PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Understanding the fundemental physics governing the evolution and dynamics of the Earth'crust and ice sheets

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THE ROYAL SOCIETY

Understanding the fundamental physics governing the evolution and dynamics of the Earth's crust and ice sheets

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There is a wide range of problems in geophysics, from earthquake prediction to the driving forces of plate tectonics, where it is necessary to understand how rocks deform. The materials science approach to understanding these geophysical processes is based on the premise that the macroscale behaviour of rock is governed by microscale interactions. Rock, as a material, will deform under an applied stress elastically, plastically, by fracturing and brittle (cataclastic) flow, and frictional sliding on a fault. The magnitude and direction of the applied stress, the rate and duration of loading, ambient pressure and temperature, the presence of fluids and previous deformation history all control the overall mechanical response. Deformation can be remotely monitored by physical measurements such as elastic wave velocities and electrical conductivity. The emerging subject of rock physics seeks to integrate the disciplines of rock mechanics with rock physical property measurements. The challenge for rock physics is to understand through experiments and modelling, microscopic rock behaviour and apply this to large-scale phenomena. The future is in holistic laboratory experiments, where a wide range of physical parameters is measured concurrently during the deformation experiment. This is important not only in a material science sense, but crucially these parameters are monitored by geophysical techniques and so laboratory experiments can be related to crustal processes.

This paper reviews how the materials science approach is applied to problems in geophysics (particularly for brittle deformation), the experimental methodologies employed and how the question of scaling from the laboratory to a planetary scale is addressed. This approach is considered in terms of the evolution and dynamics of the Earth's crust and its ice sheets as they represent the components of the solid Earth with which humankind directly interacts.

Keywords: rock physics; Earth's crust; fracture mechanics; ice mechanics; ice sheets; scaling

1. Introduction

In the evolution and dynamics of the Earth's crust, rock deformation on all scales plays a crucial role. Rock deformation is required for orogeny (mountain building) where both brittle and ductile processes are likely to be operating; crustal movements cause earthquakes; pervasive dilatant cracking in the crust enables the storage and movement of fluids important to humankind, such as petroleum and water, and pollutants. Considerable research into rock deformation has been undertaken in the last

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half century. Results from experimental rock mechanics made a significant contribution to continental drift theory when Griggs (1939) argued for convection currents in the mantle as the most likely mechanism for periodic orogeny. The concept of solid state flow of the upper mantle was introduced at an early stage in the development of plate tectonic theory (McKenzie 1968). However, many issues of the fundamental physics of rock deformation are not resolved: to understand fracturing of the crust and the associated problem of fluid flow are major challenges (figure 1). Fracturing of the crust is principally in a compressive stress field and is highly nonlinear; the crustal stress field is the least known of any of the major geophysical fields; and the crust is structurally complex. The last few years have seen the collapse of the earthquake prediction programme in the USA and the UK government's refusal to sanction the building of a new underground radioactive waste repository. Both of these can be seen as a consequence of the lack of understanding of the physics governing complex behaviour of the crust, as well as a lack of confidence in the ability of science to offer solutions (Gillott & Kumar 1995).

Ice sheets and the sea ice cover in the polar regions play key roles in the behaviour of the Earth's ocean–atmosphere system since their dynamic and radiative properties strongly influence the distribution of heat and mass within that system. In recent years there has been considerable societal concern that global atmospheric warming could cause a substantial rise in sea level. The largest source of uncertainty is the contribution of melting ice sheets (Warrick & Oerlemans 1990). There has also been concern that the break-up of smaller ice shelves on the Antarctic Peninsula is heralding the disintegration of the West Antarctic Ice Sheet (Mercer 1978). The ability to measure the precipitation-ablation mass-balance change of ice sheets is improving significantly due to satellite measurements (Wingham *et al*. 1998). However, the relationship between the growth, dynamic evolution and decay of ice sheets and the sea ice cover on the one hand and climate change on the other is interactive, while also being highly complex. It is therefore necessary to model ice dynamics, but our understanding of the fundamental physics is not sufficiently advanced to make the quantitative predictions necessary. Particular outstanding issues include the contribution of sub-glacial deformation to overall ice sheet motion (Anandakrishnan *et al*. 1998) and the role fracture plays in ice shelf break-up (Vaughan & Doake 1996) and controlling the mean thickness of the arctic sea ice cover. Investigating the plastic flow of ice has been an area of active research for the last half century (Paterson 1994), but even here there is a need to improve our understanding of the development of ice fabric anisotropy and the related question of shear localization in ice sheets.

Planetary exploration has opened up new areas of research. Meteorite impacts, planetary resurfacing and volatile controlled rheology all raise fresh questions about the fundamental physics of the planets. Some of the most interesting surface features of planets are a consequence of fracture (Solomon *et al*. 1992), such as the fractures' patterns on the volcanic plains of Venus (Banerdt & Sammis 1992) and the icy moons of Jupiter. While much research has been done into the rheological and fracture properties of rocks and ice under planetary conditions recently (e.g. Durham *et al*. 1993), considerable work remains because planetary conditions vary so widely.

The rock physics approach to understanding the dynamics of the Earth's crust and ice sheet has been strongly influenced by developments in materials science and engineering. A highly influential concept has been that of establishing the dominant

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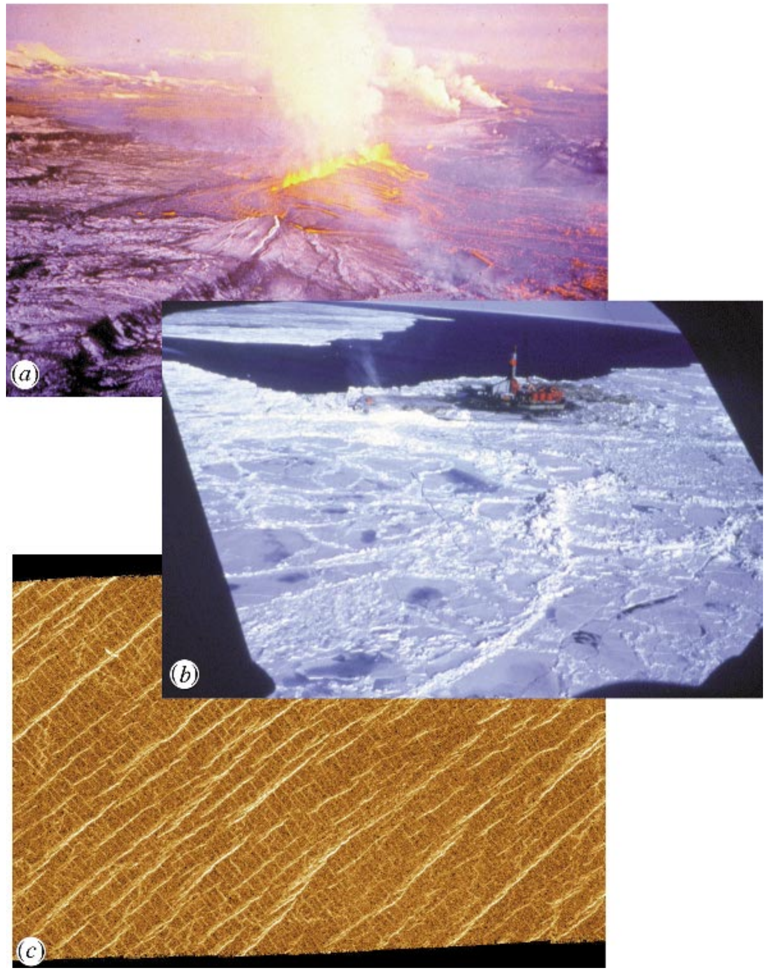

Figure 1. Fracture on all scale plays a key role in the tectonics of the Earth and planets: (*a*) opening of the Krafla fissure, Iceland; (*b*) fracturing of the arctic sea ice cover; (*c*) gridded fracture patterns of Guinevere Planitia, Venus (field of view is 80×40 km^2).

micromechanisms of flow and fracture during deformation and expressing these in terms of a constitutive law: an equation relating strain or strain rate to stress and the environmental conditions applied to a solid (Ashby & Verrall 1977). The great advantage with experiments is that dynamic variables such as the applied stress may be measured independently, whereas in the Earth even the reported crustal stress measurements (Zoback 1992) are essentially scaled from observations of strain (Main 1996). In the rock physics approach physical parameters are measured in the laboratory that are also measured by geophysical techniques in the field. New concepts from statistical physics have been introduced. This paper reviews the development of rock physics, both its theoretical and experimental approaches, and how rock physics can be applied to problems in geophysics through well-constrained models formulated in mathematical terms. This approach is considered in terms of Earth's crust and ice sheets because they represent the components of the solid Earth with

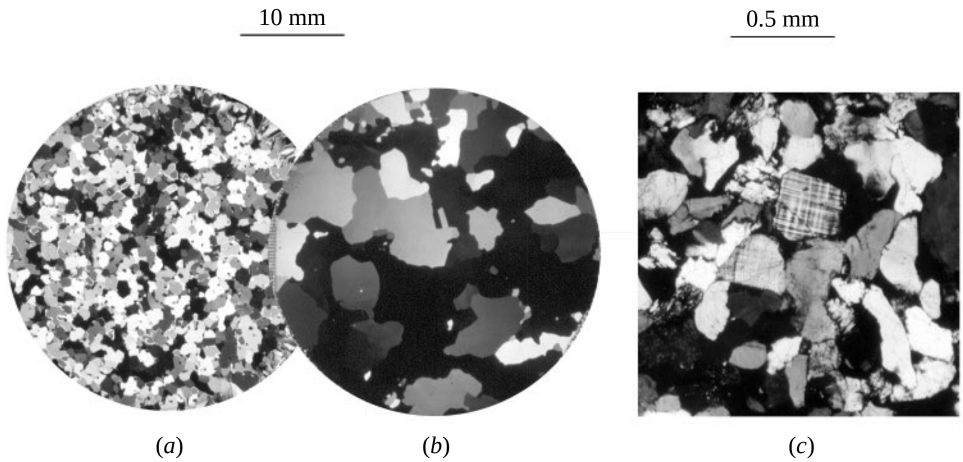

Figure 2. (*a*), (*b*) Thin sections of ice core taken from the Filchner–Ronne Ice Shelf, Antarctica. (*a*) At the top of the ice shelf the ice has formed from recent snow. The ice grains or crystals are fine and randomly oriented. (*b*) Ice at the bottom of the ice shelf has flowed off of the Antarctic continent. The ice is coarse grained and developing an anisotropic fabric. (*c*) Thin section from sandstone. Grains are principally quartz with some feldspar and mica present, cemented together by silica cement.

which humankind directly interacts. Vocadlo & Dobson (this issue) deal with the Earth's mantle and core and the different engineering necessary for very high pressure experiments.

