Transition between seismic and aseismic deformation in the upper crust

J. P. Gratier and J. F. Gamond

Geological Society, London, Special Publications 1990; v. 54; p. 461-473
doi:10.1144/GSL.SP.1990.054.01.42

Transition between seismic and aseismic deformation in the upper crust

J. P. GRATIER & J. F. GAMOND

L.G.I.T., Observatoire de Grenoble, Universite Joseph Fourier, I.R.I.G.M., BP 53X, 38041 Grenoble, France

Abstract: Displacement on faults is often accommodated by a succession of two mechanical processes: aseismic sliding mass transfer and seismic cataclastic events. Pressure solution is attested both by dissolution markers of asperities which prevent sliding on and around the fault zone, and by the mechanism of growth of the mineral fibres by aseismic crack-seal in cavities opened by sliding. The cataclastic process is attested by observations of broken and kinked fibres, and by observations of euhedral crystals in the cavities opened by fault sliding. Natural examples are given to recognize and balance the mass transfers. Finally, a model is proposed to explain seismic/aseismic transitions on a given fault based on a gradual reduction in the cataclastic strength of the fault during sliding by successive breaking of asperities, and on the principle of minimum work consumed during the slip since the energy needed either for pressure solution sliding or for the cataclastic event may vary differently with the progressive sliding. Depending on the limiting processes for pressure solution slip, stable aseismic or unstable slip with seismic/aseismic transition is predicted.

The mechanical behaviour of the upper crust is commonly considered to be cataclastic. The creep strength of rocks is presumed to increase with depth, down to a transition zone at about 15–20 km, where dislocation creep mechanisms operate (Brace & Kohlstedt 1980; Kirby 1983; Carter & Tsenn 1987). This model is supported by two facts.

(i) The maximum frequency of earthquakes is located within the upper 15–20 km of the crust, as shown, for example, by depth distribution of earthquakes in the continental crust of the United States (Sibson 1982), and focal depth of intracontinental and intraplate earthquakes (Chen & Molnar 1983).

(ii) The experimental strength of rocks indicates a transition from frictional slip to plastic creep at temperatures and pressures appropriate for depths of 15–20 km (Paterson 1978). At relatively high strain rate (greater than 10^{-8} s^{-1}), a change from a pressure-sensitive friction law (Byerlee 1968), to a strongly temperature-dependent plastic law (Poirier 1985) is observed.

Other natural and experimental data are not, however, consistent with this model for distribution of deformation mechanisms in the crust. It is well known that within the upper crust not one, but two major deformation mechanisms operate: cataclastic deformation and pressure solution creep (see Knipe 1989; Cox & Etheridge 1989 for recent reviews), with transitions in the relative intensities of these two observed mechanisms (Hadizadeh & Rutter 1983). The aim of this paper is to show that these two mechanisms may accommodate the sliding on the same fault, with transitions between seismic (cataclastic) and aseismic (pressure solution) slip, during the sliding period.

Accommodation of sliding on faults by mass transfer

The question that must be answered is how to integrate earthquakes and pressure solution in an upper crust deformation model. The problem was already discussed, for example by De Bremaecker (1987), who proposed a clear distinction between two different processes.

An earthquake is a dynamic process, and at a given instant, only part of a long fault (the active part) is in motion. This active part sweeps the length of the fault at high velocity.

After the earthquake, the barriers and asperities that prevented sliding of the fault remain under stress and may be gradually (but very slowly) dissolved by pressure solution.

The aim of this paper is to illustrate, by natural observations, the succession of these two mechanisms on the same fault. Seismic/aseismic transition markers will be described on several faults. The observations suggest that the progressive dissolution of the asperities reduces their strength, breaking them one after the

From Knipe, R. J. & Rutter, E. H. (eds), 1990, *Deformation Mechanisms, Rheology and Tectonics*, Geological Society Special Publication No. 54, pp. 461–473.

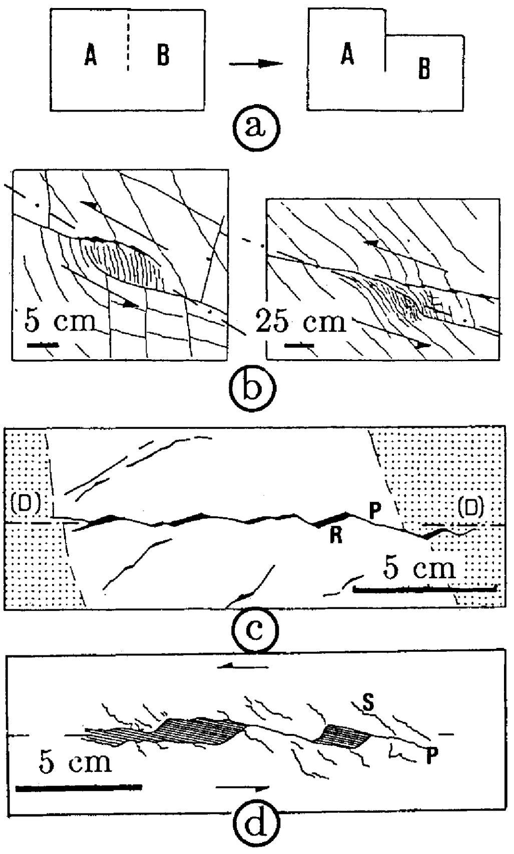

Fig. 1. Accommodation of sliding on faults by mass transfer; (**a**) internal deformation around fault (with volume change between A and B) either by change of density or by mass transfer; (**b**) compressive bridges between first generation en-echelon fractures in a fault zone (limestones in Languedoc); (**c**) pressure solution phenomena along second generation P fractures, associated with domino openings filled with quartz (black) (from gneiss in the Oisans Massif); (**d**) calcite domino opening associated with dissolution on P fractures and with diffuse dissolution around the fault (stylolites, S) (limestones in the Chaînes Subalpines).

other, leading to general fracturing of the whole fault zone i.e. earthquake. To model such behaviour the characteristics of the solution features on and around the faults have first to be discussed.

To accommodate a relatively large fault displacement, various types of internal deformation of the rocks are needed to accommodate volume changes near or along the fault (Fig. 1): such internal deformation is always needed near the termination of faults (Fig. 1a). If a fault consists of an array of en echelon fractures (Fig. 1b), first generation fractures define compressive or tensile bridges (Gamond & Giraud 1982), and internal deformation is also needed within these bridges (Segall & Pollard 1980, Rispoli 1981, Sibson 1989). Finally, if the fault surface is irregular (Fig. 1c), asperities preventing fault movement must also change shape (Elliot 1976), either by dissolution on the surface (Fig. 1c), or dissolution around the fault (Fig. 1d), to accommodate the displacement.

