward) electron flow is not connected, the circuit is not closed. Hence, with the outflow of p-holes impeded, no current can be sustained. This situation is comparable to that of a charged battery where one pole remains unconnected. The friction melt that forms coseismically during rupture, provides a conductive path downward, which closes the circuit. This allows a current to flow along the fault plane. Extrapolating from laboratory data, every km^3 of stressed igneous rocks adjacent to the fault plane can deliver 10^3–10^5 A. Hence, the current along the fault plane will not be limited by the number of charge carriers but more likely by the (electronic) conductivity of the cooling pseudotachylite vein. The sheet current will produce a magnetic field, whose vectors will lie in the fault plane and perpendicular to the flow direction.

Keywords: Earthquake physics; Friction melt; Magnetization; Pseudotachylite; Electric currents

1. Introduction

Large earthquakes can cause frictional melting (Lund and Austrheim, 2003; Warr et al., 2003) due to the conversion of mechanical energy into heat at high rates of deformation. During earthquake faulting the layer of molten rock acts as a lubricant, thereby

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limiting the amount of melt that can be formed. As soon as the motion of the two sides of a fault comes to a halt, this thin friction melt layer begins to cool, leading to glasses or partly devitrified glasses, called pseudotachylites (Sibson, 1975; Spray, 1995; Petrik et al., 2003; Bjørnerud and Magloughlin, 2004; Ferré et al., 2005a).

Pseudotachylites are recognized in the field as dark veins along the shear planes of exhumed fossil earthquake faults, similar to ultracataclasites, another form of rocks formed under conditions of extreme mechanical deformation. Main veins tend to be continuous over long distances and are generally a few centimeters wide (Austrheim and Boundy, 1994; Boundy and Austrheim, 1998). Short side veins often branch off the main vein. Occasional large blobs of pseudotachylite are observed, thought to have formed through the injection of friction melt into cracks and cavities that shortly open during the dynamically evolving events during earthquakes (Bjørnerud and Magloughlin, 2004).

The dark brown, almost black color of most pseudotachylites is due to finely grained magnetite, which precipitates during rapid cooling (Nakamura et al., 2002; Moecher and Sharp, 2004). The magnetite grains, which are mostly single domain or pseudo-single domain, i.e. of the size range 50–85 nm (Dunlop and Özdemir, 1997) are responsible for the large magnetic susceptibility of the pseudotachylites relative to that of their host rocks. Ferré et al. (2005b) reported on a pseudotachylite sample taken from a main vein in the Santa Rosa Mountains near Palm Springs, California, USA. It exhibits, in addition to a large magnetic susceptibility, an abnormally high natural remanent magnetization (NRM), up to 300:1 higher than that of the host rock (Kletetschka, 2006). Nakamura et al. (2002) drilled into the Nojima Fault after the 1995 Kobe earthquake and reported on high NRM values extending into a narrow zone in the non-melted rocks adjacent to the fault plane. Abnormally high NRM values have also been reported from massive pseudotachylites and ultracataclasites associated with the Vredeford crater in South Africa where impact melts, produced by shock waves, created situations similar to those of friction melts during large earthquakes (Carporzen et al., 2005).

The NRM freezes in the memory of the ambient magnetic field, to which a given sample was exposed while cooling through the Curie point of its constituent ferromagnetic minerals. In the case of pure magnetite, Fe_3O_4, the Curie temperature is around 570 °C a low pressure but increases to 650–660 °C at confining pressures of 4–6 GPa (Samara and Giardini, 1969; Schult, 1970). In magnetite–ulvospinel solid solutions the Curie temperatures decrease with increasing Ti content to 280–355 °C at 10% ulvospinel, depending on the confining pressure, and lower values at higher Ti contents (Schult, 1970). Rocks that cool in the Earth's main field typically acquire an NRM value of about 1/1000th of the strength of the Earth's main field. Depending on latitude, the main field ranges from about 20 to 60 μT. Therefore, most NRM values of rocks fall into the 0.02–0.06 μT range. To acquire the reported high NRM values of the Santa Rosa pseudotachylite this vein must have been exposed to a strong local magnetic field, about 1000 higher than the average Earth's main field (Ferré et al., 2005b). This strong local magnetic field must have persisted for at least as long as it took the vein to cool through the range of Curie temperatures of its magnetites. A strong local magnetic field, however, can only be generated by a strong local electrical current. This raises three questions:

(i) What is the basic current generation mechanism?
(ii) How can persistent currents be generated?
(iii) From where, to where and for how long do the currents flow?

2. Sources of currents and electromagnetic signals

Different physical processes have been proposed to account for electric currents in the ground and associated electromagnetic effects. The list includes piezo, tribo and fractoelectricity, magnetohydrodynamics, streaming potential, and fluid flash vaporization.

2.1. Piezoelectricity

Quartz is the only common rock-forming mineral that is piezoelectric. For this reason quartz has received a great amount of attention as a possible source of stress-induced electric fields and electric currents in quartz-bearing rocks (Parkhomenko, 1971; Tuck et al., 1977; Cress et al., 1987).

