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The brittle-plastic transition and the depth of seismic faulting

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With 4 figures

Zusammenfassung

Ein einfaches rheologisches Modell für Schervorgänge in der Lithosphäre, das eine weite Akzeptanz erreicht hat, ist das Zweilagenmodell mit einer oberen spröden Zone, in der Deformation über Reibungsgleitung entlang diskreter Störungsflächen stattfindet, und einer unteren duktilen Zone. Hier erfolgt die Deformation durch plastisches Fließen. Beide Zonen werden durch einen abrupten spröd-plastischen Übergang voneinander getrennt, der vermutlich durch die untere Grenze der nachweisbaren Seismizität angezeigt wird. Experimentelle Untersuchungen wie auch die Deformationsgefüge in Myloniten zeigen hingegen, daß ein breites Übergangsfeld mit semisprödem Verhalten zwischen diesen beiden Extremen liegt. Hier befindet sich ein Bereich »gemischter« Deformation, deren Ausmaß beträchtlich über den aus der Extrapolation von Hochtemperatur-Fließ-Gesetzmäßigkeiten ableitbaren Werten liegen dürfte. Für Quarz-Feldspat-Gesteine liegt das halbspröde Feld zwischen T_1, dem Beginn der plastischen Deformation des Quarzes bei ca. 300 °C, und T_2, dem Beginn der Feldspatplastizität bei ca. 450 °C. Hier wird ein Modell vorgestellt, in dem der Übergang bei T_1 nicht dem Übergang zum Gesamtfließen, sondern dem Wechsel von instabiler, geschwindigkeitsreduzierender Reibung zu stabiler, geschwindigkeitskonstanter Reibung entspricht. T_1 markiert damit die Tiefengrenze der Erdbebenbildung, stärkere Beben können dagegen in größere Tiefe reichen T_3 ($T_3 < T_2$), die dann der unteren Grenze dynamischen Reibungsverhaltens im semispröden Bereich und angenähert dem Stärkemaximum entspricht. Die Zone zwischen T_1 und T_3 zeigt alternierendes Verhalten mit plastischem Fließen während interseismischer Perioden und dynamischem Gleiten während größerer Erdbeben. Diese Zone wird charakterisiert durch Mylonite und dicht zwischengepackten Pseudotachyliten sowie anderen Anzeigern dynamischer Faltung. Hinzu kommt im Bereich des Überganges T_1 ein Wechsel im Bildungsmechanismus gestörter Gesteine. Er geht von durchgreifender Ermüdung, die zu Kataklasiten führt, bis zu adhesiver Ermüdung, die als wichtiger Bildungsmechanismus für Mylonite im oberen Abschnitt des semispröden Feldes angesehen wird.

Abstract

A simple rheological model of shearing of the lithosphere that has gained wide acceptance is a two layer model with an upper brittle zone in which deformation takes place by frictional sliding on discrete fault surfaces and a lower plastic zone in which deformation takes place by bulk plastic flow. The two are separated by an abrupt brittle-plastic transition, which is assumed to be indicated by the lower limit of seismicity. Experimental studies, however, as well as the deformation structures of mylonites, indicate that a broad transitional field of semi-brittle behavior lies between these extremes. This is a field of mixed mode deformation with a strength that can be expected to be considerably higher than that predicted from the extrapolation of high temperature flow laws. For quartzofeldspathic rocks the semi-brittle field lies between T_1, the onset of quartz plasticity at about 300 °C and T_2, feldspar plasticity at about 450 °C. A model is presented in which the transition T_1 does not correspond to a transition to bulk flow but to a change from unstable, velocity-weakening friction to stable, velocity-strengthening friction. T_1 thus marks the depth limit of earthquake nucleation, but large earthquakes can propagate to a greater depth, T_3, ($T_3 < T_2$) which corresponds to the lower limit of dynamic frictional behavior in the semi-brittle field and approximately to the peak in strength. The zone between T_1 and T_3 is one of alternating behavior, with flow occurring in the interseismic period and with co-seismic dynamic slip occurring during large earthquakes. This zone is characterized by mylonites interlaced by pseudotachylytes and other signs of dynamic faulting. The transition T_1 is also marked by a change in the generation mechanism of fault rocks, from abrasive wear above which produces cataclastites, to adhesive wear below, which is proposed as an important generation mechanism of mylonites in the upper part of the semi-brittle field.

Résumé

Le modèle rhéologique de cisaillement de la lithosphère le plus largement accepté comporte deux couches superposées: une couche supérieure cassante, siège de déformations le long de surfaces discrètes, les failles, et une zone inférieure ductile ou s'opère un fluage plastique d'ensemble. La transition entre ces deux domaines est brusque et considérée comme la limite inférieure de la séismicité.