The knowledge of the mechanisms of such a volume change is of major interest for understanding the kinematics of sliding on faults. There are two mechanisms of volume change. As elastic strains probably never exceed 1% (Molnar 1983), a difference in internal deformation, with a volume change of more than 10% between parts A & B (Fig. 1a), and linked to the displacement on the fault, must be accommodated either by a change in density or by mass transfer. The change in density may be associated with cataclastic deformation and then with fast displacement during an earthquake, whereas mass transfer by pressure solution implies a very slow displacement which cannot be associated with the dynamic motion of an earthquake. The partitioning between these two mechanisms will be discussed on several natural examples.

Mass transfer around faults

For various natural examples of faults (from decimetre to kilometre in length) it is possible to show that the total displacement along a fault may be accommodated only (or at least mostly) by mass transfer, (e.g. dissolution of asperities or barriers).

Mass transfer may be demonstrated by a variation of chemical content and density of rocks across a heterogeneously deformed zone. When two portions in an initially homogeneous rock subjected to heterogeneous stress are deformed by pressure solution, they reveal a difference in chemical composition. For example, such chemical composition differences are observed (Fig. 2b) in layer composed of both soluble and insoluble species and cut by a fault (Fig. 2a). This behaviour may be explained as follows. Before faulting, the chemical composition and density of zones A & B were the same. During the slow displacement of the fault, soluble species were removed from the B zone leading to a concentration of insoluble species in this zone and thus to a difference in chemical composition between the two zones A & B. The sector with dissolution has been called the

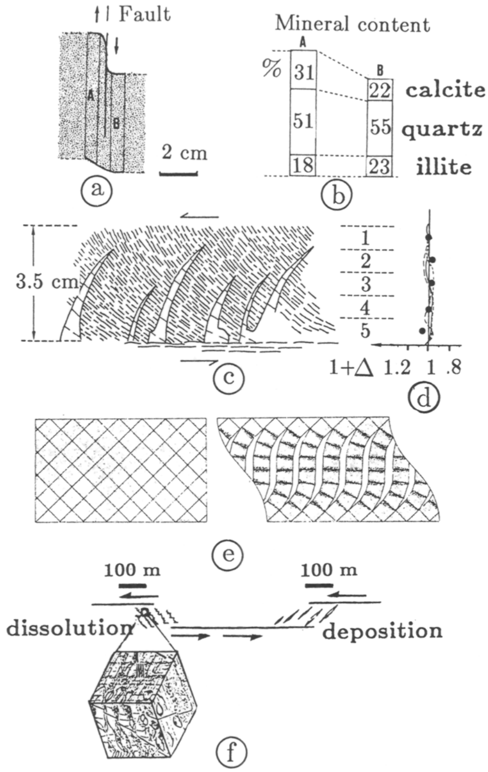

Fig. 2. Balance of mass transfers around faults. (**a**) Layer (slates) composed of both soluble and insoluble species, initially homogeneous, and cut by a fault. A difference in mineral content appears between parts A and B (**b**), giving a means of estimating the mass decrease from A to B (−22%). This value, which may be considered the same as the change in thickness (−23%, without change in density), indicates that the displacement on the fault occurred at a very low (aseismic) rate. (**c**) Shear zone (sandstones), with solution cleavage (s) and tension gashes (t). (**d**) estimated mass transfer between a reference zone (1) and some strips parallel to the displacement (2,3,4,5) by chemical (dots) and geometric (lines) analysis, indicating that each strip may be considered as a closed system. (**e**) Schematic shear zone model with mass transfer from dissolution (concentration of dots) to deposition (white veins), but in natural deformation the spacing of the zones of dissolution is smaller than those of the zones of deposition. (**f**) A 26% volume decrease (estimated by geometric analysis of various markers), within a 100 m wide compressive bridge between en-echelon faults, indicating a large open system for mass transfer (greater than 100 m), with deposition in a tensile bridge (from Morrocan marbles).

'exposed zone' and the zone either with deposition (closed system) or with the initial homogeneous composition (open system) has been called the 'protected zone' (Gratier 1983). The mass decrease of the exposed zone compared with the protected zone is defined as follows:

$$\Delta M / M_0 = (I_p / I_e) - 1,$$

where I_p and I_e are the total percentages of all insoluble minerals in the exposed zone (I_e), and in the protected zone (I_p), ΔM = change in mass, M_0 = initial mass.

This difference in composition is used to calculate the decrease in mass from A to B, and to compare this value with that for the change in thickness (Gratier 1983). In the example given here (Fig. 2a, b) these two values may be considered as the same: −22% (mass) and −23% (thickness), and this without any change in density between A and B. This demonstrates that the entire displacement along the fault was accommodated, at a very slow rate, only by mass transfer. If the volume of redeposited matter can be estimated in the protected zone, this method can also be used to establish the size of the closed system for mass transfer (size for which the dissolved mass is equal to the redeposited mass).

Another example is given in Fig. 2c, where the internal deformation around an irregular fault is accommodated by mass transfer from solution cleavage seams to tension gashes in a shear zone within a small closed system. In such a case, the balance of mass transfer may also be determined by comparative chemical analyses. For example, in Fig. 2d, the chemical analysis of strips of rock (sandstone) parallel to the shear zone displacement were compared with the mineral composition of a reference zone (this reference being away from the zone with curvature of the cleavage and tension gashes). Using the relation given above, the calculation of the volume change of each strip versus the reference zone (Fig. 2d), shows that no mass transfer occurs between these strips. The development of the tension gashes filled with quartz and calcite thus compensates for dissolution of these minerals along the solution cleavage surfaces between the veins, (see also Beach 1974). Such a closed system is confirmed by the geometric analysis of the angles between cleavage (S), veins (T) and the shear zone (Fig. 2d) suggested by Ramsay (1981). A schematic model is given for such behaviour (Fig. 2e), but in natural deformation the spacing of the zones of dissolution is always smaller than those of the zones of deposition (Gratier 1987).

These measurements clearly show that sliding on a fault may be accommodated by solution/deposition with closed mass transfer systems varying from few millimetres to several decimetres. In such a case, the mass transfer must take place by diffusion along paths with high fluid content since the temperature of deformation, deduced from fluid inclusion studies remains below 350°C (Gratier & Vialon 1980). This mass transfer mechanism is attested by other studies using stable isotope and fluid inclusion measurements (Kerrich *et al.* 1978). Various other examples confirm this mechanism for sliding (Carrio-Schaffhauser & Chenevas-Paule 1989), but the calculation of mass transfer using the difference in chemical composition is limited to samples of faults of several decimetres to one metre in size.

To investigate the volume change of rocks along fault systems with much larger closed systems, a method based on the geometric analysis of deformation markers (Ramsay & Huber 1983), is necessary. The geometric analysis of several deformation markers: stylolites, veins, deformed brachiopods in some Morrocan marble quarries (Fig. 2f) indicates a volume decrease of 26%, estimated to be a mean value for several hundred metres in size (Gratier 1976). On the contrary, rocks from other quarries, several hundred metres away show an increase in volume that is more deposition than dissolution. These two kinds of quarries are situated along a large strike alip fault, at the position of compressive and tensile bridge structures respectively. This means that mass transfer may also accommodate very slow displacement on large fault zones (dissolution and/or deposition within bridge structures several hundred metres wide). In this case, fluid displacement must take place by infiltration (Etheridge *et al.* 1984). Such a large displacement of fluids in tectonic or metamorphic regimes is attested by various types of measurements: fluid inclusion studies (Mullis 1983), carbon and oxygen isotope data (Dietrich *et al.* 1983; Rye & Bradbury 1988).
As a conclusion the observations of natural examples show two types of result.

