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Loading paths to joint propagation during a tectonic cycle: an example from the Appalachian Plateau, U.S.A.

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Abstract—Based on the timing of joint propagation during the history of burial, lithification, deformation and denudation of clastic rocks within sedimentary basins, four types of joints may be distinguished: tectonic, hydraulic, unloading and release. Tectonic and hydraulic joints form at depth prior to uplift in response to abnormal fluid pressures, whereas unloading and release joints form near the surface in response to thermal-elastic contraction accompanying erosion and uplift. Tectonic joints are distinguished from hydraulic joints in that tectonic compaction is a mechanism for achieving abnormal pore pressures leading to the propagation of the former whereas compaction by overburden loading leads to the abnormal pore pressures in the latter case. The orientation of unloading joints is controlled by either a residual or contemporary tectonic stress whereas the orientation of release joints is controlled by a rock fabric. Examples of some of these joints are found within the Devonian Catskill Delta of the Appalachian Plateau, New York. During the Alleghanian Orogeny tectonic joints (cross-fold joints) formed under abnormal pore pressure as indicated by the observation that joints propagated in the siltstones before they developed in shales and by the cross-cutting relationships of folds, cleavage and joints. This sequence is compatible with oil company hydraulic fracture data which show that the least principal stress within sandstone layers is less than that in the intercalated shale layers. Plumose structures indicate that the joints within siltstones propagated as discontinuous rupture events each of which affected less than a meter of bed length. The discontinuous rupturing is compatible with models for natural hydraulic fracturing. Release joints (strike joints) post-date the Alleghanian Orogeny as indicated by abutting relationships within the deeper parts of the Devonian clastic section. Unloading joints are orthogonal to the contemporary tectonic stress field.

INTRODUCTION

FROM field observations, Price (1966, p. 110) concluded, "it is unlikely . . . that all joints are the result of a single mechanism". Nickelsen (1976, p. 193) commented, "Fracture patterns are cumulative and persistent. Cumulative implies several episodes of fracturing . . . . Persistent means not easily erased by later tectonic events". These statements say that in sedimentary rocks joints may propagate at several different times during a tectonic cycle which includes burial, diagenesis, tectonic compression, uplift, and erosion. Joint propagation occurs when failure criteria are met; failure criteria are often specified in terms of states of stress which are calculated by considering, as a function of depth of burial, the variation of such factors as rock properties, stress history, and pore-fluid pressures. Loading path models which calculate state of stress by tracing elastic properties, temperature, tectonic stress and pore pressure during burial and erosion of a sedimentary basin confirm that the state of stress causing joint propagation occurs under several different conditions (Price 1974, Voight & St. Pierre 1974, Narr & Currie 1982).

Among the loading paths to joint propagation in sedimentary basins, four end members stand out as reasonably distinct. A fifth end member, systematic jointing in unconsolidated sediments and coal, is excluded from this discussion (Gilbert 1882, Nickelsen & Hough 1967). Although they may be applicable, the loading paths discussed here were not designed to handle the specific cases of jointing in igneous intrusions, metamorphosed and penetratively deformed mountain belts, and old basement rocks (Wise 1964). The purpose of this paper is to describe four loading paths leading to joint propagation in sedimentary basins and then to show that the propagation of joints within the Devonian Catskill Delta, New York, occurred at the end of at least three of these loading paths. This latter task is accomplished by the presentation of the facts and assumptions used to infer the conditions causing the propagation of various joints within the Devonian stratigraphic section. Of particular interest is the evidence for joints propagating as natural hydraulic fractures under the influence of abnormal pore pressure.

GEOLOGICAL BACKGROUND

The tectonic cycle affecting the Catskill Delta in western New York State consists of three stages: (1) deposition of a clastic delta during the Late Devonian; (2) tectonic compression during the Carboniferous and Permian and (3) uplift from the Mesozoic to present. Deposition started with shales of the Hamilton Group in a shallow marine basin (Fig. 1). The shales of the Hamilton and Genesee Groups are interrupted by a few thin limestones but otherwise consist of continuous sections more than 50 m thick. As the Catskill Delta prograded from east to west in the basin, the average sediment size increased. Hence, upsection, the Catskill Delta changes from largely shale to interfingered shales and siltstones to a cap of fluvial sandstone beds. In west-central New York near Watkins Glen, interfingered siltstones first appear in the upper Genesee Group. At the level of the

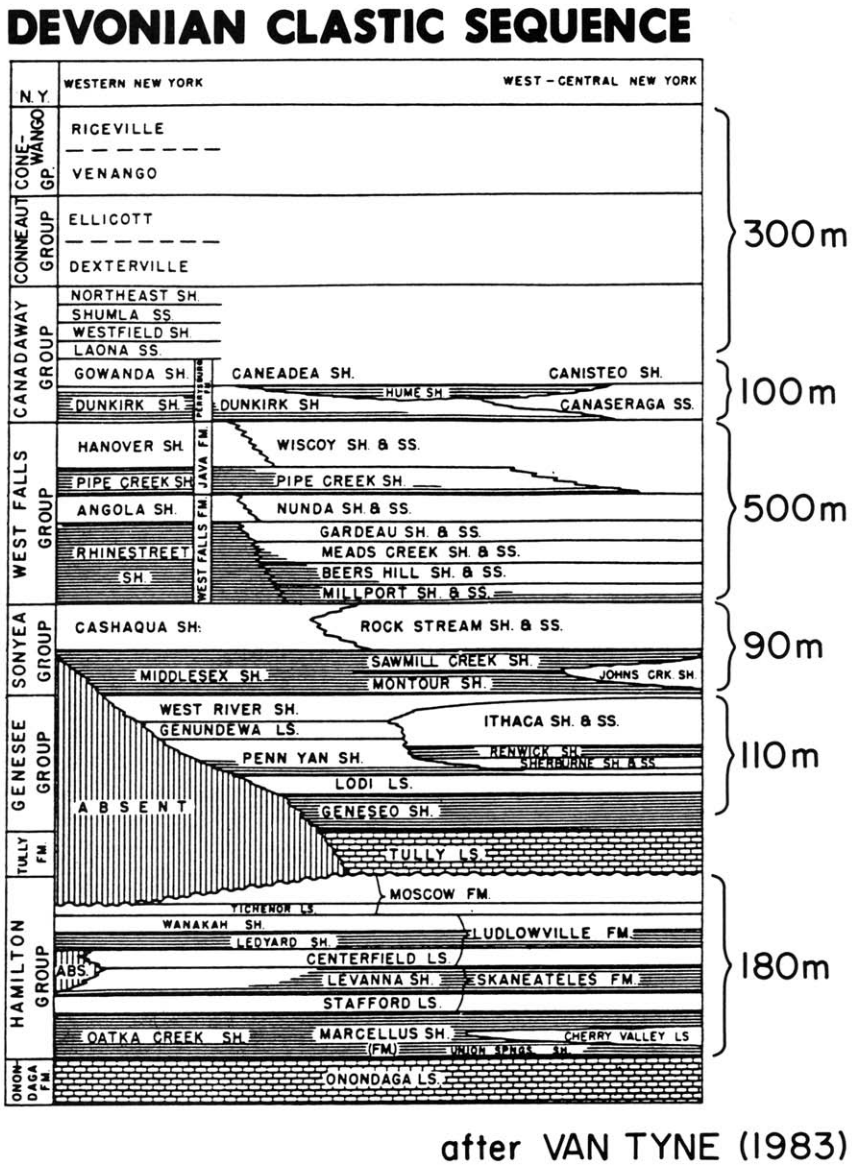

Fig. 1. Stratigraphic succession for the western portion of the Devonian Catskill Delta. Thicknesses are indicated for the delta just east of the 1 km isopach in western New York.

West Falls Group near Watkins Glen, siltstones predominate but the facies changes to black shale further west. Single beds of sandstone several meters thick occur in the Canadaway Group and higher. In the study area the Catskill Delta thins from 3 km in the east to 1 km in the west (Colton 1970, Rickard & Fisher 1970) but may have been as much as 7 km thick further to the east (Friedman & Sanders 1982) (Fig. 2).

During the Alleghanian Orogeny of Carboniferous to Permian age two phases of tectonic compression affected the Catskill Delta (Geiser & Engelder 1983). Layer-parallel shortening occurred over a broad area of the Appalachian Plateau by means of blind thrusting along detachments within Silurian salt beds. During the first phase of compression, the Lackawanna, a cleavage developed to the south in Pennsylvania but little more than a joint set formed within the study area (Fig. 3). The second or Main Phase is manifest by regional development of several mesoscopic structures including joints, a solution cleavage, a pencil cleavage, and minor folds.

