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On the Use of Regional Joint Sets as Trajectories of Paleostress Fields During the Development of the Appalachian Plateau, New York

TERRY ENGELDER

Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964

PETER GEISER

Department of Geology and Geophysics, University of Connecticut, Storrs, Connecticut 06268

To compare the orientation and development of jointing with the orientation and magnitude of finite strain recorded in the Upper Devonian rocks of the Appalachian plateau, New York, we mapped systematic joint sets on an area of 20,000 km^2. In this area, *Wedel* [1932] mapped folds with limb dips of less than a degree and axes that change strike by 30° from 090° in the east to 060° in the west. We observed two different cross-strike joint sets that maintain their approximate cross-strike position from east to west. Yet, in detail the angle between the sets is 18° ± 2° in the east and 30° ± 4° in the west. In many outcrops one joint set parallels the direction of maximum compressive strain (ϵ_f) as recorded by deformed fossils, whereas the other joint set never parallels ϵ_f. Rare calcite-filled joints are oriented parallel to the direction of ϵ_f. In addition, the calcite-filled joints both cut and are cut by solution cleavage. These observations suggest that the joint set paralleling ϵ_f formed during the deformation event represented by the deformed fossils. The joint set that does not parallel ϵ_f somehow reflects a deformational event other than that producing fossil distortion, as suggested by strain relaxation experiments. In an outcrop of the Machias formation, where the direction of ϵ_f is 15° from the trace of a cross-strike joint set, the tensor average from 14 subsurface strain relaxation tests shows in situ maximum comprehensive strain (ϵ_{oc} = 10^{-4}) parallel to cross-strike joints but not the direction of ϵ_f. Strain relaxation is coaxial with a fabric anisotropy indicated by sonic velocity tests. Our idea is that the orientations of cross-strike joints parallel to ϵ_{nc} were controlled by the same rock property that causes strain relaxation on overcoring, whereas the orientations of the cross-strike joints parallel to ϵ_f were controlled by the stress field causing the fossil distortion. Placing these structural data in a regional context allowed us to construct a dynamic and kinematic model for the structural evolution of the New York plateau. The model indicates variation in both boundry conditions and material behavior through a series of four distinct deformational events, which begin prior to lithification and end in the Recent. Thus our analysis suggests that the structural features we have used represent a set of highly sensitive tools for investigating the deformational history of the Appalachian foreland.

INTRODUCTION

Vertical joint sets form patterns on scales from individual outcrops to regions the size of orogenic belts. (To clarify the jargon associated with cracks in the earth's crust, Appendix 1 defines some of the more commonly used terms.) The complexity of joint patterns varies from simple orthogonal sets in coastal plains to repeatedly overprinted sets in old basement of the craton [*Nickelsen*, 1976]. The joint pattern developed in the sedimentary rocks of foreland fold and thrust belts and midcontinents is somewhere between these extremes in complexity. North American examples of joint patterns developed across foreland fold and thrust belts include the Appalachian plateau [*Sheldon*, 1912; *Parker*, 1942; Nickelsen and Hough, 1967], the northwest side of the Ouachita Mountains [*Melton*, 1929] and the Rocky Mountain foothills [*Muecke and Charlesworth*, 1966]. Studies of midcontinent joint patterns include those of *Ver Steeg* [1942], *Babcock* [1973], and *Hodgson* [1961]. These joint patterns are cumulative and thus represent a history that can potentially be deciphered [*Nickelsen*, 1976]. One piece of information that seems within grasp is the sequence of stress fields accounting for present joint patterns. To obtain this information, one must know the direction of propagation of joints in a deviatoric stress field and be able to date the formation of joint sets relative to each other.

The purpose of this paper is to compare the regional joint pattern with the pattern for finite strain recorded on the Appalachian plateau, New York (Figure 1). From this comparison we infer a sequence to the development of various joint sets and thus a sequence of stress fields present during deformation of the Appalachian plateau. This paper also describes two experiments illustrating the relationship between the orientation of recoverable strain and jointing. The ultimate goal is to infer the boundary conditions during the propagation of joints within the Appalachian plateau.