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L'étude des structures mylonitiques ainsi que les mesures expérimentales indiquent cependant qu'un large champ de transition à caractère «semi-cassant» s'étend entre ces deux extrêmes. Ruptures et fluage plastique sont présents dans cette zone, dont la compétence peut être considérée comme bien supérieure à celle qui résulte de l'extrapolation des lois de fluage à haute température. Pour les roches quartzo-feldspathiques, ce champ «semi-cassant» s'étend de T_1, seuil de plasticité du quartz (environ 300 °C) à T_2, celui du feldspath (environ 450 °C). Dans ce modèle, la température T_1 ne correspond pas au seuil de fluage d'ensemble, mais à la frontière entre une région de frictions «instables» qui diminuent avec la vitesse, et une région de frictions «stables» qui augmentent avec la vitesse. Elle correspond donc à la profondeur limite de genèse des séismes. Les grands séismes peuvent cependant se propager jusqu'à une profondeur T_3 ($T_3 < T_2$) qui correspond à la limite inférieure du comportement dynamique des forces de frottement dans le domaine «semi-cassant». T_3 peut être considérée approximativement comme le point de résistance maximale de cette zone de transition. La zone comprise entre T_1 et T_3 peut donc être le siège alternativement soit de processus de fluage, soit de glissements le long de surfaces de rupture à l'occasion de séismes de forte magnitude. Cette zone se caractérise par des mylonites entremêlées de pseudotachylites et autres signes de rupture *dynamique. La transition* T_1 *est également marquée* par une modification du mécanisme de transformation des roches dans les zones de faille. Audessus de cette limite, un mécanisme d'usure «abrasif» produit des cataclasites, par opposition à un mécanisme d'usure «adhésif» en-dessous. Ce dernier mécanisme est proposé comme fort probable lors de la genèse de mylonites dans la partie supérieure du champ «semi-cassant».

Краткое содержание

Описана реологическая модель для изучения феномена деформации типа сдвига (скалывания) в литосфере, в которой предусматривают наличие двух зон. Верхняя состоит из хрупкого материала, где деформация идет путем скольжения с трением вдоль скрытой плоскости нарушения, и подлежащей зоны из пластичного материала. В этой последней деформация происходит в результате пластичной текучести. Обе зоны разделены хрупко-пластичной переходной зоной, которую принимают, как нижний предел сейсмичности.

Однако, как опыты, так и изучение структур деформации милонитов разрешают считать, что переходная зона, где отмечено полухрупкое состояние пород, между этими зонами значительно шире. Здесь установлен регион «смешанной» деформации, размеры которого значительно больше размеров переходной зоны, выведенных из законов поведения материала в зависимости от температуры и текучих свойств этого материала. Для кварцо-полевошпатовых пород этот регион полухрупких пород устанавливают между T_1 – началом пластичной деформации кварца при 300°C и T_2 – началом текучести полевого шпата при 450°C. Здесь описана модель, по которой переход у T_1 соответствует не переходу к общей текучести породы, но смене нестабильного трения, замедляющего скорость движения, к стабильному трению с неизменяемой скоростью течения. Т. о. T_1 является, как бы, предельной зоной зарождения землетрясений. Но более сильные землетрясения могут возникать на большей глубине – T_3 ($T_3 > T_2$), т. е. находится на нижней границе динамического поведения трения в регионе с полухрупким состоянием материала, что примерно соответствует максимуму силы землетрясения.

Зона между T_1 и T_2 проявляет меняющееся поведение пластичной текучести в период между сейсмической активностью и динамическим скольжением во время больших землетрясений.

Эта зона характеризуется присутствием милонитов и тесно связанных с ними псевдотрахитов, а также образованием сбросов. Кроме того, в зоне перехода T_1 изменяется механизм образования брекчий трения. Это изменение выражено переходом от абразивного истирания в верхней части, в результате которого образуются катакластические породы, к пластической деформации в нижней части; при этом образуются милониты в верхней части зоны с полухрупкими породами.

Introduction

More than 50 years ago the routine determination of earthquake focal depths made it apparent that in most tectonic regions earthquakes were restricted to the shallow part of the earth's lithosphere. It was recognized then that this must result from the shallower, cooler part being brittle while deeper parts responded to large deformation in a ductile, aseismic manner (Macelwane, 1936).