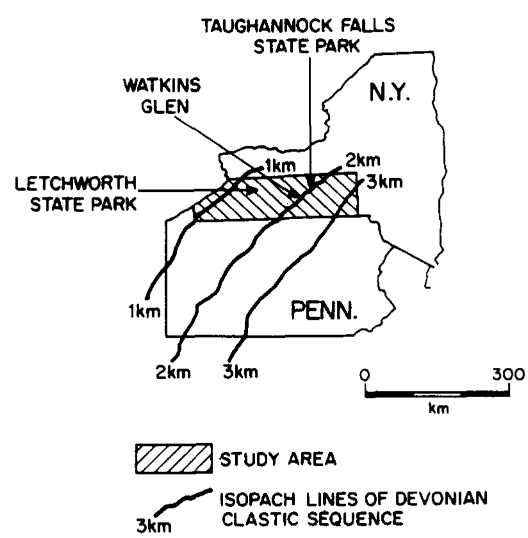

Fig. 2. Location map for the study area in western New York State

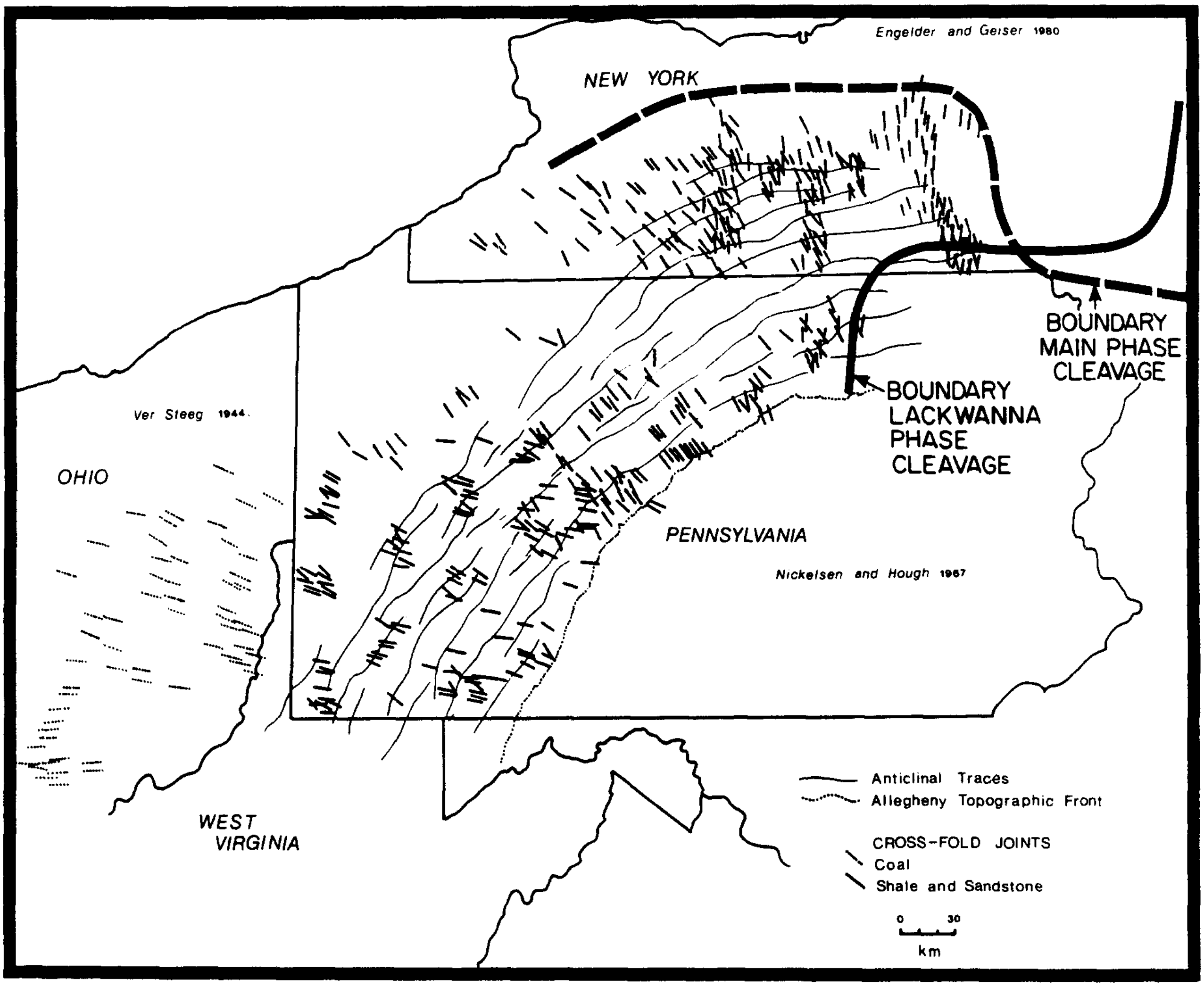

Fig. 3. Tectonic map of the Appalachian Plateau within the northeastern United States. Cleavage imprinted during the Lackawanna Phase of the Alleghanian Orogeny affected the area south of the solid line whereas cleavage imprinted during the Main Phase of the Alleghanian Orogeny affected the area south of the dashed line. The cross-fold joints are either tectonic and of Alleghanian age or later unloading joints controlled by an Alleghanian residual stress. Cross-fold joints were mapped by Ver Steeg (1942, 1944, Ohio), Nickelsen & Hough (1967, Pennsylvania) and Engelder & Geiser (1980, New York).

From Mesozoic time to the present the Catskill Delta was subject to uplift with as much as 1 km of sediment being removed by erosion. Evidence for the amount of erosion is based on a conodont color alteration index (Epstein *et al.* 1975) and xenoliths in Mesozoic ultramafic dikes (Van Tyne 1958).

Joint sets that propagated during this tectonic cycle are those identified in Parker (1942) and discussed by Engelder & Geiser (1980). These include more than one cross-fold joint set (set I; see Table 1) and a parallel to fold or strike joint set (set II). In addition, a third set (set III) is geometrically unrelated to the folds of the Alleghanian Orogeny but is orthogonal to the contemporary tectonic stress field (Engelder 1982a). These joint sets formed at different times during the tectonic cycle of the Appalachian Plateau and, hence, formed at the ends of different loading paths.

In this paper the word joint refers exclusively to an extension fracture. Evidence that all these joints formed as extension fractures (mode I cracks) includes: (1) no shear offset of fossil markers; (2) bilateral symmetry of surface morphology; (3) butting relationships and (4) low deviatoric stress during propagation (Engelder 1982b).

DEFINITION OF JOINT TYPES BASED ON LOADING PATHS

The concept that joints propagate at the ends of several loading paths largely stems from the fact that many exposures contain joints in several orientations. At' the very least, stress orientations must change between jointing events. An understanding of the development of these stress conditions comes by considering loading paths which are plots of the horizontal principal stress vs depth of burial (Fig. 4). Because the joints considered in this paper are vertical, a horizontal least principal stress is assumed to be normal to the joint which will propagate when failure conditions are met. The calculation of horizontal stresses assumes homogeneity and a laterally confined half space (i.e. $d\epsilon_x = d\epsilon_y = 0$). Assuming the rock behaves as an isotropic elastic medium, the horizontal stresses are a function of several

Table 1. Terminology for joints of the Appalachian Plateau

Systematic name	Geometric relation to folds	Timing of joint propagation	Type of fracture or crack	Location within study area	Location within stratigraphic column	Relation to other structures	References
Set I	Cross-fold	Alleghanian Orogeny or post-orogenic uplift	Tensile (Mode I)	Regional	Found throughout	Consists of as many as three joint sets at some outcrops	Parker (1942) Bahat & Engelder (1984)
Set I (Veins)	Cross-fold	Alleghanian Orogeny	Tensile (Mode I)	Eastern portion	West Falls (in cores) Genesee, Hamilton Tully Limestone	Calcite filling	Engelder & Geiser (1980) Engelder (1982b)
Set Ib	Cross-fold	Lackawanna Phase	Tensile (Mode I)	Regional	Found throughout	Cut by Main Phase cleavage	Engelder & Geiser (1980)
Set Ib	Cross-fold	Post-Main Phase uplift	Tensile (Mode I)	Western portion	Upper portion	Not orthogonal to Main Phase strain	Engelder & Geiser (1980)
Set Ia	Cross-fold	Main Phase	Tensile (Mode I)	Eastern portion	Lower portion	Contemporaneous with and orthogonal to Main Phase cleavage	Engelder & Geiser (1980)
Set II	Strike or fold-parallel	Post-Main Phase uplift	Tensile (Mode I)	Regional	Near surface	Subparallel to local cleavage	Parker (1942) Engelder & Geiser (1979)
Set III	Unrelated	Post-Main Phase uplift	Tensile (Mode I)	Regional	Near surface	Orthogonal to contemporary tectonic stress field	Parker (1942) Engelder (1982a)

variables including the Young's modulus (E), Poisson's ratio (ν), vertical stress (σ_z), temperature (T) and thermal expansivity of the rock (α). Following Voight & St. Pierre (1974) the horizontal stresses (σ_x and σ_y) may be calculated in terms of the vertical stress (σ_z)

$$\sigma_x = \sigma_y = (\nu/1 - \nu)\sigma_z + [\alpha E \Delta T/(1 - \nu)]. \quad (1)$$

The major components of the Catskill Delta include siltstones and shales which have different mechanical properties and, therefore, according to eqn. (1), should follow a different loading path to joint propagation. The greatest difference in behavior of the loading paths for shale and siltstone occurs if it is assumed that diagenesis and lithification do not take place until the maximum depth of burial, an assumption used in Voight & St. Pierre's (1974) treatment. Lesser stress differences are calculated using other approaches (Prats 1981). In this ideal case an uncemented sand aggregate and an undrained clay are taken to the maximum depth of burial (Fig. 4). Assuming a depth of burial of 1 km the effective stress, P_{σ_z}, is 15 MPa which is the total overburden stress (25 MPa km^{-1}) minus the pore fluid pressure (~10 MPa km^{-1}). For this illustration it is assumed that the temperature gradient is 25°C km^{-1}. Using the values for ν, α and E given in Table 2 for an uncemented sand and an undrained clay the horizontal effective stresses P_{σ_x} and P_{σ_y} are 4.7 MPa for the sand and 15 MPa for the clay at 1 km. Here a ν of 0.5 for the undrained clay assumes that it has no strength during burial. An uncemented sand aggregate has a relatively low E and so at depth the Poisson effect makes the major contribution to the horizontal stresses but σ_x is a fraction of the overburden σ_z (Voight & St. Pierre 1974). In contrast, the undrained clay has a Poisson's ratio of 0.5 which means that $\sigma_x = \sigma_z$ at depth (Fig. 4).

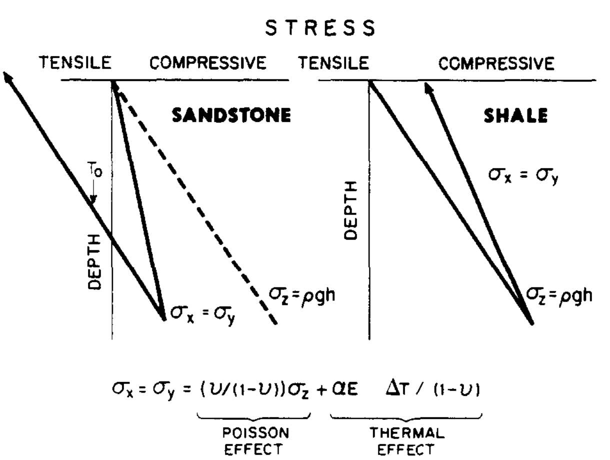

Fig. 4. Stress vs depth of burial for sandstone and shale. Lithification is assumed to occur at the maximum depth of burial with no change in the components of horizontal stress (σ_x and σ_y). Equation for calculating the horizontal stress consists of the Poisson effect and the thermal effect. This equation does not show the effect of a tectonic stress. Scale bars are left off because the illustration could apply to a variety of depths. For full details see the text.

At maximum depth of burial ν, α and E change upon cementation according to the information given in Table 2. During this lithification, as Fig. 4 shows, the horizontal state of stress does not change. This can be so only if Voight & St. Pierre's (1974) analysis is followed strictly.

Table 2. Possible mechanical properties of Catskill Delta sediments during burial and uplift

	E (GPa)	ν	α(10^{-6}°C^{-1})	Reference
Clay	small	0.5	—	Lambe & Whitman (1969)
Sand	01.0	0.21	10.0	Voight & St. Pierre (1974)
Shale*	04.9‡	0.36	10.0	Chong *et al.* (1980)
Sandstone†	16.5§	0.33	10.8	Wilhemi & Somerton (1967)

*Colorado Oil Shale.
†Average of Bandera, Berea, Boise Sandstones.
‡Uniaxial tests.
§Confined tests at 3.5 MPa.

Ordinarily, a change in elastic properties at depth should be reflected in a change in horizontal stresses through the Poisson effect. However, Voight & St. Pierre (1974) introduced a cement at zero stress and hypothesized a situation where the sand aggregate carried the entire load even after cementation. In this situation the horizontal stress is not redistributed even though there was a change in Poisson's ratio. The same effect is assumed to apply to the shale in Fig. 4.

On removal of the overburden by erosion the change in horizontal stresses may again be calculated by using equation (1). At the surface the stress change for the sandstone is 14 MPa so that during uplift the P_{σ_x} may become tensile by 9.3 MPa (4.7–14 MPa). Here the effect of cooling and decreasing vertical stress is subtracted from the horizontal stress at maximum burial. During erosion P_{σ_x} for the sandstone becomes tensile as it approaches about half the total depth of burial whereas P_{σ_x} for the shale may remain compressive throughout its unloading history (i.e. P_{σ_x} = 5.5 MPa at the surface). The tensile stress within the sandstone is large enough to induce joint propagation in the vertical plane at the depth where the tensile strength (T_o) is exceeded (Fig. 4). During accompanying erosion thermally induced tensile stresses are larger for the sandstone than for the shale mainly because of the larger E of the sandstone. This analysis suggests that both during burial and erosion there is a reduced stress and preferential jointing of the sandstones beds relative to shale.