When stress is applied to a quartz single crystal along the *x* axis (perpendicular to the polar screw axis *z*), the *x*-faces will instantly acquire charges of opposite sign. The resulting piezovoltage is proportional to the applied stress while the piezocurrent is related to the stress rate, i.e. the time-derivative of the stress. We therefore have: $P \sim \varepsilon\sigma$, where P is the electric polarization in C/m^2, ε is the dielectric constant in Coulomb [C] per Newton [N] (for quartz: $\varepsilon = 2 \times 10^{-12}$ C/N), and σ is the stress in N/m^2 or in Pa. For instance, when a quartz crystal is stressed along its piezo-axis (*x* axis) to 100 MPa (10^8 N/m^2), it instantly generates charges of opposite sign on the order of 2×10^{-4} C/m^2 on its surfaces. The polarity in a given x direction depends upon whether the z axis of the quartz crystal is left-turning or right-turning. When the stress is removed, the piezocharges instantly disappear.

Quartz-bearing rocks contain large numbers of quartz crystals, generally in random orientations, sometimes with quartz preferentially aligned along a morphological axis. While each suitably oriented quartz crystal will produce a piezovoltage when stressed, the individual piezovoltages of a large ensemble of crystals cancel when integrated over the rock volume (Ogawa and Utada, 2000) or come close to canceling (Tuck et al., 1977; Bishop, 1981). The reason is that, statistically, the ratio of quartz crystals with a left-turning z axis to quartz crystals with a right-turning z axis is always 1:1. There is no known mechanism in nature that would favor one polarity over the other. Hence the distribution of piezoelectrically active axes can never deviate from the 1:1 ratio, even in rocks with morphologically aligned quartz. Contrary to earlier statements (Parkhomenko, 1971), large piezovoltages such as those needed to produce large currents can therefore not be generated by stressing quartz-bearing rocks.

2.2. Streaming potentials

When water or brines are forced through fractured or porous rocks or through the gouge that fills many earthquake faults, there is a possibility of charge separation. Some ions, generally cations, tend to be retained along the walls of the fractures or on the surfaces of mineral grains that make up the gouge, while their counter ions are carried onward by the streaming water or brine.

Streaming potentials have been experimentally well documented and are theoretically well understood (Morgan et al., 1989; Morrison et al., 1989; Bernabé, 1998; Revil et al., 1999a,b). They play a role in some technical domains (Oommen, 1988) but have also been considered to be a source of large-scale, time-variable electric fields and associated electromagnetic emissions in the context of seismic activity (Fenoglio et al., 1995; Merzer and Klemperer, 1997; Karakelian et al., 2002) and of volcanic activity (Jouniaux et al., 2000; Revil et al., 2003). For pseudotachylites streaming potentials due to movement of liquid water or brine can be ruled out as a source of currents because any local currents that leave their imprints in the form of high NRM must flow while the vein cools through the Curie temperature of magnetite, 580 °C.

2.3. Other mechanisms

Several other processes have also been considered in the literature such as tribo or fractroelectricity (Gokhberg et al., 1982). The weakness is that tribo and fractroelectricity do not describe a single physical process but rather a series of processes that occur in the wedge and along the flanks of an opening fracture. Those processes include energetic excitations due to high but very transient electric fields appearing on opposing crack surfaces (Dickinson et al., 1986, 1987). However, these excitations are unable to produce large-scale electric fields necessary to produce large electric currents, at least not under any realistic natural conditions.

By introducing magnetohydrodynamics a concept is used that comes out of the single-fluid plasma theory (Molchanov et al., 2001). There seems to be little evidence that processes, which occur in plasmas under near-vacuum conditions, can be applied to large-scale processes in the solid Earth.

Other suggestions include sudden resistivity changes in earthquake preparation zones, which have been observed (Park et al., 1993). Resistivity changes do not provide onto themselves a mechanism to generate strong and sustained electric currents that are necessary to magnetize a pseudotachylite vein while it cools through the Curie temperature of its magnetite or magnetite–ulvospinel components.

Near-surface fluid flash vaporization of water has been proposed to explain the generation of earthquake lights (Lockner et al., 1983) but flash evaporation is not applicable to processes deep in the Earth's crust during earthquakes.

3. Positive holes

We recently uncovered a new and fundamentally different mechanism by which electrical currents can be generated from igneous rocks (Freund, 2002, 2003; Freund et al., 2006). This generation process is linked to the fact that a fraction of the oxygen anions in common igneous and high-grade metamorphic rocks are not in their usual 2-valence state (O^{2-}) but have converted to the 1-valence state (O^-). From a semiconductor perspective an O^- in a matrix of O^{2-} represents a defect electron or hole, also known as positive hole (Griscom, 1990) or p-hole for short. It represents a positive charge residing at the upper edge of the valence band. The charge can move by exchanging an electron with any O^{2-} neighbor, $O^- + O^{2-} <=> O^{2-} + O^-$.

Single O^- tend to react with other O^- to form positive hole pairs, PHP, chemically equivalent to peroxy links such as between SiO_4 tetrahedra, $O_3Si–OO–SiO_3$. In the PHPs the O^- are self-trapped and localized. Hence, they are electrically inactive. However, there are at least two known processes, which activate PHPs causing the peroxy links to break and release p-hole charge carriers: (i) mechanical stress and (ii) heating.