The state of thinking remained at essentially this level until 1980, when Brace and Kohlstedt (1980) and Kirby (1980) assembled laboratory data to produce a simple model of the rheology of the continental lithosphere. This model is illustrated schematically in Figure 1, where the scales have been omitted from the axes to avoid the well-known controversy about the effects of various parameters on these curves and to allow focussing instead on their main features. The upper part of the model is a friction law, whereas the lower part is the extrapolation of a high temperature steady-state flow law for an appropriate rock, the most well determined and commonly used being wet quartzite. The intersection of the two laws was taken to be the brittle-plastic transition in the earth. Sibson (1982) and Meissner & Strehlau (1982) compared

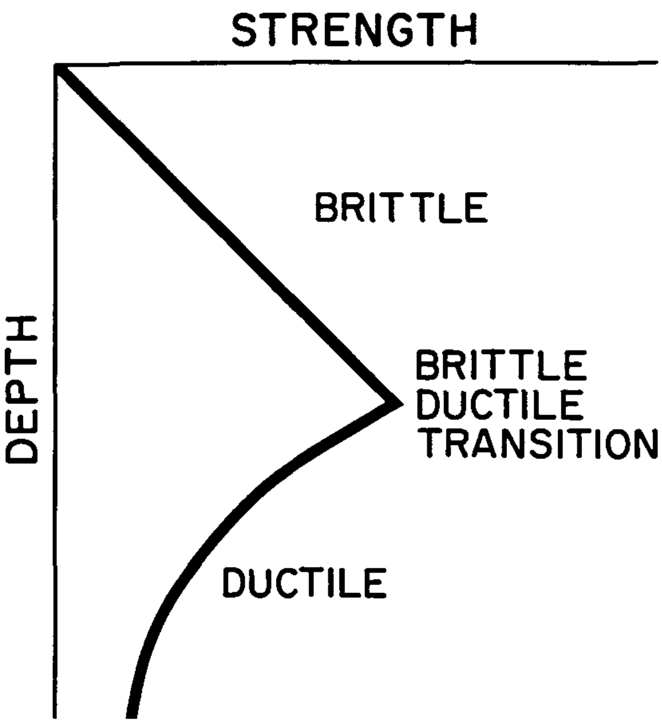

Fig. 1. A simple model for the rheology of the lithosphere.

this model with observations of the deformation mechanisms of fault zone rocks and earthquake hypocentral depth distributions and found a rough correspondence between the brittle-plastic transition predicted by this model, the depth at which evidence for crystal plasticity can be found in fault rocks, and the maximum depths of earthquakes. This seemed to provide strong confirmation for this model and so detailed refinements and applications were made, in which often minor variations in the depth distribution of seismicity were, for example, ascribed to variations in heat flow (SIBSON, 1984; DOSER & KANAMORI, 1986; SMITH & BRUHN, 1984).

More recently, a number of fundamental difficulties in this simple model have been pointed out. HOBBS et al. (1986) and STREHLAU (1986) have noted that most of the region in the lower part of this model does not correspond to the high temperature flow regime but rather to semi-brittle deformation (CARTER & KIRBY, 1978). This is a mixed mode of deformation for which the flow law is not known for any material but which is fundamentally different from high temperature flow in that it has a pressure sensitive strength (KIRBY, 1983). As a consequence, the extrapolation of high temperature flow laws into the region of the brittle-plastic transition is not valid. RUTTER (1986) has discussed in detail the problems in the use of the term »brittle-ductile transition« for what is properly a brittle-plastic transition in the model shown in Fig. 1. The term ductility refers only to the property of being able to sustain a substantial strain with macroscopically homogeneous deformation and is therefore not mechanism-specific, applying equally to semi-brittle and plastic deformation. He also points out that the transition is not abrupt: being initiated by a transition from brittle faulting to cataclastic flow (semi-brittle behavior) in which the strength is still pressure sensitive, and being concluded by a transition at greater depth to plasticity, the latter transition occurring near the peak in strength.

HOBBS et al. (1986) and STREHLAU (1986) therefore proposed three layer models in which the plastically flowing region was separated from the brittle region by a transitional layer of poorly defined, semi-brittle rheology. In an independent development, TSE & RICE (1986) explored a model based entirely on a rate and state variable friction law similar to that first described by DIETRICH (1978), in which they were able to show that the restriction of earthquakes to shallow depths could be completely explained by a transition from velocity weakening to velocity strengthening with increasing temperature, as predicted by the experimental data of STESKY (1975), without any requirement for a brittle-ductile transition. These two developments removed the underpinnings of the lower part of the model shown in Fig. 1 and lead one to suspect that the apparent success of that model was misleading, and that the physics of the transformation from seismic to aseismic faulting is far more complex. The purpose of this paper is to discuss the brittle-plastic transition in more detail, for both the case of bulk deformation and frictional sliding, and to develop therefrom a model that is more consistent with these recent findings and the relevent observational facts.