The basic premise of our paper is that vertical joints propagate normal to the least principal stress and thus follow the trajectories of the stress field present at the time of propagation. This relationship between orientation of joint propagation and the stress field is empirical [*Stearns and Friedman*, 1972]. The field check of this relationship is in part based on the hydraulic fracture experiments of *Hubbert and Willis* [1957], who showed that hydraulic fractures form approximately perpendicular to the axis of least stress. Igneous dykes are believed to fill joints propagated by hydraulic fracturing during the forced intrusion of magma. The radial pattern of syenite dykes of the Spanish Peaks region follows horizontal trajectories that are the same as stress trajectories calculated based on known geological boundary conditions [*Ode*, 1957; *Muller and Pollard*, 1977]. Likewise, flank eruptions for polygenetic volcanoes in island arcs follow regional stress trajectories inferred from the direction of convergence of the two plates bounded by the island arc [*Nakamura et al.*, 1977].

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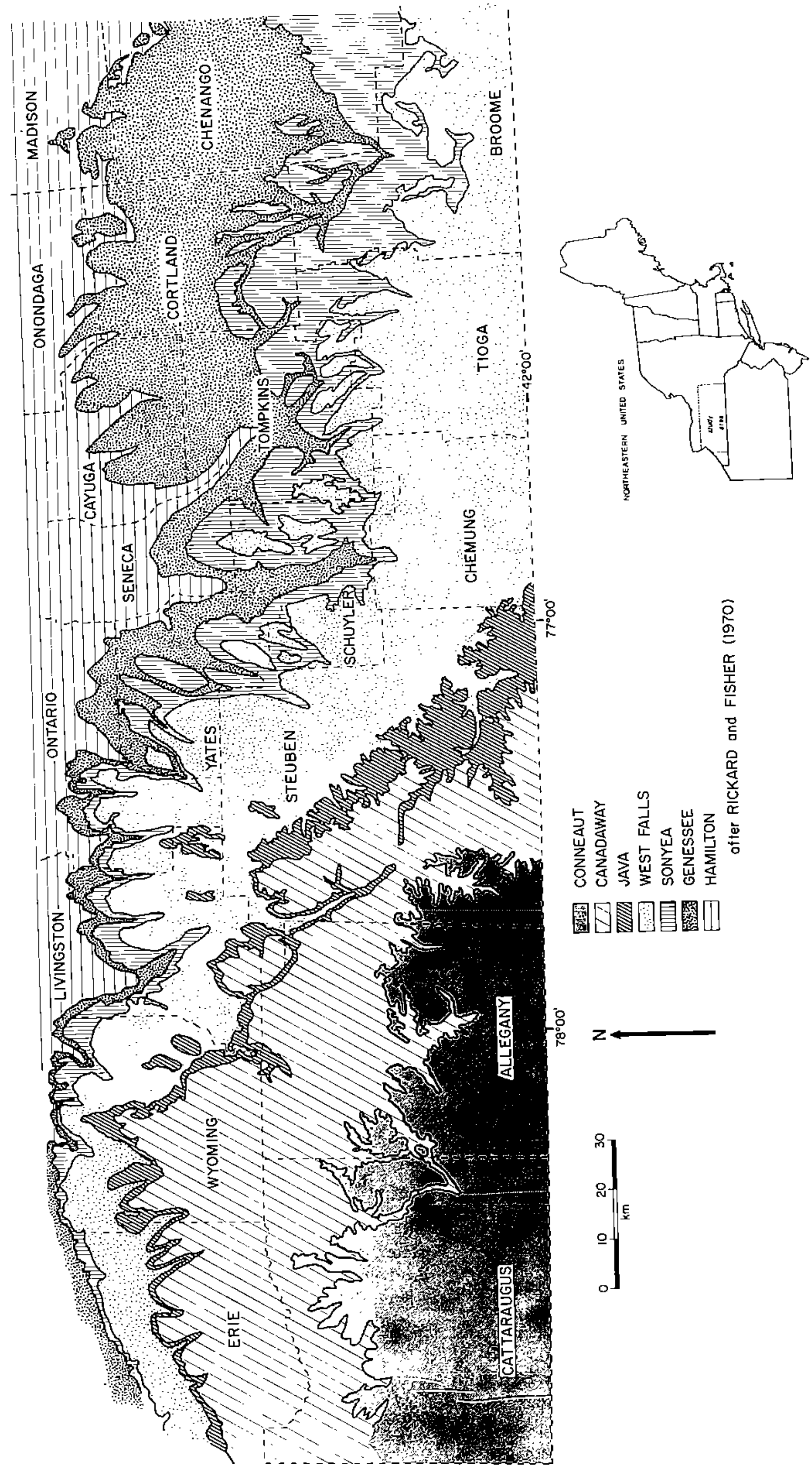
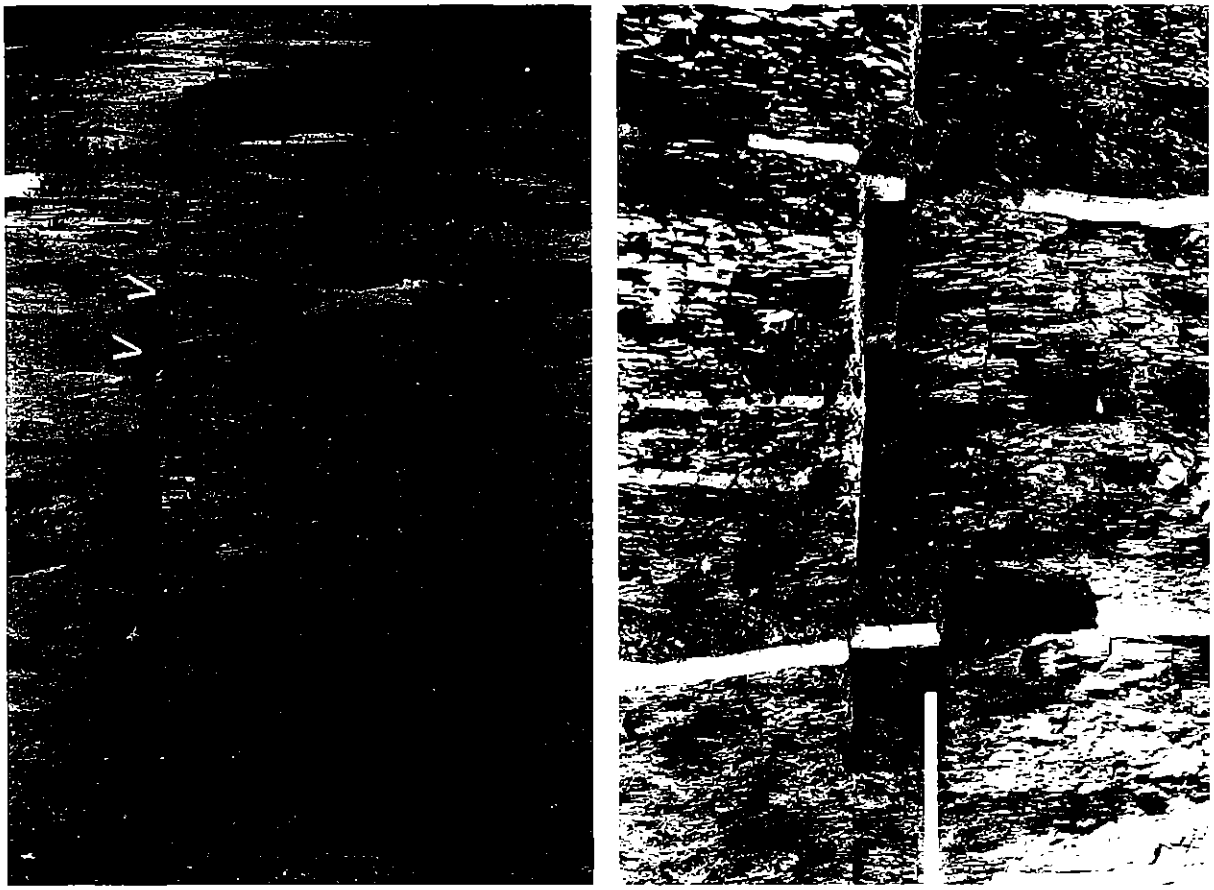


Fig. 1. The stratigraphy and geography of the southern tier of New York State.



