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The origin of tensile fracture lineaments

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Abstract—Fracture-controlled lineaments, commonly seen where brittle basement is exposed at the earth's surface, are generally restricted to a small number of sets, with angles of 45–90° between sets. The length–frequency distribution of lineaments in each set follows a truncated Poisson function. Such lineaments usually show almost no shearing offset, suggesting a tensile origin. A simple mechanical model of tensile fracturing is used to explain the spacing, directions, and length of lineaments, as well as their depth–frequency distribution. Results suggest that the penetration depth of tensile fractures which produce lineaments at the earth's surface is directly related to their length and that the fracture density is inversely proportional to fracture depth. Finally, the angles between lineament sets may be controlled by the ratio of strength of unfractured rock to that of pre-existing fractures, which might heal with time.

The most likely source of tension is tectonic uplift. Fractures due to typical uplifts of 0.5–1 km over distances of 10–100 km may penetrate as brittle fractures to several kilometres into the crust, perhaps to the depth at which seismic activity ceases.

INTRODUCTION

THE RECENT advent of satellite images and high altitude photography has led to the recognition that a multitude of linear features, ranging in length from metres to hundreds of kilometres, crisscross the surface of the earth's crust. In many places these lineaments are important components of the earth's surface morphology. Most common are the multiple linear arrays of creeks and rivers in exposed or nearly exposed basement complexes. Such drainage systems commonly follow existing fractures, apparently taking advantage of their lower resistance to erosion.

Typically, the surface of the crust is broken by 2–4 sets of sub-parallel lineaments, each lineament with a different azimuth (e.g. Blanchet 1957, Plafker 1964, Gay 1977). The direction and uniformity of this pattern may persist over areas up to 10,000 or more square kilometres. The traces of larger fractures are obscure at ground level and stand out clearly only when viewed from a great height. This obscurity is due in part to the large scale of these features and the erosion around them which is often enhanced and masks their true nature. Most significant, however, is the common absence of observable shear displacements accross these fractures. This lack of displacement, which distinguishes these fractures from faults, prevents geologists from clearly identifying such structures in the field and also obscures their mechanical origin and role in crustal tectonic processes.

Several efforts to explain the origin of lineaments involve postulating some sort of shear rupture (for review see Hodgson 1961). Extensive laboratory evidence on the failure of rock under shear stress makes this interpretation appear attractive, but it is generally unsupported by observations *in situ*.

The alternative to shear failure for the origin of the lineaments is tensile failure. This explanation was first proposed about 150 years ago in a pair of remarkable papers by Hopkins (1835, 1841) in which vertical tensile fracturing due to structural bending was held responsible for the occurrence of lineaments. The same view was held by McGee (1882), Bucher (1920), and others. More recently, Gramberg (1966) has also suggested a tensile origin for lineaments.

This study seeks to determine, first, whether sufficient tensile strain can develop in the earth's crust to produce tensile fracturing and whether such strain could create the observed pattern of intersecting sets with their widespread spatial uniformity. Secondly, estimates are made of the amount of opening during the fracturing process, the factors controlling fracture spacing, and the depth to which these fractures may penetrate. The deeper their penetration, the more important their role in crustal tectonics and applied geology. For example, a deep fracture may serve as a conduit for deep ground water flow, which in turn has major mechanical and other effects on earthquakes. Deep fractures may also provide natural permeability for potential extraction of geothermal energy and for mobility of mineralized fluids which form ore deposits. Furthermore, the implications for nuclear waste disposal can be significant.

It is shown that tensile fractures could extend deeply into the crust due to tectonic tension superposed on the lithostatic overburden. Secondly, it is shown that the spacing and the Poisson-like length-frequency distribution of lineaments is consistent with such tensile origins. The small number of directions of lineaments is also well explained by the tensile model. The angles between sets are interpreted in terms of the tensile strength of crustal rocks. Thirdly, it is shown that tectonic uplift, according to Bucher's (1920) notion, could be responsible for the tensile field required.

TENSILE STRESS IN THE CRUST

It has long been established in laboratory tests (e.g. Jaeger & Cook 1969) that rock fails in shear as long as the normal stress is compressive. When rock fails in tension, however, the tensile fracture is perpendicular to the direction of the tensile stress. The failure strength in shear is generally much greater than in tension, and it increases with confining pressure. In contrast, the tensile strength of rock is small, typically of the order of 100 bars. To assess the conditions for the development of absolute tension in the earth's crust, it is thus necessary to nullify the effective lithostatic pressure due to overburden. Assuming that the overburden pressure is approximately hydrostatic and ignoring pore pressure for the time being, the confining pressure p_c at any depth z in the crust is (see Fig. 1):

$$p(z) = g \int_0^z \rho(z')\,dz' = g\bar{\rho}z \tag{1}$$

where g is the earth's gravitational pull and $\rho(z)$ is the density. Using the approximate value of $\bar{\rho} = 3\ \mathrm{g\,cm^{-3}}$, we obtain $p(z) = \bar{\rho}g \cdot z = 300 \cdot z$, where p is given in bars and z in kilometres. Thus, the tensile stresses required to overcome the overburden pressure increase with depth at a rate of 300 bars km^{-1}.