The brittle-plastic transition

The brittle-plastic transition is usually understood to mean the transition from entirely brittle behavior to deformation that occurs entirely by crystalline plasticity. This transition, even for the simplest crystal types, cannot occur at a surface in P-T-$\dot{\varepsilon}$ space but must occur over some field within which the deformation occurs by a mixture of these processes (PATERSON, 1978). This is the semi-brittle field, referred to earlier. Also, since we are here dealing with faults, we must consider two cases of the brittle-plastic transition, the transition as it occurs in the bulk deformation of the rock, and the transition as it occurs in frictional sliding. At some shallow depth, faulting occurs primarily by frictional sliding (with, presumably, some distributed deformation within its zone of wear products) by the mechanism of brittle fracture of contacting asperities. At sufficient depth, the same deformation takes place as plastic flow in bulk as a continuum within an associated ductile shear zone. In the transition between these extremes there is likely to be a region in which the deformation takes place partly by frictional sliding and partly by bulk deformation. It is a common misconception in geology to consider friction to be synonymous with brittle behavior, whereas the brittle process was only involved in the formation of the sliding surfaces, and not necessarily with their continued slip. Friction of metals, for example, occurs almost entirely by the plastic deformation of contacting asperities. Thus the presence of ductily deformed rock in a fault zone cannot be used to rule out frictional sliding as a contributing, or even major, mechanism of transport.

The brittle-plastic transition for bulk deformation

The brittle-plastic transition for any polycrystalline material will take place in a field somewhat like that shown schematically in Figure 2, in which plasticity is favored by elevated temperature and pressure. The brittle regime is separated from the regime of full plasticity by a semi-brittle field in which the deformation occurs by a mixture of brittle and plastic processes. For monomineralic rocks, the width of the semi-brittle field will be demarked by the conditions under which the easiest slip system becomes activated on the one side and on the other those in which sufficient slip systems become activated to satisfy the von Mises-Taylor criterion for fully plastic flow (PATERSON, 1978). For multi-component rocks the semi-brittle field will typically be broader, owing to the ductility contrast of the different mineral species.

The rheology through the semi-brittle field is best illustrated for calcite marble, which has been studied across the entire field along the constant temperature path A-B by SCHOLZ (1968) and EDMOND & PATERSON (1972). At room temperature this rock typically passes into the semi-brittle field at quite low confining pressure, in which the behavior is macroscopically ductile but the internal contribution of brittle processes is revealed by pronounced dilatancy and a strong pressure sensitivity of the strength (in the experimental literature this behavior is often referred to as cataclastic flow, but we avoid this term here because it might cause confusion with the mechanism of formation of cataclastites, which are predominantly wear products of frictional sliding). As confining pressure is raised, dilatancy is suppressed and the pressure sensitivity decreases, indicating that the role of plastic processes is increasingly dominant over brittle processes. Finally, at some critical pressure both dilatancy and the pressure sensitivity vanish and the rock has entered the fully plastic field. The width of the semi-brittle field at room temperature is typically about 300 MPa.

Rheological data of this quality is not available for the more refractory silicates in the semi-brittle field. By inference from microscopic studies of deformation textures, however, we can trace the same transition in

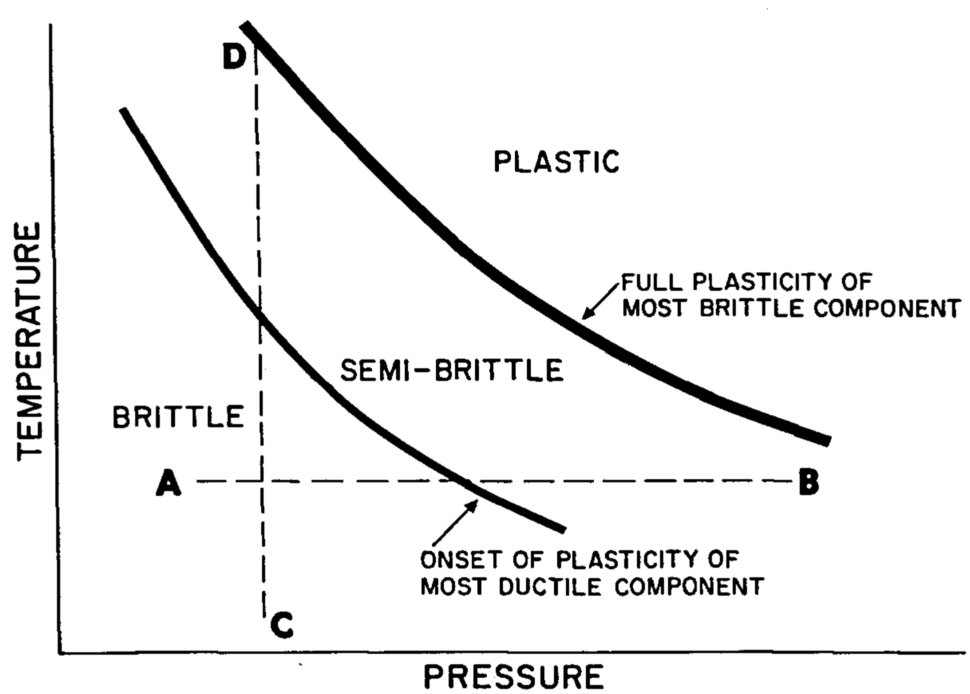

Fig. 2. Schematic diagram of the brittle-plastic transition for bulk flow.