Lithification occurs continuously during burial so that ν changes gradually with depth of burial, which is a more realistic model than sudden cementation at the maximum depth of burial as illustrated in Fig. 4. Other analyses of loading paths for the generation of joints include those of Magara (1981) and Narr & Currie (1982). To show horizontal stress variation with depth, Magara (1981) used eqn. (1) and accounted for continual cementation using Eaton's (1969) estimate of Poisson's ratio with depth in the Gulf of Mexico. Narr & Currie (1982) assumed a linear increase in E and ν with depth in the Uinta Basin, Utah. Both studies concluded that burial in the presence of a hydrostatic pore pressure does not cause jointing, whereas burial in an environment of restricted fluid circulation may lead to jointing. In addition, if aquathermal pressuring occurs (i.e. thermal expansion of water confined in a pore), a high geothermal gradient of 4.49°F 100 m^{-1} leads to an increase in fluid pressure of 4.1 MPa 100 m^{-1} compared with 1.02 MPa 100 m^{-1} for the hydrostatic pressure (Barker 1972). If fluid flow becomes restricted at a depth of 2.5 km, then burial to a depth of about 6 km will result in a situation conducive for vertical jointing as fluid pressures approach overburden pressures (Magara 1975). Jointing may also occur if an abnormally pressured reservoir leaks to an undrained higher stratigraphic level causing the effective pressure there to become tensile (Magara 1981). In general, abnormal fluid pressures occur at depths greater than 3 km so that joints associated with these pressures will not propagate at depths much less than 5 km. Without abnormal pressures an uplift from 6 km to about 3 km is required to generate the tensile stresses necessary for the onset of jointing (Narr & Currie 1982). Hence, these studies have distinguished two types of joints: those propagating while burial is in progress and those propagating during erosion and uplift. Abnormal pore pressures are required in the former case whereas thermal-elastic contaction is primarily responsible for the latter case.

Following Price (1966), Voight & St. Pierre (1974), Magara (1981) and Narr & Currie (1982), there are many loading paths leading to the failure of rock in tension with the concomitant propagation of joints. However, there are four loading paths which should be regarded as end members in the suite of all paths leading to tensile failure. The end members fall into two groups of two: hydraulic and tectonic joints which propagate during burial or at the maximum depth of burial; and unloading and release joints which propagate during uplift and erosion.

The four end-member loading paths are distinguished on three-axis diagrams where the axes are effective stress normal to the direction of the future joint plane, the pore pressure, and depth of burial (Fig. 5). The plane marked by the effective stress and the depth axes is Fig. 4 rotated counter-clockwise by 90°. Thus, compressive effective stress is plotted above the origin and the plane marked by the pore-pressure and depth axes, whereas the field of tensile-effective stress is below the origin. The envelope marking the tensile strength of rocks is the horizontal plane drawn below the origin. The pore-pressure vs depth plane is horizontal with the hydrostatic and lithostatic gradients plotted on the tensile-strength envelope. The loading path to failure in tension is shown as a solid line in three dimensions with the point of failure noted by an X. The loading path is projected on the three planes of the three-axis diagrams with dashed curves.

Hydraulic joints are those caused by abnormal pore pressure during burial (Fig. 5). The loading path includes burial under hydrostatic-pore pressures with subsequent development of abnormal-pore pressures under restricted pore-water circulation. Aquathermal pressuring may act to increase further the abnormal pressures under restricted pore-water circulation. Point A (hydraulic joints of Fig. 5) shows the depth of burial where the pore pressure rises above the hydrostatic gradient. This point is reflected by a knee in the effective-stress vs depth (point B) and effective-stress vs pore-pressure (point C) curves. The analyses of both Magara (1981) and Narr & Currie (1982) indicate that these joints form at depths in excess of 5 km.

Tectonic joints are distinguished from hydraulic joints in that they form at depth under the influence of high pore pressure which developed during tectonic compaction. The need to distinguish tectonic joints arises because abnormal pore pressures during tectonic deformation cause joints at depths of less than 3 km as will be discussed later in this paper. Pore pressure records from such basins as the Gulf of Mexico indicate that these depths are insufficient for the development of abnormal

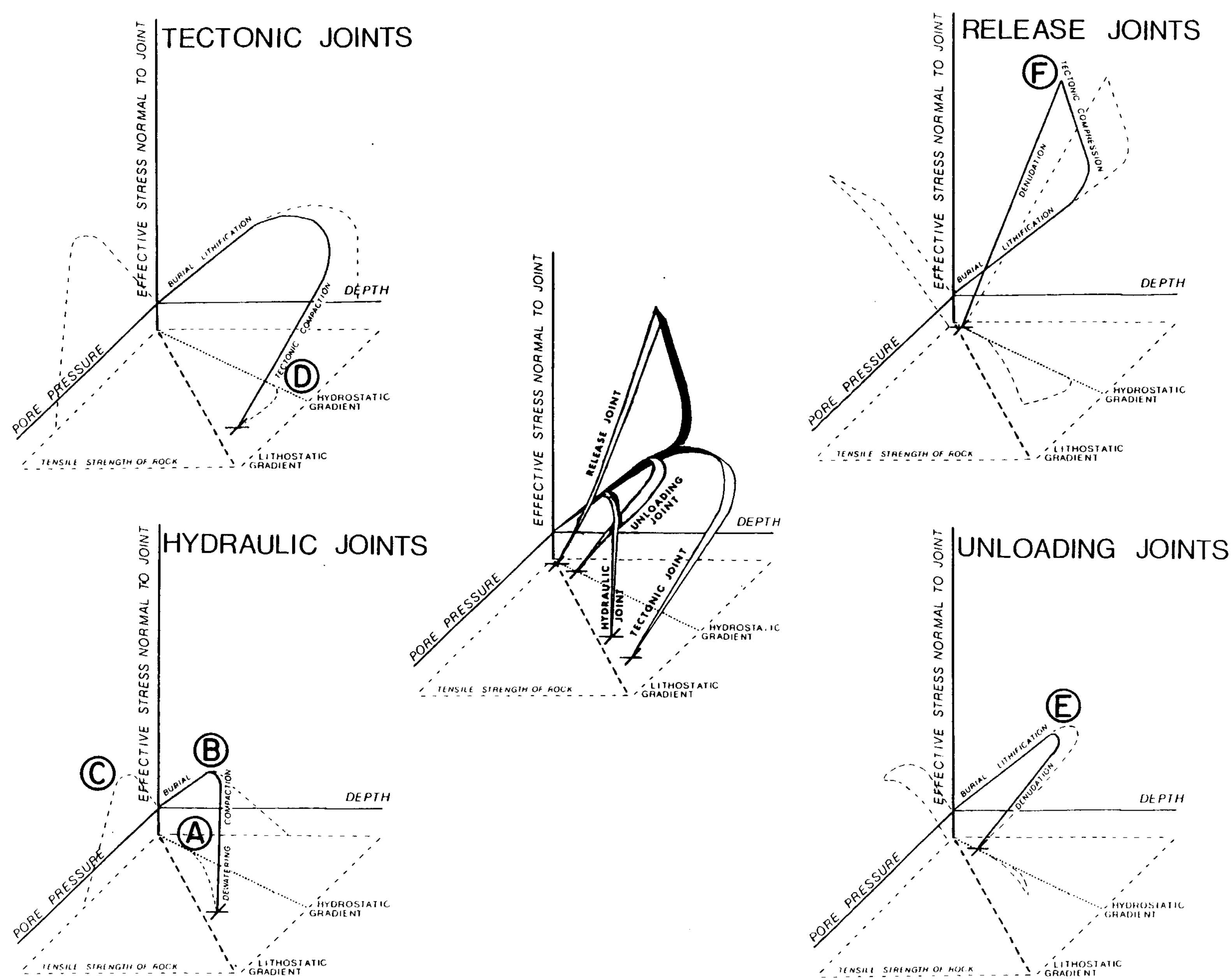

Fig. 5. Loading paths for tectonic, hydraulic, release, and unloading joints. Each diagram consists of three axes: effective stress normal to plane of jointing, depth, and pore pressure. The tensile failure envelope is shown normal to the effective stress axis. Projections of the loading path are shown by dashed lines on each of the three planes comprising the three-axis diagrams. No scale bars are shown which implies that the graphs are not necessarily drawn to the same scale. See text for details.

Fig. 6. (a) Set Ib joints within the Tully Limestone at Ludlowville Falls, New York where a later Main Phase Alleghanian cleavage cuts the joints (arrows). Coin is 1.8 cm in diameter. (b) Cross-fold joints within the Genesee Shale Formation of the Genesee Group at Taughannock Falls State Park, New York. Later joints are seen curving into earlier joints with the later joints being counterclockwise from the earlier joints. These joints are believed to belong to the same set although the variation in orientation is anomalously high. The joint spacing is about 1 m. (c) Cross-fold joints along route 414 at Watkins Glen, New York (after Bahat & Engelder 1984). The early joints formed within the thinner siltstone beds at the man's feet while the later joints formed within the thicker shale beds next to the man. The earlier joints are counterclockwise from the later joints. (d) Cross-fold joints within the upper portion of the Genesee Group at Taughannock Falls State Park, New York. The early joint (set Ib) within the thin siltstone bed is counter-clockwise from the later joints (set Ia) with the underlying shales. The siltstone bed is 19 cm thick.

Fig. 9. (a) Top of the Genesee Group at the intersection of routes 414 and 79 at Watkins Glen, New York. Here joints within a thick siltstone do not cross the interface between that siltstone and the adjacent shales. In contrast, joints originating within the underlying shales cross the overlying interface and propagated upward into the siltstone (arrows). This is an example of joint containment within the coarser grained unit which presumably is subject to a lower least principal stress at the time of joint propagation. (b) Cross-fold joints within the lower portion of the Genesee Group at Taughannock Falls State Park, New York. These joints propagated within a relatively homogeneous shale where they have a vertical dimension of as much as 50 m. The height of the outcrop is approximately 60 m. (c) The upper portion of the West Falls Group at Letchworth State Park, New York. Cross-fold joints (set Ib) cross the stream bed (parallel with the bottom of the photo), whereas strike joints (set II) parallel the stream bed. Note that the strike joints curve over distances of 50 m whereas cross-fold joints maintain parallelism. Height of the canyon wall is about 60 m. (d) Release (set II or strike) joints within the Canadaway Group at Angelica, New York. They extend from the southwest to northwest corner of the photo with 30 m along strike visible. Note that unloading (set Ib or cross-fold) joints abut the release joints.

pore pressures necessary to cause the propagation of hydraulic joints (Magara 1981, Narr & Currie 1982). Propagation of tectonic joints occurs during the active compression of the host rocks, a situation distinct from that developed in rocks whose only deformation has been overburden compaction during burial and lithification. The curve for the tectonic joints shown in Fig. 5 indicates little or no abnormal pressures developing during burial but abnormal pressure starts to develop during tectonic compaction at the maximum depth of burial (point D; tectonic joints). As was the case with hydraulic joints, the other curves on the three-axis diagram have a knee reflecting this increase in pore pressure.