The associated tensile strain ε is given by

$$\varepsilon(z) = \frac{\sigma_t(z)}{\mathrm{E}} = \frac{p(z)}{\mathrm{E}} = \frac{\bar{\rho}gz}{\mathrm{E}} \tag{2}$$

where σ_t is the tensile stress and E is Young's modulus of the rock. For a given uniform tensile strain ε_0, the depth z_c at which the applied tension is equal to the ambient lithostatic pressure is

$$z_c = \frac{\mathrm{E}\varepsilon_0}{\rho g}. \tag{3}$$

The stress required to produce tensile fractures is obtained by adding the tensile strength of rock S_t

$$\sigma_t(z) = p(z) + S_t = \bar{\rho}gz + S_t \tag{4}$$

$$\varepsilon(z) = \frac{\bar{\rho}g}{\mathrm{E}}z + \frac{S_t}{\mathrm{E}}$$

$$z_c = \frac{\mathrm{E}}{\rho g}\left[\varepsilon(z) - \frac{S_t}{\mathrm{E}}\right]. \tag{5}$$

Taking $S_t = 300$ bars and $\mathrm{E} = 10^6$ bars, (5) yields $z_c = 1, 5$ and 10 km for $\varepsilon_0 = 6 \times 10^{-4}$, 3×10^{-3}, and 6×10^{-3} respectively. In other words, an applied tectonic strain of 6×10^{-3} could, in principle, cause some kind of fracturing to a depth of about 10 km in the crust.

The gradual increase of applied tectonic tension causes surface fractures to open and extend downward as tension increases. As the tectonic tensile stress disappears, the tensile fractures cease to propagate. The net result is a set of linear surface traces without permanent shear offsets, perpendicular to the direction of the tension. Two questions are raised: (1) how are additional fracture sets induced, and (2) what controls the spacing of fractures within sets and angles between sets?

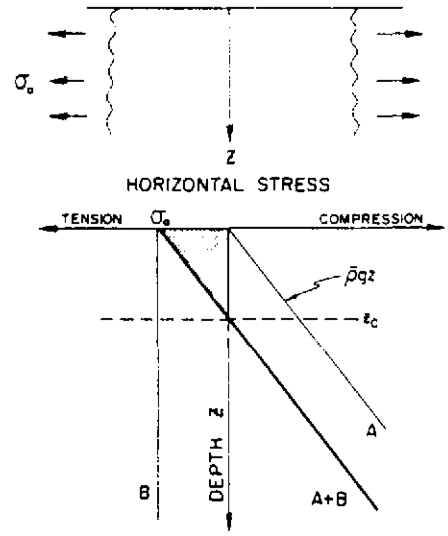

Fig. 1. Schematic presentation of uniform tension σ_0 (line B) superposed on lithostatic overburden (line A) which increases linearly with depth. A + B yields the depth z_c at which overburden is equal to the applied tension. Tensile cracks will penetrate approximately to this depth.

SPACING AND OPENING OF TENSILE FRACTURES

As mentioned above, fractures belong to sets each consisting of a multitude of subparallel planes, spaced a few metres or tens of metres apart. Furthermore, a few fractures have a much more pronounced effect on surface morphology than others. As one of many examples, Fig. 2 shows sets of fractures in the western part of Arabia, spaced several kilometres apart, which exhibit spectacular erosional control. No sizeable shear offsets have been observed on these fractures.

These observations can again be explained by considering the stress field around a tensile crack extending from the earth's free surface downward. The lateral extent over which such a crack in an elastic half space releases the tensile strain is roughly proportional to the depth of the crack. Figure 3 shows the computed strain field surrounding a stress-free crack (Head 1953) in a crust subject to a uniform tensile stress (e.g. due to uplift). Superposed on this stress is a linearly-increasing overburden pressure (due to the rock mass). In the computation it has been assumed that the crack is infinitely long and that its depth is in static equilibrium with the applied stresses. The results show that the presence of a tensile crack of depth z creates a shadow zone or a zone of strain release. The development of a second crack within this 'shadow zone' is inhibited over a distance of the order αz, where α is approximately 0.5. Thus, as a crack deepens with increasing tectonic tension, its lateral elastic 'shadow zone' increases proportionally, and pre-existing cracks cease to deepen significantly. Consequently, the spacing between induced tensile cracks of a given depth will be roughly of the order of their depth.

The above model can be used to infer the development of a set of fractures. Imagine a slow tectonic process of progressive tension leading initially to many shallow cracks (Fig. 4). As tension increases, some cracks prop-