the deformation of granite that was studied along the path C-D by TULLIS & YUND (1977, 1980). For dry granite they found that the semi-brittle field was entered at about 300-400° with the onset of flow in quartz. As temperature was increased there was a gradual transition from abundant microcracks, to a uniform density of dislocations with recrystallized grains and no microcracks, and that dislocations did not become dominant in feldspar until 550—650°. For wet granite these transition temperatures were reduced by 150—200° for both feldspar and quartz.

These experimental results are difficult to extrapolate to natural conditions, so for the present we must rely on geological observations to estimate the conditions under which the brittle-plastic transition occurs in nature. Since both quartz and feldspar have only limited slip systems, fully plastic flow in those materials cannot occur by glide alone, but must be associated with recovery and recrystallization. For quartz, this appears at about 300 °C (VOLL, 1976; KERRICH et al., 1977), and for feldspar, at about 450—500 °C (WHITE, 1976; VOLL, 1976). Thus quartz begins to flow at about the beginning of greenschist facies metamorphism, but feldspar does not begin to flow until well into the amphibolite grade. We take 300° and 450°, then, as the upper and lower bounds of the semi-brittle field for quartzo-feldspathic rocks. Between these two states these rocks behave as composite materials, with quartz flowing and the feldspar responding in a rigid, brittle manner, forming porphyroclasts, as is so commonly observed in mylonites.

Any extrapolation of a flow law from the fully plastic field into the semi-brittle field, as was done in the model illustrated in Fig. 1, will always underestimate the strength there, and that underestimation will worsen the further the extrapolation is taken. This is because, as we traverse this field from D to C, say, plastic mechanisms will gradually be eliminated, first in one component and then in another, which will increase the flow strength, and microscopic brittle processes will become increasingly necessary to accommodate the strain, and which carry with them a pressure sensitivity of strength, which was not included in the extrapolation.

The brittle-plastic transition for frictional sliding

Just as in the above, where we considered entering the transition to the semi-brittle field from below, for this problem we must also consider what occurs when the semi-brittle field is entered from the brittle side above. Only in this case we must consider how it affects frictional sliding, since that is the condition of deformation that occurs there. That is: what are the changes in friction that occur when the mechanism of shearing of asperities changes from brittle fracture, which commonly dominates friction in silicate rock at low temperature, to plastic flow, as occurs in the friction of metals, and at high temperature, for rock?

As far as it affects the gross frictional strength, the answer is: not at all. This is because it is fundamental to friction that the real area of contact, A_r is always less than the nominal area A, and is determined by a strength parameter closely related to the strength required to shear through A_r (BOWDEN & TABOR, 1964). Since A_r increases with normal stress, so must the shear stress required for slip, in a just compensating way so that the strength can be described with a simple friction coefficient. For this reason rock friction is independent of lithology and temperature (BYERLEE, 1978; SCHOLZ, 1977). We should therefore not expect that there should be an abrupt transition from friction to bulk deformation at the point where the first mineral becomes ductile, as is implied in the model shown in Fig. 1, since this transition will only take place when the friction surfaces become welded, i.e. A_r = A, which can be expected to occur only much further into the semi-brittle field. So below the point at which the first yielding and flow of minerals is observed strength should continue to increase linearly with pressure along a normal friction curve. The peak in strength should be expected well within the semi-brittle field and not at its upper boundary, as indicated in Fig. 1.

There are, however, other important effects of the transition on friction. The effect of the brittle-plastic transition on friction is shown schematically in Figure 3. The transition along the path A-B was studied by SHIMAMOTO (1986), using halite as an experimental gouge material sandwiched between sandstone surfaces in triaxial friction experiments. Since halite becomes ductile at moderate pressures and room temperature, these experiments are analogous to the marble experiments described above. At the lowest pressures he found that the friction was velocity-weakening (the velocity dependence of friction is negative) and hence unstable, exhibiting stick-slip. At higher normal stress, when the halite had entered its semi-brittle field, the rheology was still frictional, in the sense that strength continued to rise sharply with confining pressure, but the sign of the velocity dependence changed, *i.e.* friction became velocity-strengthening, and hence stable. Finally, at the highest pressures, where the halite was in its fully plastic field, the strength of the sandwich became insensitive to pressure. At that point translation of the opposing surfaces was taking place by continuous plastic shear of the halite layer, which had completely welded the surfaces (SHIMAMOTO & LOGAN, 1986).