Closed systems vary in size from few millimetres to several decimetres, with mass transfer occurring by diffusion in the shear zone or around a single fault, or, the open system is greater than 100 m in large regional fault zones with extensive area of volume change by dissolution or deposition, and with mass transfer occurring by infiltration.

It is now necessary to discover whether the displacement markers on the fault surfaces, e.g. the striae, also indicate such a very slow rate of sliding.

Mass transfer along the fault surface

Two completely different types of striae must be distinguished. One kind of striation is a mechanical in origin comparable to a mechanical scratch (Goguel 1948; Petit 1987), which may be associated with fast displacement during seismic slip. This kind of striation is well known on natural fault surfaces developed during earthquakes, and it can be reproduced easily in the laboratory when fracturing rocks at high stress level, high strain rate and low temperature (Paterson 1978). Other kinds of striae which are also very common on naturally formed faults (Durney & Ramsay 1973; Means 1987), are not scratches at all, but are rather crystal fibres, (Figs. 3 & 4). The development of such fibres may be explained as follows (Elliott 1976). If two rock parts are separated by an irregular fault surface, a slow displacement of one part of the rock may be accommodated by the dissolution of the asperities that prevent the displacement. At the same time, deposition may occur in the cavities created by the displacement on the fault. The crystals growing within these cavities sometimes have a typical fabric (fibre minerals), first described by Ramsay (1980) and explained by him as linked to a succession of microcrack openings followed by their immediate sealing. The redeposited crystalline material is derived by pressure solution from the rocks matrix (see Figs 1 & 2). On the natural example given, Fig. 3a, the mean width of each crack opening is about 10 to 50 μm and, as each crack is limited to one or two crystals, several thousand crack-events would be required to achieve a 1 cm displacement. The length of such crystal fibres may reach 20 cm on other examples (Fig. 4).

Another interesting feature is shown in Fig. 3a. Pressure solution–deposition slip on this fault is attested by crystal fibres of quartz and feldspars but some of these fibres are clearly kinked or broken and cut by cataclastic scratching faults. This means that cataclastic deformation may be found, on the same fault, associated with pressure solution markers. The presumed displacement rate on this fault has been plotted Fig. 3a with two successive processes: a very slow pressure solution slip then a fast cataclastic event.

A major problem has to be overcome. When considering microcracks within crystal fibres on a given fault, should these microcracks be thought of as part of a seismic rupture process

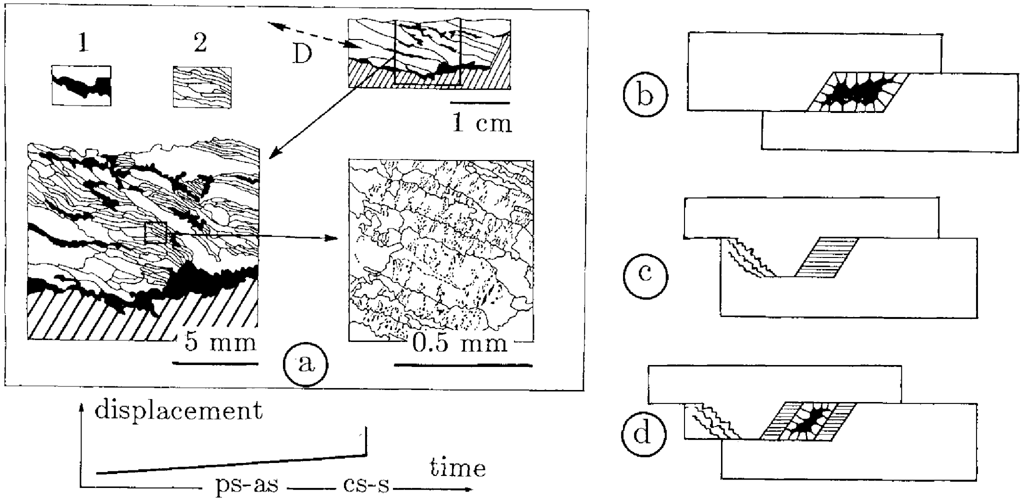

Fig. 3. Transition markers between seismic/aseismic sliding on the same fault; (**a**) Mineral fibres (aseismic slip), cut by small fractures (cataclastic event). Black area (1), cataclastic material; white area (2), crack-sealed quartz fibres; hatched area, gneiss, with general setting (top), and details of the fibres (bottom). The cavity opened by sliding is filled by mineral fibres indicating the displacement direction (D). The mechanism of such a sealing process is by successive microcracks opening (10 to 50 μm wide) immediately sealed (see Ramsay 1980; Cox & Etheridge 1983). Each microcrack is limited to only one or two crystals, each centimetre of displacement implies at least 10 000 microcrack events, during aseismic sliding with the rate limited by the solution–deposition rate (sample from gneiss in the Oisans massif). The diagram below the drawing gives the presumed sliding rate to time relation with successively pressure solution aseismic slip (ps-as) then cataclastic seismic slip (c-ss). (**b**) Euhedral radial crystals in a cavity opened by fault sliding indicate fast (wide) opening (seismic event). (**c**) Mineral fibres indicate slow (aseismic) slip. (**d**) Association of these two types of growth indicates aseismic to seismic transition.

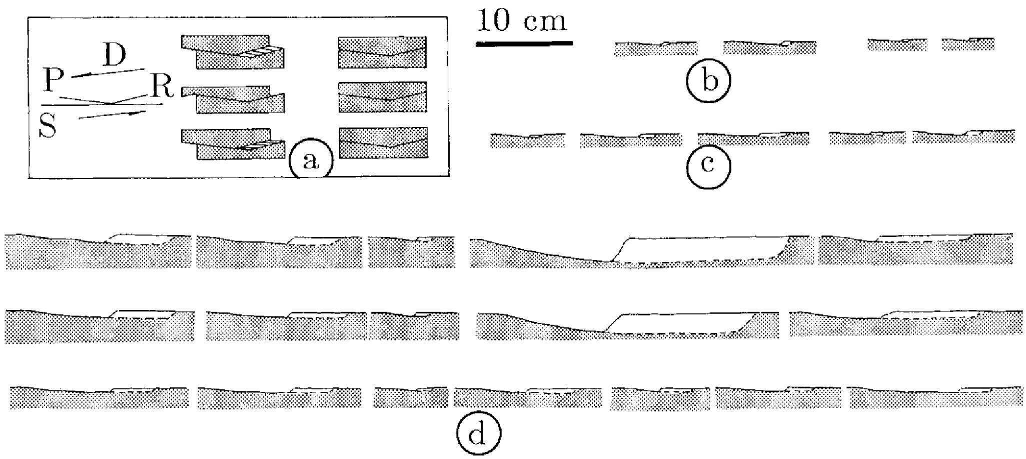

Fig. 4. (**a**) Various geometric relations between R and P fractures, displacement direction (D) and mean orientation of fault (S), illustrating various mass transfer systems: open systems with deposition (top) or with dissolution (middle) and closed systems (bottom). (**b,c,d**) Various natural geometric features in closed systems, measured on three different faults in the Chaînes Subalpines, on the same limestones (Upper Jurassic). White area, fibres; dotted area, limestones. On the faults (c) (5 fibres) and (d) (17 fibres) the length of the fibres which indicates the displacement accommodated by pressure solution, vary along the fault surface; (dotted part, limestones; white part, tectonic calcite fibres).