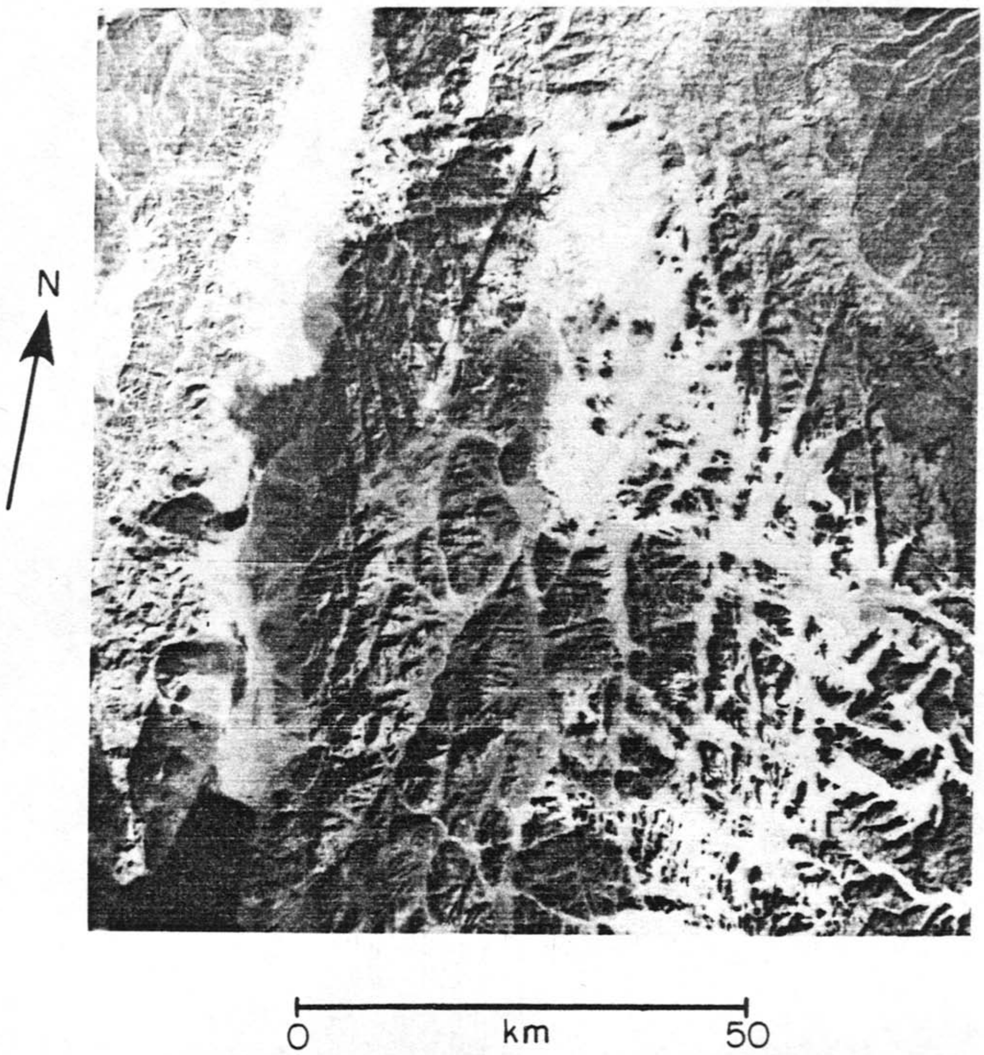

Fig. 2. A spectacular example of lineament-controlled erosion on the eastern coast of the Gulf of Agaba. (Landsat image, path 187, row 040, unknown image number or date.)

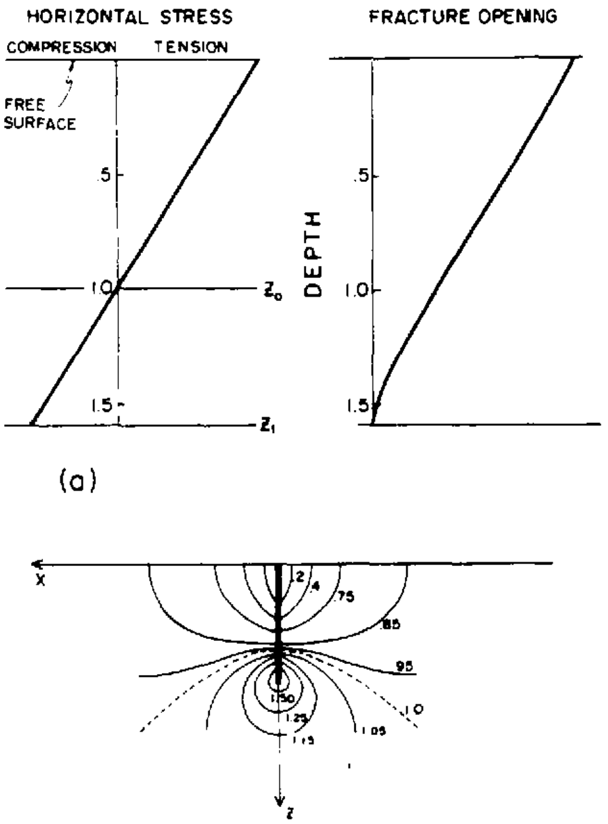

Fig. 3. The elastic field of a tensile crack in the crust. (a) The distribution of tensile stress vs depth of the crack exceeds by about 50% the depth z_c at which the tensile stress vanishes. (b) Strain field surrounding a stress-free crack defines a shadow zone created by a tensile fracture. Numbers indicate the residual tension. The shadow zone is defined here as the region over which tension is reduced by 50%.

agate to greater depth, creating their corresponding tensile shadow zones. Neighbouring fractures thus cease to propagate. As this process goes on, fewer and fewer fractures penetrate to increasingly greater depth, leading to a fracture density/depth distribution given roughly by

$$n(z) \sim 1/z \tag{6}$$

where z is the depth and $n(z)$ is the number of fractures extending to that depth.

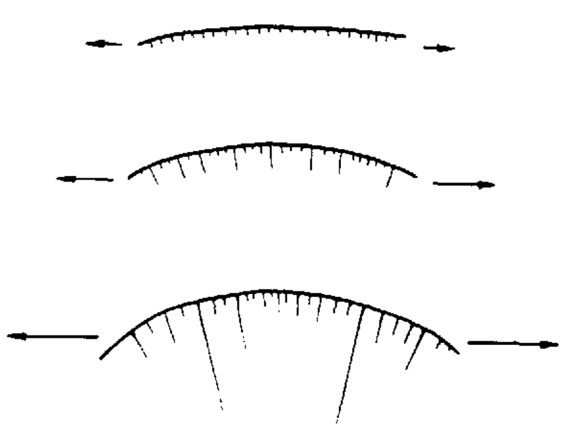

Fig. 4. Schematic development of fewer and deeper tensile fractures with increasing tension, taken here as being due to upheaval.