In this paper we deal mostly with the effect of mechanical stress, but also mention the effect of heating in the context of modeling a cross section through the Earth's crust from the cool upper and middle crust down to the hot lower crust.

3.1. Stress-activation of p-hole currents

Stress-induced changes in the electric conductivity of rocks have been measured in the past (Kariya and Shankland, 1983; Glover and Vine, 1992; Hyndman et al., 1993; Glover and Vine, 1994; Tyburczy and Fisler, 1995; Shankland et al., 1997; Fuji-ta et al., 2004). Laboratory experiments were typically set up in such a way as to apply stress to the entire cross section of a given rock sample, generally of cylindrical shape, while applying voltages across the rock to measure the conductivity.

We have taken a very different approach to measure electric currents in rocks. With the benefit of knowledge from prior work (Freund, 2002) we designed the experiment in such a way as to respond to the specific characteristics of p-hole charge carriers. Instead of rock cylinders to be loaded over their entire cross section we take large samples, either slabs with a rectangular cross section or square tiles about $30\times30\times1$ cm^3. We apply stress to only a small subvolume. We attach one electrical contact to the volume to be placed under stress and another contact to an unstressed part of the rock. We connect ammeters to each of these two contacts and measure the currents that flow to ground.

Note that, with such an experimental set-up, we are not measuring the electrical conductivity proper, i.e., not the amount of current that can be forced through the rock under externally applied voltages. Instead we measure the self-generated electrical currents that flow out of the stressed rock volume on their own, without externally applied voltage. The underlying process is akin to that in a battery. The thermodynamic driving force is produced by the stress gradient between the stressed and unstressed portions of the rock, which translates into a concentration gradient of p-hole charge carriers.

Fig. 1a shows an experiment with a 1.2 m long slab of Sierra White granite, 10×15 cm^2 cross section, fitted with two Cu electrodes of equal size, 30×15 cm^2, wrapped around both ends. The Cu tape has a graphite-loaded conductive adhesive. One end of the slab is placed

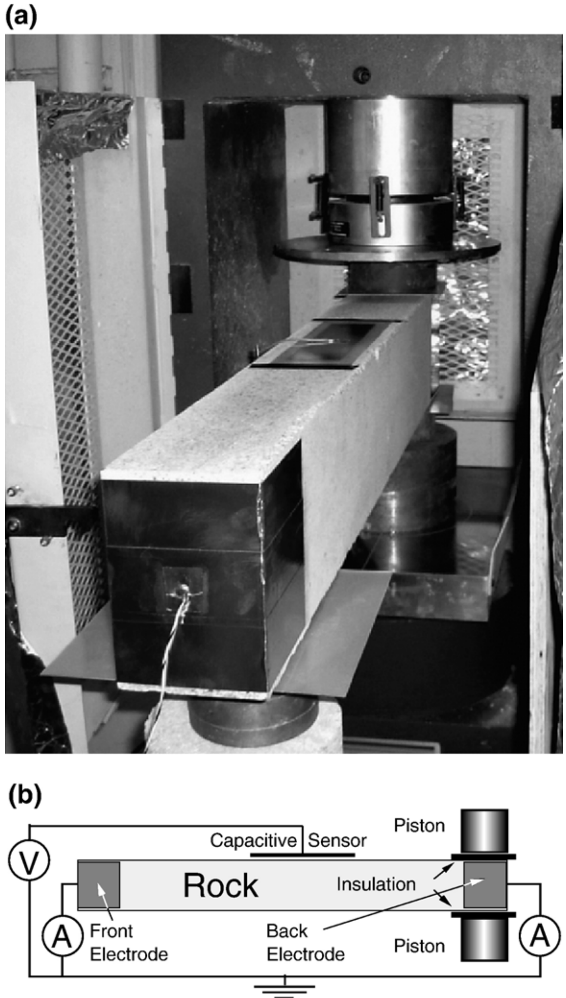

Fig. 1. (a): Granite slab placed in the press, ready for the uniaxial compression tests. The granite slab (1.2 m long, 10×15 cm^2 cross section) is fitted with two Cu electrodes (30×15 cm^2), one at the back end and one at the front end, plus a non-contact capacitative sensor for measuring the surface potential. The rock is insulated from the pistons and the press by 0.8 mm thick polyethylene sheets ($>10^{14}$ Ω cm). (b): Block diagram of the electric circuit for allowing the self-generated currents to flow out of the stressed rock volume.

between the pistons of a hydraulic press, about 11.5 cm diameter, and electrically insulated from the pistons by 0.8 mm thick sheets of polyethylene ($>10^{14}$ Ω cm). In addition a 10×20 cm^2 metal plate is placed on top of the rock, separated by a 0.8 mm thick, 5 mm wide stripes of polyethylene to serve as a capacitive sensor for measuring surface charges. Fig. 1b shows the fully passive electrical circuit that contains two Keithley 487 ammeters to measure currents and a Keithley 617 electrometer to measure voltages.

A volume of approximately ~1500 cm^3 at one end of the block was loaded at a rate of ~0.11 MPa/s to 67 MPa,

~ 1/3 the failure strength of unconfined granite. The load was released suddenly and reapplied after 30 min. The loading/unloading cycles were repeated 6 times.

