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

The Nature and Determination of Stress in the Accessible Lithosphere [and Discussion]

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The nature and determination of stress in the accessible lithosphere

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Results of stress determination show that stress state is characteristically heterogeneous and apparently unpredictable. Characterization of *in situ* stress state requires the determination of stress at individual locations and then spatial if not also temporal extrapolation. A variety of different measurements are used as a basis to determine stress state. Thus an adequate understanding of both the nature and origin of rock stress is essential. Representation of the lithosphere as a non-equilibrium, dissipative, dynamical system is shown to be consistent with observations of stress state and fluctuation of crustal displacement. The evolution of rock cores subject to a variety of perturbations in the laboratory can similarly be shown to be consistent with the evolution of a dissipative, dynamical system, driven in part by stored strain energy. These observations are inconsistent with the assumption that rock stress can be adequately represented by superposed traction and internally balanced stresses arising from quasi-static processes. The potential value of analyses of the dynamics of rock stress evolution is emphasized as a means to simplify the apparent complexity arising from present perceptions.

1. Introduction

The motivation to characterize the distribution of forces in the lithosphere stems from geological hazards, engineering activities and resource exploration. Because of the coupled nature of geological processes, these forces profoundly influence not only the capacity for work, but also a wide range of transport phenomena, for example fluid flow.

For reasons of cost and practical difficulty, determinations of stress are often both incomplete and sparse. An ability to characterize the distribution of stress throughout a given volume of rock on the basis of a very limited suite of measurements is therefore essential. A predictive capability based on extrapolation from point measurements to neighbouring volumes would be very valuable.

The intent of this paper is to present a perspective of current practice and perceptions and a basis for future developments in the characterization of stress state from individual measurements. Inevitably, this involves appraisal of the nature and genesis of stress, being a logical basis for interpreting individual measurements and extrapolating from a sparse data set. We focus on the accessible lithosphere, by which we mean that part of the lithosphere to which we can potentially gain access for measurement purposes. In practice, this is usually less than 5 km, but here we

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assume a depth limit of 10 km. Other than for purposes of illustration, we make no attempt to document previous determinations or characterizations of stress.

With the exception of the term residual stress in place of initial stress, the terminology of Timoshenko & Goodier (1982) is used. The adjective dynamical is used to describe systems in which the configuration at a later time is a function only of the configuration at an earlier time. Taking an example from Arrowsmith & Place (1982) as an illustration, if the state of a system is given by the vector $\boldsymbol{x}(t)$ at time t then the time-development of states is governed by the dynamical equations (or equations of motion) $\dot{\boldsymbol{x}} = \boldsymbol{X}(\boldsymbol{x})$.

2. Current practice and perspective

(a) *The nature and origin of stress*

Early in the history of rock stress determination, it was recognized that stress differences are characteristic of *in situ* stress state (Hast 1967). Hast noted differences between all three principal stresses at a point. In 1978, following a review of a large number of stress determinations, McGarr & Gay concluded that it is not possible to generalize concerning the magnitudes of horizontal stresses except that they usually increase with depth. They may be lesser or greater than the vertical stress. Indeed, the vertical stress may be less than or greater than that of the calculated lithostatic value (McGarr & Gay 1978; Barton 1982; Martin 1990). McGarr & Gay (1978) concluded that the orientation of horizontal stress may be either relatively homogeneous, with local anomalies, or completely incoherent from site to site. The same authors concluded that no reference state of stress exists that represents any particular region and also has worldwide applicability. The degree of consistency of the direction of principal stresses projected onto the horizontal plane is evident primarily at large scales, notably those which are a significant fraction of plate dimensions. However, bearing in mind that these directional data provide extremely limited information concerning the stress tensor, the inevitable conclusion is that stress state in the upper lithosphere is characteristically heterogeneous. (Apparent exceptions may occur locally in weak rock unable to support a shear stress, e.g. in highly overpressured formations.) Because stress state is unpredictable (McGarr & Gay 1978; Hiltscher *et al.* 1979), it is necessary to perform measurements to determine the stress state in any given volume of rock.

A summary of the origin of stresses in the lithosphere is given by Ranalli (1986). Forces promoting translation in the plane of the plate are body forces acting in the vicinity of divergent and convergent margins. Those resisting motion are surface forces including mantle drag, transform fault resistance, frictional resistance associated with collision, viscous resistance of the mantle of slab sinking and bending forces in the vicinity of trenches. Intraplate stress is strongly influenced by gravitational forces. All these forces maintain a reservoir of strain energy in the lithosphere which is relieved at approximately the same rate as the rate of input (Bott & Kusznir 1984). These authors note that a concentration of stress in the upper lithosphere can be inferred from the variation of rheology with depth.

All these forces are normally viewed as quasi-static, the influence of the dynamics of change and characteristics of the dynamics, such as fluctuation, having received little or no attention.

Many investigations requiring determination of stress state are concerned with volumes of rock comprising a very small part of a tectonic plate. Many are limited

Table 1. *Bases for stress determination*

basis	measurement variable
earthquake	seismic radiation
hydraulic fracture	fluid pressure
borehole	geometry of finite deformation, reloading strains
core	relief strains, evolution of relief strains, reloading strains, finite deformation state

to the accessible lithosphere. At this much smaller scale, rock stresses are also typically treated as resulting from external forces (surface forces and body forces) alone. However, it has long been recognized that strain relief measurements in isolated bodies of rock, free of surface forces, characteristically reveal the presence of residual strains (Emery 1963). Consideration of the possible modes of genesis (Voight & St Pierre 1974; Holzhausen & Johnson 1979), suggests that residual stresses are characteristic of the majority of rocks derived from the accessible lithosphere. However, with few exceptions, they have been ignored. This might be acceptable if the scale of balanced domains is much larger than the volume of rock of interest or the influence of residual strain and stress are otherwise insignificant. If this is not the case, as the strain relief measurements and genetic considerations suggest, then treatments of stress in terms of surface and body forces alone may be misleading.

(*b*) *Determination of stress at a point location*

Table 1 presents a simple grouping of the principal methods of stress determination. Although the distinction between different methods may not be as straightforward as suggested by the table, it serves to emphasize that the measurement basis is not common to all methods. If the results of different methods are to be comparable (i.e. quantitatively relatable to the single parameter stress), this implies a knowledge of the relationship between wave propagation from earthquakes responding to a long-term strain energy accumulation; the opening and closing pressures of a fracture; almost instantaneous strain relief; time-dependent strain relief; and, progressive failure. All of these measurements are interpreted in terms of tractions (external forces). The question of whether we have adequate understanding to compare the results of different test methods will be considered in §3 (apart from that relating to focal plane analyses). It is also apparent that the four bases of stress determination (table 1) represent three length scales extending to approximately three orders of magnitude. For results to be comparable implies either a scale-independence of stress state or a knowledge of any scale-dependence.

For many geological and engineering purposes, stress is determined using hydraulic fractures, boreholes or rock core. Theoretically, these methods all have the potential for complete determination of the stress tensor from a set of measurements at a point location but in practice this only applies when rock coring is undertaken.

The creation and re-pressurization of hydraulic fractures is a well-tried and widely accepted technique of stress determination. Combined with a technique for identifying fracture orientation, multiple boreholes are normally required to determine the stress tensor. The conventional method is limited to sections of borehole devoid of natural fractures but Cornet (1982) developed a technique specifically based on the re-pressurization of mutually inclined natural fractures.