Unfortunately, no definite data exist at present to test this frequency–depth distribution of vertical fractures *in situ*. Available data (e.g. Snow 1968) are highly scattered and too shallow to test the above prediction. Perhaps the distribution could be determined in the future either directly from observations in deep boreholes, or indirectly through seismic-velocity or electrical-resistivity tests.

The opening and subsequent closure of tensile cracks may involve crumbling of the fracture surfaces, leading to 'shatter zones' (Spencer 1977, Phillips 1972). As the fracture is partly filled with rock fragments and displaced blocks, closure may be incomplete. This partial healing may lead to a narrow zone up to about 30 m wide which is relatively porous, permeable to fluids, mechanically weak, and liable to erosion. As a result, more pronounced erosional effects may be associated with the deeper and more brecciated wider-spaced fractures.

The width of an open fracture Δu at the earth's surface is related to its depth z_0. It can be shown that

$$\Delta u \simeq \sigma \cdot z_0/E \tag{7}$$

where σ, the tensile stress, is related to depth through equation (1) $\sigma = \rho g z_0$. Hence

$$\Delta u = z_0^2 \frac{\rho g}{E}. \tag{8}$$

Taking, as before, $E = 10^6$ bars and $z_0 = 3, 5$ and 10 km, we obtain $\Delta u = 3, 8$ and 30 m, respectively. An idealized fracture to a depth of 10 km has an opening of a few tens of metres at most. This value will be further reduced to about half when the opening is distributed among all fractures with differing depth. These values of Δu are in reasonable agreement with widths of observed breccia *in situ*.

DEPTH AND LENGTH DISTRIBUTIONS

Although the actual depth distribution of macrofractures is unknown, reasonable data are available on their length distribution, as measured at the earth's surface. Although no deterministic relation of depth to length is known, we may consider a statistical relation between the two.

Figure 5 shows a typical lineament length/frequency data set (from Kowalik & Gold 1976). The frequency $n(l)$ of fractures of length l over the range of measured length is described remarkably well by the (simple) truncated exponential relation:

$$n(l) = n_0 e^{-kl} \quad \text{for} \quad l < l_0. \tag{9}$$

n_0 and k are constants within a given area, and l_0 is the shortest sampled length. To explain this simple distribution it is necessary to find what controls the length of a fracture.

It can be shown that the stress required to open a crack in elastic solids is controlled by the short dimension of the crack (e.g. Eshelby 1957). Thus, the influence of fractures in reducing tensile stress is optimal when their length is equal to or greater than their depth. Consequently, it is very unlikely that a crack will have a length l smaller than its depth z. Though l may be greater than z by any amount, very long fractures $l \gg z$ will be less probable statistically than shorter ones.

Imagine that tensile fractures are initiated randomly at the crust's surface. As straining increases, a decreasing number of these fractures grow deeper and wider while most others cease growing. The length of each of these elementary fractures is roughly equal to its depth. In order to create a fracture which is substantially longer, several fractures must coalesce.

Assuming that cracks with length $l = z$ are all roughly parallel and have random locations, then a truncated Poisson distribution (Bracewell 1965) represents the probability of a composite fracture having the length $l > z$.

$$p(l; z) = A \cdot e^{-(l-z)} \qquad l > z \tag{10}$$

where A is a constant. For $l < z$ we assume $p(l; z) = 0$. Integrating over the distribution of crack depths as given in (6) yields the frequency distribution of fracture lengths

$$n(l) = A^* \int_{z_{min}}^{\infty} \frac{1}{z} p(l; z)\,dz \quad = A \int_{z_{min}}^{\infty} \frac{1}{z} e^{-(l-z)}\,dz \tag{11}$$

where z_{min} is the smallest depth considered and A^*, A are constants. Integration of (11) yields

$$n(l) = C_0 \cdot e^{-l} \qquad l \geq z_{min} \tag{12}$$

where C_0 is a constant. This distribution is identical with the one observed in (9).

The statistical consideration above does not prove that the observed exponential distribution of fracture lengths (equation 6) results from the suggested depth distribution (equation 6). They do imply, however, that the observed distribution of lineament lengths is compatible with the theoretical depth distribution.

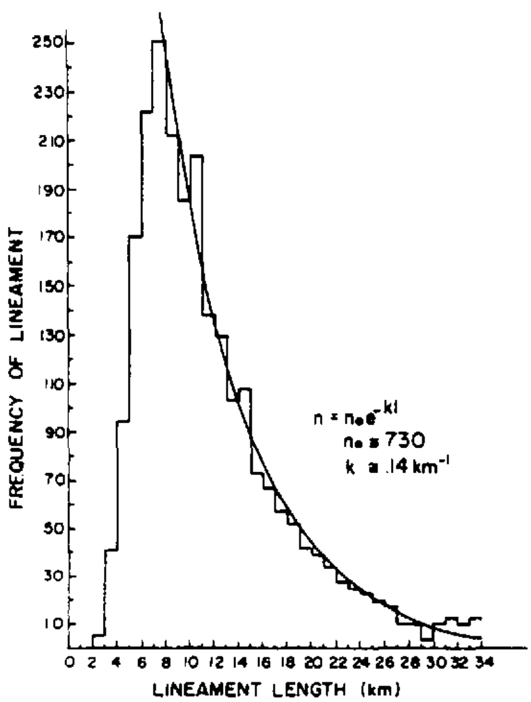

Fig. 5. The exponential dependence of the frequency of lineaments on their length. Modified from Kowalik & Gold (1976).