Fig. 2 (top) shows the results of the 4th loading/unloading cycle, which are very similar to those of the first three and following two cycles. As soon as stress is applied, two currents of the same magnitude but opposite sign flow out of the stressed rock volume, the "source" (S). This is depicted in Fig. 2(bottom), which shows the cross section of the experimental set-up and the electric circuit. The two outflow currents are part of one closed loop divided into two currents. One current, carried by electrons, e′, flows from the source into the Cu electrode attached to the stressed part of the rock and thence to ground. The other current, carried by p-holes, h˙, flows from the source into and through the unstressed rock. It traverses the full length of the granite slab and reaches the front end Cu electrode, where it meets electrons that come up from ground.

3.2. Two outflow currents — one current loop

The electric circuit shown at the bottom of Fig. 2 illustrates the battery situation described here. Except for small differences in the outflow currents at the beginning

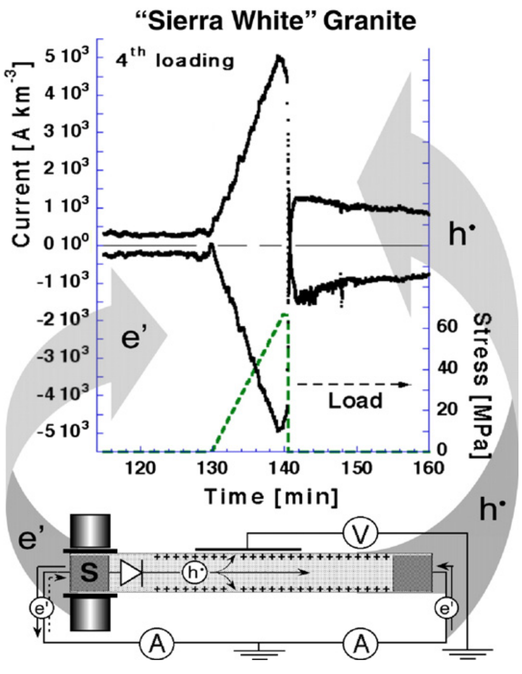

Fig. 2. Two currents flowing out of the stressed rock volume, the "source" S, and a schematic representation of the current flow through the external circuit and inside the rock with holes passing through the boundary between stressed/unstressed rock. This boundary lets p-holes pass through but blocks electrons. It acts as a barrier symbolized here as a diode.

of loading and after unloading, which are due to polarization currents and will be discussed elsewhere (Freund et al., 2006), we are dealing with one and the same current loop. What we call outflow currents are in fact two legs of the same current leaving the stressed rock volume in opposite directions and via different conduction mechanisms.

Obviously the mechanical stress that we apply to the rock activates two types of electronic charge carriers, electrons and p-holes. Before application of stress these charge carriers must already have existed in the rock, albeit in the form of non-active, dormant precursors. Mostly likely these precursors consist of PHP or peroxy links $O_3Si–OO–SiO_3$ as outlined above. The peroxy links become activated when dislocations sweep through the minerals as a result of the applied stress, straining the SiO–OSi bond angle and destabilizing the peroxy bond. The peroxy link is thought to then take over an electron from one of its O^{2-} neighbors, thereby converting this O^{2-} neighbor into O^-. This p-hole is able to propagate (as an electronic state) away from the site where it was generated (Freund et al., 2006).

As electronic states associated with O^- in a matrix of O^{2-}, the p-holes occupy energy levels at the upper edge of the valence band. Though silicate minerals are considered to be insulators, p-holes are able to spread through the O^{2-} sublattice jumping from O^{2-} to O^{2-}. They can cross grain boundaries and traverse ~ 1 m of granite. In the experiment depicted in Fig. 1 the thermodynamic driving force for this outflow of p-holes is provided by their concentration gradient between stressed and unstressed rock. At low to moderate temperatures, when p-holes dominate the conductivity response, our data suggest that rocks such as granite are always p-type conducting (Freund, 2003).

The electrons that are co-activated with the p-holes behave differently. Though they are probably only loosely bound to the reconstituted peroxy links (Freund et al., 2006), evidence provided by Fig. 2 suggests that the electrons are unable to leave the stressed rock volume by flowing into the unstressed rock. The reason is that, in order to become mobile charge carriers, electrons need higher energy levels, which are available at or close to the lower edge of the conduction band. At low to moderate temperatures the thermal energy of the system is insufficient to promote electrons to such high levels. However, at elevated temperatures, typically above 550–600 °C, a small number of electrons begins to populate higher levels. Once the electrons occupy high energy levels and become mobile, they soon start to dominate the conductivity response, because electrons generally have a significantly higher mobility than holes

(Pascoe, 1973). As a result, with increasing depth and increasing temperature along the geotherm, the conductivity response of the rocks will change from p-type to n-type (Freund, 2003). In our laboratory experiment we substitute the contact to n-type hot rock with the Cu tape attached to the stressed rock volume in Figs. 1 and 2 or with the steel pistons in Fig. 3 below.

3.3. Outflow currents as a function of the stress rate

In our experiment with the granite slab, both outflow currents increase approximately linearly with stress as shown in Fig. 2. The granite delivers +7 and −7 nA out of a source volume of 1500 cm^3, equivalent to ~5×10^3 A/km^3. However, this linear current-stress relation is most surely an artifact caused by the slab geometry. The reason is that the constant cross section of the slab limits the outflow of p-hole charge carriers regardless of how many p-holes are generated in the stressed rock volume.