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Major advantages of the hydraulic fracturing technique are the capability to operate at most drilled depths, the relative simplicity of the measurement principle and the considerable body of experience gained to date. The fluid pressure required to open and close the created fracture, assumed to be a mode 1 fracture, can be measured. (See Hickman & Zoback (1982) for a more detailed account.) In contrast to the majority of other stress determination methods, measured pressure is directly equatable to a traction representing one component of the stress tensor. Analysis of fracture initiation pressure in previously unfractured borehole additionally allows interpretation of the maximum traction acting normal to the borehole axis.

Borehole deformation, or breakout, is now widely used to identify the directions of horizontal stress. Caliper logs are often run for reasons other than stress determination, making the method inexpensive as well as practical. In contrast to hydraulic fracturing, the method relies entirely on inelastic processes (failure of the borehole wall). Determination of stress magnitude would require an ability to relate the finite change of borehole wall geometry to the perturbed stress state around the borehole by means of a quantified knowledge of the progressive failure process. Despite research efforts, the potential for determination of stress magnitudes from breakout logs is unpromising, in part because of the ill-defined complications of mud chemistry, and undocumented transient mud pressures and temperatures. Indeed, based on numerical simulations, Zheng *et al.* (1988) concluded that breakout geometry is both stress history-dependent and non-uniquely related to *in situ* stress. Other techniques based on borehole deformation are less used. Some involve re-loading of the borehole wall. Other, technically attractive research techniques are based on sidewall coring (Schmitt *et al.* 1985). No attempt is made here to comprehensively review all techniques.

Determination of stress based on rock core, initially confined to overcoring techniques (Hooker & Bickel 1974; Leeman 1969), has been extended to a range of other core-based methods (Teufel 1981). Use of overcoring methods, although able to determine both magnitude and direction of stress and even the complete stress tensor in a single determination (Leeman & Hayes 1966), has been constrained by depth limitations. The greatest depth capability recorded, 500 m (Hallbjorn *et al.* 1990), is substantially less than that of many oil field and geothermal operations. The principle of overcoring is based on the strain induced when a core is created, i.e. when the volume of rock which becomes the core is relieved of the applied stress. This strain relief process is assumed to be elastic (reversible) so that the core may be subsequently reloaded to determine suitable moduli from which to relate strain relief to stress relief. However, another technique (anelastic strain relief (Teufel 1981)) is entirely reliant on the time-dependency of core strain evolution. No allowance is made by the overcoring technique for this characteristic time-dependency. This would only be serious if the deformation properties of the core are irreversible and significantly different to *in situ* properties before overcoring, or if significant inelastic strains develop during a test (assuming that representation of time-dependent behaviour is not achieved by additional testing).

Overcoring strain measurements, in common with hydraulic fracturing pressure measurements, are interpreted as *in situ* tractions. Therefore, no allowance is made for any irreversible change of internal forces. One notable exception is the work of Hiltscher *et al.* (1979). These authors reported residual stresses in free boulders of 1–15 MPa, comparable with *in situ* stress magnitudes typically less than 30 MPa.

Attempts to deduce stress magnitudes from time-dependent strains observed when core is brought to the surface are also reported only in terms of tractions (Teufel 1985) and the potential influence of internal strains, inferred to be present from experiments involving residual strain release, are again not taken into account. It could be argued that the residual strains are solely a result of the unloading to a traction-free condition and cooling to ambient conditions and not present *in situ*. However, not only would this be unsubstantiated, but consideration of the genesis of residual strains in the context of well-known geological processes including changes of depth of burial (change of confining pressure and temperature), cementation by diagenetic fluids, pressure solution, and fracturing, strongly suggests that this is unlikely.

Use of differential strain curve analysis for stress determination (Strickland & Ren 1980) is based upon hydrostatic reloading of cubes cut from core. Stress directions are inferred from directional differences of reloading strains. Attempts have been made to deduce stress magnitudes from changes of slope of the stress–strain curves, inferred to correspond to the closure of microcracks formed during core strain relief. It is thus assumed that strain relief is essentially reversible, that the effect of a lack of shear stresses and other effects during core reloading are insignificant, that core dilation relates only to microcrack opening and that the effect of creating new free surfaces and the shape of the cube are insignificant. Most, if not all, of these assumptions have not been justified.

This necessarily brief review of methods of stress determination is neither comprehensive nor exhaustive. We have deliberately avoided a focus on details of test procedure. Readers wishing to learn either more of each of the techniques mentioned or of alternative techniques not mentioned here are strongly urged to examine the independent literature. Our purpose in this scant review is to highlight some of the potentially important assumptions and apparent inconsistencies in order to assess the present ability to determine rock stress state at a point location. First, we ask whether the fundamental nature of rock stress is adequately understood. Because the role of internal forces and their relationship to surface forces has not been defined, the essentially universal tendency to ignore internal strains remains, at best, unjustified. At worst, we must conclude that the nature of rock stress is not adequately understood. Second, in view of the fact that the various test techniques do not yield the same information, is there an adequate basis for comparison of the results achieved by different methods? If the role of internal forces is significant, or if there exists an undefined scale-dependency within the length scale spanned by the various forms of measurement, then the answer to this question is negative. Indeed, if a given test is influenced by internal forces in an ill-defined manner, even comparisons of the results of the same test conducted on different rock types may be misleading.

(c) *Characterization of stress distribution*

Given a group of individual stress determinations it is necessary to interpolate and extrapolate these in space, and sometimes in time, to characterize the stress distribution. Viewing the literature as a whole, we note a pronounced tendency to assume a uniform 'regional' or 'far-field' stress state (tractions) which is derived by assuming that directional data may be averaged and that depth variation may be reduced to a linear increase of each stress component. The linear increase with depth is either explicitly or implicitly related to gravitational body force or frictional criteria. Any departure from this uniformity may be either ignored, or related to local structure or contrasts of present-day material properties responding to 'far-