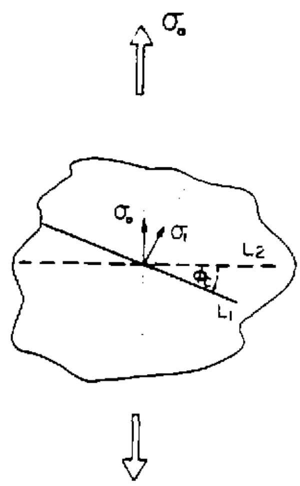

Fig. 6. The geometrical relation between a pre-existing fracture (L_1) and a newly created one (L_2). L_1 is oblique to the applied tensile stress σ_0 and L is normal to σ_0. L will develop when ϕ_c is sufficiently large such that it is easier to form a new fracture than to overcome the lower tensile strength of a pre-existing fracture in an unfavorable orientation. See text for details.

AZIMUTHAL DISTRIBUTION OF TENSILE LINEAMENTS

One of the most common properties of fractures and lineaments is their finite number of sets *in situ*; typically there are 2–4 sets with angles of 45–90° between sets. The literature shows the most common case is that of two, mutually perpendicular sets. There are many cases of three sets with angles of 60–90° and a few cases of four sets with angles ranging from a few degrees to 45° (e.g. Hobbs 1911, Hodgson 1961, 1976, Muehlberger 1961, Gay 1973, Hancock & Kadhi 1978). Why are the directions and number of sets so few?

Suppose that, in a portion of the crust which already contains one pre-existing fracture set (L_1), horizontal tension σ_0 is induced at an angle 90–ϕ° with the direction of tension, as shown schematically in Fig. 6. As tension increases sufficiently, either a new crack set (L_2) will develop perpendicular to the direction of tension, or the pre-existing cracks, oblique to the tensile direction, may reopen. The actual outcome depends on the ratio of tensile strength of the pre-existing fracture to the strength of the unfractured rock.

The tensile stress σ_0 required to create a new fracture perpendicular to the direction of tension is equal to the tensile strength S_r of the unfractured rock (r)

$$S_r = \sigma_0. \tag{13}$$

On the other hand, the stress σ_i required to reopen a pre-existing fracture (f) is equal to the tensile strength of the fracture S_f, which can be anywhere from zero to the rock strength, $0 \leq S_f \leq S_r$. Such fracture strength may result from cementation or other fracture healing processes. The tensile stress component σ_i acting normal to, and required to open a pre-existing fracture is

$$\sigma_i = \sigma_0 \cos^2 \phi = S_f. \tag{14}$$

The critical angle ϕ_c at which the stress necessary to open a new fracture is equal to the stress required to reopen an existing crack is then given by

$$\phi_c = \cos^{-1}(R)^{1/2} \tag{15}$$

where $R = S_f/S_R$. When $\phi < \phi_c$, the pre-existing fracture will open and no new fracture will develop. When $\phi > \phi_c$ a new fracture will open perpendicular to the direction of the tensile stress.

In the course of geological time, each episode of tension in various directions may cause either the opening of a new fracture or the reopening of a pre-existing one, depending on the strength ratio R and the angles between the directions of applied stress and the pre-existing fracture set. For $R > 0$, there is a finite angle ϕ_c, measured from the direction of the pre-existing set, in which no new fractures can be induced. This azimuth discretization process leads to a final state of few sets, aligned in a discrete number of directions. When both rock and fracture strength $R = S_f/S_R$ are known, the angular spacing between sets is simply equal to ϕ_c° as given by (15). Thus, the number of sets n for a uniform S_f is $n = 180°/\phi_c^\circ$, as shown in Figs. 7(a) and (b). For fracture strength $S_f = 0$ there are only two sets at right angles to each other, which is the most frequent case *in situ*. For $S_f = S_R/4$, $\phi_c^\circ = 60$ and $n = 3$; for $S_f = S_R/2$, $\phi_c = 45°$ and $n = 4$. If the fractures acquire strength with time, different sets may have different strengths, depending on the relative time of their formation. In this case the angular spacing between tensile sets could be uneven, as is sometimes observed (Gay 1973).