2. Basic concepts: flow, fracture and friction

(*a*) *Mechanics of solids*

The nature of interatomic bonding and crystal structure together determines the physical properties and characteristic strength of crystalline solids (Nye 1957). Ice (figure 2*a*, *b*) is an example of such a crystalline solid. At the top of an ice sheet grains formed from recent snow are fine and randomly oriented: the ice is isotropic. At depth, grain growth and deformation result in coarse, aligned grains forming an anisotropic fabric (Castlenau *et al.* 1996). Ice on Earth is a unique rock because it is mono-minerallic and of atmospheric origin. More usual rocks are crystalline solids composed of mineral assemblies with different crystal structures (figure 2*c*). Grain growth and bonding between grains are strongly dependent upon thermal history and chemical composition (Covey-Crump 1997). However, the physical properties of rocks are not those of individual mineral grains, but are averaged properties and strongly dependent upon the intergranular bonding and the presence of any cracks. The elastic response is consequently highly nonlinear. It has therefore been argued that rocks form a separate class of materials, together with concrete, ceramics and soil (Guyer & Johnson 1999). Considerable research has been undertaken by crystallographers, mineral physicists and petrologists on the structure, bonding and physical properties of rocks through experiment, analytical techniques and computer modelling, which is beyond the scope of this paper to review. However, some findings which impinge directly on rock physics I discuss here.

Table 1. *Isomechanical groups of crystalline solids*
(Murrell 1989; Frost & Ashby 1982)

group	crystal structure	examples	easy glide systems
metals	face-centred cubic	Pb, Cu, Ni	5
	body-centred cubic	α-Fe, W	5
	hexagonal-close pack	Zn, Mg	2
ionic solids	rock salt cubic	alkali halides, NaCl	2
	rock salt cubic	simple oxides, MgO	2
ionic-covalent solids	rhombohedral	oxides, Al_2O_3	2
	orthorhombic	silicates, Mg_2SiO_4	2
covalent solids	diamond cubic	elements, Si, Ge, C	5
	hexagonal	compounds, Sic, Si_3N_4	2
hydrogen-bonded solids	hexagonal	ice, H_2O	2

The constitutive relations governing the elastic deformation of solids are (from Hooke's law):

$$\sigma = E\varepsilon, \qquad \tau = G\gamma,$$

where σ is the applied normal stress, E is the stiffness or Young's modulus, ε is normal strain, τ is the shear stress, G is the rigidity or shear modulus and γ is shear strain. The velocity of elastic waves in a rock, the compressional P-wave and shear S-waves are dependent on the elastic properties. Measuring P- and S-wave velocities in the laboratory, and comparing these with the velocity profile of the Earth determined from earthquake seismology, was important in determining the likely composition of the crust (Birch 1960). However, a developed fabric or the presence of cracks can strongly influence the velocities and frequencies of waves propagating in a rock. Full treatment of the physics is difficult and may involve using nonlinear wave equations (Guyer & Johnson 1999). Stress, strain and elastic modulus are tensor quantities, with many components depending upon the symmetry. Even calculating the number of measurement directions needed to measure stress-induced anisotropy is an exercise in mathematical physics. Solutions presently exist only for common lattice symmetries and loading configurations (Nikitin & Chesnokov 1981). However, these very complexities are also the reason why elastic wave measurements are such a powerful tool in determining rock properties and hence why they have been used extensively in the development of rock physics.

There are two distinct ways a solid can fail (Kelly 1966): it may either cleave in a brittle manner when the applied tensile stress exceeds the theoretical cohesive strength of the crystal lattice, or it may flow apart when the shear stress exceeds the theoretical shear strength. These theoretical strengths are only attained notionally by perfect crystals (Griffith 1920). Measured strengths are orders of magnitude lower than their theoretical maximum strengths because solids contain imperfections. Plastic flow of a solid, below the theoretical shear strength, is caused by the shear stress acting on dislocations, or line defects, and point defects in the crystal lattice. Fracture below the theoretical cleavage strength occurs because of microcracks or flaws present in the solid.

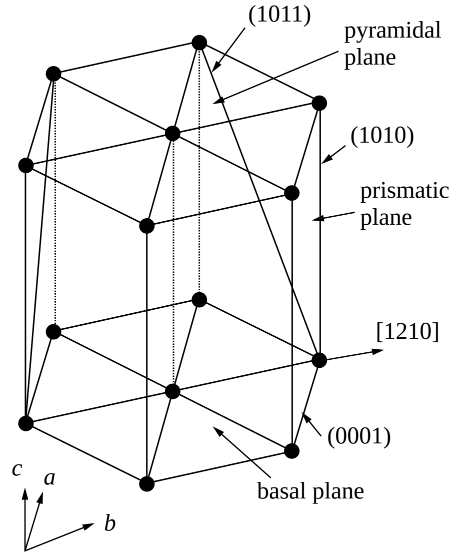

Figure 3. Hexagonal crystal lattice. The basal (0001), prismatic (1010) and pyramidal (1011) planes are shown.

Crystalline solids may be broadly classified into isomechanical groups (table 1). Ductility depends on the number of easy-glide slip systems there are available; that is to say the number of planes in the lattice along which dislocation glide can occur at low stresses (Nye 1957). The von Mises criterion states that there need to be five independent slip systems operating to maintain continuity at grain boundaries during ductile deformation (Kelly 1966). Ductile metals have five, while silicate rock and ice only have two and this is the reason why they are brittle materials. For the hexagonal crystal lattice of ice and silicate rocks the two easy slip systems are on the basal plane (figure 3). The prismatic and pyramidal planes are hard glide systems. For rocks to be ductile and flow plastically without cracking high temperatures are required so that slip can occur on hard-glide slip systems.

(*b*) *Flow of solids*

At high temperatures solids will flow in a manner analogous to fluids (Frost & Ashby 1982). For pure metals the temperature needed may be as low as 30% of the melting temperature, i.e. $0.3T_{\mathrm{m}}$. For silicate rocks, temperatures above $0.6T_{\mathrm{m}}$ are needed. Ice on Earth exists at temperatures above $0.85T_{\mathrm{m}}$ and commonly above $0.95T_{\mathrm{m}}$, hence it will flow plastically. Ice in the Solar System, being much colder, will be brittle. I will address the flow of solids in terms of the creep of ice as it has probably been characterized better than any other non-metallic solid.

When a load is placed on a specimen of ice, it will deform elastically, but also immediately begin to creep. Plastic flow of solids is driven by the shear stress alone and hence is independent of pressure, except at very high pressure (above 1% of the bulk modulus (Ashby & Verrall 1977)). Numerous laboratory creep experiments have established an empirical constitutive law describing high temperature plastic flow:

$$\dot{\gamma} = A\tau^{n} \exp(-Q/RT),$$

where $\dot{\gamma}$ is the shear strain rate, A is a scalar constant dependent upon rock fabric and impurities, n is the power law exponent, T is temperature in kelvins, Q is the activation energy for creep and R is the molar gas constant. This equation is the power-law creep equation. It is analogous to the equation for Newtonian viscous flow of a fluid, $\dot{\gamma} = \tau/\eta$, with viscosity η. However, the power-law dependence on stress means that the effective viscosity is stress dependent. Therefore, this is often referred to as non-Newtonian flow. Power law creep is thermally activated, hence the Arrhenius exponential dependence on temperature. The micromechanism of power law creep is thermally activated dislocation glide or dislocation glide plus climb. Other important mechanisms of plasticity and flow can also operate (Ashby & Verrall 1977). The power-law creep equation is extensively used for creep of metals and non-metals. Often it is forgotten that it is only an empirical equation. Frost & Ashby (1982) state that the present theoretical models for creep are still unsatisfactory since they are unable to explain the wide range of power law exponents found experimentally. Clearly, much work still needs to be done.

One area of active research is the question of plastic anisotropy as plastic anisotropy makes rocks and ice significantly softer. The plasticity of ice is highly anisotropic. As ice is buried in an ice sheet and moves along a flow line, its degree of anisotropy changes and so does its strength (figure $2a, b$). Ice fabric can be characterized by optical microscopy in thin section by a mean Schmid factor,

$$S = \frac{1}{N}\sum \cos\chi_i \sin\chi_i,$$

where χ_i is the angle of the crystallographic c-axis to the vertical and N is the number of grains. Azuma (1994) has assumed that flow occurs only by basal slip (but with some grain boundary accommodation) and has shown that anisotropy may be incorporated into the flow law through the Schmid factor: $\dot{\gamma} = A_0 S^{n-1}\tau^n \exp(-Q/RT)$.