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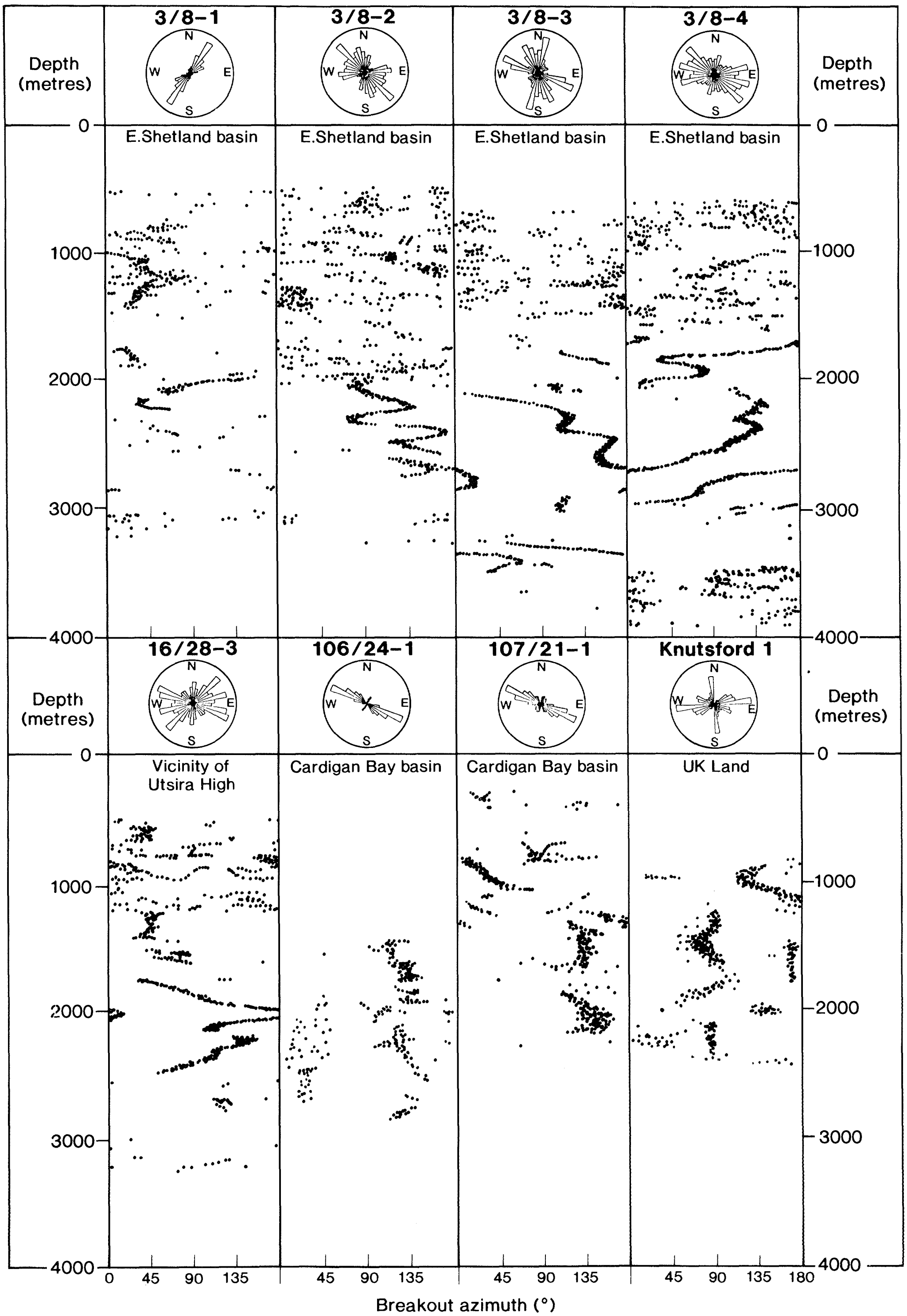

Figure 1. For description see opposite.

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field' tractions. The distribution of stress in time is normally ignored and therefore implicitly assumed to be unimportant.

Many records demonstrate that averaging of directional data can be misleading and helpful only when applied to localized volumes of rock with a demonstrated consistency of stress direction or if the variability concealed by the averaging process is of no consequence to the characterization of stress state (McGarr & Gay 1978; Martin 1990; Shamir *et al.* 1990; Dames & Moore 1978; Hiltscher *et al.* 1979, fig. 1). Often the assumption that averaging is justified is only implicitly stated.

Stress contrasts may be present across major stratigraphic discontinuities (Haimson & Lee 1980), and stress gradients and stress contrasts can be associated with faults, whether or not these are recognizably active (Anderson *et al.* 1983; Harper & Szymanski 1981*a*; Martin 1990). As noted by Hickman *et al.* (1988) spatial fluctuations in stress magnitude and orientation have been documented in homogeneous granitic rock. Kry & Gronseth (1982) concluded that the detailed nature of the *in situ* stress state in the Deep Basin of Alberta and British Columbia cannot simply be related to lithology. This conclusion is also demonstrated by the results compiled by Szymanski (1979) for tuff and crystalline rock in the Great Basin and reported by Dames & Moore (1978) for sediments in the Northern Appalachian Basin. These observations demonstrate that stress is not solely a function of lithology, present-day structure, and conventional rock mechanical properties.

Indeed Hast (1967), Voight & St Pierre (1974), Harper & Szymanski (1981*b*) and Szymanski (1989) and others have emphasized the time-dependency of stress. A combination of the object of the investigation and the rate and nature of energy transfer in the lithosphere determine the timescale of stress changes which may be relevant. For example, analysis of stress changes occurring during periods of only hundreds or thousands of years would be warranted for assessment of an active fault hazard, or for safety analyses of candidate high level waste repositories, particularly those sited in areas subject to high rates of energy transfer. Szymanski (1989) concluded that the rate of stress evolution at Yucca Mountain, Nevada, is sufficient to potentially influence such a facility. By numerically simulating the influence of a change of 'far-field' stress state on a jointed but otherwise uniform rock mass, Lemos *et al.* (1985) emphasized the influence of stress history on non-uniformity of stress distribution. These reports suggest that an understanding of the evolution of stress may be an essential prerequisite to an adequate interpretation of stress state. Static models may be insufficient.

3. A nonlinear systems perspective

(*a*) *Introduction*

The preceding review suggests that current conceptualizations of both the nature and origin of stress provide an incomplete basis for the characterization of stress

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Figure 1. Examples of the variation of direction of maximum horizontal stress with depth inferred from borehole breakout records in the North Sea, Cardigan Basin and onshore U.K. All boreholes are deviated less than 5° and defined by an ellipticity threshold of 0.05 in (*ca.* 13 mm). (Note that borehole ellipticity can be influenced by factors other than stress. The potential influences of bottom hole assembly stiffness contrasts, logging tool rotation, steeply dipping thin beds of contrasting strength and the effect of varying well trajectory have been assessed in interpretation of the pronounced progressive depth variation encountered in the E. Shetland wells. Only the effect of well trajectory was not eliminated (for lack of data) but since the deviation from the vertical is so slight, it is extremely unlikely that well trajectory effects are significant.)

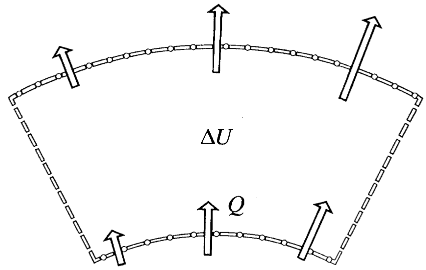

Figure 2. Conceptual thermodynamic model of the lithosphere. ⊐ ⊏, Mechanical work (time- and space-dependent), W; ⊐○⊏, mechanical work and heat flux (both time- and space-dependent). $\Delta U = Q - W$ is the change of internal energy. Note that the boundaries of the conceptual model are fixed in space and time.

state. Ideally, we would be better able to relate the different measurements which provide the basis for individual stress determinations. As a minimum, we must be able to relate the process which yields the measurement, the values represented as stress which are computed from this measurement and the engineering or geological process of interest. To translate these individual observations into a characterization of stress distribution, it is sometimes required to explicitly address the evolution of stress as a function of time as well as space. Unless we can be confident that stress distribution is independent of time, that is, independent of geological history and dominated by present-day quasi-static surface and body forces, it is an implicit requirement to consider stress evolution.

(*b*) *A conceptual thermodynamic model*

Figure 2 represents a generalized conceptual model of a section of the lithosphere. It is convenient to consider all of the lithosphere, but the model could equally well be confined to the accessible lithosphere. The model is based upon energy transfer and transformation (i.e. it is thermodynamic). A flux of energy and matter occurs across all boundaries, varying as a function of both time and space. For simplicity, the radial boundaries are assumed to be adiabatic, otherwise energy transfer across boundaries is by both work and heat transfer. At the lower boundary, energy is transferred from the mantle by time- and space-variable heat transfer or by work associated with shear and normal stresses acting on the base of the lithosphere. Mantle drag forces vary spatiotemporally in part because the profile of the base of the lithosphere is irregular (Wan *et al.* 1989) and varies with time, but this is not a necessary condition for this spatiotemporal variability of energy transfer. There is no reason to assume that forces acting on the lateral boundaries, promoting circumferential plate translation, are time-invariant, because they are associated with Rayleigh convection in the mantle and the lithosphere deforms internally in a time-dependent manner. Incessant heat flux and redistribution of matter occur at the Earth's surface. The latter, in conjunction with density changes caused by phase changes both in the mantle and in the lithosphere in the vicinity of the aesthenosphere–lithosphere boundary, and by other effects, results in changes of gravitational body force. The lithosphere is thus perceived as a non-equilibrium (in the thermodynamic sense), open system.