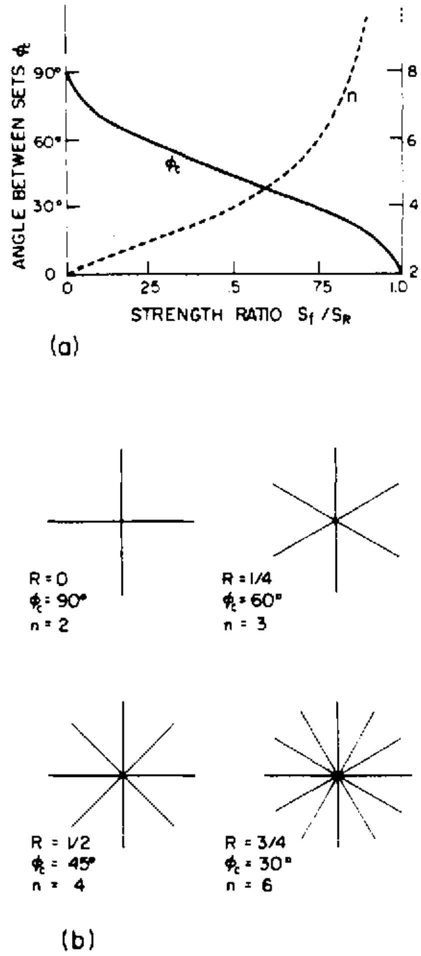

Fig. 7. The azimuthal distribution of fracture sets as a function of the ratio between pre-existing fracture strength (S_f) and rock strength (S_r). (a) ϕ_c is the angle between sets as a function of $R = S_f/S_r$ and $n = 180°/\phi_c$ is the number of sets. (b) Synthetic rose diagrams for some simple strength ratios.

THE SOURCE OF TECTONIC TENSILE STRAIN

There are about six well-understood sources for tensile strain in the earth. These will be evaluated on the basis of their relevance to the formation of tensile fractures.

Planetary strains

Typical planetary strains (compressive and tensile) such as solid earth tides, ocean tides and rainfall, range from 10^{-8} to 10^{-7} for solid earth and ocean tides, respectively. Rainfall flooding up to several metres causing mostly compression may cause local tensile strain as large as -5×10^{-7}. According to equation (5) and neglecting S_t/E, this tension will overcome the lithostatic pressure at negligible depths ranging from 3 cm to 3 m. These strains are therefore much too small to play an important role in deep tensile fracturing in the crust.

Strains associated with faulting

Typically, stress drop during earthquakes is from 1 to 100 bars. Assuming pure shear stress, the tensile component would also be 1–100 bars with a corresponding tensile strain of 10^{-6} to 10^{-4} (for E = 10^6 bars). This tension is overcome by the lithostatic pressure at maximum depths of a few hundred metres. Although near-surface tensile fracturing may result from this shallow tensile regime, this strain cannot create deep and widespread sets of fractures.

Glacial and fluvial loading

Localized crustal loading due to glaciers and large lakes may persist over several thousands of years. The stress difference associated with the Fennoscandian glacial loading at a height of about 2–3 km is of the order of 200–300 bars (for ice density of $\rho \simeq 1\ \mathrm{g\,cm^{-3}}$). This type of loading may lead to tensile fracturing to a depth of 1 km or so. Since surface loads tend to cause compression almost anywhere (e.g. Love 1911, Bell & Nur 1978), their actual role in tensile fracturing cannot be very significant, even though the rebound from the removal of such loads may cause broad tension. Finally, the few localized loads of this kind can hardly explain the ubiquity of fractures on the surface of the crust.

Thermal contraction and sediment removal

Tensile fractures caused by the cooling of hot rock to surface temperature have been discovered recently in new oceanic crust with abnormal heat flow due to convection through thermally induced tensile fractures that penetrate into the ocean floor. There is less understanding, however, of the thermal history of the continental crust, and one cannot, therefore, determine whether ubiquitous fractures are caused by thermal contraction. Typical contraction fractures, such as columnar joints in basalt, show hexagonal patterns in which roughly equidimensional blocks develop, and fracture sets elongated parallel to a migrating cooling surface are observed. It is possible that when cooling occurs under non-hydrostatic stress conditions, hexagonal patterns may be replaced by a single preferred direction of jointing. Whether this actually happens in the crust remains to be investigated. For example, Spears (1961) found that joints in a sill, which had been formerly interpreted as columnar, are part of the regional tectonic system. Haxby & Turcotte (1976) believed that erosion, uplift, and cooling may cause net horizontal tension in the crust. When previously buried basement is exposed, it may be subject to sizeable tensile stresses which might produce tensile jointing (e.g. Price 1966, 1974). It is likely, however, that exhumed basement has been well fractured before it subsided in the first place. Further work is needed to estimate the magnitude of these stresses more accurately.

Large-scale tectonic movements

Over long periods of time the earth's surface undergoes large horizontal and vertical displacements. The former can create large tensile stresses; Turcotte (1974) has considered the membrane stresses which develop in plates as they move over the ellipsoidal surface of the earth. In particular, large tensile stresses may develop as plates bend, moving from regions near the pole, where the radius of the earth's curvature is smaller. The tensile stresses may reach several kilobars, counteract the overburden compression, and cause fracturing to a depth of several kilometres, according to (3). Whether such large-scale deformation can cause the observed small-scale spacing between lineaments is unclear.

Localized bending

Typically, differential vertical displacements to a few kilometres occur over distances of tens to hundreds of kilometres (e.g. Isachsen 1975). Such deformations are commonly large, gentle anticlines or anticlinoriums. These features have various origins: in some situations, bending of the upper crust results from large-scale intrusions in the lithosphere; in other situations, uplift results from an abnormally hot asthenosphere (e.g. Crough & Thompson 1977). Whatever the cause, crustal bending that extends over a few hundred kilometres in scale may lead to large tensile strains, as shown in the next section.