To test new types of anisotropic flow laws such as this, simple creep tests are not sufficient. Figure $4a, b$ shows a new ice deformation cell where an ice specimen is loaded biaxially, but also confined by high pressure oil to suppress cracking. A large ice specimen is necessary because of the coarse grain size found at depth in ice sheets. Simultaneously with deformation, computer-controlled P- and S-wave velocity transducers scan the specimen to map the development of fabric anisotropy. This is possible because there is a small difference in wave velocities in different directions in the ice crystal (figure 3). The cell is capable of operating at $-40\,^{\circ}$C and pressures up to 50 MPa, which encompass ice sheet conditions on Earth. It will be used to test ice recovered from the deep borehole drilling programmes in Antarctica and Greenland.

(*c*) *Fracture*

Griffith established the basic principles of fracture in 1920. However, it was only in the 1950s that the science of fracture mechanics was developed, driven by the requirements of jet aircraft design and nuclear power engineering and made possible by the development of the complex algebra capable of handling elasticity theory for cracks in analytical form. (Cherepanov (1998) provides a historical account of the development of fracture mechanics). Griffith proposed that fracture initiates in brittle solids from pre-existing cracks or flaws once the local tensile stresses at the tip of the crack are sufficient to overcome interatomic bonding. In the Irwin–Orowan

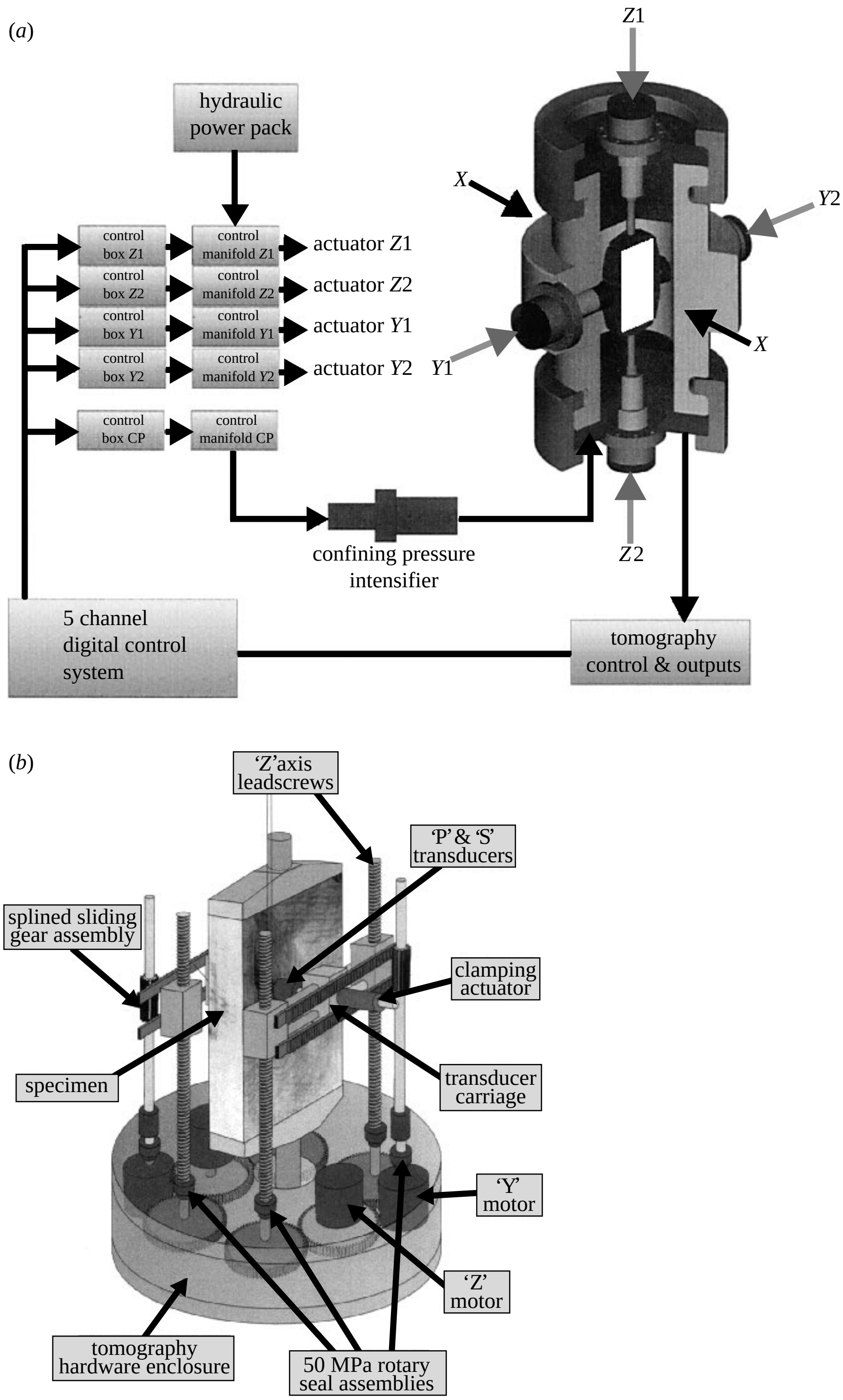

Figure 4. For description see opposite.

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formulation of linear-elastic fracture mechanics a scale-independent stress intensity factor K describes the magnitude of stresses around the crack tip,

$$K = \sigma(\pi c)^{1/2},$$

where σ is the applied stress and $2c$ is the crack length. The precise formulation depends on crack geometry but propagation always occurs at a critical value of the stress intensity factor, the material's fracture toughness, K_{IC}. The critical value of stress intensity can be achieved through an increase in remotely applied stress or an increase in crack length: at very low stresses a long enough initial crack will propagate. Fracture toughness is important because it is scale independent. It is the same for laboratory specimens or for faults running hundreds of kilometres. The crude notion of fracture strength is scale dependent because it depends upon flaw size.

Considerable recent effort has been devoted to measuring fracture toughness in rocks (Atkinson 1987). A modern rock fracture mechanics apparatus is shown in figure 5. This is designed to measure the fracture toughness of rock under simulated conditions of hydrothermal circulation at mid-ocean ridges: a highly corrosive seawater environment, at pressures up to 70 MPa (corresponding to a depth of a few kilometres) and temperatures of up 500 °C. A large cylindrical 'short-rod' test specimen is used, cut with a V-notch. As the specimen is jacked apart in tension, a flat-fronted crack grows incrementally along the V-notch from the tip. The crack grows in a stable manner because the crack front width increases with each increment of crack growth.

In a biaxial stress field, a crack will both open at the crack tip and shear along the crack plane: it is subject to a normal stress, σ_{n}, and a shear stress, τ, whose magnitudes depend upon the crack orientation, β, with respect to the principal stresses, σ_1 and σ_2 (figure 6*a*). For closed cracks sliding on the crack will be opposed by friction. The crack is subjected to both tensile opening stress intensity (mode I), $K_{\mathrm{I}} = \sigma_{\mathrm{n}}(\pi c)^{1/2}$ and mode II shearing stress intensity, $K_{\mathrm{II}} = \tau(\pi c)^{1/2}$. Erdogan & Sih (1963) proposed that fracture will initiate from a pre-existing flaw in a critical direction, θ, when

$$\cos \tfrac{1}{2}\theta_0 (K_{\mathrm{I}} \cos^2 \tfrac{1}{2}\theta_0 - \tfrac{3}{2} K_{\mathrm{II}} \sin\theta_0) \geqslant K_{\mathrm{IC}}.$$

(Murrell & Digby (1970) have developed a similar criterion for a 'Griffith' ellipsoidal crack in an arbitrary triaxial stress field.) The fracture initiation envelope in principal stress space is shown in figure 6*b*. For stress states outside the fracture initiation envelope the flaw will propagate, but for those inside the envelope the flaw will be stable. How the principal stress map may be used to describe creep-brittle behaviour (for sea ice) is discussed by Sammonds *et al.* (1998).

Figure 4. (*a*) A new design of ice deformation cell. The cell consists of biaxial loading actuators (X- and Y-directions) mounted on a low-temperature pressure vessel. A large ice specimen 200 mm × 100 mm × 40 mm is used. The figure shows schematically the servo-hydraulic control system. Hydraulic power is provided by a power pack. Digital controllers control oil flow to the manifolds, which in turn control the loading of the actuators. (*b*) During deformation the specimen is scanned by internally mounted P- and S-wave transducers, driven by stepper motors, to monitor anisotropic fabric development. (Designed by Sammonds, Boon & Hughes.)