Energy is transferred within the system. In common with most natural systems heat generation accompanies work, whether mechanical, nuclear, electrical or other, and by mechanisms such as friction and thermoelastic effects, particle bombardment or current flow. From the second law of thermodynamics, we note that the system is dissipative. If $X_1, \ldots, X_n$ denote a complete set of macroscopic variables of such a system, their time evolution will take the form (Nicolis & Prigogine 1989).

$$\partial X_i/\partial t = F_i(X_1, \ldots, X_n, \boldsymbol{r}, t),$$

in which F_i may be complicated functions of the Xs and their space derivatives, as well as explicit functions of space $\boldsymbol{r}$ and time t. This evolution equation implies that a variability in time and space, with associated gradients, is characteristic of lithospheric systems for reasons additional to variable boundary flux.

The energy transformation and transport phenomena occurring within the system are characteristically nonlinear. Feedback loops occur in processes such as flow and chemical reaction (Nicolis & Prigogine 1990) and strain localization (Cundall 1990). Coupled processes with feedback relationships are widespread, such as fracturing and fluid pressure change, variation of rheology with strain rate and temperature, Rayleigh convection, diagenesis and flow of mineralizing solutions.

A remarkable number of useful conclusions can be drawn from this simple conceptual model. First, the model suggests that gradients are a characteristic of the lithosphere. This is consistent with the observations of stress heterogeneity (§2) and with the concept of internal forces. Second, the genesis of stress (or strain energy) occurs in the context of an evolving system; treatment of stress as an isolated variable or by superposition, such as linear thermoelasticity, may be unhelpful or misleading. Third, as a nonlinear, open, dissipative dynamical system, at least two very influential characteristics should be expected of the lithosphere: fluctuation and self-organization (Nicolis & Prigogine 1989). It seems that many geoscientists have traditionally viewed geological processes as basically monotonic. Quite in contrast, the simple dynamical model discussed here suggests that fluctuation, perhaps superimposed upon a monotonic trend, is incessant. The existence and perhaps surprisingly high rates of fluctuations are demonstrated by vertical crustal movements deduced from geodetic levelling (Whitten *et al.* 1977), horizontal strain measurements (Smith & Kind 1972), and by geomorphologic, stratigraphic and structural (Grant 1990) records. The work done during these fluctuations, conventionally neglected, can be expressed as a coherent, unidirectional effect such as heat transfer (Kurzweg 1985) and mass transfer (Chatwin 1975; Nilson & Lie 1990). Moreover, the conceptual model suggests that treatment of stress genesis as a quasi-static process (Warpinski 1986; Prats 1981) has limited potential.

Recognition of the truly thermodynamic nature of geological systems implies the daunting task of analysing the many coupled processes. However, one characteristic of dissipative systems offers at least the hope of remarkable simplification of the task of characterizing stress distribution. In contrast to conservative systems, dissipative systems are capable of eliminating the effect of perturbations and returning to a reference or attractor state. (They may also amplify small perturbations.) Fractal attractor states have been reported for a variety of geological features (Turcotte 1989; Main *et al.* 1990; Velde *et al.* 1990). The concept of self-organized criticality (Bak *et al.* 1988), having power law spatial and temporal correlations, has been applied to earthquakes (Bak & Tang 1989). Keilis-Borok (1990), recognizing that quantitative description of all the nonlinear, coupled relationships governing lithosphere behaviour is unrealistic, has emphasized the need for a generalized description. With the exception of Sornette *et al.* (1990), these concepts have yet to be applied specifically to stress evolution.

(c) *Laboratory tests*

(i) *Introduction*

The recognition that geological processes occur within the context of a system (Harper & Szymanski 1981*b*), such as outlined in §3, implicitly questions the applicability of laboratory tests staged such that many variables are constrained to be constant. This severely reduces or eliminates the interdependence characteristic of geological systems. Nevertheless, they are helpful providing that the effect of the imposed experimental conditions is fully recognized.

Laboratory tests can be used to indicate the potential significance of internal forces. In addition, they can be used to show that simple superposition of forces corresponding to residual stresses (stresses which exist in the presence of traction-free boundaries) and the hypothetical forces instantaneously arising within the material from the static application of tractions, cannot be used to describe the internal force distribution of rock *in situ*. The phrase interaction forces is therefore introduced here as a category of internal forces. It is used to describe the force distribution which occurs for given mechanical and thermal boundary conditions at a given time in the evolution of the rock, minus those forces which would result from a hypothetical static response to the application of the prevailing tractions (and thermal boundary conditions) if the rock behaved as a conventional continuum. This is certainly not a rigorous description. Moreover, it is of limited helpfulness because the absolute nature of such stresses cannot be quantitatively determined by experimentation. We shall see that it is very probable that even the 'residual' stress state in rock cannot be uniquely determined because a perturbation imposed to investigate residual stress causes change. However, interaction forces, although a loose description, at least refer to *in situ* geological conditions where tractions would normally form part of any description of stress state. The term is intended to emphasize the critical role of the evolving interactions which occur in a system of particles (grains, crystals, mineral groupings) or particles and cement. (Evolving interactions between much larger volumes of material are also implied, such as strata of differing properties or blocks bounded by discontinuities.) This new terminology is necessary because none of the established terms appear to be appropriate. From the subsequent review of laboratory experiments it may be inferred that the internal force distribution in many rocks evolves irreversibly when subject to perturbations such as the application or removal of external forces or temperature change. Consequently, neither initial stresses nor residual stresses, which represent a condition of zero external force, are identical to that component of the internal force distribution under load which does not simply reflect transmission of the external forces. Nor are these interaction forces solely self-balancing in the presence of external forces so the term self-equilibrating is also inappropriate.

(ii) *Observations*

The evolution of interaction forces and their relationship to surface forces, or tractions, can be inferred by subjecting suitable rock core, such as a sandstone, to a range of perturbations. Acoustic emission during creep experiments implies irreversible changes involving interaction force redistribution. Microseismicity occurs not only during loading but also during creep recovery (Harper 1970). That is, application of surface forces influences interaction forces. This is consistent with the torsion loading experiment conducted on granite by Tan & Kang (1980), who, after

Figure 3

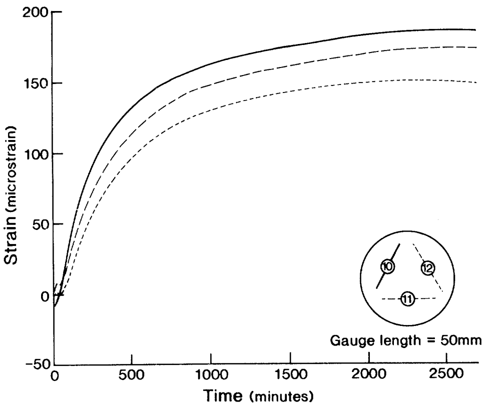

Figure 4

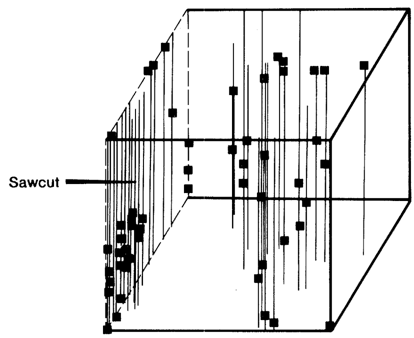

Figure 3. Time-dependent strain relief of Permian sandstone cored at a depth of 3042.4 m from vertical well 42/30-7. Strains were induced in the 100 mm diameter core of original length 150 mm by a sawcut normal to the core axis and 60 mm from the strain gauges placed on the flat surface normal to the core axis. (Note that the magnitudes of core strains induced by the single sawcut represent the upper limit of the spectrum of magnitudes observed in the Lower Leman sandstone in this area.)