Unloading joints are formed by a loading path similar to that proposed by Narr & Currie (1982) and illustrated in Fig. 4. This path involves little or no abnormal pore pressure during burial and subsequent erosional events. Although rocks in which unloading joints form may have been affected by a tectonic compression during burial, the compression has no bearing on the final propagation of these joints and, therefore, is not included in the loading path of Fig. 5. The key knee in the curve for the unloading joints (point E, Fig. 5) is the reversal in depth of burial caused by the change from active sedimentation to active erosion. These joints propagate after more than half of their overburden has been removed as calculated from the data in Table 2. Either a contemporary tectonic stress during erosion or a residual stress may act to control the propagation direction as will be elaborated later in the paper. Vertical unloading joints indicate that the effective stress in the horizontal plane became tensile, which is a stress condition that develops from thermal cooling and Poisson contraction (Price 1966, Haxby & Turcotte 1976). To achieve a tensile stress condition the effective stress gradient with depth must be steeper during denudation than during burial and such may happen with a change in thermal expansivity and Poisson's ratio during lithification (Voight & St. Pierre 1974).

Release joints, like unloading joints, form in response to the removal of overburden during erosion. Here a distinction is made because a tectonic compression and the fabric it leaves do have a bearing on the orientation of these joints. The orientation of release joints is fabric controlled whereas the other three joints are stress controlled. In the case of the release joints the orientation of the future joint plane is normal to the tectonic compression. Following burial and lithification, tectonic compression further increases the stress normal to the future plane of these joints. Hence, the normal stress becomes higher than for any of the other three joints (point F, release joints). On erosion these joints open in much the same manner as release joints in a triaxial compression experiment. Here the orientation of the joint may be controlled by some rock fabric such as solution cleavage planes rather than the contemporary tectonic stress at the time of propagation. In a fold and thrust belt these joints form parallel to the axes of folds but post-date active folding.

SEQUENCE OF JOINT DEVELOPMENT IN THE CATSKILL DELTA

Although joints that propagated under different stress conditions may have distinctive surface morphologies, the orientation of the joints relative to other structures including other joints is the most useful tool for determining the ages of the joints. A byproduct of age determinations is an inference about stress conditions at the time of joint propagation.

The relative age of joints was determined from the abutting relationships of the various joints within several exposures over a region. The general rule is that younger joints propagate up to but do not cut across older joints if the older joint had no tensile strength at the time of propagation of the younger joint. The older joint with no tensile strength stops the propagation of the younger joint because the tensile stresses at the tip of the advancing, younger joint cannot be maintained beyond the discontinuity of the initial joint (Kulander *et al.* 1979, Grout & Verbeek 1983). Hence, younger joints will abut against older joints.

Concerning the sequence of cross-fold joints in the Catskill Delta, Engelder & Geiser (1980) observed that more than one cross-fold joint set had formed in many outcrops and that one set propagated early in the tectonic cycle whereas another set propagated late within the tectonic cycle. Hence, the cross-fold joints were divided into two sets: Ia and Ib (Table 1). Although Engelder & Geiser (1980) identified two different cross-fold joint sets, Bahat & Engelder (1984) found evidence for more than two sets. Here, set Ia refers to those joints that strike parallel to the direction of compression of fossils and cleavage developed during the main phase of the Alleghanian Orogeny. If it appears within the same exposure, joint set Ib strikes counter-clockwise from the strike of set Ia. Engelder (1982a) concluded that the earliest cross-fold joints (more than one set) formed within the deeper, shalier portions of the Catskill Delta whereas the youngest cross-fold joints formed within the shallower, sandier portions of the delta.

The best evidence for the relative age of early cross-fold joints is found within the Tully Limestone where older set Ib joints are cut by a later spaced cleavage (Fig. 6a). Here set Ia joints propagated normal to and formed synchronously with the cleavage planes as indicated by cutting relationships. From this it is seen that some set Ib joints propagated before set Ia joints. In terms of orientation, the joint set (Ib) whose strike is counter-clockwise from the other set (Ia) propagated first. In the deeper portion of the Catskill Delta set Ib joints correlate with the earlier phase of the Alleghanian Orogeny, the Lackawanna Phase, whereas the set Ia joints correlate with the later Main Phase of the Alleghanian Orogeny (Geiser & Engelder 1983).

Evidence for abutting of joints from one cross-fold set against joints from another is rare in the deep portions of the delta and is not found in the shallow portions. The most common situations are either for two cross-fold joint sets to form in adjacent beds without intersecting

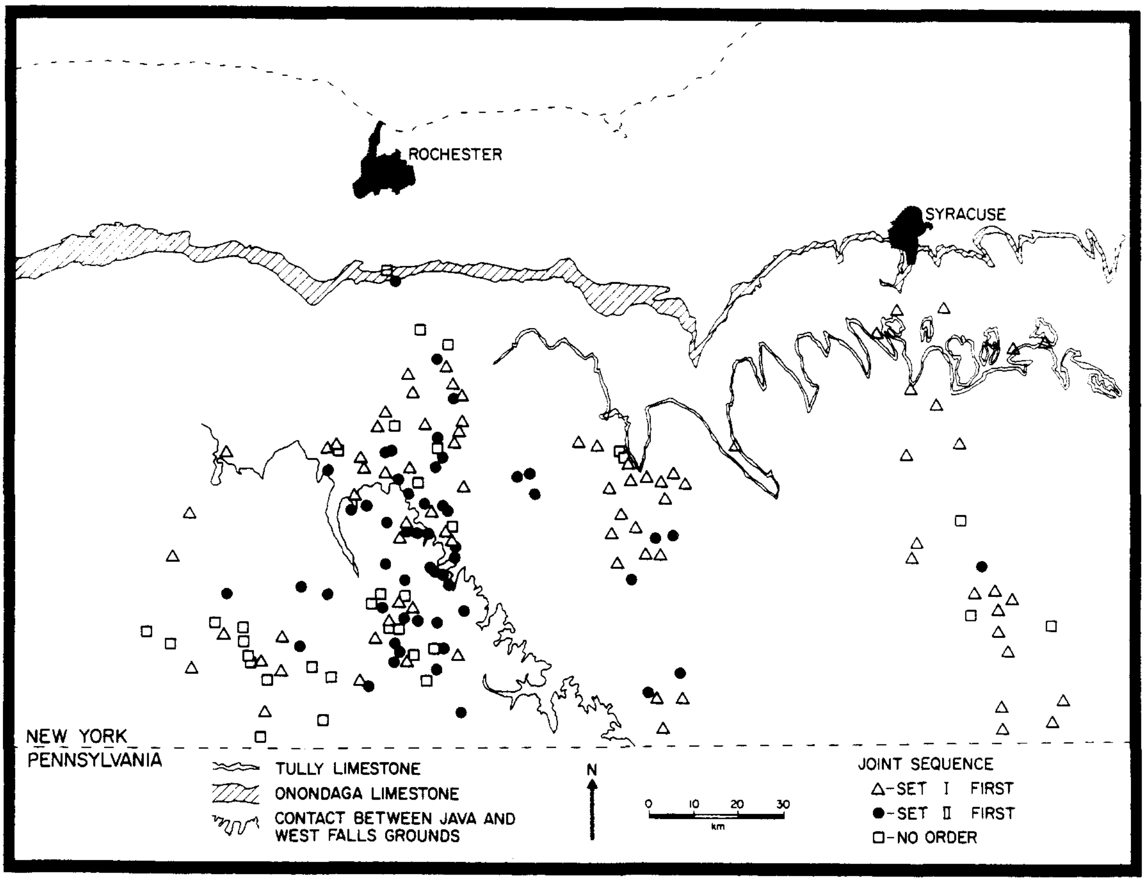

Fig. 7. Map of western New York showing the abutting relationships between set I (cross-fold) joints and set II (strike) joints (after Engelder 1982b).

or for the joints in the same bed to mutually crosscut. In the latter case the joints commonly lack mineral filling which could be used to distinguish the age of the joints. An example of abutting cross-fold joints is found in the Genesee Shale Formation of the Genesee Group at Taughannock Falls (Fig. 6b). This is an unusual outcrop because the cross-fold joints curve and are less ordered or parallel than is typical. In other exposures, cross-fold joints tend to be parallel and planar in outcrop (Figs. 6c & d). Because the joint set Ib abuts joint set Ia, joint set Ib is taken to be post-Alleghanian and correlates with those set Ib joints found in and above the West Falls Group.

Evidence that cross-fold joints formed late within the upper, sandier units of the Catskill Delta is the abutting relationships between the strike joints and cross-fold joints (Fig. 7). Engelder (1982a) noted that in the deeper portions of the Catskill Delta, almost all cross-fold joint sets propagated prior to the strike joints. This includes both sets Ia and Ib seen within the Tully Limestone but not set Ib seen within the Genesee Shale Formation. In contrast, in and above the West Falls Group, where set Ia joints are missing the opposite sequence of development is more common but not the rule. In the upper portions of the Catskill Delta some outcrops show both abutting sequences so that no order of formation may be inferred. Thus, it is likely that late-forming set I joints (the Ib joint set of Engelder & Geiser, 1980) and set II joints formed at the same time within and above the West Falls Group. These joint sets formed after the Alleghanian Orogeny as they are not cut by the Alleghanian age structures. From these observations it may be stated that the set Ib joints identified by Engelder & Geiser (1980, fig. 4a) do not all have the same mechanism of formation as set Ia.

To infer the time of propagation of set III joints based on abutting relationships is difficult. Generally, set III joints propagated as isolated joints rather than as a closely spaced set. This joint set may be observed cross-cutting other joint sets as well as abutting them. Occasionally, joints from other sets abut the set III joints. Based on these age relationships, the sequence of jointing in the Catskill Delta of the Appalachian Plateau was: (1) set Ib joints below the West Falls Group; (2) set Ia joints below the West Falls Group and (3) set Ib joints above the West Falls Group plus set II and set III joints. Evidence from abutting relationships does not allow the latter three sets to be dated relative to each other.

In a sedimentary basin with the four idealized loading paths, the sequence of development of joint types would be hydraulic joints, tectonic joints, release joints, and unloading joints. In the Catskill Delta none of the joint sets may be considered as candidates for hydraulic jointing as the depth of burial was insufficient for this process (Narr & Currie 1982, Magara 1981). However, the oldest joints (set Ia and set Ib formed below the West Falls Group) are tectonic joints because of their depth of burial at the time of propagation and because of their mutually cross-cutting relationships with Alleghanian cleavage (Engelder & Geiser 1980). The joints that propagated after the Alleghanian Orogeny are most likely to have propagated during a phase of uplift and erosion, in which case shallow set Ib joints would be unloading joints whereas shallow set II joints would be release joints. The orientation of the shallow set Ib joints was controlled by an Alleghanian-age residual stress (Engelder & Geiser 1980) whereas the orientation of the set II joints was controlled by a solution cleavage related fabric within the rocks. Again it is important to emphasize that joints originally lumped into set Ib by Engelder & Geiser (1980) consist of more than one type of joint. It will be argued later in the paper that set III joints are also unloading joints with an orientation controlled by the contemporary tectonic stress field.