When we take a square rock tile and load the center, the currents generated in the stressed rock volume can flow out radially within the plane. This geometry gives a larger ratio cross section between the stressed and unstressed rock, plus the cross section increases from the center to the edge. As a result the currents become nonlinear with stress. Fig.3 shows the outflow currents from a gabbro tile 30×30×1 cm^3, loaded at the center with 3.2 cm diameter pistons stressing a volume of ~10 cm^3 at 0.1 MPa/s to 50 MPa. The pistons serve as one electrical contact, while a 12.5 mm wide Cu tape along the edge of the tile serves as the other contact. At the bottom of Fig. 3 we show a cross section through the square tile and the schematic of the outer circuit.

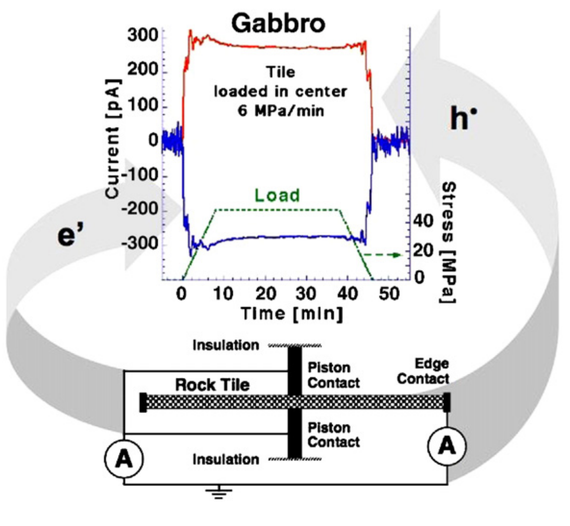

Fig. 3. Outflow currents from a gabbro tile, 30×30×1 cm^3, loaded at the center (~10 cm^3) moderately fast and kept at constant load for 30 min. The outflow currents increase very rapidly at the beginning of loading, reach a stable value and then decay very slowly.

We observe a fast initial current increase, even when the stress increases at a constant rate. When we hold the stress constant, the outflow currents remain high, decaying only slowly with a halftime on the orders of hours to days. This suggests that the charge carriers activated within the stressed rock volume have long lifetimes. In addition we note that the gabbro delivers higher currents than granite, in this case ~3×10^4 A/km^3. We observe that

(i) both outflow currents increase rapidly with increasing stress, especially between 0 and 10 MPa;
(ii) both currents remain high under constant stress;
(iii) upon unloading both currents do not decrease until the stress decreases below ~10 MPa.

The fact that the battery currents from quartz-free gabbro exceed the outflow currents from granite is a strong indication that the piezoelectricity of quartz does not play a role in this process.

Upon loading and unloading granite, gabbro and anorthosite tiles repeatedly, up to 30–50 cycles, the outflow currents do not decrease. This suggests that loading to 1/4–1/3 of the failure strength does not noticeably damage the rocks.

To further investigate the stress-current relationship we took a gabbro tile and varied the stress rates over 5 orders of magnitude, from 2×10^{-4} MPa/s to 20 MPa/s, going from 0–58 MPa. The p-hole outflow currents are plotted in Fig. 4a in units of A/km^3 versus stress. At slow loading rates, 2×10^{-4} to 2×10^{-3} MPa/s, the currents increase approximately as a function of t^2 (t=time) approaching saturation values of about 28,000–31,000 A/km^3 above 10–30 MPa. At faster rates the character of the curves changes: (i) the saturation currents achieved above 10–30 MPa increase from ~28,000 A/km^3 at the slowest rate to over 50,000 A/km^3 at the fastest rate; (ii) an initial current spike develops with increasing loading rate, approaching 100,000 A/km^3 at 20 MPa/s.

4. Modeling the outflow currents

Eq. (1) emulates the stress-generated outflow currents as a function of time during stress ramp up:

$$I(t) = K_0 \cdot e^{\frac{-t}{\tau_0}} \cdot \cos(\omega_d \cdot t + \phi) + K_1 \cdot \left(1 - e^{\frac{-t}{\tau_1}}\right) + K_2 \tag{1}$$

where τ_0 is the decay time constant of quasi-frequency ω_d and phase φ and where τ_1 is the growth or charging