Figure 4. Approximate distribution of acoustic emission accumulated in a core of Lower Leman sandstone during a period of 10.35 h by a sawcut normal to the core axis. The distribution of events is approximated by recording only those events detectable at all six transducers. The cylindrical core is represented as a cube. Each event is denoted by a rectangular box with a vertical line to indicate the relationship to the boundaries of the core.

unloading, observed a finite angular strain opposite in sense to the applied torsional load. Records of the release of internal strain energy to generate surface forces are rare. However, Price (1966) reported a time-dependent progression of extensional strain during the first loading stage of a compressional creep experiment. This implies that, for a period greater than 50 days, the nodular, muddy limestone sample was exerting a stress in excess of that applied (approximately 36 MPa). This rare example suggests that internal forces can influence surface forces. It is well known that changing sample geometry (cutting a free surface) results in strain and displacement of the sample boundaries, some examples of which are noted by Tullis (1977). Figure 3 records the time-dependent strain relief observed by strain-gauging one end of a North Sea sandstone reservoir core and sawing a slice of material from the other end. Successive cuts of this type have been found to initiate successive periods of strain relief, the strains in each case being extensional and appearing to stabilize in a period of approximately half a day. Acoustic emission is observed throughout the core, not simply in the vicinity of the saw cut (figure 4). Strains have been observed at a distance more than 10 times the largest dimension of the created free surface (figure 5), revealing that measurable affects can in some tests be observed for very small geometry changes and that perturbations may propagate to distances substantially greater than can be explained by elastic behaviour. Displacements of sample surfaces, accompanying irreversible sample strains, can also be induced by imposing minor temperature changes. Figure 6 shows the response of reservoir sandstone, revealing that finite extensional strains develop in response to cyclic minor temperature change. Again, imposition of successive temperature cycles results in successive increments of time-dependent extensional strain. At the

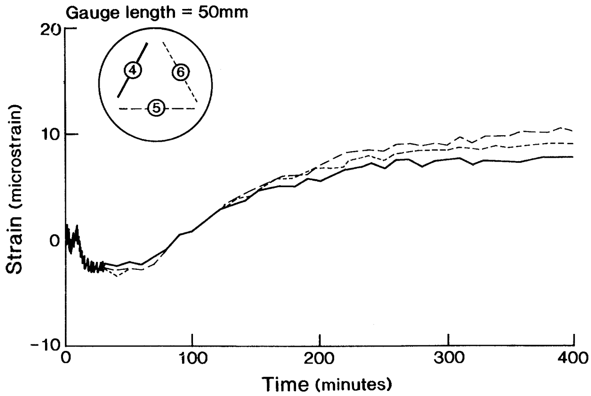

Figure 5. Strain relief remote from perturbation. Strains observed 235 mm from the base of an 8 mm diameter, 20 mm deep undercore in Lower Leman sandstone cored at 3031.42–3031.67 m in vertical well 48/6-30, North Sea. The strains were measured in a plane normal to the core axis.

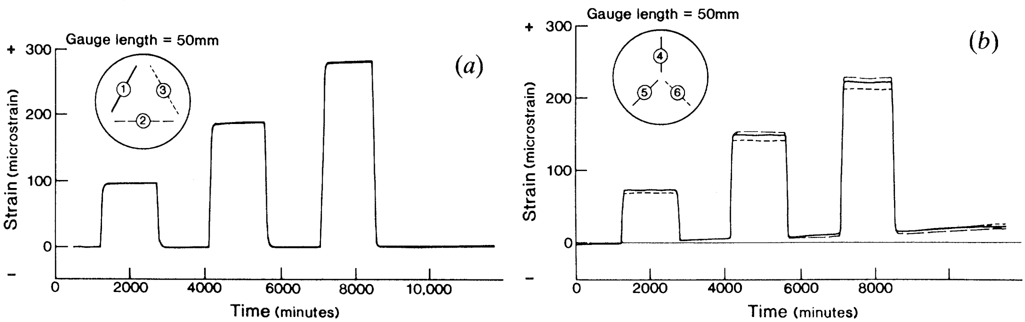

Figure 6. Response of sandstone to incremental heating cycles. Carboniferous sandstone cored at 3143 m from North Sea well 42/30-7. The 100 mm diameter core sample was strain-gauged normal to the core axis and heated with a steel control sample in a temperature-controlled oil bath.

conditions of zero confining pressure pertaining in the tests reported in figures 3–6, a distinct asymmetry is evident in that the finite strains developed in a time-dependent manner after returning clean sandstone cores to the reference state are always extensional.

(iii) *Interpretation*

Although more experiments of a different nature are required for a more convincing demonstration, these experiments strongly infer that surface forces (and/or displacements) and interaction forces (and/or displacements) are interdependent. The remote acoustic emissions and strains occurring in response to a local geometric perturbation reveal that the whole core may evolve. The effect of a local perturbation is propagated in a time-dependent irreversible manner away from the perturbation. A consequence of the local perturbation is a finite displacement at the sample boundaries. Contrary to the conclusions drawn by Holzhausen & Johnson (1979), but consistent with the hypothesis of Voight & St Pierre (1974), it is concluded that microscopic complexity, approaching apparent disorder, can be converted to macroscopic order and therefore coherently perform work. Interpreting the acoustic emission to represent crack initiation, propagation and sliding, and noting also that stress changes in elastic materials give rise to heat transfer (the thermoelastic effect),

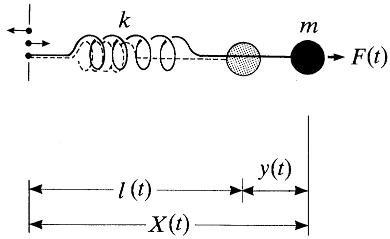

Figure 7. Single and mass system with variable equilibrium position $l(t)$. Modified after Dilts (1985).

the evolution of rock cores is seen to be consistent with the evolution of a dissipative dynamical system, and therefore also qualitatively consistent with the model of much larger lithospheric volumes developed in §3*b*.

Dilts (1985) represented processes such as microcracking by analogy to the elastic–plastic behaviour of a mass-spring system. This model can be used to illustrate the significance and evolution of the equilibrium state of the rock cores. Figure 7 shows Dilts's (1985) single mass-spring system, subject to an inhomogeneous force $F(t)$, in which the equilibrium position $l(t)$ can be varied independently of the motion of the mass m. The motion of the system may be described by

$$m\ddot{y} = -ky - m\,\mathrm{d}^2l/\mathrm{d}t^2 + F(t),$$

in which the term $m\mathrm{d}^2l/\mathrm{d}t^2$ represents the additional force generated by the changing equilibrium position. Now suppose that the equilibrium position is neither fixed nor arbitrarily variable but varies depending on the equations of motion, to represent yield of the spring when subject to either compressive or tensile forces. The motion can now be represented by

$$m\ddot{y} = -ky - m\,\mathrm{d}^2l(y,\dot{y})/\mathrm{d}t^2 + F(t).$$

Not only does the value of the equilibrium position $l(y,\dot{y})$ depend on the equations of motion, it in turn influences the equations of motion, so that the preceding expression represents a dynamical interplay between the two. Equilibrium becomes a dynamical quantity (state parameter). However, the actual value of the equilibrium position of mass m is irrelevant to the equations of motion because it merely introduces a constant force. The motion is influenced by the displacement from equilibrium y and the rate of change of this displacement, $\dot{y}$. The dynamical viewpoint illustrated by Dilts's analogy thus illustrates that the motion of the system is of overriding significance relative to the absolute value of the equilibrium state. This differs markedly from the more usual assumption that the state of a geological system is best described by quantifying the equilibrium state.