A B

Fig. 2. (*a*) Set Ib joints in the Machias formation at Belmont, New York. Spacing is about 2 m. Note the steps in the outcrop trace of the joint set indicating a zone of slightly offset joints (arrows). (*b*) Set Ib joints in a vertical face of the Genessee Group at Letchworth State Park, New York. Joints in western New York are often restricted in a vertical direction by lithologic changes and thus have a horizontal dimension far greater than a vertical dimension. Camera perspective gives beds an apparent offset. A 15-cm-long rule is visible.

The caveat to our premise is that other phenomena may control the direction of fracture propagation under very low deviatoric stresses. Simulated hydraulic fracture tests in the laboratory show that at low stress the propagation direction is parallel to the plane of weakness [*Haimson and Avasthi,* 1975]. In point load tests fractures propagate parallel to the direction of maximum compressive residual strain [*Friedman and Logan,* 1970].

TECTONIC SETTING

The Appalachian plateau in western New York is an allochthonous sheet detached within the Silurian salts [*Rodgers,* 1963; *Prucha,* 1968]. Slip on this decollement is, in part, accommodated by a series of splay faults that cut the Devonian Oriskany sandstone and Onondaga limestone and end in the overlying shales and siltstones as blind thrusts [*Wedel,* 1932; *Bradley and Pepper,* 1938; *Gwinn,* 1964]. These splay faults are reflected at the surface by folds with limbs which dip less than 2°. Slightly deformed fossils throughout the Upper Devonian siltstones and shales indicate that a compressive strain accommodated as much as 20% layer-parallel shortening (ϵ_f) approximately normal to the fold axes [*Engelder and Engelder,* 1977]. The lower part of the allochthonous sheet shortened by a combination of splay faulting and strain, whereas the upper part of the sheet shortened primarily by layer-parallel strain, as indicated by fossil distortion and a minor component of folding. Mechanisms for fossil distortion include intragranular deformation by the mechanical twinning of calcite and pressure solution along solution cleavage planes [*Engelder,* 1979*a*]. Intragranular deformation accounts for 1–5% of the layer-parallel shortening, whereas pressure solution accounts for 4–18% of the layer-parallel shortening. Strain drops to less than 5% farther north where the throw on the splay faults decreases to a few meters. Here the rocks also contain a recoverable strain imposed by tectonic stresses during the development of the Appalachian foreland fold and thrust belt to the southeast [*Engelder,* 1979*b*]. This recoverable strain is manifested by residual strain which is relieved upon overcoring, and an elastic distortion of quartz grains which can be detected by X ray techniques.

The Appalachian plateau is most suitable for our study of the relationship between jointing and finite strain for several reasons. First, where exposed the major folds have limb dips of less than 2°. Present erosional levels indicate that these folds developed within one kilometer of the earth's surface. With these two facts we assume that one principal axis of the stress field was always vertical, as is the case at a free surface and thus the Upper Devonian rocks of western New York were subjected to horizontal principal stresses that were never more than 2° from parallel to bedding. Secondly, within the Upper Devonian rocks, finite strain parallel to bedding is well documented [*Engelder and Geiser,* 1979]. The orientation of