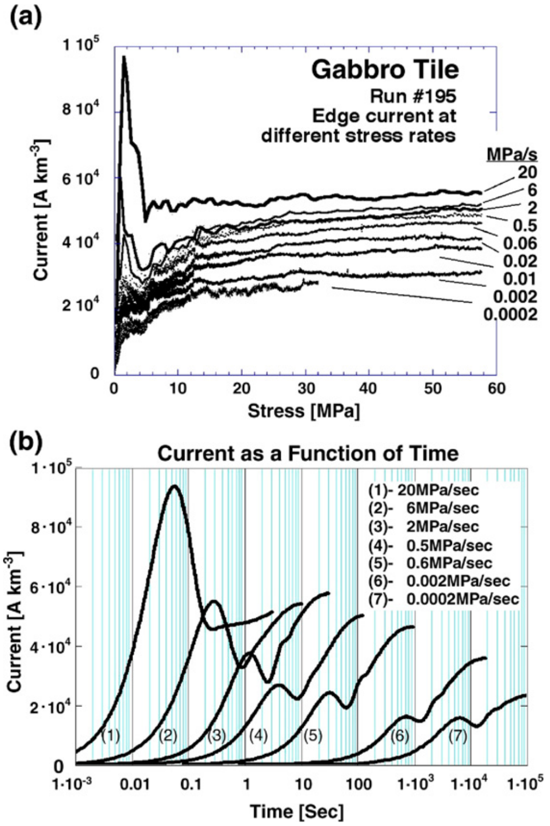

Fig. 4. (a): Stress-current relationship for a gabbro tile from 2×10^{-4} to 20 MPa/s between 0 and 50 MPa. The time range covers about 5 decades, from 3 s to 133,100 s (37 h). Shown are the p-hole outflow currents. The electron outflow currents are close mirror images. (b): Approximated currents as a function of time for gabbro for the same loading rates as 4(a), using Eq. (1).

time constant. These parameters along with scale constants K_0–K_2 (in A/km^3), when solved to minimize the difference between Eq. (1) and the actual data, build a general formula that can be used to depict a noise-free version of the currents.

The result of applying the curve-fit Eq. (1) to the data from Fig. 4a is seen in Fig. 4b, where we plot the current versus time spanning 8 decades in time. The p-hole currents are offset-corrected for the different stress rates. All $I(t)$ curves are similar. They differ only in their relative contributions to the initial peak versus the slow rise toward the saturation values. Faster loading rates produce higher initial peaks.

The data shown in Fig. 4a and the curve fits in Fig. 4b suggest that two mechanisms operate in this gabbro during loading, both generating p-holes and electrons though at different rates. Once these charge carriers are activated, they both have long lifetimes, consistent with the measured current outflow activity up to ∼37 h. Similar results have been obtained with other igneous rocks such as granite and anorthosite.

5. Modeling a fault rupture

In the experiments described so far the circuit was always closed during loading and unloading with the metal contacts providing the electrons and p-holes the opportunity to continuously flow to ground through the two ammeters. Of course such conditions are artificial, typical of the laboratory experiment, and different from conditions assumed to exist inside the Earth's crust.

A highly simplified, but geophysically relevant scenario is shown in Fig. 5 with the geometry akin to that of our laboratory experiment depicted in Fig. 1a/b using the symbol of a diode to indicate the presence of a barrier that lets p-holes pass but rejects electrons. We consider the crust to consist of a single block of rock as shown in Fig. 5a subjected to increasing temperatures with increasing depth along the geotherm (disregarding the effect of isostatic pressure increase). We consider the tectonic forces to act over the entire cross section of the crust, i.e. from the primarily p-type conducting upper and middle crust to the n-type conducting lower crust (Freund, 2003). The tectonic forces generate a horizontal stress gradient in the relatively cool and brittle upper portion of the crust, where a large number of charge carriers, both p-holes and electrons, will be activated. We postulate two incipient outflow currents: (i) p-holes that spread into the p-type conductive unstressed or less stressed rock as shown by the dotted horizontal arrows in Fig. 5b, and (ii) electrons that flow from the stressed rock volume into the n-type conductive lower crust as indicated by the dotted, downward pointing arrows.

However, as long as the outflow of electrons does not reconnect to the outflow of p-holes, i.e. does not include a return current for the electrons flowing upward, the circuit remains open as suggested above when we drew the analogy to an open-circuit battery.

Fig. 5c depicts the next step in this model. We assume that a major earthquake occurs leading to a rupture along an inclined fault plane. The rupture spreads upward to the Earth's surface as well as downward. We assume that the rupture extends from the p-type conductive upper portions of the crust all the way down into the n-type conductive lower crust. We further assume that friction causes a thin sheet of rock to instantly melt along the fault plane. Electrically, the thin sheet of molten rock and thin layers of hot rock on either side provide an n-type conductive pathway through the entire crust.