The approach to equilibrium involves fluctuation. Tan & King (1979) provided apparent experimental observations of such fluctuation. Dilts (1985) simulated fluctuation, including non-uniformity of the time-series, by calculations of the motion of multiple, connected mass spring elements. Multiple elements would better represent rock core, as does addition of a dissipative function (which was also incorporated by Dilts). It would also be appropriate to incorporate viscosity of the individual elements, but this is not a necessary condition for time-dependence. These additions would render the model more representative. However, the single mass-spring model provides a simpler basis to contrast the fundamental difference between descriptions which emphasize the equilibrium state of a geological system and those which emphasize the motion.

4. Conclusions

A number of different parameters and a range of forms of perturbation are used to determine rock stress state at a point location. Some aspects of the stress state determined by the different methods, notably the directions of the maximum and minimum horizontal stress, may be common to the different methods. However, based on the considerations given here, the stress magnitudes deduced from the different methods of measurement currently in use should not be assumed to be identical (even assuming precise, error-free measurements). The magnitude of the differences in a given situation may appear insignificant in engineering terms if treated simply as an error bar, or if members of individual determinations are averaged, thereby masking fluctuations. Nevertheless, it is desirable if not essential to develop a basis for reconciling the differences, or for selecting the measurement such that the measurement results most closely relate to the engineering or geological process in relation to which the stress measurements are conducted. The process giving rise to the quantitative value derived from a stress determination must be relatable to the relevant engineering or geological process. For example, a hydraulic fracture stress determination is ideal for designing pumping capacity and leak-off calculations for a stimulation treatment. However, it provides little or no information regarding the time-dependent closure of the fracture which can be inferred to result from evolution of the interaction forces in the walls of the stimulation fracture.

An understanding of the genesis of rock stress or some as yet unidentified generic relationship is necessary to the characterization of stress state from point determinations. Such determinations are often sparse and incomplete and consequently considerable extrapolation is required. Although any one of these factors may exert a dominant influence in a specific situation, observations suggest that stress is not simply related to lithology, structure or present-day conventional rock mechanical properties. Heterogeneity and gradients are characteristic of *in situ* stress state.

It is apparent that interaction forces should not be ignored. Field and laboratory results and theoretical considerations consistently suggest that interaction forces and surface forces are interdependent and not subject to superposition, which requires linearity. Interaction forces are characteristic of many rocks and probably of the entire accessible lithosphere. Internal strain changes and surface displacements are interdependent. It can be inferred that interaction forces influence the time-dependency of stress, strain and displacement and capacity for propagation of perturbations. Indeed, these considerations re-emphasize that rock stress is inextricably linked with rock constitutive properties. The two are interdependent expressions of the evolution of a nonlinear system. In practice they are separated, but the degree to which we can describe or predict the evolution of forces in geological systems may be seriously limited by this working approach.

For a number of reasons, but primarily because measurement involves perturbation which induces change, we may not be able to quantify internal force distributions. The conventional target of rock stress determination programmes, a quantification of surface forces or tractions which are assumed to be time-invariant, linearly related to internal forces and to represent the equilibrium state, is typically an approximation of unspecified inaccuracy. The considerations presented here,

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which accept the role of interaction forces, suggest that the equilibrium state is, strictly speaking, indeterminable from quasi-static analysis. Stress determinations may involve change to or from a reference state, typically the stress-relieved state. Yet the dynamics suggests that the new equilibrium state (stress-relieved overcore or hydrofracture on the point of closure) is a nonlinear function not only of the pre-test equilibrium state but also of the distance and history of distance from equilibrium during and after perturbation (by stress-relief or fracture opening/closing).

More attention to the dynamics of change rather than solely the assumed equilibrium state is justified for a number of reasons. First, the dynamics may be more relevant to both engineering design and our understanding of geological processes. The former is concerned with a response to perturbation, the latter a manifestation of system evolution. Understanding stress genesis in a nonlinear system is almost certainly necessary not only for extrapolation of individual stress determinations in time but almost certainly also in space. Second, recognition of the significance of the dynamics implies that new techniques of stress determination based on the dynamics of change may be feasible. Given the practical and economic constraints of existing techniques, any sound additions to the range would be welcome (and, as noted, may more closely relate to some of the engineering processes of interest and to geological processes). Third, the dynamics are presently or potentially observable in the field and the laboratory. Finally, the accessible lithosphere is a non-equilibrium nonlinear dissipative system. There is therefore the potential for self-organization and a considerable simplification of the apparent complexity.

The core experiments were conducted by a team from BP Research led by Dr J. T. Hagan. In addition to this valuable contribution, and the support of BP Research, we acknowledge stimulating discussions with Dr M. I. Asgian, Dr W. B. Bradley and Dr K. J. Heffer. The data in figure 1 were provided by R. J. Klein.

References

Anderson, R. N., Zoback, M. D., Hickman, S. H. & Healy, J. H. 1983 *In situ* stress and physical property measurements in the Cleveland Hills fault zone, Oroville, California. *Eos Trans. AGU* **64**, 834.

Arrowsmith, D. K. & Place, C. M. 1982 *Ordinary differential equations*. London and New York: Chapman & Hall.

Bak, P. & Tang, C. 1989 Earthquakes as a self-organised critical phenomenon. *J. geophys. Res.* B **94**, 15635–15637.

Bak, P., Tang, C. & Wiesenfeld, K. 1988 Self-organised criticality. *Phys. Rev.* A **38**, 364–373.

Barton, N. 1982 Hydraulic fracturing to estimate minimum stress and rock mass stability at a pumped hydro project. *Proc. Workshop XVII on hydraulic fracturing stress measurements, U.S. Geol. Surv. Open-File Rep.* 82-1075.

Bott, M. H. P. & Kusznir, N. J. 1984 The origin of tectonic stress in the lithosphere. *Tectonophys.* **105**, 1–13.

Chatwin, P. C. 1975 On the longitudinal dispersion of passive contaminant in oscillatory flows in tubes. *J. Fluid Mech.* **71**, 513–527.

Cornet, F. H. 1982 Analysis of injection tests for *in situ* stress determination. *Proc. Workshop XVII on hydraulic fracturing stress measurements, U.S. Geol. Surv. Open-File Rep.* 82-1075.

Cundall, P. A. 1990 Numerical modelling of jointed and faulted rock. *Proc. Int. Conf. on Mech. of jointed and faulted rock, Vienna, April.* (In the press.)

Dames & Moore 1978 *Geologic investigation, Nine Mile Point Nuclear Station Unit 2*, vol. 3. *Rock stresses*. Niagara Mohawk Power Corporation.

Dilts, G. A. 1985 The dynamical equilibrium: a mathematical model for the dynamical description of the permanent deformation of an elastic material caused by microfracture or plastic flow. Ph.D. thesis, University of Colorado.

Emery, C. L. 1963 Strain energy in rocks. In *State of stress in the Earth's crust. Proc. Int. Conf., June 13–14, Santa Monica, California*.