Some cross-fold joints propagated during the Alleghanian Orogeny when the Catskill Delta was at maximum depth of burial. At these depths natural hydraulic fracturing under abnormal pore pressure is the only mechanism that comes to mind for joint propagation. Joint propagation did not occur in the shallow parts (above the West Falls Group) of the Catskill Delta during the Alleghanian Orogeny. In this part of the section, abnormal fluid pressures from tectonic compaction leaked before becoming high enough for joint propagation. Strain markers indicate a tectonic compaction of 10% adding to the likelihood that abnormal pressures developed during tectonic deformation in the deeper part of the Catskill Delta (Engelder & Engelder 1977). The deeper part of the delta is the only location where tectonic joints propagated parallel to layer-parallel shortening as indicated by deformed fossils (Engelder & Geiser 1980). Joints of the same age, further to the south in the Appalachian Valley and Ridge, are interpreted as having propagated under high pore pressure prior to development of the major Alleghanian folds (Nickelsen 1979). The following is the case that supports a hydraulic fracture mechanism for the tectonic joints of the Catskill Delta.

HYDRAULIC FRACTURING

Industrial hydraulic fractures

An understanding of both the timing and mechanisms of the propagation of tectonic joints within the Catskill Delta may be derived from recent developments in the use of hydraulic fractures to enhance the recovery of hydrocarbons. The idea behind a hydraulic fracture treatment is to drive a vertical fracture from the wellbore into a stratigraphic horizon containing hydrocarbons and hence, provide a conduit for directing the flow of hydrocarbons into the wellbore (Hubbert & Willis 1957). Generally, it is desirable to drive a vertical fracture as far as possible into the pay zone which usually has a finite thickness but is of unlimited lateral extent.

As early as the 1950s it was recognized that under certain circumstances vertical hydraulic fractures do have limited height while propagating a great distance laterally. In an interbedded sandstone and shale sequence the hydraulic fractures characteristically grow vertically to a boundary (generally a horizontal bedding plane) that prevents further vertical growth (Daneshy 1978). At this point, additional pumping of fluid into the wellbore drives the hydraulic fracture outward from the wellbore without vertical growth into the overlying bed. This is called a confined-height fracture or a contained hydraulic fracture. Nordgren (1972) predicted that for a Newtonian fluid driving a confined-height fracture outward at a constant pumping rate, the fluid pressure at the wellbore increases proportionally to time of propagation raised to an exponent:

$$p(t) \sim t^e \qquad (2)$$

where $1/8 < e < 1/5$. In a plot of wellbore pressure vs log of time this curve has a small positive slope.

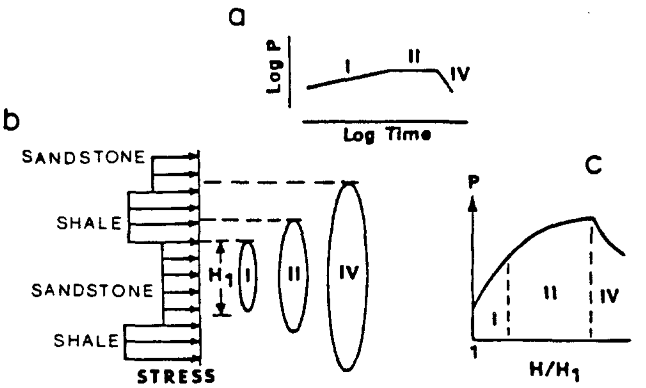

Fig. 8. (a) Pressure-time curve for driving a hydraulic fracture into a pay zone from a wellbore (log-log plot) (after Nolte & Smith (1981). P is the pumping pressure. The three stages of pumping are indicated as I, II and IV. (b) The relative magnitudes of the horizontal stresses in shale and sandstone beds. Pumping stages I, II and IV on the pressure-time curve are representative of hydraulic fracture heights: I, confined to the pay zone; II, breaking upward into the overlying shale and IV, breaking through the shale into the top sandstone. H_1 is the thickness of central sandstone bed. (c) Pressure-fracture height curves through stages I, II and IV of fracture growth. H is the height of the fracture.

In examining records from hydraulic fracture treatments, Nolte & Smith (1981) recognized that after a certain amount of pumping many log pressure vs log time curves deviated from Nordgren's (1972) prediction of a small positive slope by becoming horizontal and on occasion negatively sloping (Fig. 8). Nolte & Smith (1981) concluded that these slope changes reflected the penetration of the hydraulic fractures into vertically adjacent beds. A major assumption in this model is that the sandstones have a lower least principal stress than the adjacent shales. In this interpretation the fracture initially propagates vertically in sandstone to a bedding interface with shale but stops because the fluid pressure in the hydraulic fracture is not sufficient to reduce the effective least principal stress within the shale above the interface to the point of fracturing. Additional pumping drives the vertical fracture outward and increases fluid pressure slowly, overcoming the viscous friction between fracture surfaces and fracture fluid flowing towards the fracture tip. When the slope of the curve for pressure vs log time becomes zero, Nolte & Smith (1981) suggested that the wellbore pressure is sufficient to initiate fracturing in the overlying shale bed. The fracture is then driven through the overlying shale without an additional increase in pumping pressure; hence, the pressure–time curve has a zero slope. If the fracture continues to grow vertically and climbs into another sandstone where there is a lower least principal stress, the fluid pressure necessary for fracture propagation is less, which is represented by a negative slope in the curve.

Nolte & Smith (1981) used strain-relaxation techniques to measure stress variations within the Second Frontier Shale and the Muddy-J Formation from Colorado to demonstrate the higher least principal stress within shales relative to sandstones. In another test a 50% increase in effective stress was required to match correctly the pressure behavior in shale relative to sandstone within a sequence, the location of which was not reported (Nolte & Smith 1981). Higher stresses may occur within the shale relative to sandstone layers because stress relaxation in the shale transfers vertical to horizontal stresses. This interpretation is supported by Abou-Sayed *et al.* (1981) who showed that in general shales have a higher ratio of least horizontal to vertical stress compared to sandstones at the same depths of burial. During burial and diagenesis the Poisson's ratio for the shale must become relatively large as reported by Eaton (1969) for the shales of the Gulf of Mexico.

Hydraulic fracture measurements within the Appalachian Basin suggest that even to this day sandstone layers have a lower least principal stress than adjacent shale beds. Voegel *et al.* (1981) reported that within the Devonian shales which were penetrated by Columbia Gas Company Well No. 20402, West Virginia, stresses vary among layers at depths between 870 and 1330 m. In the case of the Benson Sandstone within Columbia Gas Company well No. 20538, West Virginia, at about 1320 m of depth, the sandstone layers have a least stress of 12.4 MPa which is less than the least stress of 15.7 MPa in the surrounding shales (Abou-Sayed *et al.* 1978). In an experiment within interfingered sandstones and shales in British Columbia, Canada, Gronseth & Kry (1983) report a minimum 6 MPa difference between the least stresses in sandstones and shales.

Application to joint propagation within the Catskill Delta

The upper portion of the Genesee Group, the Sonyea Group and the lower portion of the West Falls Group of the Catskill Delta are characterized by interfingered siltstone and shale where the data and interpretation from hydraulic-fracture treatments may be applicable. Based on the treatments, there are several possible observations: (1) natural hydraulic fractures should propagate in siltstones prior to shales; (2) early joints in siltstone should not penetrate shale beds but rather stop at shale–siltstone interfaces: (3) later joints in shales may well propagate into siltstone layers and (4) joints in thick shale beds should have a greater vertical extent than joints in interfingered siltstone–shale beds where there are relatively thinner beds and many bedding interfaces. All of these characteristics are observable within the Catskill Delta.

A key exposure is that reported by Bahat & Engelder (1984) where the Genesee Group has siltstones between 20 cm and several meters thick interfingered with shale beds of the same thickness range (Figs. 6c & d). Set Ib joints striking 335° cut siltstones and stop at siltstone-shale interfaces whereas set Ia joints striking 345° cut the shales. The principal shortening direction as indicated by deformed fossils is oriented parallel to the joints in the shales. This shortening was caused by compression during the Main Phase of the Alleghanian Orogeny (Geiser & Engelder 1983). Joints cutting the siltstone correlate geometrically with folds and cleavage of the Lackawanna Phase of the Alleghanian Orogeny, an earlier and weaker compressional event affecting rocks mainly to the southeast of the study area (Fig. 3). Set Ib joints also correlate with those within the Tully Limestone that are cut by and hence, predate the Main Phase cleavage (Fig. 6b). These correlations indicate that the set Ib joints within the siltstone beds propagated prior to the Ia joints within the shale beds, the same sequence predicted by industrial hydraulic fracturing.

Joints which propagate within shales do not necessarily propagate into the siltstone layers as would be predicted using Nolte & Smith's (1981) analysis (Fig. 6d). However, there are examples that conform with the prediction, such as the outcrop of the Genesee Group at the intersection of routes 414 and 79 at Watkins Glen (Fig. 9). Early joints cut a four meter siltstone bed and stop at the siltstone–shale interfaces, whereas joints from the lower shale beds are seen cutting across a siltstone–shale interface and propagating up into the siltstones. The joints within the siltstones tend to be more planar compared with joints within the shales.

Combining this information with that developed during the discussion concerning sequence of joint formation, it is evident that the cross-fold joints within the siltstone layers formed prior to those cross-fold joints within the thicker shales. This sequence may also be expected by referring back to the very simple model for joint formation pictured in Fig. 4, in which it is shown that the least principal stress in sandstones during burial may be less than the least principal stress within shales. Hence, smaller abnormal pressures are required to assist joint propagation in the siltstone relative to the shale.

The thick shales in the Catskill Delta occur in the Hamilton and the lower Genesee Groups. The best vertical exposure of these shales occurs in the Genesee Group at Taughannock Falls State Park at Ithaca, New York, where 50 m of shale may be viewed in continuous vertical exposure. Single cross-fold joints may be traced vertically for much of the 50 m exposure (Fig. 9b). Also the cross-fold joints within these shales are less regular or planar than the planar joints cutting thinner siltstone beds higher in the stratigraphic section (Fig. 6a vs Fig. 6d). These joints may also be traced within the stream bed of Taughannock Creek where their horizon-