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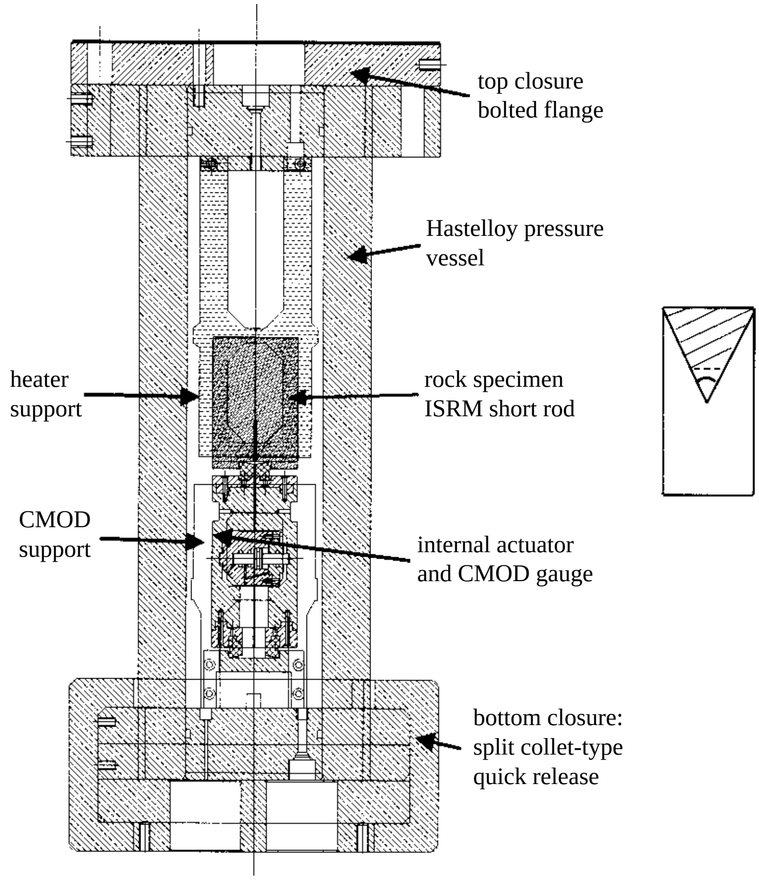

Figure 5. A modern rock fracture mechanics apparatus designed to measure the fracture toughness of rock simulating conditions at mid-ocean ridges. The apparatus consists of a pressure vessel with an internal heater inside of which is a small hydraulic tensile actuator or jack. An 80 mm diameter by 160 mm long 'short-rod' test specimen is used. As the jack loads the specimen in tension, a 'crack mouth opening displacement' (CMOD) gauge measures the tensile opening. The extreme conditions require the latest materials and measurement technology. (Designed by Jones, Boon & Meredith.)

(*d*) *Friction*

The surface of a solid is made up of asperities. When two surfaces are pressed together they do not contact over the entire area, the apparent area of contact, but touch only at a number of asperity contacts. The sum of the areas of all the asperity contacts constitutes the real area of contact, which is much less than the apparent area of contact. At these contacts strong adhesive bonds are formed between the surfaces. It is these adhesive (or cohesive) bonds which constitute the resistance to frictional sliding (Bowden & Tabor 1954). Surfaces adsorb molecules of water vapour and atmospheric oxygen, which forms a film on the surface, a few molecules thick. The film interrupts the adhesive contact between asperities, which is the reason why frictional strength is in general much lower than fracture strength (Ohnaka 1996).

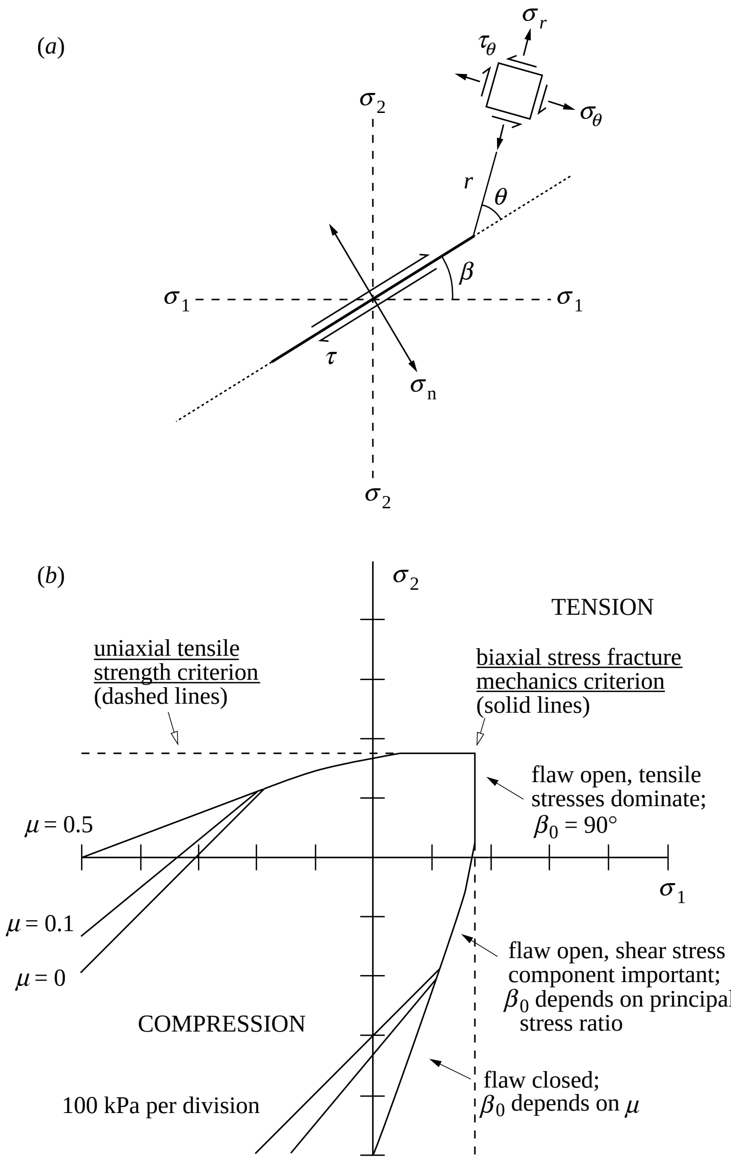

Figure 6. (a) Flaw in a biaxial stress field. Normal and shear stresses act on the crack. Even when both principal stresses are compressive, there will be tensile stresses acting at the crack tip. (b) Fracture initiation criterion plotted in principal stress space along with a simple fracture criterion based on uniaxial tensile strength. For stress states outside the fracture initiation envelope the flaw will propagate, but for those inside the envelope the flaw will be stable. Note the dependence of the envelope on the coefficient of friction, μ, used. (After Rist *et al.* 1999.)

Amonton's law defines the coefficient of friction, μ, by $\tau = \mu\sigma_{\rm n}$, where τ is the shear stress acting along the surface and $\sigma_{\rm n}$ is the normal stress acting across it.

Friction can be dependent on pressure, contact time and sliding speed, and be highly temperature sensitive at temperatures approaching the melting point. Friction of ice at low stresses and sliding speeds is primarily due to very effective adhesion between contacting asperities on the opposing surfaces (Bowden & Tabor 1954). The frictional resistance of ice increases with decreasing temperature, as the strength of the adhesion increases. At high sliding speeds and high temperatures, frictional heating generates a thin water film at the surface, greatly decreasing frictional resistance and making skiing possible. For rock, frictional resistance is usually expressed in terms of a Coulomb-type law,

$$\tau = \mu\sigma + C,$$

where C is a cohesive strength term. Many experimental rock friction data fit this equation, which is frequently referred to as Byerlee's law (Paterson 1978). Frictional sliding may be stable, or unstable when repeated sticking then slipping occurs. Stick-slip has been proposed as the earthquake mechanism (Brace & Byerlee 1966) and hence considerable effort has gone into formulating stick-slip frictional constitutive laws (Scholz 1998). The transition between stable and unstable sliding is sensitive even to small changes in the coefficient of friction and therefore depends on rock type, surface roughness and the presence of fault gouge material, pressure temperature and sliding rate (Marone 1998). How μ evolves with slip displacement, the surface, and time is crucial. In rate and state-dependent laws of frictional sliding (Dieterich 1979) time-dependent healing between the contacting surfaces is viewed as key to the onset of instability. Hydrothermal circulation has an important effect on this (Olsen *et al.* 1998).

In high-resolution friction experiments (Ohnaka & Shen 1999) the velocity of slip propagation along a simulated fault in a large laboratory specimen has been measured by strain gauges placed along the fault. These experiments show that there is a small but finite slip-strengthening phase followed by slip weakening. Slip strengthening is caused by an increase in the real contact area of mating asperities up to maximum frictional resistance. In this first nucleation phase slip propagation is slow, but accelerates up to high-speed slip propagation. The strength eventually degrades to a residual stress with ongoing slip displacement on the surfaces. This process is called shear rupture. It is seen as being directly analogous to the nucleation and propagation of a crack in shear and can be treated through a fracture mechanics approach. The patch where slip nucleates is analogous to an existing flaw. Once a critical amount of slip has occurred propagation will become dynamic. The theory behind this approach is based on a cohesive zone model, which used the concept of a cohesive zone at the crack tip (Palmer & Rice 1973).

(*e*) *Fracture in compression*

Mechanical testing of metals is usually done with simple test specimens loaded in one direction in tension, or compression–uniaxial loading. Triaxial tests are common on rocks. This reflects the fact that failure of brittle solids depends on both normal and shear components of stress, although to some extent this arises from a desire to simulate practical situations. Plastic flow is driven by the shear stress alone. The simplest and most commonly used procedure for achieving a triaxial stress state in the laboratory is to superimpose a confining pressure on a uniaxial stress—the conventional triaxial test. In figure 7 a modern triaxial cell for rock deformation is shown, which operates at confining pressures up to 400 MPa and temperatures up to 400 °C, encompassing the conditions found in the upper crust of the Earth. A rock