Friction of granite was studied along the path C-D to 700 °C by STESKY (1975, 1978). He found the same pattern, in which at low temperature the friction showed velocity-weakening and stick-slip, but that above 300 °C, where quartz begins to flow, it became velocity-strengthening and stable. He found that the rheology was frictional, with a strong pressure effect on strength, at all temperatures studied, but that this effect was diminished at the highest temperatures. The model of TSE & RICE (1986) was based on Stesky's results.

The onset of plasticity in the semi-brittle field is thus seen to coincide with a change from velocity-weakening to velocity-strengthening in friction and hence to a transition from an unstable frictional field to a stable frictional field. When asperity deformation is largely brittle, velocity-weakening can arise from a class of environmentally aided processes of the type discussed by SCHOLZ & ENGELDER (1976) and DIETRICH (1978). When asperities become ductile, however, they deform according to a flow law with an intrinsic positive strain rate dependence and the corresponding friction becomes velocity-strengthening (RABINOWICZ, 1965; SHIMAMOTO, 1986). The transition to continuous flow, however, occurs at much higher temperature and pressure where the surfaces become welded, nearer the transition to full plasticity for bulk deformation.

The transition from unstable to stable friction thus appears to correspond to the transition from brittle to semi-brittle behavior of the rock. There is also an important change in the mechanism of wear at this boundary. At low temperatures wear occurs by the brittle fracture of asperities, resulting in the production of loose wear particles. This is called abrasive wear (RABINOWICZ, 1965), and is the primary mechanism of generation of cataclastites. Wear still occurs when the asperities deform plastically, as they would, in part, in the semi-brittle field, but this type of wear, which is commonly observed for the ductile metals and is called adhesive wear, occurs by ductile shearing out of asperities, often with transfer of material to the opposite surface. This wear mechanism would thus result in rocks with mylonitic fabrics and microstructure, even though the macroscopic behavior of the fault was still frictional. Thus the first appearance of mylonitic rocks cannot be taken as firm evidence that the relative motion across the shear zone is taken up entirely by bulk shear as a continuum in the shear zone. Often, for example, the rocks in shear zones in the semi-brittle field are S-C mylonites, which contain discrete planes (the C planes) which offset markers and along which slip has thus occured (e.g. LISTER &

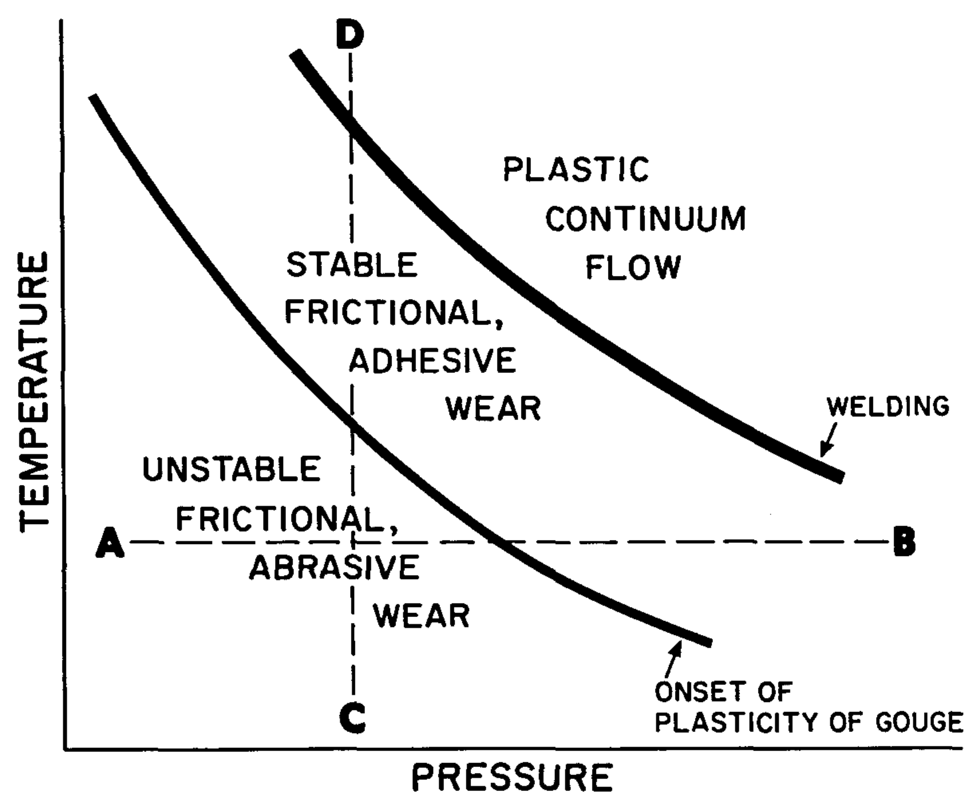

Fig. 3. Schematic diagram of the brittle-plastic transition for frictional sliding.