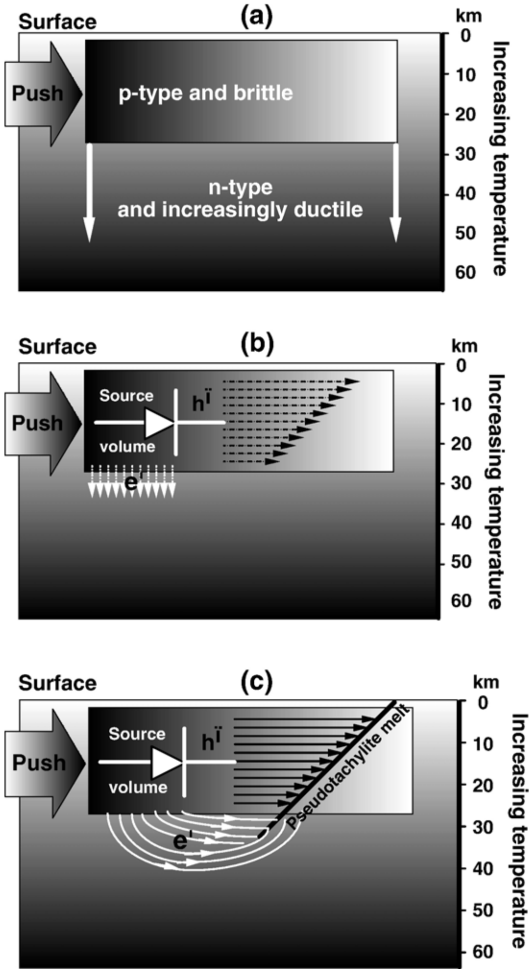

Fig. 5. (a): Schematic of the crust divided into the cool, brittle, p-type conductive upper to middle crust and the hot, ductile, n-type conductive lower crust. The depth scale in km is for orientation purposes only (b): Assuming tectonic forces acting on the crust laterally from left to right, a "source" volume develops in which p-holes and electrons are activated. While the p-holes flow out laterally, the electrons are assumed not to be able to flow downward, connecting to the n-type conductive lower crusts. (c): The sheet of molten rock along the fault plane electrically connects the upper to the lower crust allowing a current to flow.[1]

Thus electrons from the infinitely large source of the hot lower crust flow upward along the fault plane providing a return current. The electrons use high-lying energy levels that are thermally populated at temperatures above 500 °C, defining a p–n transition in the conductivity character of the rocks (Freund, 2003). The electrons flowing upward along the fault plane meet the p-holes that precipitate into the plane from the adjacent rocks and recombine. This current thereby closes the circuit of the "battery".

6. Simulating the current loop closure in the laboratory

We can conduct laboratory experiments that simulate the circuit closure. Using the geometry sketched in cross section at the bottom of Fig. 3 we took a gabbro tile and loaded its center, applying stress to a volume of ~10 cm^3. Both pistons and the Cu tape around the periphery were set up to be connected to ammeters and, hence, to ground but only the piston contact was connected. The edge contact was left open. We loaded the tile from 0 to 62 MPa at a rate of 6.2 MPa/s, kept the load constant for 10 min as shown in Fig. 6 and then removed it again at 6.2 MPa/s. We repeated the procedure three times, connecting the edge contact about 2 min, 5 min and 7 min after applying the load.

Fig. 6 shows that, under open circuit conditions no current flows. We observe a small transient current each time we apply the load. This current is due to the outflow of p-holes from the stressed rock volume into the unstressed portion of the rock, i.e. a one-time polarization current, which sets up a counteracting electric field and then stops. Though the rock is now fully loaded and the battery "charged", the current returns to zero. In the moment we close the circuit by connecting the edge contact to the second ammeter, we observe an instant current pulse. Its rise time is about 50 ms. It

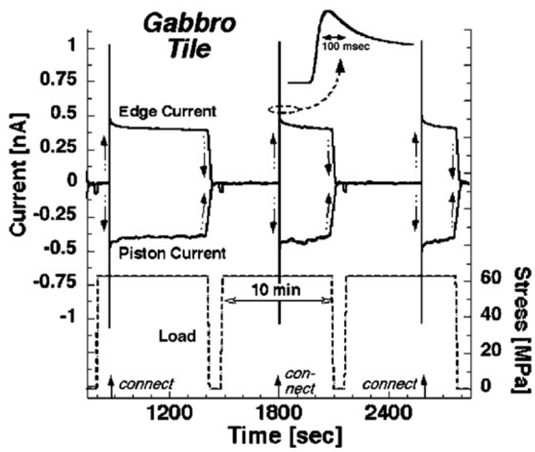

Fig. 6. Experiment with gabbro tile simulating the sudden current surge and persisting current after closure of the electric circuit at constant stress level (see text).

[1] Pressure certainly is also a factor though nothing is known at present about the pressure effect on the thermal stability of the peroxy link with respect to bond breakage and release of a p-hole charge carrier. Most likely pressure has a stabilizing effect on the peroxy bond, shifting its break-up temperature upward.

exceeds 1 nA, equivalent to $\sim 10^5$ A/km^3, and decays rapidly to about 0.5 nA equivalent to $\sim 5\times10^4$ A/km^3. Thereafter the current stabilizes and decays at a much slower rate with a halftime on the order of hours.

The most important information from Fig. 6 is that, when a thin sheet of rock instantly melts along the fault plane and closes the "battery" circuit, a mechanism exists that can inject a large electric current into the fault plane. The magnitude of this current will not be limited by the number of charge carriers that can be supplied from the adjacent rocks, capable to deliver as much as 10^4–10^5 A/km^3. The limitation of the current will rather come from the limited current carrying capacity of the thin, rapidly cooling layer of hot rock along the fault plane. However, because the charge carriers are electronic in nature, the conductivity of this thin sheet of hot rock can be expected to be significantly higher, most likely by several orders of magnitude, than the conductivity obtained from a downward extrapolation of high temperature conductivity data which are routinely measured in some laboratories but normally reflect only the ionic conductivity (Freund, 2003).

In addition, our laboratory data suggest that the stress-activated currents are not only potentially large but can also persist for a long time after a given stress level has been reached, at least on the order of minutes, possibly on the order of hours, the current carrying capacity of the vein remains high. Since the pseudotachylite vein is thought to cool through the Curie temperature of magnetite around 580 °C within tens of seconds (Ferré et al., 2005b), we have reasons to believe that the vein can easily be subjected to the effects of strong local electric currents still flowing along the fault plane. Hence, the pseudotachylite will record the resulting local strong magnetic field in the form of a high NRM value.