Grant, N. T. 1990 Episodic discrete and distributed deformation: consequences and controls in a thrust culmination from the Central Pyrenees. *J. struct. Geol.* **12**, 7, 835–850.

Haimson, B. C. & Lee, C. F. 1980 Hydrofracturing stress determinations at Darlington, Ontario. In *Underground Rock Eng: Proc. 13th Can. Symp. on Rock Mech.*, pp. 42–50.

Hallbjorn, L., Ingevald, K., Martna, J. & Strindell, L. 1990 A new automatic probe for measuring triaxial rock stresses in deep bore holes. *Tunnelling Underground Space Technol.* **5** (1/2), 141–145.

Harper, T. R. 1970 An investigation of stressed rock by observations of acoustic emission. M.Sc., Imperial College, University of London.

Harper, T. R. & Szymanski, J. S. 1981*a* Structural control of rock mass stress distribution. In *Mechanics of structured media. Proc. Int. Symp. on the Mechanical behaviour of structured media, Ottawa, May*, Pt. B, pp. 283–299. Elsevier.

Harper, T. R. & Szymanski, J. S. 1981*b* A perspective of geological characterizations required for subsurface nuclear waste disposal. In *Subsurface space* (ed. M. Bergman). *Proc. Int. Symp. Stockholm*, vol. 3, pp. 1345–1352. Pergamon.

Hast, N. 1967 The state of stresses in the upper part of the earth's crust. *Eng. Geol.* **2**, 1, 5–17.

Hickman, S. H. & Zoback, M. D. 1982 The interpretation of hydraulic fracturing pressure–time data for *in situ* stress determination. *Proc. Workshop XVII on hydraulic fracturing stress measurements, U.S. Geol. Surv. Open-File Rep.* 82-1075.

Hickman, S. H., Zoback, M. D. & Healy, J. H. 1988 Continuation of a deep borehole stress measurement profile near the San Andreas Fault. 1. Hydraulic fracturing stress measurements at Hi Vista, Mojave Desert, California. *J. geophys. Res.* **93**, 15183–15195.

Hiltscher, R., Martna, J. & Strindell, L. 1979 The measurement of triaxial rock stresses in deep boreholes and the use of rock stress measurements in the design and construction of rock openings. *Proc. 4th Int. Congr. Rock Mech., Montreux,* vol. 2, pp. 227–234.

Holzhausen, G. R. & Johnson, A. M. 1979 The concept of residual stress in rock. *Tectonophys.* **58**, 237–267.

Hooker, V. E. & Bickel, D. L. 1974 Overcoring equipment and techniques used in rock stress determination. *U.S. Bur. Mines Rep. Invest.* 8618, 32 pp.

Keilis-Borok, V. I. 1990 Introduction: non-linear systems in the problem of earthquake prediction. *Phys. Earth planet. Int.* **61**, 1–7.

Kry, P. R. & Gronseth, J. M. 1982 *In situ* stresses and hydraulic fracturing in the Deep Basin. Preprint, *33rd Ann. Tech. Mtg of Pet. Soc. of CIM, Calgary, June*.

Kurzweg, U. H. 1985 Enhanced heat conduction in fluids subjected to sinusoidal oscillations. *J. Heat Transfer* **107**, 459–463.

Leeman, E. R. 1969 The 'doorstopper' and triaxial rock stress measuring instruments developed by the CSIR. *J. S. Afr. Inst. Min. Metall.* **69**, 305–339.

Leeman, E. R. & Hayes, D. J. 1966 A technique for determining the complete state of stress in rock using a single borehole. In *Proc. 1st Cong. Int. Soc. Rock Mech., Lisbon,* vol. 2, pp. 17–24.

Lemos, J. V., Hart, R. D. & Cundall, P. A. 1985 A generalized distinct element program for modelling jointed rock mass. *Proc. Int. Symp. on fundamentals of rock joints, Bjorkliden, Centek, Lulea*.

Main, I. G., Peacock, S. & Meredith, P. G. 1990 Scattering attenuation and the fractal geometry of fracture systems. *Pageoph.* **133**, 283–304.

Martin, C. D. 1990 Characterising *in situ* stress domains at the AECL Underground Research Laboratory. *Can. Geotech. J.* **27**, 631–646.

McGarr, A. & Gay, N. C. 1978 State of stress in the Earth's crust. *A. Rev. Earth planet. Sci.* **6**, 405–436.

Nicolis, G. & Prigogine, I. 1989 *Exploring complexity*. New York: Freeman.

Nilson, R. H. & Lie, K. H. 1990 Double-porosity modelling of oscillatory gas motion and contaminant transport in a fractured porous medium. *Int. J. numer. analyt. Methods Geomech.* **14**, 565–585.

Prats, M. 1981 Effect of burial history on the subsurface horizontal stresses of formations having different material properties. *Soc. Petrol. Engng J.* 658–662.

Price, N. J. 1966 *Fault and joint development in brittle and semi-brittle rock*. Pergamon.

Ranalli, G. 1987 *Rheology of the Earth*. Allen & Unwin.

Schmitt, D. R., Bass, J. D. & Ahrens, T. J. 1985 Application of holographic interferometry to *in situ* stress measurement. *26th U.S. Symp. on Rock Mech., Rapid City, 26–28 June*, pp. 1055–1056.

Shamir, G., Zoback, M. D. & Cornet, F. H. 1990 Fracture-induced stress heterogeneity: Examples from the Cajon Pass scientific drillhole near the San Andreas Fault, California. In *Rock joints* (ed. N. Barton & O. Stephansson). Rotterdam: Balkema.

Smith, S. W. & Kind, R. 1972 Regional secular strain fields in Southern Nevada. *Tectonophys.* **14**, 57–69.

Sornette, D., Davy, P. & Sornette, A. 1990 Structuration of the lithosphere in plate tectonics as a self-organised critical phenomenon. *J. geophys. Res.* **95**, 17353–17361.

Strickland, F. G. & Ren, N.-K. 1980 Predicting the *in situ* stress state for deep wells using differential strain curve analysis. *Pet. Soc. CIM* Pap. no. 80-31-33.

Szymanski, J. S. 1989 Conceptual considerations of the Yucca Mountain groundwater system with special emphasis of the adequacy of this system to accommodate a high-level nuclear waste repository. U.S. Dept of Energy, Las Vegas.

Tan, T.-K. & Kang, W.-F. 1980 Locked in stresses, creep and dilatency of rocks, and constitutive equations. *Rock Mech.* **13**, 5–22.

Teufel, L. W. 1981 Strain relaxation method for predicting hydraulic fracture azimuth from orientated core. *Proc. SPE/DOE Low permeability gas reservoir symp.*, SPE/DOE 9836 Denver, 81–86.

Teufel, L. W. 1985 *In situ* stress state at the Mounds test well as determined by the anelastic strain recovery method. *SPE/DOE low permeability gas reservoirs symp.*, Denver, SPE/DOE 13896, 467–474.

Timoshenko, S. P. & Goodier, J. N. 1982 *Theory of elasticity*. McGraw-Hill.

Tullis, T. 1977 Reflections on measurement of residual stress in rock. *Pageoph.* **115**, 57–68.

Turcotte, D. L. 1989 Fractals in geology and geophysics. *Pageoph.* **121**, 171–196.

Velde, B., Dubois, J., Touchard, G. & Badri, A. 1990 Fractal analysis of fractures in rocks: the Cantor's dust method. *Tectonophys.* **179**, 345–352.

Voight, B. & St Pierre, B. H. P. 1974 Stress history and rock stress. *Proc. 3rd Congr. Int. Soc. Rock. Mech., Denver, A*, vol. 2, 580–582. Pergamon.