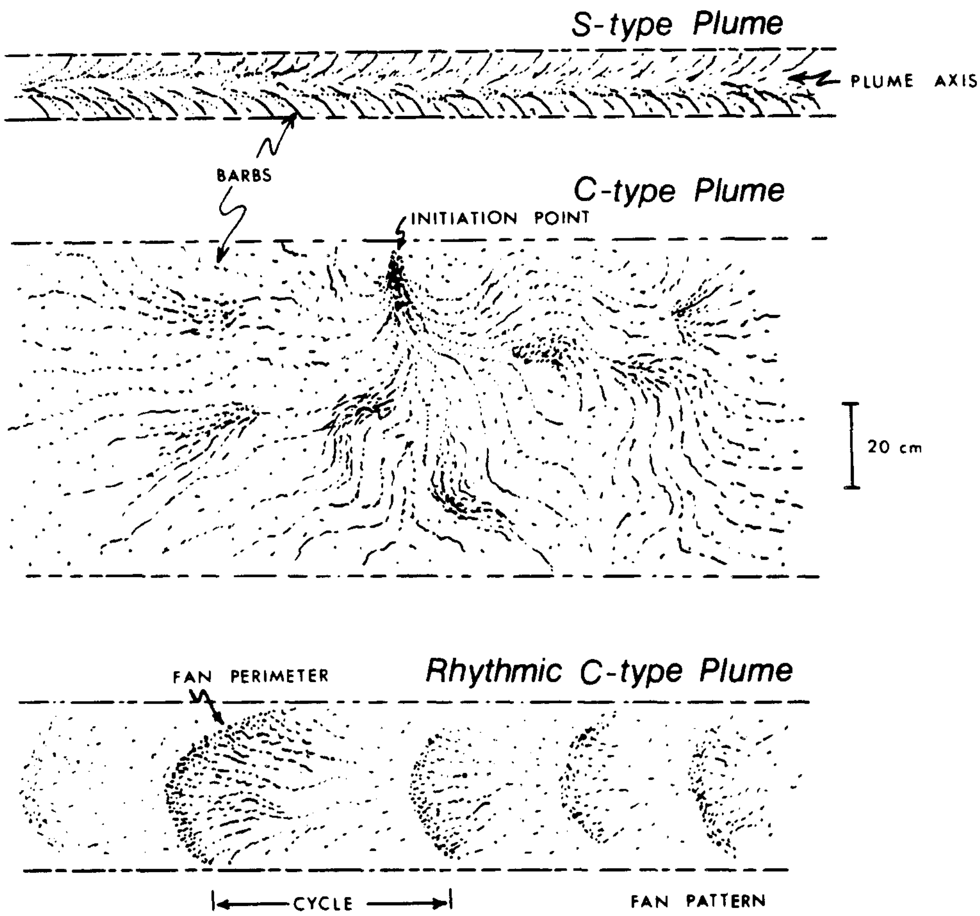

Fig. 10. The surface morphology of cross-fold joints at Watkins Glen, New York (after Bahat & Engelder 1984). The *c*-type plumes are found on the earlier formed joints within the siltstone beds whereas the *s*-type plumes are found on the later formed joints cutting thin siltstone beds within the thick shale beds.

tal dimension is about the same as their vertical dimension in the cliff face. This is in contrast to cross-fold joints cutting bedded siltstone–shale sequences where the horizontal dimension is much larger than the vertical dimension.

The idea from the experience of the petroleum industry is that fractures will propagate in the direction of a gradient of decreasing least stress or, failing that, propagate in the direction of no gradient rather than propagate towards a gradient of increasing least stress. In a homogeneous shale bed least principal stress decreases in a vertical direction by virtue of there being less overburden in that direction. Thus once a fracture initiates in a homogeneous bed its tendency is to propagate in a vertical direction. At Taughannock Falls, the equal horizontal and vertical dimension suggests the vertical gradient in stress was about equal to the horizontal gradient.

Surface morphology of tectonic joints

Using surface morphology, some inferences may be drawn about the loading history of cross-fold joints. Bahat & Engelder (1984) described the surface morphology of cross-fold joints that formed within the interfingered siltstone–shale portion of the Genesee Group near Watkins Glen, New York. Briefly, their observations were that the two different cross-fold joint sets cutting siltstones and shales had different types of surface morphology (Figs. 6c and 10). *S*-type plumes form on the more northerly striking cross-fold joints (345°, set Ia) cutting thin siltstone beds embedded in thicker shale formations. These plumes have straight plume patterns with axes parallel to bedding. *C*-type plumes form on more westerly striking cross-fold joints (335°, set Ib) cutting thick siltstone beds. These plumes have curved plume patterns with axes that either curve or show fan-like rhythmic patterns that alternately increase and decrease in intensity. Joints cutting only shales exhibit no distinct surface morphology other than long arcuate arrest lines. Arrest lines reflect the point at which fracture propagation stopped often with the rupture rotating out of the plane of propagation. Other pertinent characteristics of joint morphology includes Parker's (1942) observation that throughout the present study area plumose markings are rare on strike joints but more common on cross-fold joints.

The fan-like rhythmic patterns of the *c*-type plumes on some cross-fold joints (set Ib) in the thick siltstones indicate that the joints formed by cyclic propagation rather than by one massive rupture. Cyclic propagation, which is a process discussed by Secor (1969), indicates that the driving force for these fractures diminished as the rupture front migrated and hence, the propagation

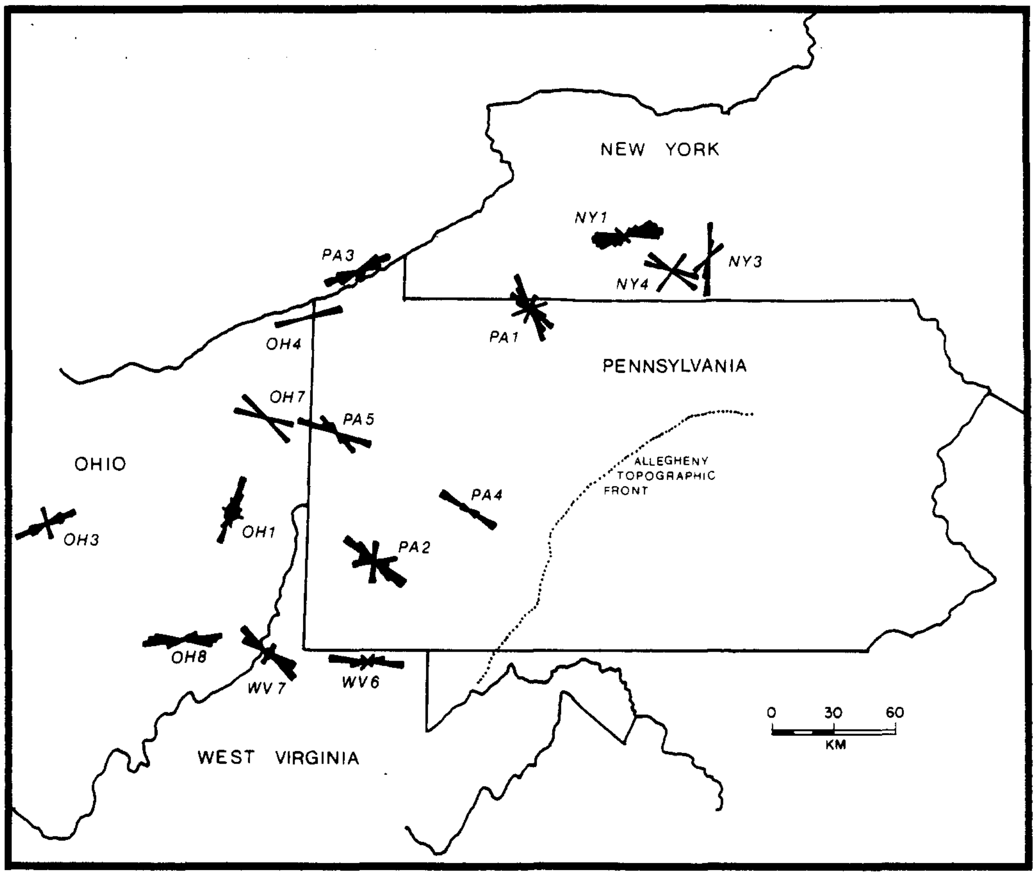

Fig. 11. The location of 15 wells drilled by the Department of Energy for their eastern gas shales program. The rose diagrams show the strike of natural joints found within core taken from each of the 15 wells. Comparing the strike of joints at depth with the orientation of cross-fold joints shown in Fig. 3, it is apparent that most joints from wells near the Allegheny Topographic Front are in the cross-strike orientation whereas those joints from wells most distant from the Allegheny Topographic Front are parallel to the contemporary tectonic stress field which has a maximum horizontal stress oriented ENE (Engelder 1982a). This apparent geographic distribution is a function of depth from which the cores were taken as shown in Fig. 12.

stopped within several tens of centimeters from the initiation point. If the effective stress for failure was generated by abnormal water pressure within the crack, then the opening of the crack on rupture would immediately reduce the pressure within the crack. Hence, the effective stress would increase without a change in differential stress, causing crack propagation to cease. Only multiple recharges of fluids from the adjacent shale beds or the pores of the siltstone would cause repeated propagation through several cycles of fracture growth.

This evidence for natural hydraulic fracturing suggests that the fluid pressure along the crack was continuously recharged until fluid pressure exceeded the least principal stress and forced the crack open. In order for the fluid pressure along the crack to exceed the least principal stress, a decreasing fluid pressure gradient is required from the rock matrix to the crack. This is possible only if the fluid in the pores of the adjacent rock was at a higher pressure. One possibility is that the adjacent shale at a higher least principal stress can sustain a high pore pressure without fracturing while slowly draining into and recharging the siltstone (Fig. 4). Another possibility is that the pore pressure in the matrix of the siltstone is equal to the average stress on the siltstone $P_p = (\sigma_1 + \sigma_2 + \sigma_3)/3$ rather than the least principal stress, σ_3. This case would give a positive pore pressure gradient from the matrix to the crack.

The joints (set Ia) within the shales appear to have formed by massive ruptures. This notion is supported by the *s*-type plume on the thin siltstone beds interfingered with the thick shales. The *s*-type plumes show no arrest lines in more than 50 m of outcrop length. By analogy with industrial hydraulic fracturing, these joints in shale must have been driven with larger fluid reservoirs compared with the limited reservoir driving single rupture events in siltstone layers. Yet, the stress condition for driving the rupture is unlikely to involve fluid pressure within the crack at a distance from the rupture front, because the rate of fluid migration into the crack would probably be slow relative to the propagation rate of the joints. Thus, the fluid pressure would decrease within the crack during crack propagation. With this condition the rupture could not be driven by effective stress conditions back within the open joint and must be sustained by some other mechanism. If the fluid pressure was critical for maintaining the effective stress conditions necessary for crack propagation, effective stress right at or in front of the crack tip must be important to the process.

UNLOADING AND RELEASE JOINTS

Joints related to Alleghanian structures and residual stress

The joint distribution within the upper and westward portion of the Catskill Delta differs from that in the lower and eastward portion. This difference may be observed within Letchworth State Park where the upper West Falls Group is exposed. In the vicinity of Letchworth State Park the Catskill Delta thinned to about 1.3 km compared to a total thickness of about 2 km in the vicinity of Watkins Glen (Fig. 2). Overburden on the West Falls Group during the Alleghanian orogeny was less than 1 km. Within the Genesee River gorge the West Falls Group is a thinly bedded siltstone–shale sequence in which joints are restricted to neither the siltstones nor shales and, yet, the joints have not grown as tall as the joints in the Genesee Group at Taughannock Falls (Fig. 9c). These cross-fold (set Ib) joints in the West Falls Group formed late as unloading joints with their orientation being controlled by a residual stress (Engelder & Geiser 1980). Because the joints do not favor siltstones or shales, it may be suggested that variation in least principal stress between siltstones and shales is smaller or non-existent at the time of propagation which is inferred to be during unloading as indicated by the butting sequence for joints in western New York (Fig. 7). Unloading joints do not require abnormal pore-fluid pressure to drive the joint propagation (Fig. 5). The result is that a joint prefers neither the sandstones nor the shales and does not climb vertically more than a few meters.