7. Modeling the local magnetic field

We now further develop the model depicted in Fig. 5c to identify the conditions, under which a magnetic field can be generated that is larger, potentially much larger, than the Earth's dipole field.

To produce a significant NRM during cooling of a pseudotachylyte vein the current responsible for the magnetizing field must still flow while the temperature passes through the Curie temperature of the micro to nanosized magnetite crystals, ≤ 580 °C. Ferré et al. (2005b) modeled a 2 cm wide vein in a 400 °C host rock, assuming flash-melting during an earthquake and cooling by heat conduction to the adjacent rock. Under these conditions, it takes the vein 10–20 s to cool to below 500 °C.

As sketched in Fig. 7 we have to consider a uniform planar, upward-flowing sheet current. It is assumed to be infinite in two dimensions. It generates a magnetic field ***B*** with field vectors lying in the plane perpendicular to the direction of the current. The magnetic field is highest at the two surfaces of the sheet current but drops to zero in the center plane. The magnetic induction or field density is:

$$\boldsymbol{B}_{//} = (\mu_0/2)K \tag{2}$$

where μ_0 is the permeability of free space ($4\pi\times10^{-7}$-Henry/m) and K is the current density at the surface of the sheet. It can be represented by I/L or the current per cross-section length. This estimate does not take into account the magnetic susceptibility χ_m of the medium, i.e., of the rocks in which the current flows. If the medium is diamagnetic, we have $\chi_m<0$. If it is paramagnetic, we have $\chi_m>0$. To take the effect of the medium into account, we have to modify Eq. (2) to:

$$\boldsymbol{B}_{//} = [\mu_0(1+\chi_m)/2)]K \tag{3}$$

The average intensity of the associated magnetic field is given by:

$$\boldsymbol{B}_{//} \approx 2\pi\times10^{-7} I/L \tag{4}$$

At the surface of the current sheet the magnetic field is twice that high. For $B_{//}$=450 μT (10 times a typical Earth's dipole field), a current intensity I on the order of 2–3×10^5 A per km fault length would be required. Taking the pseudotachylite vein to be about 2 cm wide, the current density during passage through the Curie temperature of magnetite, 580 °C, would then have to be on the order of 1 A cm^{-2}.

Due to the non-uniformity of the stress gradient field in realistic geophysical situations and the expected uneven pre-rupture distribution of p-holes in the rock

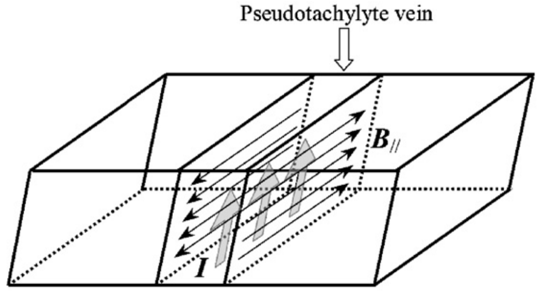

Fig. 7. Sketch of a uniform planar current (shadowed arrows) flowing upwards through a n-type conductive pseudotachylite vein and its associated in-plane magnetic field (solid arrows).

volumes next to the fault, we expect the current flowing along the fault plane to be highly variable in time and space. Consequently any coseismic and post-rupture magnetic fields will also be highly variable. In particular we can expect the sheet current along the fault plane to be irregular following curved paths and pinched at many places. In this case the magnetic field vectors and the direction of the frozen-in NRM would not be restricted to lie within the plane. In fact, Ferré et al. (2005b) reported that at least one pseudotachylite sample from a vein in the Santa Rosa Mountains exhibits an NRM vector forming a steep angle with respect to the fault plane. So far no large-scale NRM mapping of pseudotachylites has been reported that would allow the reconstruction of the current flow patterns along these old earthquake faults.

8. Conclusions

The discovery of mobile electronic charge carriers that are activated by stress allows us to propose a mechanism to generate large electric currents flowing along the fault plane of major earthquakes following the formation of a thin sheet of friction melt. These currents are not driven by piezoelectric voltages or any other mechanism discussed so far in the literature. They are driven by p-holes and electrons, which are activated in large rock volumes during the build-up of the stresses prior to the earthquake. The p-holes are assumed to spread laterally in the relatively cool upper and middle crust, while the electrons are assumed to flow downward into the hot lower crust. The stressed rock volume with potentially two outflow currents forms a "battery". At first, before fault rupture and formation of a friction melt, the circuit of the "battery" is open. When the rupture occurs and a friction melt appears along the fault plane, the sheet of molten rock will provide a conductive path spanning the upper and middle crust and reaching into the lower crust, thereby closing the circuit. This allows a sheet current to flow along the fault plane. The magnitude of this sheet current is not limited by the supply of p-hole charge carriers from the rocks adjacent to the fault plane or by the availability of electrons in the hot lower crust. The current will persist as long as the (electronic) conductivity along the fault plane remains high. The current will magnetize the nanometer-sized magnetite crystals, which seem to be a characteristic feature of most pseudotachylites, while they cool through their Curie temperature around 580 °C, leaving a permanent imprint in the form of a high natural remanent magnetization, NRM.

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