which occurred on this fault? According to the approach described here, the answer is no. The difference between seismic and aseismic deformation must be determined with reference to a ratio between displacement of an event and size of the asperities along the fault. If an event breaks several asperities, it is a cataclastic-seismic event for the fault. If an event only occurs within an asperity without significant associated displacement, it is not. This process of successive microcracks with immediate sealing, associated with dissolution, must then be considered as an aseismic slip. Of course some of these microcracks may be induced by the regional seismicity but the seismic/aseismic transition must be defined for a given fault.

Another indicator of such a transition between seismic and aseismic deformation is the observation of the change in the type of crystal growth within the cavities opened by sliding. If sliding occurs at very low rate, successive microcracks affecting only some grains may be immediately sealed with the typical fabric described above (Fig. 3c). On the contrary, if a relatively large displacement occurs, large cavities are opened and filled after their opening, by euhedral crystals with radial growth (Fig. 3b). An association of the two types of growth may be observed, attesting to a aseismic to seismic transition (Fig. 3d).

Volume transfer balance along faults may also be discussed through a simple geometrical plane-strain model in which volume loss is assumed to combine with sliding on the first and second-generation fractures (Gamond 1987). Left-lateral discontinuity (Fig. 4a) is considered to have two types of en echelon fractures, P & R, with respectively compressive and tensile behaviour. The volume transfer balance can be determined between the volume dissolved under transpression (on P fracture) and the volume deposited in the gaps (on R fracture) opened by sliding on the fault. On a section perpendicular to the fault surface, the change in area depends upon the geometry (P and R) of the asperities and their spacing, and upon the angle between the displacement (D indicated by the fibres) and the mean surface of the fault (F). Simple theoretical examples are given on Fig. 4a: open system with only dissolution (D parallel to R), or only deposition (D parallel to P), and closed system (D parallel to F).

In the case of closed systems (fibres parallel to the major discontinuity) various measurements were performed on natural faults, using a shape tracer in order to determine typical geometric features of the asperities. Measurements on a same fault surface in limestones, for example Fig. 4d, show that the length of the asperities (A) varies from 5–30 cm, as the length of the fibres (F) varies from 1.5–20 cm. The F/A ratio expresses the portion of the asperity dissolved during aseismic slip. Its mean value is 0.4, most values being between 0.5 and 0.35, with extreme values of 0.7 and 0.25. Measurements on other faults (Fig. 4b & c) in the same limestones (Upper Jurassic), but in different areas, show lower range of values for the asperity and fibre length, with lower F/A ratio (0.3). The spacing between the asperities is more difficult to estimate, but is probably of the same order of magnitude as the size of the asperities. The angle between dissolution surface and mean displacement on fault remains low, with a narrow range (from 5 to 15°). Conversely, the angle between the deposition surface and the fault varies more widely from 30 to 80°. On the given examples the change of length of the fibres along each fault surface is either random (Fig. 4c & 4d) or constant (Fig. 4b). Other measurements (Gamond 1983) indicated progressive decrease in the length of the fibres near the termination of a fault.

These features may be explained as follows. The variation in asperity size is linked to the geometry of the first and second-generation fractures along the fault zone and will be discussed later. A variation in fibre length may be due to different processes:

(i) some cavities may begin to form while sliding is still active, within bridge zones which were undergoing internal deformation (Fig. 5a)
(ii) some asperities may break during the sliding (Fig. 5b)
(iii) a decrease in displacement along the fault always appears near the end of the fault (Fig. 5c), associated with a heterogeneous volume change in the vicinity (see Fig. 1a).

From these observations, a model is derived which predicts why cataclasis versus pressure solution takes place, and the factors which affect or control these deformation processes can be determined.

Change in sliding mechanisms with time

The cataclastic shear fault strength is considered to be the yield stress needed to fracture a large part of the fault (earthquake). This cataclastic shear strength is proportional to the total length of the asperities along the fault. Progressive breaking of some asperities during aseismic slip (Fig. 5b), will lead to progressive decrease in this shear seismic strength.

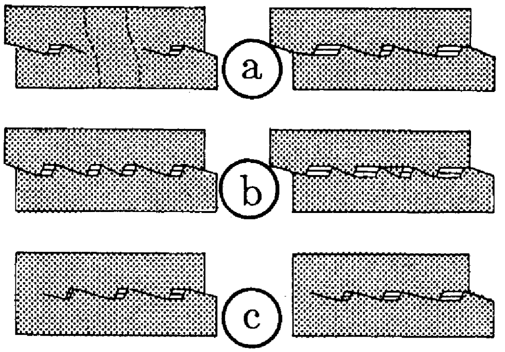

Fig. 5. Some explanations of the variation in the fibre length which can be observed on a given fault (see Fig. 4d). (**a**) Development of a new cavity in an internal deformation zone, while sliding is going on. (**b**) Breaking of an asperity during sliding. (**c**) Progressive decrease in displacement near the end of the fault (see Figs 1a and 2a).

Since the total length of the asperities array also decreases by progressive dissolution, this cataclastic shear strength value is also linked to the dependence of the two sliding mechanisms (cataclastic or pressure solution) on the asperity geometry. To study this effect, a simple model of sliding by pressure solution may be constructed for asperities of very simple shape. Pressure solution is a mechanism which implies successive processes: dissolution, transfer (by diffusion or infiltration), and deposition. A variety of thoeretical pressure solution creep laws have been established in which the overall rate is controlled by the slowest process (Raj & Ashby 1971; Rutter 1976 & 1983; Elliott 1976; Raj 1982; Etheridge *et al.* 1984; Gratier & Guiguet 1986; Gratier 1987). The same analysis was used to establish a theoretical pressure solution sliding law (Rutter & Mainprice 1979). With the different types of limiting processes, the sliding rates ($\dot{\gamma}$) are as follows:
when infiltration rate is the limiting process:

$$(I)\text{: } \dot{\gamma} \propto Kic$$

when kinetics of reaction is the limiting process:

$$(R)\text{: } \dot{\gamma} \propto k_c \sigma_n v / RT$$

when diffusion rate is the limiting process:

$$(D)\text{: } \dot{\gamma} \propto Dc\sigma_n w / RThd\alpha$$

where kc = kinetics of reaction; v = molar volume; R = gas constant; T = temperature (K); c = concentration of soluble component; σ_n = difference in normal stress between dissolution and deposition sites; K = permeability coefficient; i = pressure gradient; w = width of path transfer; D = diffusion coefficient; h = asperity height; d = asperity length, α = numerical coefficient.