As indicated in the discussion on jointing sequence, the upper part of the Catskill Delta differs from the more shaly lower portion in that strike joints are more likely to have propagated before cross-fold joints. The strike (set II) joints frequently curve and are not parallel with adjacent joints of the set (Figs. 9c & d). This is in contrast with all cross-fold (set I) joints that are parallel regardless of position within the Catskill Delta. The strike joints are considered to be release joints by virtue of post-dating the Alleghanian Orogeny and opening normal to the compression direction of the Alleghanian Orogeny. Also, as will be discussed in the next section, strike joints are not common at depths that are greater than 500 m. This near surface distribution further supports the implication that the strike joints opened during uplift and removal of overburden.

Joints related to the contemporary tectonic stress field

The orientation of unloading joints is controlled by either the tectonic stress field at the time of denudation and uplift or residual stress left from some previous tectonic event. A set of joints (set III of Parker 1942) on the Appalachian Plateau is aligned with the contemporary tectonic stress field and because of this relationship Engelder (1982a) proposed that the joints were genetically related to the contemporary tectonic stress. These joints are related to neither a residual stress nor a structural fabric left by the Alleghanian tectonic compression that affected the Devonian rocks of the Appalachian Plateau. Set III unloading joints on the Appalachian Plateau are distinct from cross-fold joints of Alleghanian age (tectonic joints of Fig. 5), later cross-fold joints (unloading joints following an Alleghanian residual stress), and strike joints of post-Alleghanian age (release joints of Fig. 5) (Engelder & Geiser 1980).

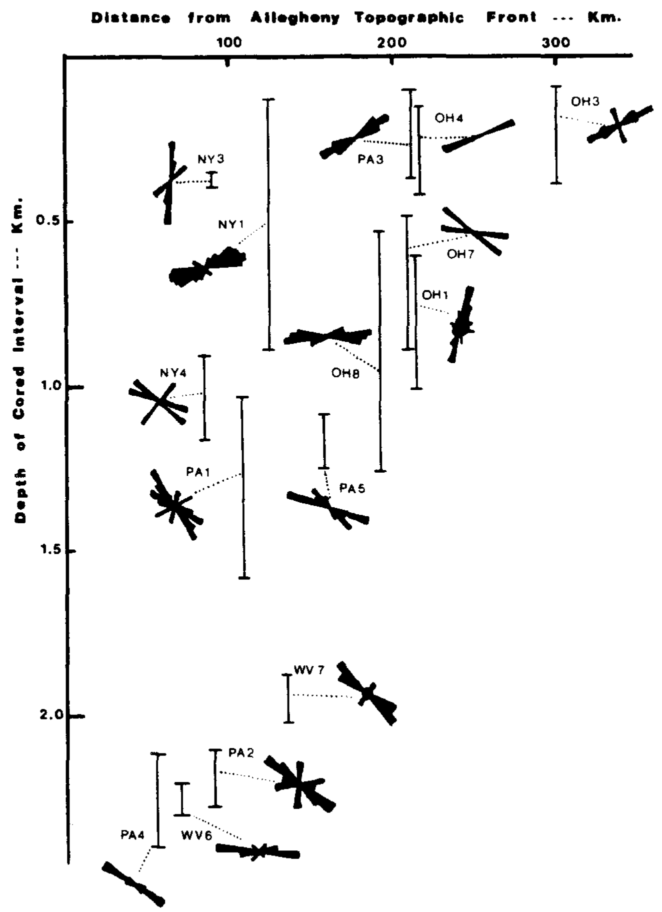

Fig. 12. A plot of depth of cored interval vs distance from the Allegheny Topographic Front. As in Fig. 11 the strike of the natural joints is plotted in map view with north toward the top of the figure. Here it is seen that the deep cores contain cross-fold joints whereas shallow cores contain joints parallel to the contemporary tectonic stress field. An interpretation is that deep joints are tectonic whereas the shallow joints are unloading structures.

From 1975 to 1981 selected intervals of thirty-three wells were core-drilled for the Eastern Gas Shales Project in the Appalachian Basin. The cores were oriented so that natural joints intersected by the well could be logged and correlated with mechanical characteristics, structural position and stratigraphic interval. Compilations of these data appear in Morgantown Energy Technology Center report DOE/MC/14693-1296 (Cliffs Minerals 1982). The locations of 15 of these wells are plotted on a map (Fig. 11) which should be compared with Fig. 3 showing the orientation of cross-fold joints as mapped by Ver Steeg (1944), Nichelsen & Hough (1967) and Engelder & Geiser (1980). Joints at depth are divided into those that correlate with cross-fold joints and those that correlate with the ENE direction of maximum horizontal compression in the contemporary tectonic stress field (Fig. 12). The most common joint

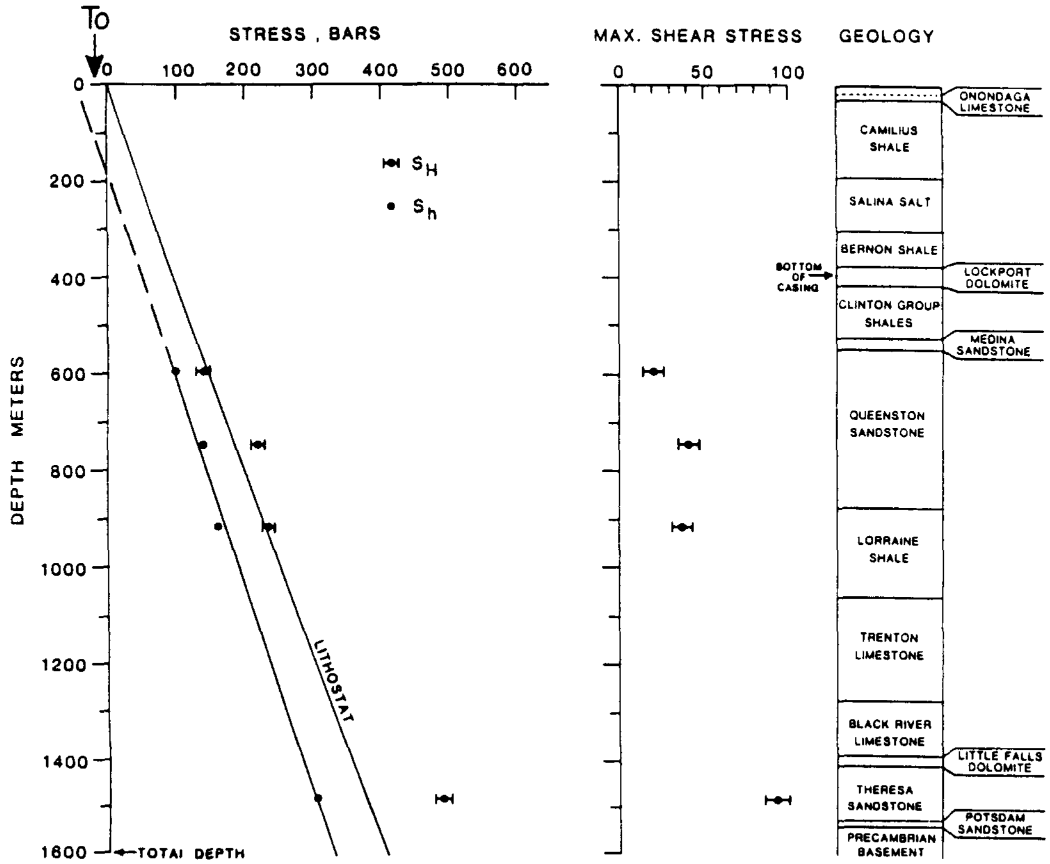

Fig. 13. A plot of horizontal stress vs depth for the hydraulic fracture measurements taken at the Auburn, New York geothermal wall (after Hickman *et al.* 1984). An extrapolation of the least principal stress back toward the surface shows that the tensile strength of the local rocks should be exceeded at a depth of about 200 m.

within the top 0.5 km strikes subparallel to the contemporary tectonic stress field whereas those observed below the top 0.5 km are almost exclusively cross-fold joints. Joints within the core from OH-1 are the only joints that do not fit into the scheme of unloading joints at <0.5 km depth and tectonic joints between 0.5 and 2.5 km. In fact the joints within OH-1 are oriented parallel to strike joints and may be release joints.

The deepest set of stress measurements from the Appalachian Basin was taken with a 1.6 km deep well at Auburn, New York (Hickman *et al.* 1984). An extrapolation of the magnitude of the least principal stress towards the surface suggests that tensile stresses might develop at about 200 m depth (Fig. 13). If normal pore-fluid pressure extends from the surface to 400 m depth, the zone of effective tensile stresses extends down to about 400 m. The orientation of the least principal stress is about 5° west of north. These tensile stresses would be relieved by the propagation of unloading joints oriented slightly north of east. Although stress data from this well do not come from the Devonian sequence, the depth at which tensile stresses develop is within the depth range of Devonian core containing joints striking parallel to the contemporary tectonic stress field.

Discussion

Unloading joints require the removal of overburden equal to more than 50% of their initial depth of burial depending on the change in Poisson's ratio during lithification (Fig. 4). Estimates for the removal of overburden from the Appalachian Plateau of western New York vary from 500 m (Van Tyne 1983) to 2 km (conodont isograd index of Epstein *et al.* 1975). This denudation is compatible with the propagation of unloading joints at a few hundred meters to 1 km in depth. Data from both Hickman *et al.* (1984) and Cliffs Minerals (1982) suggest that the depth of propagation of unloading joints is between 200 and 500 m. These observations and this interpretation further reinforce the proposal of Engelder (1982a) that there is a genetic relationship between the contemporary tectonic stress field and set III joints.

Haimson & Doe (1983) report that joints of the ENE orientation are the most common within crystalline basement to a depth of 1.6 km in a deep well drilled in northern Illinois. If these joints are also unloading joints then they formed at depths indicating less than 50% of the overburden has to be removed before propagation starts. Much less than 1.6 km of cover rock has been removed during the period in which the midcontinent has been subjected to the contemporary tectonic stress field. According to equation 1, rocks with a higher Young's modulus such at those within a crystalline terrain may develop tensile stresses with the removal of much less overburden than is required by the more compliant sedimentary rocks.