Wan, T., Tang, Y. & Zheng, W. 1989 Thermal structure of the lithosphere in Fujian, China. *J. geophys. Res.* **94**, 1888–1894.

Warpinski, N. R. 1986 Elastic and viscoelastic calculations of stresses in sedimentary basins. *Unconventional Gas Tech. Symp. of Soc. of Pet. Eng.*, Louisville, SPE 15243, 409–422.

Whitten, C. A., Green, R. & Meade, B. K. (eds) 1977 Recent crustal movements. *Tectonophys.* (Special Issue) **52**, 1–4.

Zheng, Z., Myer, L. R. & Cook, N. G. W. 1988 Borehole breakout and stress measurements. *Key questions in rock mechanics* (ed. P. A. Cundall, C. L. Starling & A. M. Starfield). *Proc. 29th U.S. Symp.* Rotterdam: Balkema.

Discussion

P. Hancock (*Bristol University, U.K.*). Accepting Dr Harper's observations as an account of what happens today, particularly in the context of variations in vertical stress of both magnitude and direction, why does he think that so many brittle geological structures in regions of simple tectonics in horizontal rocks are so uniformly orientated throughout such large areas? What does he understand by saying 'a long period of time'? Is this is a long period to a geologist or a long period to an experimentalist?

T. R. Harper. I am aware of Dr Hancock's observations not only of the uniformity of joint direction but also that this direction appears to correspond to the direction of maximum horizontal stress obtained by averaging a large number of *in situ* determinations. In this context we discussed the Syracuse–Oswego area in New York State, where a pronounced variation of stress magnitude and direction near the surface was recorded within a region of approximately 1 km^2 near Oswego. More than 100 successful determinations were made, and the variation was attributed to local structural evolution. I understand that when all these directional results were subsequently averaged, the result was a direction subparallel to the strike of joints in the region of Syracuse.

It may be relevant to recall the borehole breakout data we showed (figure 1). A substantial variation of the direction of horizontal stress with depth is inferred. When all the data points are represented as a rosette diagram (figure 1), they typically indicate a pronounced preferential direction (in some cases with approximately symmetrically disposed secondary directions). At larger scales, such as the U.K. North Sea region, some overall consistency, with local variation, is observed. From the non-equilibrium system viewpoint, which I have emphasized, fluctuation is to be expected not only in space but also in time. From this perspective, drawing such observations together, one can speculate that if the brittle geological structures developed progressively over a long period of time, then their finite resulting configuration may approximate to an averaged value of the stress directions which fluctuated in space and time throughout the period of growth of the structures.

I have made no attempt to quantify the period of time involved, but I used the adjective 'long' not in a geological context but to denote periods of typically 10^4–10^6 years. In high-energy flux areas, the lower bound would be reduced by an order of magnitude or more. Please note, however, that these numbers are only a generalization, not only because they are not rigorously derived, but also because they have little meaning without specifying accompanying magnitudes. Fluctuation of geological systems can be observed at very much shorter timescales (less than 10 years), such as by geodetic relevelling. The motion of geological systems appears to occur at a wide range of scales, and one is tempted to speculate that the spatiotemporal fluctuations may be characteristically self-similar.

A. Batchelor (*Geoscience Ltd, U.K.*). Dr Harper made the general conclusion about the vertical stress not being a function of integrated density and showed an example from Canada. Does he have any more evidence for that? My experience of measuring stresses at depth is that once you get below 400 or 500 m in an area of not severe topography the vertical stress can indeed be approximated by the density. In the North Sea, in a block not so far from the one that he mentioned, I would argue that we saw evidence that vertical stress could be estimated from integrated density.

T. R. Harper. Although for many engineering purposes it is often adequate to approximate the vertical stress by depth-integration based on density logs, I have attempted to emphasize that it is not always helpful to assume that the vertical stress can be accurately represented by such an assumption. I have in mind two particular situations in which this assumption would be unhelpful. The first is high-energy transfer rate environments, in which the stress state is evolving relatively rapidly and one is concerned with quantifying the implications, such as any fluid flow driven by evolving deformation gradients. This situation can be inferred to pertain at Yucca mountain, Nevada, a candidate nuclear waste repository site with active faults and high heat-flow gradients.

The second situation is that in which the objective is to understand lithospheric processes. Perhaps the presence of gradients of vertical stress departing from that calculated from overburden weight is easiest to appreciate by reference to the structural evolution of an evolving high-angle normal fault, for which an evolving gradient of fault-parallel normal stress may be readily inferred on genetic grounds.

Observations suggest a spatial variability of stress which, when averaged, suggests some consistency. Regrettably, determinations of vertical stress are far less numerous than of horizontal stresses and so (bearing in mind also the problems of measurement error) it is not an easy matter to test the proposed variability in relation to vertical stress. Nevertheless, Dr Batchelor will recall that the overcoring measurements with which he was involved at South Crofty Mine showed precisely this effect: I believe six CSIRO measurements at a depth of about 800 m yielded values of the near-vertical intermediate principal stress which varied substantially (*ca.* 9 MPa), yet the average of the six values, which was aligned vertically, yielded, as you have implied, an acceptable approximation of the magnitude which would be obtained from the integrated density approximation.

Hiltscher *et al.* (1979) also observed a variability of vertical stress. Other references citing a departure from the calculated lithostatic value are given in the written paper. So, to summarize, it appears that integration of density logs is adequate for many but not all engineering purposes, but that there is sufficient evidence to warn us that it is neither accurate, helpful nor supportable on genetic grounds to assume that the vertical stress is always equal to the integrated weight of overburden.

S. Murrell (*University College, London, U.K.*). In the thermal cycling experiment in the laboratory, was allowance made for the slow diffusion of heat in rock samples where one would expect there to be substantial strain lag effects? Could this be the possible explanation for the difference between steel and rock?

T. R. Harper. The 100 mm diameter core and control samples were maintained at the elevated and reference temperatures for sufficient time for temperatures to become uniform, typically not less than one day. Obviously transient temperature and stress gradients are imposed, but we confirmed that the cores were uniformly heated for the majority of each temperature cycle by continuously recording the output from thermocouples with sensors located in the centre of the samples.

N. Kusznir (*University of Liverpool, U.K.*). One of the issues concerning many of the people here is to what extent we can use local measurements of stress to actually get a feel for the horizontal force distribution in the lithosphere on a global scale. What is the implication of the variability, both temporally and spatially, that is observed for our mapping of what is often called 'global stress', but what is probably an approximation to global lithosphere force?

T. R. Harper. In addition to the variability of the results of stress determinations that I have discussed, geological, geomorphological and geodetic observations demonstrate that spatiotemporal fluctuation extends across a wide range of scales. These data are consistent with the concept of the lithosphere as a non-equilibrium thermodynamic system, and suggest that this variability should apply throughout the lithosphere. Therefore, we should be cautious of extrapolations based on the assumption of steady-state conditions.

Near-surface stress determinations are potentially a useful guide to horizontal force distribution in the lithosphere only if first we make allowance for the spatial variability in the stress direction determinations and, second, develop a genetic model upon which to base the extrapolation to depth. Also, the concept of stress may be inappropriate when gradients are characteristic of a wide range of scales. This becomes readily apparent at the microscopic level, at least in clastic and crystalline rocks, at which scale the use of force is more helpful. It is also particularly apparent at the other extreme of spatial scale, involving large volumes of lithosphere, largely inaccessible, at which scale the lack of resolution potentially achievable also suggests that the use of force may be more suitable than stress.

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