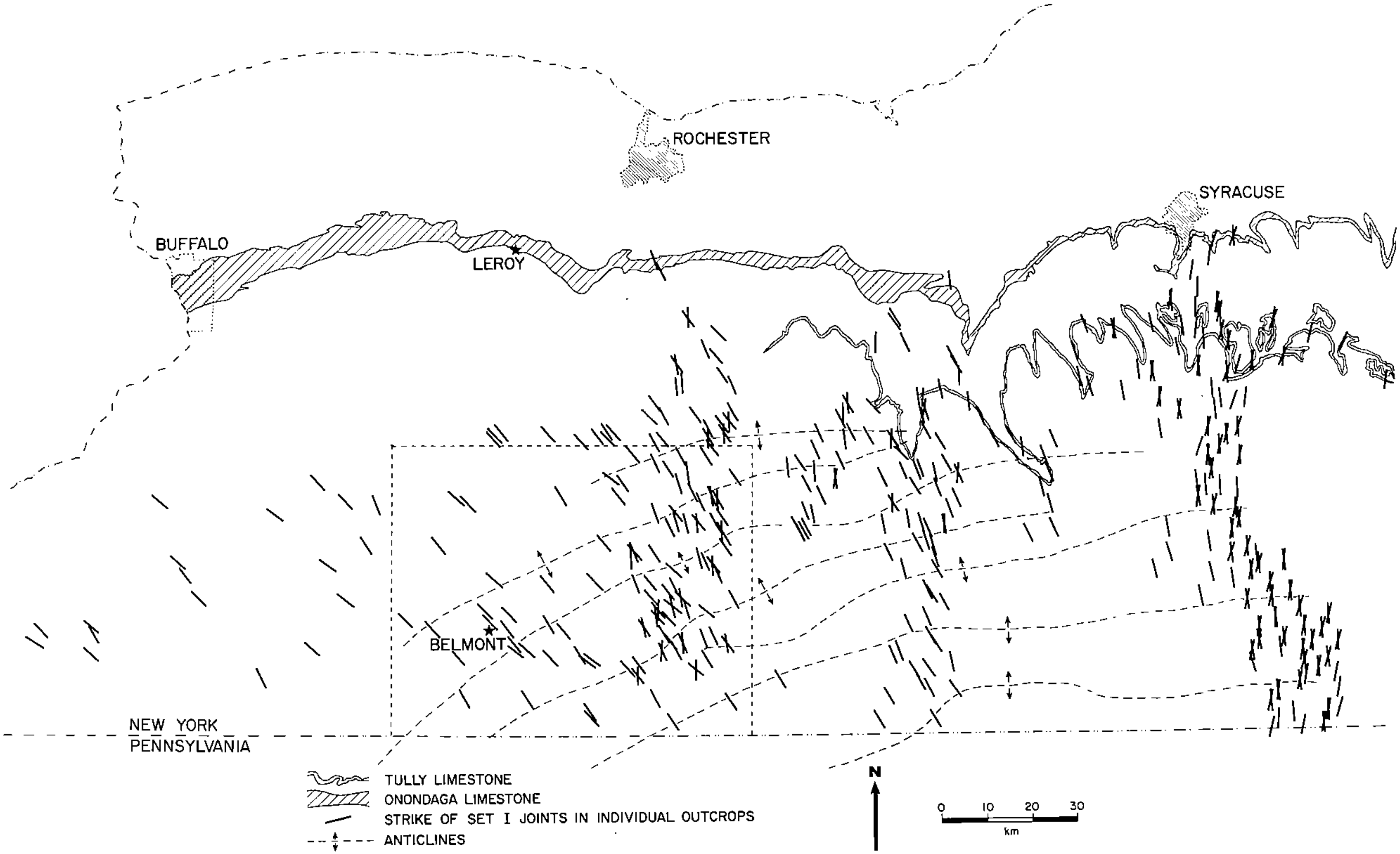


Fig. 3. A plot of the strike of cross-strike joints in the Middle and Upper Devonian rocks of the southern tier of New York State. Each datum represents an outcrop.

joints relative to the finite (ϵ_f) strain is unambiguous. Third, reasonably simple joint patterns allow regional variations in orientation of one particular joint set to be distinguished.

JOINTING WITHIN OUTCROPS OF THE APPALACHIAN PLATEAU

The Upper Devonian rocks of the Appalachian plateau consist of interbedded shales, siltstones, and massive fine grained sandstones. With the exception of massive sandstone beds 2 m thick, individual beds are less than 0.5 m thick. Joints in the siltstones and sandstones have a regular spacing (Figure 2*a*), whereas joints in shales tend to be less regularly spaced. In vertical sections joints often terminate at lithologic boundaries such as shale-siltstone interfaces (Figure 2*b*). Some beds contain as many as three joint sets whereas beds either below or above may contain only nonsystematic joints. The three predominant joint sets on the Appalachian plateau either cut across folds at high angles or strike subparallel to the fold axes. The nomenclature for these joint sets is given in Appendix 2.

Within individual outcrops we identified all systematic joint sets and nonsystematic joints and noted the nature of the joint surface as either planar or nonplanar. Joints in both subvertical faces and subhorizontal pavements were measured. Other measurements included the vertical and horizontal extent of individual joints and joint spacing both within zones and between zones. A description was made of the joint filling, joint surface, and degree of weathering. Features such as gouge and slickensides commonly associated with shear offset along reactivated joints would have been noted if present but were not.