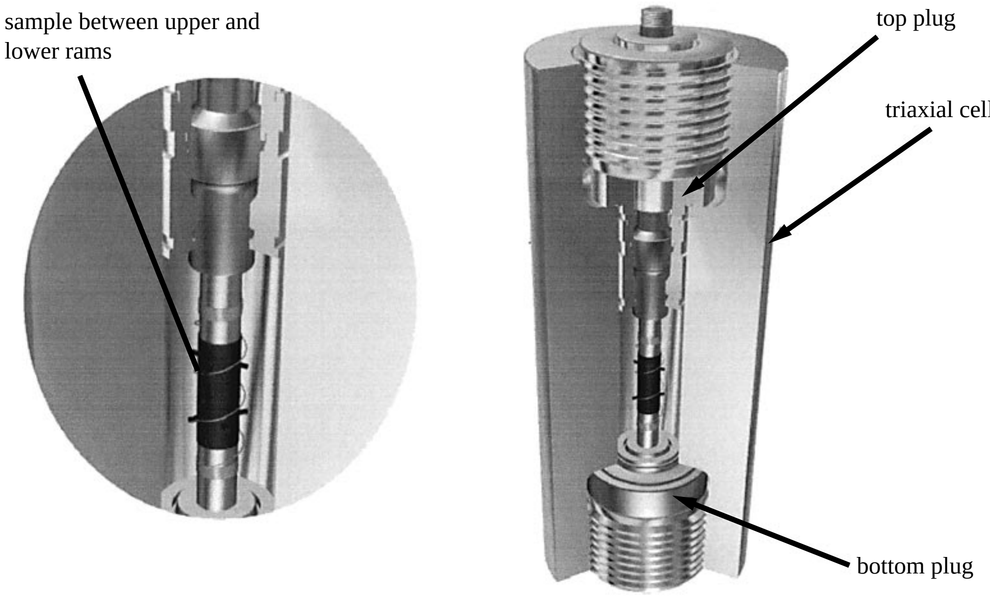

Figure 7. Triaxial testing cell for rock deformation. The rock sample is jacketed in a multiple-instrumented plastic sleeve to keep out the confining fluid. Twelve electrical and/or acoustic measurements can be made simultaneously on the sample. The sample is deformed by upper and loading rams. These in turn are loaded by an axial loading piston that is pressure balance so only the differential stress (the total axial stress minus the confining pressure) need be applied, greatly improving the accuracy of load measurement and experimental control. The confining fluid is oil. An internal heater allows temperatures up to 400 °C to be used. (Designed by Sammonds.)

sample 40 mm in diameter by 100 mm long is used. Axial deformation and confining pressure are servo-controlled. Pore fluid pressures and volume in the specimen can be controlled and measured, and permeability measured during a deformation experiment. The plastic sleeve, which jackets the specimen to keep out the confining fluid from cracks and pores, can be used to make 12 electrical/acoustic measurements on the rock.

Triaxial apparatus have been used at pressures up to 1400 MPa (about the pressure at the depth of the crust) and temperatures in the range −250 to 1600 °C. A wide variety of physical properties measurements has been made concurrently with rock deformation: P- and S-wave velocities and seismic attenuation (e.g. Ayling *et al*. 1995), complex electrical conductivity (Glover *et al*. 1994, 1996), electrical potential (Yoshida *et al*. 1998) and acoustic emissions emitted by growing cracks (Sammonds *et al*. 1992). Modern apparatus can use sophisticated instrumented jackets so that seismic tomography and acoustic emission (AE) locations can be done. They are fully servo and computer controlled. For instance, the rate of AE has been used to control the rate of advance of the loading piston, allowing close control of fault formation to be attempted (Lockner *et al*. 1991).

In uniaxial tension a rock fails by an extension fracture, where there is a clean separation with no offset between the surfaces. In uniaxial compression, failure is by

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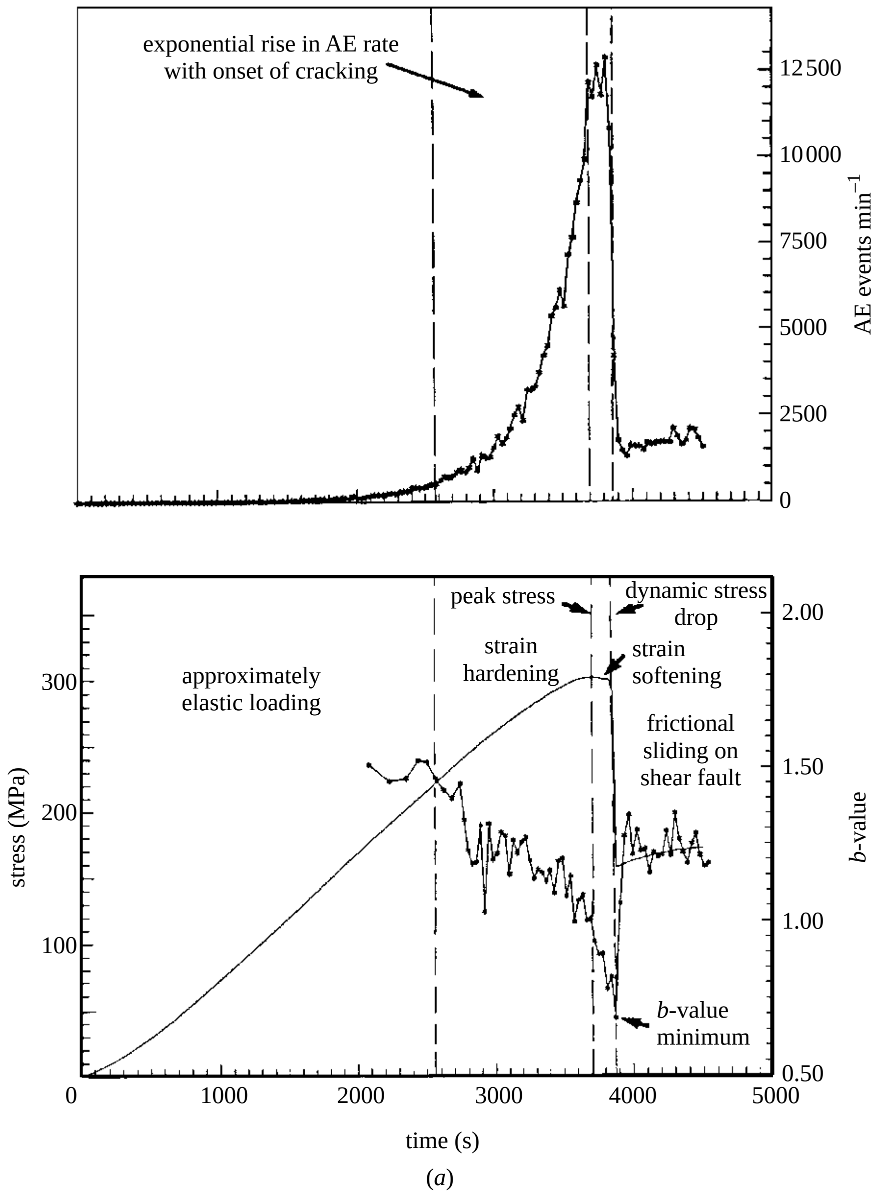
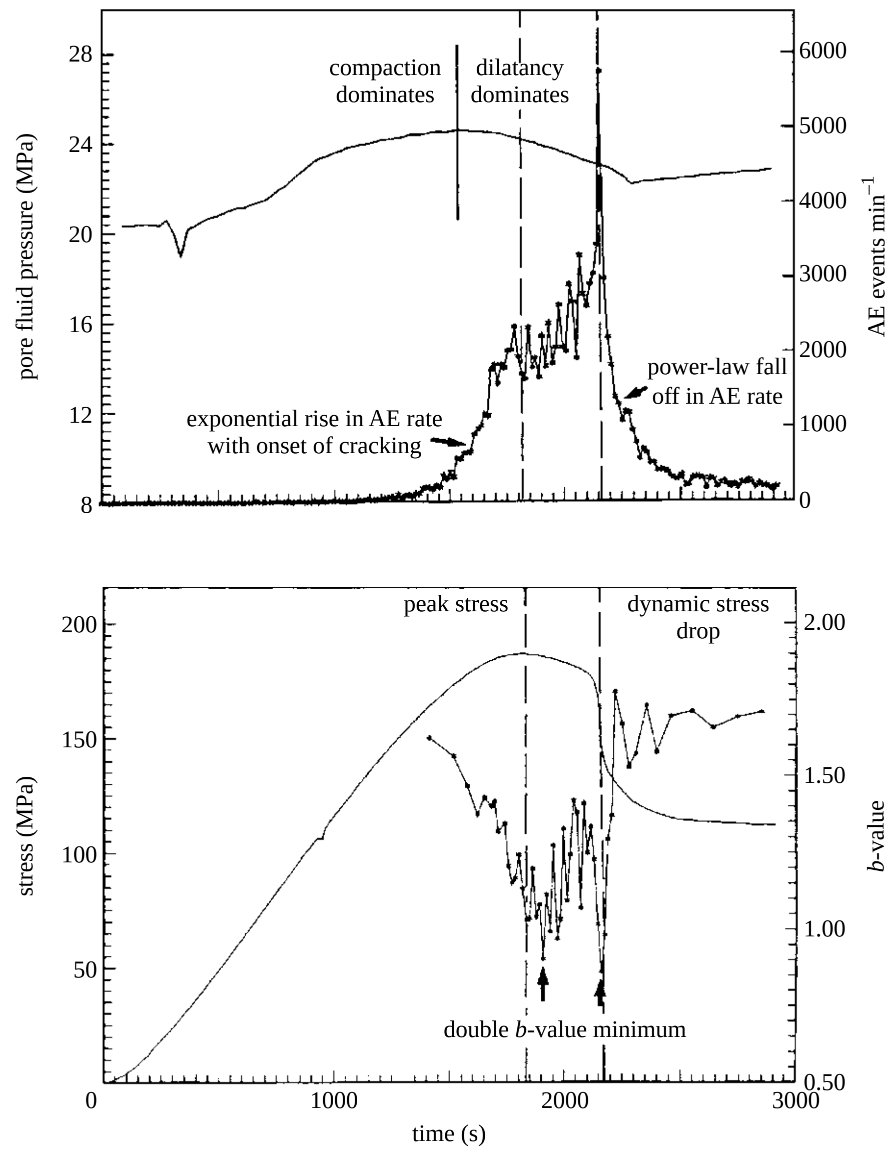