CRUSTAL AND LITHOSPHERIC BENDING

The tensile strains and stresses associated with bending deformation can be estimated by means of a two-dimensional bending theory (e.g. Timoshenko & Woinowsky-Kreiger 1959, pp. 4–6). Consider the bending of a plate subject to uniform load, as shown in Fig. 8. The deflected portion of the plate can be considered, in a first approximation, as a section of a cylindrical shell. If $2z_0$ is the thickness of the plate, xy is the neutral middle plane, and y lies along the cylindrical axis, then the curvature is approximately

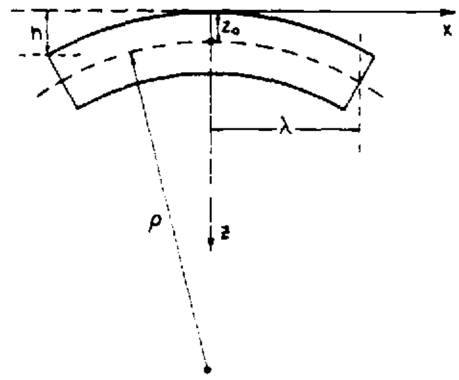

Fig. 8. Cross-sectional geometry of an elongated bend of a crustal or lithospheric segment. The coordinate y is perpendicular to the plane of the page, xz.

$$\rho \simeq [-\mathrm{d}^2\omega/\mathrm{d}x^2] \quad (16)$$

where $\omega(x)$ is the deflection of the vertical z direction measured from the free surface. The lateral strain ε_x is

$$\varepsilon_x = -(z_0 - z)\frac{\mathrm{d}^2\omega}{\mathrm{d}x^2} = -(z_0 - z)\rho^{-1} \quad (17)$$

which yields tensile strain for $z < z_0$ and compressive strain for $z > z_0$ where z_0 is the depth to the neutral plane.

Assuming linear elastic behavior, the lateral stresses are

$$\sigma_x(z) = \frac{\mathrm{E}}{1-\nu^2}\varepsilon_x = -\frac{\mathrm{E}}{1-\nu^2}\cdot(z_0 - z)\rho^{-1} \quad (18)$$

and

$$\sigma_y = \nu\sigma_x < \sigma_x$$

where E is Young's modulus and ν is Poisson's ratio.

Denoting λ as the half width and h as the maximum height of the uplift the radius of curvature is given by

$$\rho = [\lambda^2 + (\rho - h)^2]^{1/2}. \quad (19)$$

For small h this becomes

$$\rho \simeq \frac{\lambda^2}{2h} \quad (20)$$

and the stress

Table 1. Values of radius of curvature $\rho = \lambda^2/2h$ for reasonable ranges of geological uplifts h and half wave length λ

	Half width λ (km)		
Height h (km)	50	100	150
0.5	2.5×10^3	10^4	2.25×10^4
1.0	1.25×10^3	5×10^3	1.2×10^4
1.5	0.8×10^3	3×10^3	7×10^3

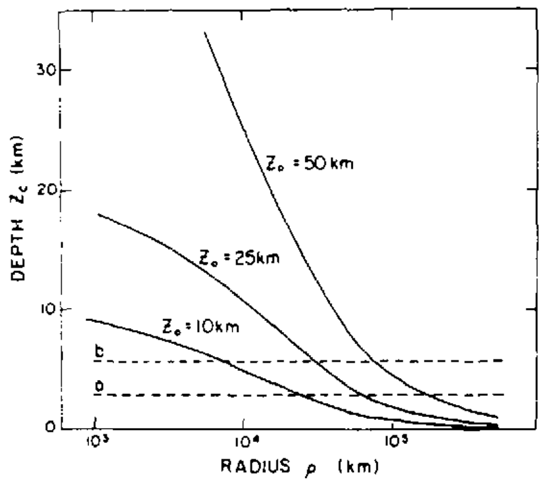

Fig. 9. Approximate penetration depth of tensile fracture (z_c) in a bending layer vs the ratio of uplift to half-width of the uplift. Tensile fracture may penetrate to 10 km or more. Z_c is the depth at which the applied tension is equal to the ambient lithostatic pressure; z_0 is the half thickness of the layer; ρ is the radius of curvature of the bend in the layer; λ is the half width; and h is the uplift of the band. Line (a) represents the estimated maximal depth according to Brace (1964); and line (b) is based on the laboratory strength of intact granite. High pore pressure can increase the depth of tensile fractures substantially over (a) or (b).

$$\sigma_x = -\frac{E}{1-v^2}(z_0 - z)\frac{2h}{\lambda^2}. \quad (21)$$

The tensile stress of (21) is superposed on the overburden pressure of (3). Equating the two stresses and using (3), we obtain

$$\rho g z_c = \frac{E}{1-v^2}(z_0 - z)\frac{2h}{\lambda^2} \quad (22)$$

where z_c is the depth at which the tectonic tension is exactly cancelled by the overburden pressure. Solving for z_c yields

$$\frac{z_c}{z_0} = \frac{1}{1 + \left[\frac{\rho g(1-v^2)}{E} \times \frac{\lambda^2}{2h}\right]}. \quad (23)$$

If $\rho g = 300$ bars km^{-1}, and $2E/(1 - v^2) \simeq 3 \times 10^6$ bar

$$\frac{z_c}{z_0} \simeq \frac{1}{1 + 10^{-4}\rho/\rho_0} \quad \rho = \frac{\lambda^2}{2h} \quad \text{and} \quad \rho_0 = 1\text{ km}. \quad (24)$$

Table 1 indicates the values of h and λ for several geometries of uplifts. Typically, h is of the order of hundreds of metres, and λ is tens to a few hundred kilometres. Typical values of the radius of curvature $\rho = \lambda^2/2h$ are therefore 10^3–10^4 km.