SNOKE, 1984). In these cases, though the deformation is clearly not brittle, the shear zone may still be partially frictional in its character, which in terms of its rheology means that its strength is still pressure dependent.

A synoptic shear zone model

We are now ready, based on the points made above, to incorporate some improvements in the simple model of Fig. 1 and to develop a somewhat more realistic, though necessarilly more complicated, model of a shear zone, which is shown in Figure 4. This model is intended for quartzo-feldspathic rocks, so the principal fiducial points, shown on the left axis, are 300 ° and 450 °C, taken as the onset of quartz and feldspar plasticity. The depth axis is based on the assumption of a commonly used geotherm model for the San Andreas fault (LACHENBRUCH & SASS, 1980, model B). This depth scale should not be regarded too seriously since even for much of the San Andreas fault it underestimates the depth of the transitions by 3—5 km.

The 300° isotherm is taken as T_1, the transition from brittle to semi-brittle behavior. The semi-brittle field is bounded by T_1 at the top and by the transition to full plasticity at T_2, at the bottom, where feldspar becomes plastic. The transition from cataclastites to mylonites (SIBSON, 1977) occurs at the upper boundary, T_1, which, however, is not a transition from friction to bulk flow but a transition from abrasive to adhesive wear, in which the latter is dominated by ductile flow within a frictional regime. The asymmetric hourglass shape of the shear zone is somewhat fanciful, and is based on a wear model (SCHOLZ, 1987) in which wear rate and thus fault thickness increases with normal stress, and hence depth. The neck in the hourglass is caused by the change in wear mechanism, since usually adhesive wear is associated with much lower wear rates than abrasive wear (RABINOWICZ, 1965). There is no data at present to support this idea, and even in the upper part there are observations of both thickening and thinning of fault zones with depth (FENG & MCEVILLY, 1983; ANDERSON et al., 1983). The hypothetical shape shown would only occur if the rock was of uniform composition over all depths (SCHOLZ, 1987) and would only be preserved in the case of strike-slip faulting.

The seismic behavior of the fault is described using the friction rate parameter A-B (RUINA, 1983). If A-B is negative, friction is velocity-weakening and instable; if it is positive, it is velocity-strengthening and stable. This parameter is plotted schematically in the middle frame of Figure 4, and follows closely the input model b of TSE & RICE (1986), which was based on the results of STESKY (1975). We have, in addition, added a region of velocity-strengthening near the surface. We expect this behavior because of the presence there of unconsolidated and/or clay rich fault gouge. Such materials are not observed to stick-slip in the laboratory, and this provides an upper stability boundary, T_4, to explain why earthquakes are not observed to nucleate at depths less than ca. 2 km on well developed faults. The lower stability boundary, where A-B becomes positive at depth, is approximately at T_1, as independently determined from Stesky's results. Thus