Figure 8. (*a*) Triaxial deformation experiment on air-dried sandstone at 50 MPa confining pressure. Acoustic emissions are measured concurrently with deformation. The *b*-value exhibits a single minimum.

multiple axial splitting. With the imposition of a confining pressure on the specimen, compressive fracture occurs by a shear fault along an inclined plane, on which frictional sliding may take place. The very earliest experiments of von Karman (Paterson 1978) showed how strongly the fracture strength of rock increases with increasing pressure. Figure 8*a*, *b* shows examples of rock deformation leading to shear faulting where concurrently the AE event rate (which is proportional to the cracking rate)



(*b*)

Figure 8. (*b*) Triaxial deformation experiment where pore volume has been kept constant during deformation. Acoustic emissions are measured concurrently with deformation. The *b*-value exhibits the double minimum predicted by the fracture mechanics model. (Sammonds *et al*. 1992.)

and their amplitudes (which are proportional to crack sizes) have been measured (Sammonds *et al*. 1992). The changes in the microseismic *b*-values (the log-linear slope of the AE frequency–amplitude distribution) with deformation and time are plotted. These AE measurements therefore provide a means of interpreting the development of shear faulting. For the dry specimen (figure 8*a*), microcracking initiates at *ca.* 50% of peak stress. During the strain hardening portion there is a steady decline in b-value as microcracks grow, with shear fracture occurring at a b-value minimum as cracks interact to form an inclined shear fault. Sliding takes place along the fault plane resisted by friction. This decline in b-value has also been observed in direct shear frictional sliding experiments just before frictional instability (Sammonds & Ohnaka 1998) and in nature prior to major earthquakes (Smith 1981).

The fracture criteria of Griffith (1920) and Erdogan & Sih (1963) are fracture initiation criteria. In tension, fracture initiation will usually lead directly to fracture propagation and failure. However, as rock physics measurements demonstrate, fracture initiation will not usually lead to failure in compression. Initial cracks grow stably before interacting to form a through going shear fault. Crack growth has been modelled using wing crack models; however, these models do not adequately describe shear fault development and formation in rock. An outstanding fundamental problem for rock physics remains the three-dimensional modelling of cracks, crack dynamics and crack interactions, producing physical models of shear rupture, faulting and the development of physical anisotropy.

(*f*) *Fluids in rocks*

The presence of fluids in pores and cracks in rocks has both strong mechanical and chemical effects. The coupling of rock deformation and fluid flow is currently a strong area of research interest (Zhu & Wong 1997). The mechanical effect of fluid can be expressed in terms of the pressure of the pore fluid, p_{p}, reducing the confining pressure, p, to an effective pressure, $p - p_{\mathrm{p}}$. However, the effectiveness of this pressure reduction is dependent upon how well connected the pore space is in the rock, or how permeable it is. The presence of pore fluids can have a strong influence on rock deformation as strain softening can be promoted by local dilatancy hardening caused by a decreasing pore-fluid pressure (Sammonds *et al.* 1992). Chemically, the addition of water can increase the kinetics for plastic flow and pressure solution. Also cracks can grow below the critical stress intensity by a process called stress corrosion, or subcritical crack growth. The movement of chemical components by fluid flow from one chemical environment to another causes chemical reactions between the minerals and the fluids. Dissolution, precipitation, ion exchange and sorption occur. The effect can be the opening and blocking of fluid pathways, thereby changing rock permeability (Tenthorey *et al.* 1998). Therefore there are complex interactions and feedbacks between rock deformation, rock permeability, fluid flow, chemical reactions and heat, which are as yet poorly understood.

In the nineteenth century some carefully constructed laboratory experiments produced a set of empirical laws that expressed the flux of mass and energy in terms of a driving force and material properties (Bredehoeft & Norton 1990). These laws include Fick's law of diffusion, Fourier's law of heat conduction, DeDonder's law of affinity and Darcy's law. Darcy's law describes the conservation of fluid momentum during flow through porous media,

$$v = \frac{k}{\eta\varphi}\frac{\partial}{\partial x}(\rho g h),$$

where v is the fluid velocity, k is the rock permeability, η is fluid viscosity, φ is rock porosity and $\rho g h$ is the driving head of pressure in the x-direction. Rock permeability

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is a tensor quantity. Permeability measured in the laboratory may be two to three orders of magnitude lower than those measured in the crust (Brace 1980). This brings us to the unresolved questions: how to model fluid flow through deforming rock, and how does permeability evolve and scale with size? A number of recent approaches such as employing the theory of mixtures (Morland 1992), percolation theory (Sahami 1994) and network models (Zhu & Wong 1999) have gone some way to addressing these problems.

In modern laboratory experiments, a servo-controlled pore fluid intensifier controls fluid in a rock specimen. This type of apparatus can be used to control and measure the pressure and volume of fluids present in a rock. It can be used to measure rock permeability by applying a transient pressure pulse at one end of a specimen and measuring its decay, by applying a sinusoidally varying pressure to one end, or using two intensifiers to flow fluid through the rock. Associated physical parameters such as electrical conductivity and potential are also measured. Experiments show that even though crack density may only change by 50% (calculated from wave velocity measurements) the permeability may change by two orders of magnitude.

3. Applications

To model deformation of tectonic plates and ice sheets it is necessary to employ continuum mechanics. Continuum mechanics is the extension of Newton's principles of mechanics to the behaviour of fluids and solids that was successfully made in the 18th century. In some cases the laws of thermodynamics are also required. The medium is treated as being continuous: free of cracks and voids, and the discrete nature of crystalline and molecular structure is ignored. The continuum approach can be used to calculate a stress field, and then a fracture criterion introduced to predict fracture.

(*a*) *Ice shelf crevassing*

Rist *et al*. (1999) have used the approach of investigating experimentally the mechanical and physical behaviour of Antarctic shelf ice taken from cores drilled through the thickness of the Ronne ice shelf, and use these measurements in fracture criteria describing surface crevassing, spatial distribution of surface crevasses, bottom crevassing and tidal flexure. The natural range and variability of Antarctic ice, including porosity, anisotropy, impurity content and fabric, is therefore taken into account. The velocity field of flow of the ice shelf can be determined from satellite altimeter measurements of ice shelf topography from which a strain-rate field may be calculated by finite-element computer numerical modelling. A laboratory-derived flow law for ice can be used to turn this strain-rate field into a stress field. The stress field is then combined with the laboratory-derived fracture criterion to predict the spatial crevasse distribution and depth of crevasse penetration. At the surface of a large ice mass, away from its boundaries, plane stress conditions apply since the principal stress normal to the surface is zero. Rist *et al*. applied the Erdogan & Sih criterion to ice shelf fracture. Figure 9 shows a prediction for surface crevassing on the Filchner–Ronne ice shelf, and for comparison a Landsat-prepared crevasse map is shown. There is good agreement between the two, demonstrating the success of the methodology. The chief discrepancies arise because the model predicts

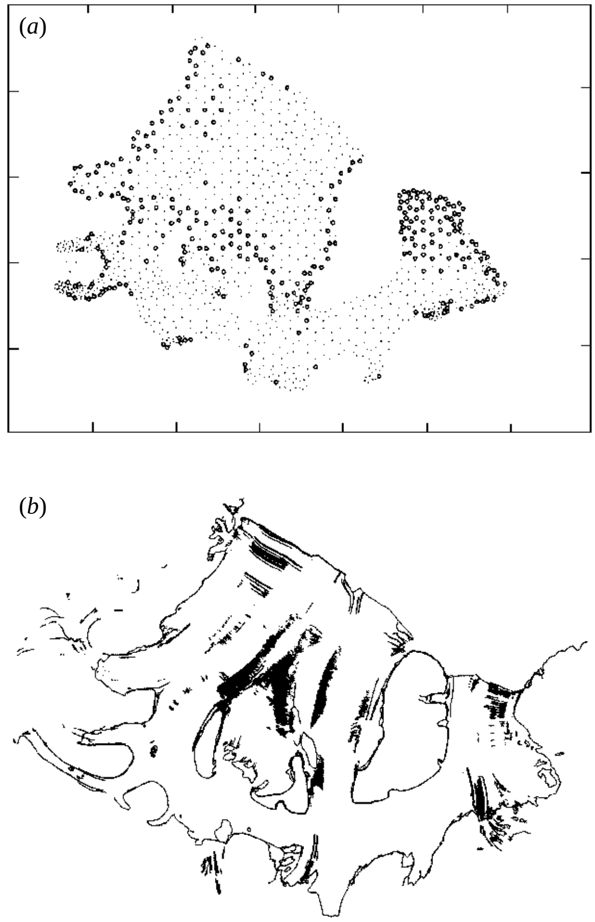

Figure 9. (*a*) Areas of potential crevassing derived from a finite-element model of the Filchner–Ronne ice shelf. Open circles represent nodes at which stresses are such that the biaxial fracture criterion of Erdogan & Sih (1963) is exceeded. (*b*) Fracture patterns (crevasse and rifts) on the Filchner–Ronne identified on visible Landsat and SAR satellite imagery. (Rist *et al.* 1999.)

fracture initiation. Once initiated, crevasses will flow with the ice shelf. There is also a lack of visible crevassing at the ice shelf land margin; however, field surveys indicate that buried crevasses do exist. The principal improvement that could be made to this model is to improve the stress field derived from the flow field through higher-resolution computer simulation.

This methodology could also be applied to similar problems, such as the fracture distributions on Venus, and fracturing of lava flows for volcanic hazard assessment (Kilburn 1996).