The depth z_0 to the neutral plane is more uncertain. The half-depth z_0 of the brittle part of the crust is about 10 km, for the entire crust $z_0 \simeq 25$ km, and for the lithosphere as a whole $z_0 \simeq 50$ km. Figure 9 gives the computed depth z_c to which tension would penetrate as a function of $\rho = \lambda^2/2h$, for given z_0. For the reasonable values, $h = 0.5$–1 km and $\lambda = 50$–100 km, the tensile stress penetrates to a depth comparable with the brittle part of the crust in which most earthquakes occur. These values suggest that extension fractures can develop to substantial depth in the crust due to local tectonic uplifts.

TENSION, PORE PRESSURE AND DEPTH OF FRACTURES

Whether tensile fractures do actually penetrate to depths of 5–10 km is uncertain. Classical laboratory determinations of failure condition of brittle basement rocks (e.g. Brace 1964) suggest that the maximum depth of tensile cracking is about 2.5–3 km. At greater depth, failure should be of the normal faulting type. In fact, Griggs & Handin (1960) believe that tensile fractures are unlikely to be deeper than just a few hundred metres. Secor (1965, 1969) and Price (in Fyfe *et al.* 1978) have shown, however, that high, super-hydrostatic pore pressure may actually counteract the overburden effect and permit deep tensile fracture. Secor's fractures do not originate at the free surface but rather at depth, from which they may propagate towards the free surface (Secor & Pollard 1975). The actual depth of tensile fracturing here depends critically on the ratio of pore pressure to overburden pressure. The origin of the high pore pressure remains enigmatic, however. In fact, Brace (1979) suggests that pore pressure in the crust must normally be equal to or less than hydrostatic pressure because of the high crustal hydraulic permeability which he infers from a wide range of observations. Thus, because of the lack of more definitive data, the application of simple, static failure criteria with pore pressure do not provide a good estimate of maximum depth of tensile fractures.

Some sort of depth estimate may be provided by the topography of rugged basement terrains; for example, deep canyons or steep cliffs such as in Yosemite Park, California, or on the Norwegian Coast tower to 1–2 km above base level. Consequently, tensile fracture may penetrate at least to these depths.

The problem is further complicated by the fact that the static failure envelopes, referred to above, cannot deal with the process of fracture growth. A tensile fracture tip, propagating downwards from the earth's surface does not 'know' when it enters a regime of normal faulting. Therefore, the Mohr's type failure criterion, which is a static one and at a point, may not be valid for rupture propagation processes.

Until some of these dilemmas are resolved, it remains unclear to what depth tensile fractures actually penetrate. They certainly can penetrate to 2–3 km and perhaps to significantly greater depth.

CONCLUSIONS

Simple considerations of rock failure in tension suggest that many lineament and macrofracture sets may be the result of transient tensile episodes in the earth's crust. In particular, the proposed tension model yields the following main features of fracture-controlled lineaments and joints: (1) the orientations of lineament sets, the small number of sets, and the horizontal angles between sets; (2) the spacing of fractures; (3) the Poisson distribution of fracture lengths and (4) the lack of significant shear offsets and the narrowness of shatter zones.

The fracture density is proportional to the intensity of deformation; a greater ratio of uplift to width of uplifted zone leads to a greater density of fractures in a given set.

The small number of sets is controlled by the ratio of the tensile strength of pre-existing fractures to that of unfractured rock. For reasonable ratios, the angles between sets are 45–90°. These angles imply the development of 2–4 sets. The model of the tensile process predicts that the fracture density $n(z)$ of a given set decreases with depth z as $n(z) \simeq 1/z$. Fractures as deep as 10 km may involve a 10–30 m opening at the free surface, well within reasonable range for observed 'shatter zones'. Finally, the exponential frequency-length distribution of lineaments is consistent with the tensile model and is statistically related to the frequency distribution of the depths of fractures.

The most likely origin of the transient tensile stress is tectonic updoming, of the order of hundreds to thousands of metres of crustal portions, over distances of tens to hundreds of kilometres. This process may also be strongly linked with erosional unloading as suggested by Haxby & Turcotte (1976). Less certain sources are thermal contraction and glacial rebound.

Several major questions related to the proposed mechanisms remain. For example, there is no direct proof that tensile opening of lineaments always occurs, as in the form of 'shatter zones' (Spencer 1977). Careful field work would provide a test for this feature. Transient geological uplifts must be tested in detail to establish the relevance of this process. Other sources for tensile episodes, in particular, thermal origins and global plate motions, are yet to be studied. Finally, there is at present no direct proof for the depth distribution of fractures. Systematic geophysical and borehole investigations may eventually determine the actual depth distribution and thus confirm or disprove the simple model proposed here. However, using reasonable values for the elasticity, tensile strength, and crust density, it is likely that fractures penetrate at least to 2–3 km and may penetrate deeper, possibly providing deep conduits for fluids (Fyfe *et al.* 1978). These fluids may have a profound effect on the mechanical behaviour of the crust, its effective strength at depth, the occurrence of earthquakes, and the potential for geothermal energy and its efficient extraction. The existence of deep fractures containing water is consistent also with the conclusion by Bell & Nur (1978) that deep seismicity induced by reservoir impounding is due to the presence of pore water at great depth. This conclusion may be further supported by the low electrical resistivity which is commonly observed down to mid-crustal depths.

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