Details of these laws will not be discussed here. For the purpose of the work, the most interesting relation is that between the sliding rate and the asperity length (d). In the first two relations (I and R), the sliding rate is not dependent on the asperity length. In the third one (D) the sliding rate depends upon the distance of mass transfer from dissolution to deposition, more or less linked to the asperity length. Variations in distance of mass transfer during the sliding also depends on the crystal growth mechanism and on the distribution of dissolution between the two parts of the fault (hatched upper part and shaded lower part, Fig. 6). If only one part of the fault is dissolved e.g. the upper part (Fig. 6), the distance of mass transfer decreases with the progressive sliding if each successive growth increment occurs at the fibres/upper part limit, and this distance remains constant if each growth increment occurs at the fibres/lower part limit (Fig. 6). In these cases, the numerical coefficient α varies from $1/d$ to 1, d being the length of the asperities on the upper part of the fault. The variation in the distance of mass transfer with the progressive sliding is more complex if dissolution occurs on both parts of the fault, and/or if fibres grow by the syntaxial mechanism (each new increment located in the median part of the fibres, see Durney & Ramsay 1969).

Assuming a linear relation between stress (or pressure gradient) and sliding rate (Rutter 1976; Urai *et al.* 1986), the stress values needed to drive a constant sliding rate may or may not change during the progressive sliding, depending on the sliding laws. As the strain energy consumed during pressure solution sliding depends on this stress value (Elliott 1976), at constant sliding rate, the change in rate of work with time differs depending on the laws considered. The energy needed to break the asperities, however, is always dependent upon the asperity length.

The energy values needed to accommodate either cataclastic or pressure solution sliding may be compared. When the two energy values vary in a similar manner during the progressive sliding, stable sliding is expected (e.g. pressure solution aseismic sliding Fig. 6), whereas, when the two energy values vary in a different manner, an unstable process is expected (Fig. 6), with successive mechanisms using minimum energy (aseismic/seismic slip during the progressive sliding).

The geometry of the asperities and fibres

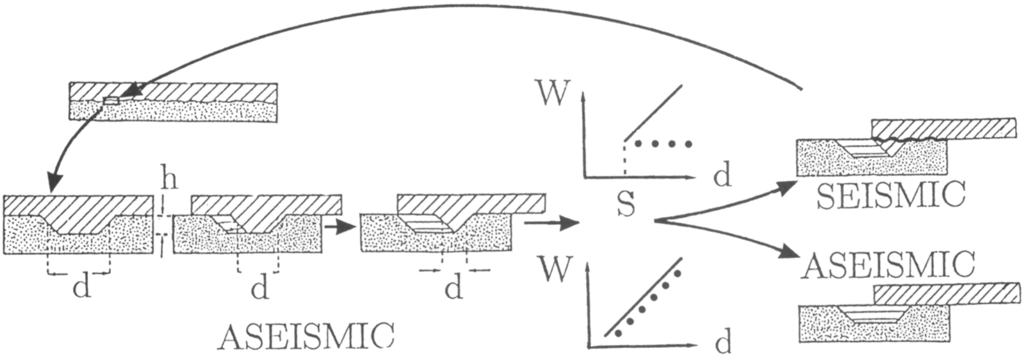

Fig. 6. Theoretical model of pressure solution sliding with two types of change. Stable aseismic sliding (total dissolution of the asperity), when the energy needed for pressure solution (dotted line) or breaking (solid line) of the asperity is the same during progressive sliding (bottom diagram), or unstable sliding (aseismic/seismic transition at S point), when the energy needed for pressure solution (dotted line) is not dependent upon asperity length and when the energy for breaking the asperity (solid line) becomes lower than the energy needed for pressure solution (top diagram).

measured on several faults may be used to estimate the ratio between displacement by pressure solution (length of fibres) and total (initial) length of asperities (0.3 to 0.4, Fig. 4).

Change in sliding mechanisms with depth

By way of example, we will examine the results of deformation in the upper crust through a large crustal thrusted region, such as the Alpine chain (Fig. 7), where significant internal deformation was associated with thrust motion. In this example, horizontal contraction of the Subalpine sedimentary cover varies from 5% to 30% when homogeneous deformation (folds and minor faults) of large elements (15 × 15 km of initial horizontal extent) is considered (Gratier *et al.* 1989). Cataclasis (zones where rocks are broken into lithic fragments) is fairly rare, and is generally located near fault zones. This mechanism alone cannot account for the whole internal deformation of the 15 × 15 km elements of the upper crust. Large diffuse deformation associated with folds, shear zones or faults is accommodated by pressure solution. The markers of this mass transfer deformation are pitted pebbles in molassic basins (McEwen 1981), stylolites in limestones, or solution cleavage in slates, metamorphic rocks and granites (Gratier & Vialon 1980). These dissolution markers are always associated with veins (deposition), and the study of the fluid inclusions trapped in these veins (Bernard *et al.* 1977; Jenatton 1981) indicates the depth of deformation from 1 to 10 km (Fig. 1).

Cataclasis and pressure solution are clearly associated in thrust sheet motion (e.g. limestones in the Chaînes Subalpines; Fig 8b), and in other orogenic belts (see for example Geiser 1988). Cataclasis is located near the ramp of the basal discontinuity whereas the deformation of the sheet moving over this ramp is accommodated by pressure solution (stylolites, solution cleavage and veins). As this internal deformation of the sheet is always necessary to accommodate its displacement this means that the sheet motion was slow enough and continuous to be associated with a diffuse creep process at depths lower than 5 km. In slates of the sedimentary cover of the Oisans massif the amount of strain accommodated by pressure solution over a large region (several km) may reach high values such as 0.4 (contraction) × 2.5 (extension), at depth from 5 to 10 km (deduced from fluid inclusion studies as for the preceding example, Gratier & Vialon 1980). Following the finite strain/time relations given by Pfiffner & Ramsay (1982), and taking into account the regional geological constraints, such deformation could take place over 3×10^6 years at a strain rate of $10^{-14}s^{-1}$. Other examples in the Alps indicate pressure solution creep at depths up to 15–20 km.