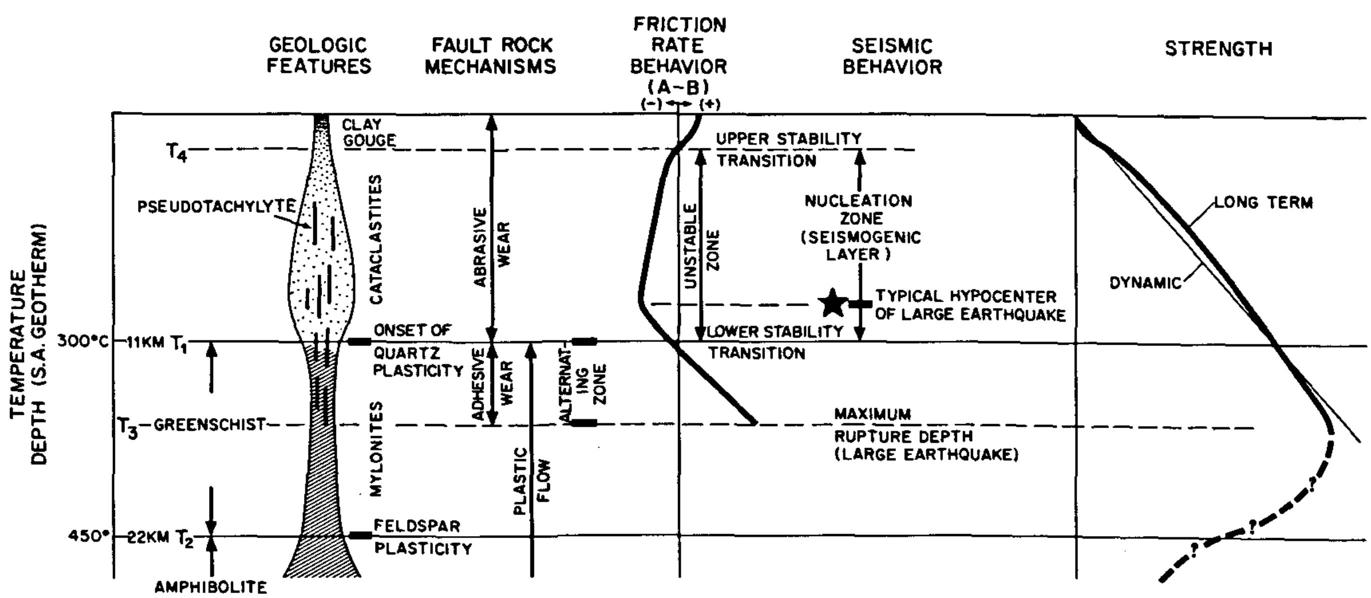

Fig. 4. A synoptic shear zone model, illustrating the major geological and seismological features.

earthquakes can only nucleate between T_4 and T_1, which defines the seismogenic layer. Large earthquakes, which rupture the entire seismogenic layer, and may also commonly break through the thin stable region above T_4 to the surface, are most likely to originate near the base of the seismogenic layer where A-B is a minimum (Tse & Rice 1986.) This same expectation is obtained from considering the dynamics of rupture propagation in a situation where strength and stress drop increase with depth (Das & Scholz, 1983). That large earthquakes do indeed typically originate from near the base of the seismogenic layer has been known for a long time, and this fact was formerly the source of some consternation (Macelwane, 1936).

If T_1 is a boundary between velocity-weakening and velocity-strengthening friction, as is the thesis being explored here, then modeling has shown that during a large earthquake rupture will extend dynamically well into the semi-brittle field below T_1 (Das, 1982; Tse & Rice, 1986). As Strelau (1986) has noted, seismological evidence is either ambiguous or circular on this point. The depth of the deepest aftershocks will be limited by T_1, since that is the depth limit of nucleation, but they do not necessarily, as is so commonly assumed, delineate the total depth of the mainshock. Neither seismic nor geodetic methods have had sufficient resolution to define the maximum depth of faulting. For example, Thatcher (1975) pointed out that the geodetic data for the 1906 San Francisco earthquake are insufficient to resolve slip beneath 10 km at the 1 m level. The most recent seismological techniques appear to bear out this prediction, however. J. Nabelek (pers. comm., 1987) has found, in the case of the 1983 Borah Peak earthquake, significant moment release down to a depth of 16 kms, whereas the aftershocks cut off abruptly at 12 kms.

We thus conclude that large earthquakes will propagate below T_1 to some greater depth, T_3 and that the region between T_1 and T_3 is one of alternating behavior, with co-seismic dynamic slip and interseismic semi-brittle flow. There is a growing body of geological evidence for such an alternating zone, consisting of shear zones in which mylonites are found interlaced with pseudotachylytes and breccias (Sibson, 1980; Stel, 1981, 1986; Wenk & Weiss, 1982; Passchier, 1984); Hobbs et al. (1986) developed a model of ductile instability in an attempt to explain this phenomena. Their model, however, shows that it is unlikely for such an instability to initiate within the alternating zone. Assuming the rheology of wet quartzite and a strain rate of $10^{-12}sec^{-1}$, they found that flow was stable at temperatures greater than about 200 °C, i.e. well above T_1. Thus one might not expect *nucleation* within the alternating zone, but one can use their model to show that an earthquake should be expected to *propagate* well into this zone. An earthquake impinging on T_1 from above will impose there a strain rate some 10^8 times the interseismic rate, so that the co-seismic strain rate will be about $10^{-4}sec^{-1}$, which from their model will allow progagation to much greater depth.

The consequences of this model on the strength of the lithosphere is shown on the right frame of Fig. 4. The shear zone strength is frictional from the surface to T_3, so it increases with depth over that interval. T_1 represents a change in the micromechanism of friction from brittle to ductile which produces a change from velocity-weakening to velocity-strengthening and from abrasive wear to adhesive wear, but no significant break in the strength profile. Hobbs et al. (1986) and Strelau (1986) both introduced transitional regions in their strength profiles but for unexplained reasons assumed that the strength was constant below T_1. As was pointed out above, however, as long as the behavior is essentially frictional, regardless of the micromechanism, strength should increase approximately linearly with normal stress, and hence depth. Thus the peak in strength should not occur at T_1 but at some greater depth where welding has occurred. We place this approximately at T_3. Below this depth we expect the strength to sag, joining a high temperature flow law at T_2. The welding and strength decrease below the peak is probably not produced by plastic flow alone, but is aided by »pressure solution« mechanisms resulting from the intrusion of fluids into the fault zone (Etheridge et al., 1984; Kerrich, 1986).

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