(*b*) *Strength of the lithosphere*

Application of research into the fundamental physics of rocks to the Earth's crust is done within the framework of modern plate tectonics theory, which is the latest in

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a succession of theories of continental drift (Le Grand 1988). When Wegener (1915) developed the theory of continental drift he envisaged rigid continental blocks, which had been 'permanent throughout the Earth's history', drifting across the surface of the Earth. In plate tectonics theory (McKenzie & Parker 1967), it is considered that the Earth's outer layer consists of small number of rigid lithospheric plates in relative motion to each other. Narrow belts of earthquakes define the boundaries between lithospheric plates where the relative motion and deformation of the plates is accommodated. At the plate boundaries major orogeny occurs (Dewey & Bird 1970).

The lithosphere consists of crust and upper mantle, which are chemically, compositionally, distinct: the boundary between the two is marked by the Moho seismological discontinuity. Lithospheric plates are mechanically decoupled from the underlying asthenosphere, which flows in a ductile manner allowing isostatic readjustment to occur as the plates move. The lithosphere–asthenosphere transition is a thermal-mechanical boundary: it is not compositional, nor marked by a seismological discontinuity (O'Nions 1992). There are considerable differences between oceanic and continental lithosphere. The oceanic lithosphere and crust are geologically young (less than a few hundred million years old) and formed largely by thermal processes during sea-floor spreading at the mid-ocean ridges with some metamorphism due to cooling and reaction with seawater. By contrast, the continental crust has been formed and transformed over prolonged thermal, tectonic and sedimentary processes over much of Earth's history. The oceanic crust is relatively homogeneous, while the continental crust is highly heterogeneous. The thickening and thinning of the continental crust, the concomitant burial and exhumation of rock and changes in thermal gradients, along with rock deformation itself leads to metamorphism, where rocks may recrystallize existing minerals, grow new minerals and dehydrate. The oceanic crust averages *ca.* 6 km thick, while the continental crust averages 40 km in tectonically stable regions, but may be much thicker locally. The continental crust contains larger quantities of radioactive isotopes than the oceanic crust, with the result that the temperatures in the upper mantle beneath the continents may be considerably higher than at the same depth in the oceanic lithosphere.

In mechanical terms the most important differences between the oceanic and continental lithospheres are that the continental lithosphere is less dense and weaker than the oceanic lithosphere. In the past 20 years there has been considerable progress made in estimating the strength of the lithosphere with depth based on the results of laboratory experiments (Kohlstedt *et al*. 1995). The strength of the lithosphere cannot be greater than rock strength measured in the laboratory, but might be considerably weaker. Figure 10 shows a schematic illustration of the maximum rock strength as a function of depth for oceanic and continental lithosphere. Kohlstedt *et al*. describe the profiles. In the uppermost part where the temperature and lithostatic pressure (due to the rock overburden) are relatively low, frictional sliding governs mechanical behaviour. At greater depth, due to increasing temperature and pressure, plastic deformation controls the strength. Between these two regions, brittle and plastic processes interact in a transition zone. These strength profiles reflect the experimental observations that frictional strength increases approximately linearly with increasing pressure and is relatively insensitive to temperature, while plastic strength decreases rapidly with increasing temperature but is relatively insensitive to pressure. There is considerable present effort to refine these strength profiles. In

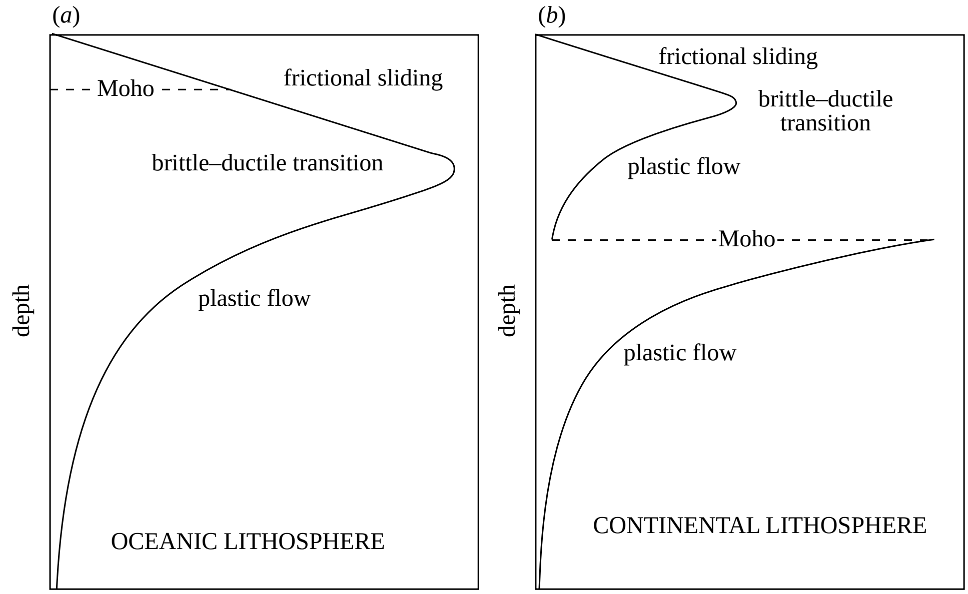

Figure 10. Schematic of maximum rock strength as a function of depth for (*a*) oceanic lithosphere and (*b*) the continental lithosphere.

particular much work is being done on the influence of water and the presence of partial melts (Rutter & Neumann 1995). These tend to weaken rocks, but their roles have not been properly quantified.

4. New approaches

We have seen in the preceding discussion that materials science has successfully explained the brittleness of materials and that fracture mechanics can be applied successfully to tensile fracture and fracture initiation in compression. Fracture mechanics may be used to model large-scale behaviour such as crevassing in ice shelves. But this approach stops short of modelling ice shelf break-up. Laboratory experiments have been used to derive oceanic and continental lithospheric strength profiles. But strength profiles alone cannot tell us about earthquakes, as they are probably instability phenomena. At a laboratory scale there are also problems. Shear faulting in compression is associated with exponentially increasing cracking activity, and has not been modelled successfully. Rock fracture permeability is highly sensitive to crack porosity and does not appear to conform to Darcy's law. Questions remain about scaling from the laboratory scale to the Earth.

Methods from statistical physics are being introduced to rock physics. There is not the space to review all these developments here, and Main (1996) has provided a recent review. However, these developments have important implications for the design of future rock physics experiments, and I will review some of the major features of these.

Many geological fabrics and fault patterns are scale invariant, otherwise it would not be necessary to include the ubiquitous geological hammer on photographs of rock outcrops (Main *et al*. 1990). The only distribution of features with characteristic

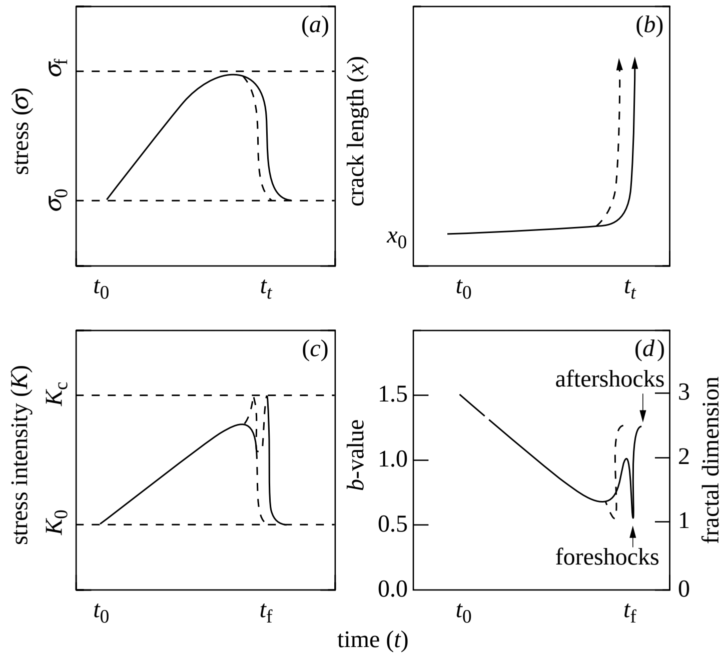

Figure 11. The fracture mechanics model of the b-value anomaly for the earthquake source (Main *et al.* 1989). (a) Stress–time behaviour for rock deformation. (b) Acceleration of crack tip to failure. (c) Combined effects of the stress and crack length on the stress intensity factor, K. (d) Because b is negatively correlated with and linearly related to K, the model predicts a double b-value minimum.

length-scale that is scale-invariant is a power law. This takes the form,

$$N(l) = Cl^{-D},$$

where N is the number of faults of length exceeding l, C is a constant and D is the scaling exponent of the length distribution (the fractal dimension). This fractal interpretation has been widely validated for frequency–length distribution of shear faults in the crust and seismological studies of faults. The Richter–Gutenberg log-linear frequency–magnitude distribution for earthquakes is a fractal distribution,

$$\log N = a - bm,$$

where N is the number of earthquakes of magnitude m, a is a constant and b is the seismic b-value. The same distribution is observed for acoustic emission from rock fracture in laboratory experiments, except that m is the AE amplitude. If some basic assumptions about earthquakes are true, such as stress drop being constant, then the seismic b-value is related to the fractal power law exponent, D, by (Main *et al.* 1990)

$$D = 3b/c.$$

For most common cases of earthquakes, $c = 3/2$, so $D = 2b$. There is therefore a direct relation between the fractal dimension and b-value, and a means of scaling between laboratory AE experiments and crustal scale deformation.

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Main *et al*. (1989) proposed a fracture mechanics model of the earthquake source, which explained the fluctuations in the seismic b-value observed in nature in terms of the underlying physical processes of time-varying applied stress and crack growth (figure 11). The model predicts two minima separated by a short-lived maximum. Laboratory experiments, under conditions of controlled pore fluid conditions were able to reproduce this prediction, thereby validating the model (figure 8*b*).