Using the theoretical sliding laws, and both natural observations and experimental results (see also Cox & Etheridge 1989), the order of magnitude of the pressure solution shear strength of rocks may be estimated through a typical section (sediments with high porosity, limestones, slates, metamorphic rocks (gneiss then schists), then granitic rocks), for a strain rate of $10^{-14}s^{-1}$. Classical values were used for transfer coefficients (Rutter 1976), with values of the size of the closed system for mass transfer

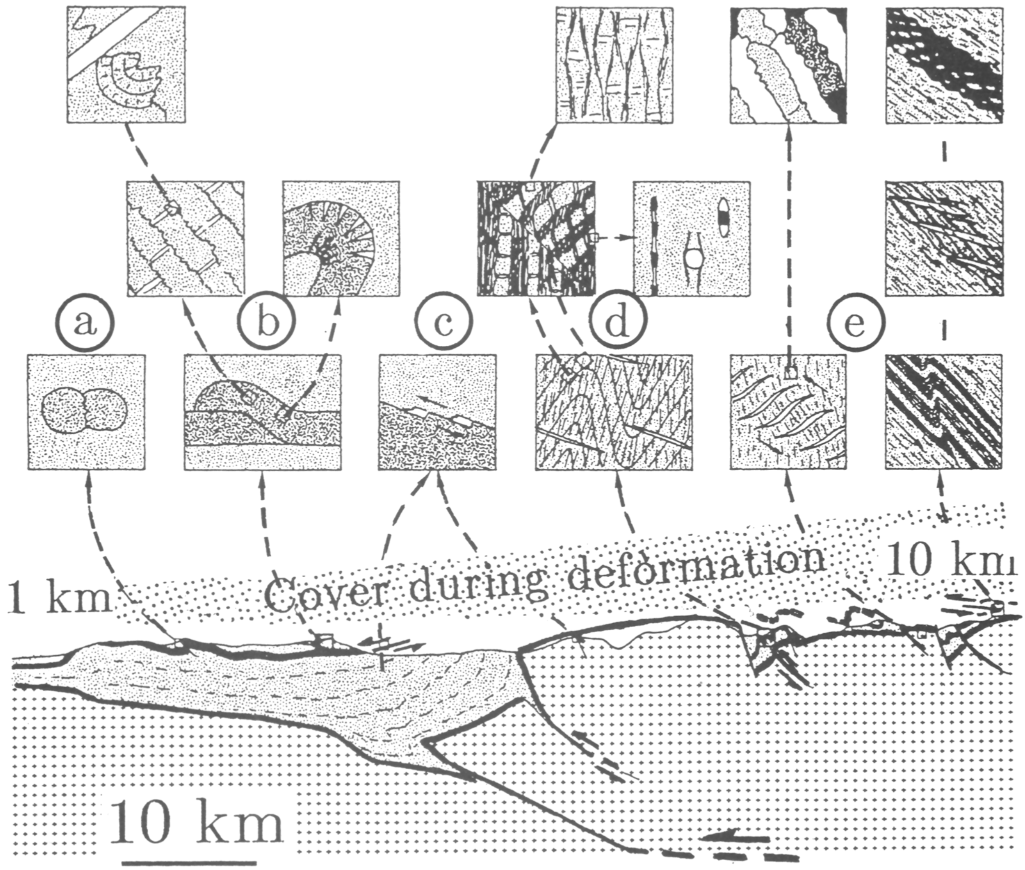

Fig. 7. Internal deformation of a thrust domain in the Alpine Chain; cross section from thrusting basement (dotted area, Oisans) to deformed sedimentary cover (shaded area, Chaînes Subalpines) near the passive margin. The diffuse deformation associated with folds (**a,b,d**), shear zones (**e**), and fault (**c**), is accommodated by pressure solution, attested by pitted pebbles (a, molassic basins), stylolites (b, limestones), solution cleavage (d,c, slates, metamorphic and granitic rocks), and deposition in tectonic veins. The process occurs at all scales from macro (bottom) to micro (top) scale. The thickness of the cover during the deformation was deduced from fluid inclusion studies (in tectonic veins).

from 0.1–10 mm (increasing values from granite-gneiss to sediments, shales, and limestones). A wide range of values was obtained, with zones of very low strength (which may be associated with flat shear zones either in extensive or compressive deformation), and with a mean value more than one order of magnitude below the cataclastic shear strength (Molnar & Tapponnier 1981; Sibson 1982). Following this approach, pressure solution creep is equivalent to a Newtonian fluid with orders of magnitude of viscosity from 10^{19} to 10^{21} Pa s, with at least one or two orders of magnitude of uncertainty linked to the meagre knowledge of the behaviour of a fluid phase trapped between two stressed solids. For example, from experimental results (Gratier & Guiguet 1986) a value of 10^{17} Pa s has been extrapolated for deformation of a fine-grained quartz aggregate in water at about 350°C. The two profiles (pressure solution and cataclastic shear strength) are drawn on Fig. 8. The strength of the upper crust is probably between these two profiles: this strength may change both with time and with space depending on the fluid content and on the mean strain rate of the upper crust.

Conclusions

A single fault surface often displays evidence of coupled seismic and aseismic slip processes.

For a given fault zone, the observations

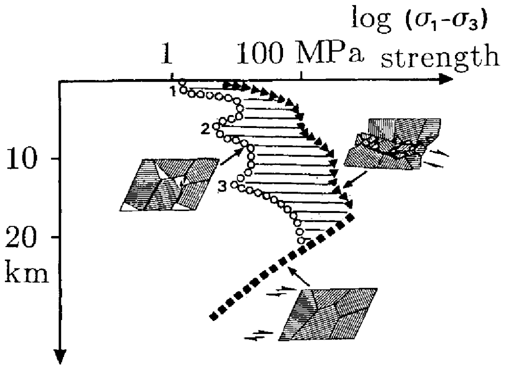

Fig. 8. Strength profile through the upper crust taking into account the two (observed) deformation mechanisms: cataclastic deformation (law derived from Mohr–Coulomb behaviour, triangles), and pressure solution creep (laws derived from theoretical, experimental and natural observations, circles), at geological strain-rate: 10^{-14} s^{-1}, (equivalent to a change in length of 3 mm a^{-1} for an element of 10 km size). Dislocation creep mechanisms (squares) operate within the lower crust. The model of the typical crustal section is composed of different rocks, (see Fig. 7): from top to bottom, sediments with high porosity (1), limestones, slates (2), metamorphic rocks (gneisses then schists (3)), granitic rocks. The strength values vary considerably with depth, depending on various parameters (possibility of mass transfer, solubility of solid in fluid, mean displacement rate at the boundaries of the system, etc.).

suggest the following sequence which integrates earthquakes and pressure solution mechanisms in an upper crust deformation model.

(a) The development of first generation fractures generally leads to arrays of *en-echelon* fractures, either P or R fractures, depending on the deformation conditions, (Fig. 9a).

(b) Second generation fractures appear in the bridges between the previous ones. These second generation fractures generally only need an incremental displacement to achieve the connection with the first ones (Fig. 9b).

(c) As the rate of displacement is generally imposed at the boundaries of the system, the behaviour of the zone depends on various parameters classified by order of importance: fluid content, solubility of certain minerals of the rocks in this fluid, geometry of mass transfer path and value of the coefficient of mass transfer (by diffusion or infiltration), size of the closed system, type of limiting process (reaction kinetics, mass transfer rate), stress level, temperature and pressure conditions. If dissolution–deposition mechanism can accommodate the imposed displacement rate, the motion on the fault is limited by the rate limiting solution-deposition process (Fig. 9c). If accommodation by mass transfer is not sufficient, the stress level increases and an earthquake is likely to occur (Fig. 9c′).