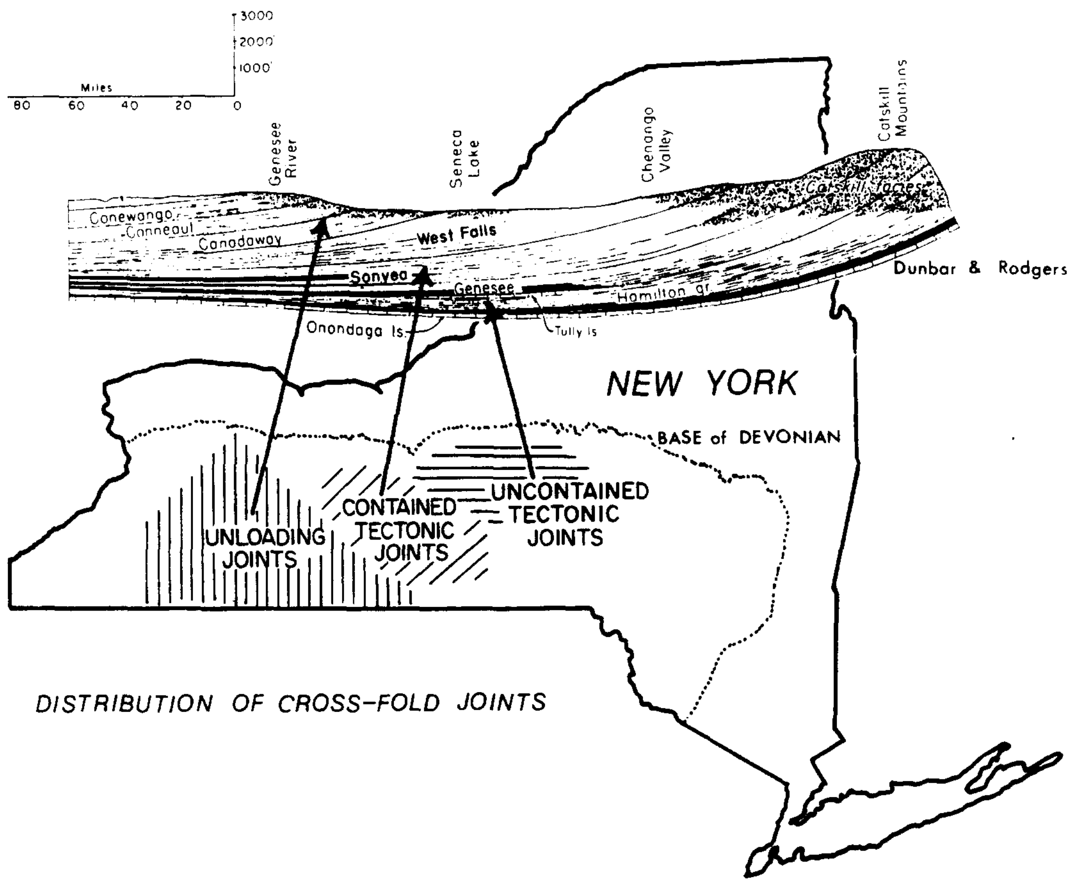

Fig. 14. Schematic map and section of the location of the various cross-fold joints within the Catskill Delta.

CONCLUSIONS

Four types of joint have been distinguished based on their loading paths: tectonic, hydraulic, unloading and release. Of these, three types can be identified within rocks of the Catskill Delta of the Appalachian Plateau. These three include tectonic joints that require abnormal pore pressures to achieve effective tensile stresses, plus unloading and release joints that propagate once erosion and uplift have generated a state of effective tensile stress. The type of cross-fold joint varies with vertical position within the Catskill Delta. Uncontained tectonic joints propagated within the deepest portions, contained tectonic joints propagated at intermediate depths, and unloading joints propagated within the shallow portions (Fig. 14). During the burial and tectonic compaction phases of a tectonic cycle joints do not form within the shallow portions of the Catskill Delta.

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REFERENCES

Abou-Sayed, A. S., Brechtel, C. E. & Clifton, R. J. 1978. In situ stress determination hydrofracturing: a fracture mechanics approach. *J. geophys. Res.* **83**, 2851–2862.

Abou-Sayed, A. S., Ahmed, U. & Jones, A. 1981. Systematic approach to massive hydraulic fracturing design. *Petrol. Engng Soc.* Paper 9877.

Bahat, D. & Engelder, T. 1984. Surface morphology on joints of the Appalachian Plateau, New York and Pennsylvania. *Tectonophysics* **104**, 299–313.

Barker, C. 1972. Aquathermal pressuring: role of temperature in development of abnormal-pressure zones. *Bull. Am. Ass. Petrol. Geol.* **56**, 2068–2071.

Chong, K. P., Uenishi, K., Smith, J. W. & Munari, A. C. 1980. Non-linear three dimensional mechanical characterization of Colorado oil shale. *Int. J. Rock Mech. Min. Sci.* **17**, 339–347.

Colton, G. W. 1970. The Appalachian Basin—Its depositional sequences and their geologic relationships. In: *Studies of Appalachian Geology Central and Southern* (edited by Fisher, G. W., Pettijohn, F. J., Reed, J. C. & Weaver, K. N.). John Wiley, New York, 5–47.

Cliffs Minerals, Inc. 1982. Analysis of the Devonian Shales in the Appalachian Basin. U.S. DOE contract DE-AS21-80MC14693, Final Report. Springfield Clearing House, Washington.

Daneshy, A. A. 1978. Hydraulic fracture propagation in layered formations. *J. Petrol. Technol.* **18**,

Eaton, B. A. 1969. Fracture gradient prediction and its application in oil field operations. *J. Petrol. Technol.* **21**, 1353–1360.

Engelder, T. 1982a. Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America? *Tectonics* **1**, 161–177.

Engelder, T. 1982b. Reply to a comment on 'Is there a genetic relationship between selected regional joints and contemporary stress within the lithosphere of North America" by A. E. Scheidegger. *Tectonics* **1**, 465–470.

Engelder, T. & Engelder, R. 1977. Fossil distortion and decollement tectonics on the Appalachian Plateau. *Geology* **5**, 457–460.

Engelder, T. & Geiser, P. 1980. On the use of regional joint sets as trajectories of paleostress fields during the development of the Appalachian Plateau, New York. *J. geophys. Res.* **85**, 6319–6341.

Epstein, A. G., Epstein, J. B. & Harris, L. B. 1975. Conodont color alteration—An index to organic metamorphism. *U.S. Geol. Surv. Open File Rept.* **75-379**, 54.

Friedman, G. M. & Sanders, J. E. 1982. Time–temperature burial significance of Devonian anthracite implies former great (~6.5 km) depth of burial of Catskill Mountains. *Geology* **10**, 93–96.

Geiser, P. & Engelder, T. 1983. The distribution of layer parallel shortening fabrics in the Appalachian foreland of New York and Pennsylvania: evidence for two non-coaxial phases of the Alleghanian Orogeny. *Mem. geol. Soc. Am.* **158**, 161–175.

Gilbert, G. K. 1882. Post-glacial joints. *Am. J. Sci.* **23**, 25–27.

Gronseth, J. M. & Kry, P. R. 1983. Instantaneous shut-in pressure and its relationship to the minimum in situ stress. In: *Hydraulic Fracturing Stress Measurements* (edited by Zoback, M. D. & Haimson, B. C.). National Academy Press, Washington, 55–60.

Grout, M. A. & Verbeek, E. R. 1983. Field studies of joints, insufficiencies and solutions, with examples from the Piceance Creek Basin, Colorado. *Proc. 16th Oil Shale Symposium* (edited by Gray, J. H.) Colorado School of Mines, Golden, Colorado, 65–80.

Haimson, B. C. & Doe, T. W. 1983. State of stress, permeability, and fractures in the Precambrian granite of northern Illinois. *J. geophys. Res.* **88**, 7355–7372.

Haxby, W. F. & Turcotte, D. L. 1976. Stress induced by the addition or removal of overburden and associated thermal effects. *Geology* **4**, 181–185.

Hickman, S. H., Healy, J. H. & Zoback, M. D. in press. *In situ* stress, natural fracture distribution, and borehole elongation in the Auburn Geothermal Well, Auburn, New York. *J. geophys. Res.*

Hubbert, M. K. & Willis, D. G. 1957. Mechanics of hydraulic fracturing. *Trans. AIME*, **210**, 153–168.

Kulander, B. R., Barton, C. C. & Dean, S. L. 1979. The Application of fractography to core and outcrop fracture investigations. Tech. Rpt. for US Dept. of Energy Contract EY-77-Y-21-1391, METC/SP-79/3, 174.

Lambe, T. W. & Whitman, R. V. 1969. *Soil Mechanics*. John Wiley, New York.

Magara, K. 1975. Importance of aquathermal pressuring effect in Gulf Coast. *Bull. Am. Ass. Petrol. Geol.* **59**, 2017–2045.

Magara, K. 1981. Mechanisms of natural fracturing in a sedimentary basin. *Bull. Am. Ass. Petrol. Geol.* **65**, 123–132.

Narr, W. & Currie, J. B. 1982. Origin of fracture porosity—Example from Altamont field, Utah. *Bull. Am. Ass. Petrol. Geol.* **66**, 1231–1247.

Nickelsen, R. P. & Hough, V. D. 1967. Jointing in the Appalachian Plateau of Pennsylvania. *Bull. geol. Soc. Am.* **78**, 609–630.

Nickelsen, R. P. 1976. Early jointing and cumulative fracture patterns: 1974. New Basement Tectonics contribution n. 23. *Utah Geol. Ass.* **5**, 193–199.

Nickelsen, R. P. 1979. Sequence of structural stages of the Alleghany orogeny at the Bear Valley Strip Mine, Shamokin, Pennsylvania. *Am. J. Sci.* **279**, 225–271.

Nolte, K. G. & Smith, M. B. 1981. Interpretation of fracturing pressures. *J. Petrol. Technol.* **33**, 1767–1772.

Nordgren, R. P. 1972. Propagation of a vertical hydraulic fracture. *Petrol. Engng Soc.* **12**, 306–314.

Parker, J. M. 1942. Regional systematic jointing in slightly deformed sedimentary rocks. *Bull. geol. Soc. Am.* **53**, 381–408.

Prats, M. 1981. Effect of burial history on the subsurface horizontal stresses of formations having different material properties. *J. Soc. Petrol. Engng* **40**, 658–662.

Price, N. J. 1966. *Fault and Joint Development in Brittle and Semi-brittle Rock*. Pergamon Press, London, 1–176.

Price, N. J. 1974. The development of stress systems and fracture patterns in undeformed sediments. *Proc. 3rd Rock Mech. Congr. ISRM*, **1**, 487–496.

Rickard, L. V. & Fisher, D. W. 1970. Geologic map of New York. New York Museum and Science Service Map and Chart Series No. 15.

Secor, D. T. 1965. Role of fluid pressure in jointing. *Am. J. Sci.* **263**, 633–646.

Secor, D. T. 1969. Mechanics of natural extension fracturing at depths in the earth's crust. *Geol. Surv. Pap. Can.* **68-52**, 3–48.

Van Tyne, A. 1958. Petrology of the Euclid Avenue Peridotite Intrusive. M. S. thesis. Syracuse Univ., Syracuse, New York.

Van Tyne, A. 1983. Natural gas potential of the Devonian black shales of New York. *Northeastern Geol.* **5**, 209–216.

Ver Steeg, K. 1942. Jointing in the coal beds of Ohio. *Econ. Geol.* **37**, 503–509.

Ver Steeg, K. 1944. Some structural features of Ohio. *J. Geol.* **52**, 131–138.

Voegel, M. D., Abou-Sayed, A. S. & Jones, A. H. 1981. Optimization of stimulation designed through the use of in situ stress determinations. *Soc. Petrol. Exp.* 10308, R Fall Ann. Mtg.

Voight, B. & St. Pierre, B. H. P. 1974. Stress history and rock stress. *Proc. 3rd Rock Mech. Congr. ISRM* **2**, 580–582.

Wilhemi, B. & Somerton, W. H. 1967. Simultaneous measurement of pore and elastic properties of rocks under triaxial conditions. *J. Soc. Petrol. Engng* (AIME) **7**, 283–294.

Wise, D. U. 1964. Microjointing in basement, Middle Rocky Mountains of Montana and Wyoming. *Bull. geol. Soc. Am.* **75**, 287–306.