5. Conclusion

The future for experimental rock physics is in holistic laboratory experiments, where a range of physical parameters (e.g. elastic wave velocities, permeability and electrical conductivity) are measured concurrently with controlled application of stress and fluid conditions that simulate crustal conditions. This will require the development of more sensitive electrical and acoustic transducers, development of micro-circuitry and the use of the latest developments in electronics such as digital filtering and control. Improvements in construction materials, such as high-strength, high-temperature alloys and plastics will allow further extension of pressure and temperature operating conditions. Measurements of physical parameters are not only important in a materials science sense, but crucially these parameters are also routinely monitored by geophysical techniques in regions of active tectonism and in sub-surface reservoirs. There have already been successes using this approach, such as reproducing the pattern of seismicity observed before major earthquakes. Experimental rock physics is underpinned by the theory of fracture mechanics. However, the application of fracture mechanics needs to be further developed to encompass the dynamic behaviour of crack ensembles with distributions of physical parameters; to employ percolation statistics with physically realistic crack interactions; and introduce novel methods from physics to calculate effective rock physical properties. The goal will be to combine holistic rock physics laboratory experiments with crustal stress measurements, borehole drilling and investigation, satellite remote sensing, crustal geophysics, mathematical and computer numerical modelling, to understand and predict the behaviour of the Earth's crust, which is the home of humankind, over millennia.

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References

Anandakrishnan, S., Blankenship, D. D., Alley, R. B. & Stoffa, P. L. 1998 *Nature* **394**, 62–65.

Ashby, M. F. & Verrall, R. A. 1977 *Phil. Trans. R. Soc. Lond.* A **288**, 59–95.

Atkinson, B. K. (ed.) 1987 *Fracture mechanics of rock*. Academic.

Ayling, M. R., Meredith, P. G. & Murrell, S. A. F. 1995 *Tectonophys.* **245**, 205–221.

Azuma, N. 1994 *Earth Planet. Sci. Lett.* **128**, 601–614.

Banerdt, W. B. & Sammis, C. G. 1992 *J. Geophys. Res.* **97**, 16 149–16 166.

Birch, F. 1960 *J. Geophys. Res.* **65**, 1083–1102.

Bowden, F. B. & Tabor, D. 1954 *The friction and lubrication of solids*. Oxford: Clarendon.

Brace, W. F. 1980 *Int. J. Rock Mech.* **17**, 876–893.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Brace, W. F. & Byerlee, J. D. 1966 *Science* **153**, 990–992.

Bredehoeft, J. D. & Norton, D. L. 1990 In *The role of fluids in crustal processes*, pp. 27–41. Washington, DC: National Academy Press.

Castelnau, O., Duval, P., Lebensohn, R. A. & Canova, G. R. 1996 *J. Geophys. Res.* **101**, 13 851–13 868.

Cherepanov, G. P. (ed.) 1998 *Fracture.* Malabar, FL: Krieger.

Covey-Crump, S. J. 1997 *Contrib. Mineral Petrol.* **129**, 239–254.

Dewey, J. F. & Bird, J. M. 1970 *J. Geophys. Res.* **75**, 2625–2647.

Dieterich, J. H. 1979 *J. Geophys. Res.* **84**, 2161–2168.

Durham, W. B., Kirby, S. H. & Stern, L. A. 1993 *J. Geophys. Res.* **98**, 17 667–17 682.

Erdogan, F. & Sih, G. C. 1963 *J. Basic Engng* **85**, 519–527.

Frost, H. J. & Ashby, M. F. 1982 *Deformation-mechanism maps.* Oxford: Pergamon Press.

Gillott, J. & Kumar, M. 1995 *Science and the retreat from reason.* London: Merlin Press.

Glover, P., Meredith, P. G., Sammonds, P. R. & Murrell, S. A. F. 1994 *J. Geophys. Res.* **99**, 21 635–21 650.

Glover, P. W. J., Gomez, J. B., Meredith, P. G., Boon, S. A., Sammonds, P. R. & Murrell, S. A. F. 1996 *Surv. Geophys.* **17**, 307–330.

Griffith, A. A. 1920 *Phil. Trans. R. Soc. Lond.* A **221**, 163–198.

Griggs, D. 1939 *Am. J. Sci.* **237**, 611–50.

Guyer, R. A. & Johnson, P. 1999 *Physics Today*, April 1999, pp. 30–36.

Kelly, A. 1966 *Strong solids.* Oxford: Clarendon Press.

Kilburn, C. R. J. 1996 In *Monitoring and mitigation of volcano hazards* (ed. R. Scarpa & R. Tilling), pp. 491–537. Springer.

Kohlstedt, D. L., Evans, E. & Mackwell, S. J. 1995 *J. Geophys. Res.* **100**, 17 587–17 602.

Le Grand, H. E. 1988 *Drifting continents and shifting theories.* Cambridge University Press.

Lockner, D. A., Byerlee, J. D., Kuksenko, V., Ponomarev, A. & Sidorin, A. 1991 *Nature* **350**, 39–42.

McKenzie, D. P. 1968 In *The history of the Earth's crust* (ed. R. A. Phinney), p. 28. Princeton University Press.

McKenzie, D. P. & Parker, R. L. 1967 *Nature* **216**, 1276–1280.

Main, I. G. 1996 *Rev. Geophys.* **34**, 433–462.

Main, I. G., Meredith, P. G. & Jones, C. 1989 *Geophys. J.* **96**, 131–138.

Main, I. G., Meredith, P. G., Sammonds, P. R. & Jones, C. 1990 In *Deformation mechanisms, rheology and tectonics* (ed. R. J. Knipe & E. H. Rutter). Geol. Soc. Spec. Publ. no. 54, pp. 81–96.

Marone, C. 1998 *A. Rev. Earth Planet. Sci.* **26**, 643–696.

Mercer, J. H. 1978 *Nature* **271**, 321–325.

Morland, L. W. 1992 *Surv. Geophys.* **13**, 209–268.

Murrell, S. A. F. 1989 In *Deformation processes in minerals, ceramics and rocks* (ed. D. J. Barber & P. G. Meredith), pp. 109–137. London: Unwin Hyman.

Murrell, S. A. F. & Digby, P. J. 1970 *Geophys. J. R. Astr. Soc.* **19**, 309–334.

Nikitin, L. V. & Chesnokov, E. M. 1981 *Izvestiya Earth Phys.* **17**, 174–183.

Nye, J. F. 1957 *Physical properties of crystals.* Oxford University Press.

Ohnaka, M. 1996 *Proc. Natn. Acad. Sci. USA* **93**, 3795–3802.

Ohnaka, M. & Shen, L.-f. 1999 *J. Geophys. Res.* **104**, 817–844.

Olsen, M. P., Scholz, C. H. & Leger, A. 1998 *J. Geophys. Res.* **103**, 12 223–12 237.

O'Nions, K. 1992 In *Understanding the Earth: a new synthesis* (ed. G. C. Brown, C. J. Hawkesworth & R. C. L. Wilson), pp. 145–163. Cambridge University Press.

Palmer, A. C. & Rice, J. R. 1973 *Proc. R. Soc. Lond.* A **332**, 527–548.

Paterson, M. S. 1978 *Experimental rock deformation—the brittle field.* Springer.

Paterson, W. S. B. 1994 *The physics of glaciers*, 3rd edn. Elsevier.

Rist, M. A., Sammonds, P. R., Doake, C. S. M., Oerter, H., Murrell, S. A. F., Meredith, P. G. & Matsuki, K. 1999 *J. Geophys. Res.* **104**, 2973–2987.

Rutter, E. H. & Neumann, D. H. K. 1995 *J. Geophys. Res.* **100**, 15,697–15,715.

Sahami, M. 1994 *Applications of percolation theory.* London: Taylor & Francis.

Sammonds, P. R., Meredith, P. G. & Main, I. G. 1992 *Nature* **359**, 228–230.

Sammonds, P. R., Murrell, S. A. F. & Rist, M. A. 1998 *J. Geophys. Res.* **103**, 21 795–21 815.

Sammonds, P. R. & Ohnaka, M. 1998 *Geophys. Res. Lett.* **25**, 699–702.

Scholz, C. H. 1998 *Nature* **391**, 37–42.

Smith, W. D. 1981 *Nature* **289**, 136–139.

Solomon, S. C. (and 10 others) 1992 *J. Geophys. Res.* **97**, 13 199–13 255.

Tenthorey, E., Scholz, C. H., Aharonov, E. & Leger, A. 1998 *J. Geophys. Res.* **103**, 23 951–23 967.

Vaughan, D. G. & Doake, C. S. M. 1996 *Nature* **379**, 328–330.

Warrick, R. & Oerlemans, J. 1990 In *Climate change—IPCC scientific assessment* (ed. J. T. Houghton, B. A. Callander & S. K. Varney), pp. 262–281. Cambridge University Press.

Wegener, A. L. 1915 *Die Entstehung der Kontinente und Ozeane.* Brunswick: Friederich Viewag.

Wingham, D. J., Ridout, A. J., Scharroo, R., Arthern, R. J. & Shum, C. K. 1998 *Science* **282**, 456–458.

Yoshida, S., Clint, O. & Sammonds, P. R. 1998 *Geophys. Res. Lett.* **25**, 1577–1580.

Zoback, M. L. 1992 *J. Geophys. Res.* **97**, 11 703–11 728.

Zhu, W. & Wong, T.-f. 1997 *J. Geophys. Res.* **102**, 3027–3041.

Zhu, W. & Wong, T-f. 1999 *J. Geophys. Res.* **104**, 2963–2971.

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