(d) During this very slow sliding, local fracturing of some asperities may occur without

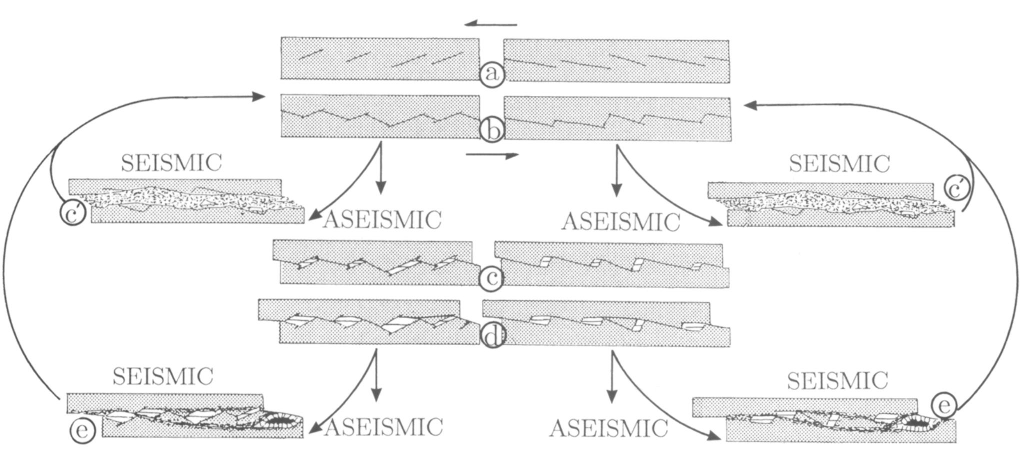

Fig. 9. Seismic/aseismic transition on two fault zones. (**a**) First generation fractures, either R (left) or P (right) fractures. (**b**) Second generation fractures (within compressive or tensile bridges), leading to an irregular fault surface. (**c**) Pressure solution sliding, with eventually local (aseismic) breaking of some asperities (**d**). (**e**) General fracturing of a large part of the fault (cataclastic displacement greater than asperity length, earthquake). If the two parts of the rock remain imbricated after seismic displacement, pressure solution sliding may be reactivated (b). An earthquake may occur (c′) at any time, if pressure solution sliding is not able to reduce the stress level linked to the constant displacement rate at the boundaries of the system.

general fracturing (Fig. 9d) but this local fracturing gradually lowers the cataclastic shear strength of the fault zone.

(e) The system behaviour depends on the variation in energy needed either to dissolve the asperities or to break them. As this energy change does not always have the same dependence on the geometry of the asperities (Fig. 6), the system may behave either as a stable system (dissolution of all the asperities) or as an unstable system (general breaking of a large part of the fault, earthquake, Fig. 9e). This cataclastic deformation may be followed by a new imbrication of the two parts of the fault, the process being reinitialised at the step corresponding to Fig. 9b.

The model implies alternately fast, localized processes (seismic-cataclasis) and large slow creep processes (aseismic-pressure solution). The same cataclastic/pressure solution coupled processes may occur to accommodate the deformation around the fault. Successive aseismic/seismic slip cycles probably lead to softening of the fault, since the size and strength of the asperities decrease. As the deformation of rocks by pressure solution is always associated with the development of a tectonic differentiation (Gratier 1987), softened and hardened bands are developed which may change the location of sliding deformation.

Finally, one of the major problems about earthquakes is to know whether such a phenomenon can be avoided. This study shows that natural sliding on faults is often accommodated by an aseismic mechanism. The question is: is it possible to increase this aseismic rate of sliding artificially in order to release the stress concentration? Following experimental results on pressure solution, the best strain rate activation is obtained when using a very good solvent, either water with salts (Rutter 1976; Raj 1982; Pharr & Ashby 1983) or NaOH and NH_4Cl solutions with quartz or calcite respectively (Gratier & Guiguet 1986). One method of preventing earthquakes could be to inject such good solvents of at least one of the minerals of the rocks, through a fault zone (at very low pressure of course to avoid hydraulic fracturing).

We thank D. Prior, A. McCaig, P, Molnar and an anonymous reviewer for their careful revision of this paper.

References

BEACH, A. 1974. A geochemical investigation of pressure solution and the formation of veins in a deformed Greywacke. *Contributions to Mineralogy and Petrology*, **46**, 61–68.

BERNARD, D., GRATIER, J. P. & PECHER, A. 1977. Application de la microthermométrie des inclusions fluides des cristaux syncinématiques à un problème tectonique. *Comptes-Rendus Sommaires de l'Académie des Sciences Paris*, **5**, 284–288.

BRACE, W. F. & KOHLSTEDT, D. L. 1980. Limits on lithospheric stress imposed by laboratory experiments. *Journal of Geophysical Research*, **85**, 6248–6252.

BYERLEE, J. D. 1968. Brittle-ductile transition in rocks. *Journal of Geophysical Research*, **73**, 4741–4750.

CARRIO-SCHAFFHAUSER, E. & CHENEVAS-PAULE, F. 1989. Quantification géométrique de la dissolution-cristallisation liée à une déformation cassante. *Bulletin de la Société Géologique de France*, **8**, 597–604.

CARTER, N. T. & TSENN, M. C. 1987. Flow properties of continental lithosphere. *Tectonophysics*, **136**, 27–63.

CHEN, W. P. & MOLNAR, P. 1983. Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere. *Journal of Geophysical Research*, **88**, 4183–4214.

COX, S. F. & ETHERIDGE, M. A. 1983. Crack-seal fibre growth mechanisms and their significance in the development of oriented layer silicate microstructures. *Tectonophysics*, **92**, 147–170.

—— & —— 1989. Coupled grain-scale dilatancy and mass transfer during deformation at high fluid pressures: examples from Mount Lyell, Tasmania. *Journal of Structural Geology*, **11**, 147–162.

DE BREMAECKER, J. Cl. 1987. Thrust sheet motion and earthquake mechanisms. *Earth and Planetary Science Letters*, **83**, 159–166.

DIETRICH, D., MCKENZIE, J. A. & SONG, H. 1983. Origin of calcite in syntectonic veins as determined from carbon-isotope ratios. *Geology*, **11**, 547–551.

DURNEY, D. W. & RAMSAY, J. G. 1973. Incremental strains measured by syntectonic crystal growths. *In*: DE JONG, K. A. & SCHOLTEN, R. (eds) *Gravity and tectonics*. J. Wiley, New York, 67–96.

ELLIOTT, D. 1976. The energy balance and deformation mechanisms of thrust sheets. *Philosophical Transaction of the Royal Society*, **283**, 289–312.

ETHERIDGE, M. A., WALL, V. J., COX, S. F. & VERNON, R. H. 1984. High fluid pressure during regional metamorphism and deformation: implications for mass transport and deformation mechanisms. *Journal of Geophysical Research*, **89**, 4344–4358.

GAMOND, J. F. 1983. Displacement features associated with fault zones: a comparison between observed examples and experimental models. *Journal of Structural Geology*, **5**, 33–45.

—— 1987. Bridge structures as sense of displacement criteria in brittle fault zones. *Journal of Structural Geology*, **9**, 609–620.

—— & GIRAUD A. 1982. Identification des zones de failes à l'aide des associations de fractures de second ordre. *Bulletin de la Société Géologique de France*, **7**, 24, 755–762.

GOGUEL, J. 1948. *Introduction à l'étude mécanique des déformations de l'écorce terrestre*. Memoires du Service de la Carte Géologique France.

GEISER, P. A. 1988. Mechanisms of thrust propagation: some examples and implications for the analysis of overthrust terranes. *Journal of Structural Geology*, **10**, 829–845.

GRATIER, J. P. 1976. Deformation et changement de volume dans un marbre à stylolites de la ràgion de Rabat (Maroc). *Bulletin de la Société Géologique de France*, **7**, 18, 1461–1469.

—— 1983. Estimation of volume change by comparative chemical analyses in heterogeneously deformed rocks (folds with mass transfer). *Journal of Structural Geology*, **5**, 329–339.

—— 1987. Pressure solution-deposition creep and associated tectonic differentiation. *In*: JONES, M. E. & PRESTON, R. M. (eds) *Deformation of Sediments and Sedimentary Rocks*. Geological Society, London, Special Publication, **29**, 25–38.

—— & GUIGUET, R. 1986. Experimental pressure solution on quartz grains: the crucial effect of the nature of the fluid. *Journal of Structural Geology*, **8**, 845–856.

—— & VIALON, P. 1980. Deformation pattern in a heterogeneous material: folded and cleaved sedimentary cover immediately overlying a crystalline basement (Oisans, France). *Tectonophysics* **65**, 151–180.

——, MENARD, G. & ARPIN, R. 1989. Strain-displacement compatibility and restoration of the Chaines Subalpines of the Westerns Alps. *In*: COWARD, M. P., DIETRICH, D. & PARK, R. G. (eds) *Alpine Tectonics*. Geological Society, London, Special Publication, **45**, 65–81.

HADIZADEH, J. & RUTTER, E. H. 1983. The low temperature brittle-ductile transition in a quartzite and the occurrence of cataclastic flow in nature. *Geologische Rundschau*, **72**, 493–509.

JENATTON, L. 1981. *Microthermométrie des inclusions fluides des cristaux associés à l'ouverture des fentes alpines*. Thesis, University of Grenoble.

KERRICH, R., BECKINSALE, R. D. & SHACKLETON, N. J. 1978. The physical and hydrothermal regime of tectonic vein systems: evidence from stable isotope and fluid inclusion studies. *Neues Jahrbuch fur Mineralogie Abhandlungen*, **131**, 225–239.

KIRBY, S. H. 1983. Rheology of the lithosphere. *Reviews of Geophysics and Space Physics*, **21**, 1458–1487.

KNIPE, R. J. 1989. Deformation mechanisms: recognition from natural tectonites. *Journal of Structural Geology*, **11**, 127–143.

MCEWEN, T. J. 1981. Diffusional mass transfer processes in pitted pebble conglomerates. *Journal of Structural Geology*, **3**, 25–38.

MEANS, W. D. 1987. A newly recognized type of slickenside striation. *Journal of Structural Geology*, **9**, 585–590.

MOLNAR, P. 1983. Average regional strain due to slip on numerous faults of different orientations. *Journal of Geophysical Research*, **88**, 6430–6432.

—— & TAPPONIER, P. 1981. A possible dependence of tectonic strength on the age of the crust in Asia. *Earth and Planetary Sciences Letters*, **52**, 107–114.

MULLIS, J. 1983. Transported metamorphism of the external parts of the central Alps, shown by fluid inclusions. *Symposium of European Current Research on Fluid Inclusions, Orleans*.

PATERSON, M. S. 1978. *Experimental rock deformation: The brittle field*. Springer-Verlag, Berlin.

PETIT, J. P. 1987. Criteria for the sense of movement on fault surface in brittle rocks. *Journal of Structural Geology*, **9**, 597–608.

PFIFFNER, A. O. & RAMSAY, J. G. 1982. Constraints on geological strain rates: arguments from finite strain states of naturally deformed rocks. *Journal of Geophysical Research*, **87**, 311–321.

PHARR, G. M. & ASHBY, M. F. 1983. On creep enhanced by a liquid phase. *Acta metallurgica*, **31**, 129–138.

POIRIER, J. P. 1985. *Creep of crystals, high-temperature deformation processes in metals, ceramics and minerals*. Cambridge Univ. Press.

RAJ, R. 1982. Creep in polycrystalline aggregates by matter transport through a liquid phase. *Journal of Geophysical Research*, **87**, 4731–4739.

—— & ASHBY, M. F. 1971. On grain boundary sliding and diffusional creep. *Metallurgical Transactions*, **2**, 1113–1128.

RAMSAY, J. G. 1980. The crack-seal mechanism of rock deformation. *Nature*, **284**, 135–139.

—— 1981. Shear zone geometry: a review. *Journal of Structural Geology*, **2**, 83–99.

—— & HUBER, 1983. *Modern structural geology, vol. 2*. Academic Press.

RISPOLI, R. 1981. Stress fields about strike-slip faults inferred from stylolites and tension gashes. *Tectonophysics*, **75**, T29–36.

RUTTER, E. H. 1976. The kinetics of rock deformation by pressure solution. *Philosophical Transactions of the Royal Society*, **A283**, 43–54.

—— 1983. Pressure solution in nature, theory and experiment. *Journal of the Geological Society, London*, **140**, 725–740.

—— & MAINPRICE, D. H. 1979. On the possibility of slow fault slip controlled by diffusive mass transfer process. In: STILLER, H. (ed.) *Symposium on Physics of Earthquake sources*. Gerlands Beiter Geophysik, Leipzig, **88**, S154–162.

RYE, D. M. & BRADBURY, H. J. 1988. Fluid flow in the crust: an example from a Pyrenean thrust ramp. *American Journal of Science*, **288**, 197–235.

SEGALL, P. & POLLARD, D. D. 1980. Mechanics of discontinuous faults. *Journal of Geophysical Research*, **85**, 4337–4350.

SIBSON, R. H. 1982. Fault zone models, heat flow, and the depth distribution of earthquakes in the continental crust of the United States. *Bulletin of the Seismological Society of America*, **72**, 151–163.

—— 1989. Earthquake faulting as structural process. *Journal of Structural Geology*, **11**, 1–14.
URAI, J. L., SPIERS, C. J., ZWART, H. J. & LISTER, G. S. 1986. Weakening of rock salt by water during long-term creep. *Nature*, **324**, 554–557.