For each joint set we measured the orientation of three to ten joint surfaces with a Brunton compass. A mean joint plane was determined for each joint set at a particular outcrop using the averaging procedure of *Cox and Doell* [1960]. With very few exceptions the standard deviation of the orientation of one set of joint traces within an outcrop was less than two degrees. Sources of this variation include the precision of a Brunton compass which is about $\pm 1°$ and non-planar joint surfaces which can vary considerably more than $\pm 2°$. When a non-planar joint surface was encountered the average trace on a bedding plane was measured.

Figure 3 shows the orientation of all cross-strike joints mapped across the southern tier of New York from Binghamton to Jamestown, a distance of 300 km. This joint pattern is similar to that mapped by *Sheldon* [1912] and *Parker* [1942] in an area that includes the eastern portion of Figure 3. In some areas two cross-strike joint sets (sets Ia and Ib) intersect at less than 30° with no obvious distinction between them other than their orientation. Besides orientation, the only observed distinction between set Ia and set Ib joints at the outcrop scale is that some set Ia joints in the Tully limestone are calcite filled whereas no calcite-filled set Ib joints have been observed. In Pennsylvania, *Nickelsen and Hough* [1967] also observed that cross-strike joints in the same outcrop could intersect at low angles.

In order to best visualize the regional variation of both cross-strike joint sets as well as to distinguish members of one set from the other in regions where the angular differences are small, a form line map was prepared for each joint set using the data in Figure 3. A form line map is a means by which the continuity of a given surface is best visualized by extending the strike line of the surface in a continuous curve from data point to data point [see *Ragan,* 1973, pp. 161–163]. By starting in the western portion of the map area where the angular differences are largest, it is possible to systematically differentiate between the cross-strike sets by working a single set of strike lines eastward. In addition, a form line map of the orientation of maximum compressional strain (ϵ_f) as indicated by solution cleavage, pencil cleavage, and fossil distortion [*Engelder and Geiser,* 1979] was also constructed. Superposition of the form line map for each cross-strike joint set on the form line map for ϵ_f (Figures 4*a* and 4*b*) revealed that the angular difference between the two sets changes from $18° \pm 2$ in the east to $30° \pm 4$ in the west and that joint set Ia is coincident with the orientation of ϵ_f in the east whereas set Ib diverges widely from the orientation of ϵ_f everywhere.

For a second test of a regional variation in joint orientation the outcrop data were divided into several groups by stratigraphy and geography. The outcrop pattern for the western 70% of Figure 1 shows the stratigraphic boundaries striking from NW to SE with the youngest units in the southwestern corner. These stratigraphic divisions are the several groups recognized in western New York by *Rickard and Fisher* [1970]. The geographic division is by county with Cattaraugus County as the westernmost and Broome County as the easternmost. Table 1 lists the mean orientations of both set Ib and set Ia for county and stratigraphic divisions. In general the orientations of both joint set Ib and joint set Ia rotate counterclockwise from east to west for both geographic divisions of a single stratigraphic groups and for adjacent stratigraphic groups. The counterclockwise rotation in joint orientation from east to west is consistent with the curvature of fold axes from east to west. However, both Figure 3 and Table 1 show that the acute angle between the trace of sets Ia and Ib increases from east to west.

As is also described by *Parker* [1942] and *Nickelsen and Hough* [1967], our data show that the set I joints do not change strike to conform with local variation in fold axes developed on the Appalachian plateau. Although there appears to be a rough correlation between the joints and folds (i.e., they both curve around the salient), in fact, the strike of individual joint sets is less variable than that of the associated folds, and there is no consistent orientation relative to the folds (Figure 4).

The angle between cross-strike (set I) and strike (set II) joints is rarely 90° where mapped by *Parker* [1942] and *Sheldon* [1912]. Figure 5 shows the relationship of set I and II joints mapped in Allegany and Steuben counties, New York. The normal to ϵ_f is also shown in Figure 5. Of 26 outcrops where the orientation of ϵ_f could be measured using deformed fossils, 16 outcrops contained set II joints whose strikes were consistently rotated counterclockwise more than 3° from the normal to ϵ_f. The other 10 outcrops included 5 with set II joints whose strikes were within 2° of the normal to ϵ_f, one outcrop with set II joints whose strikes were rotated clockwise more than 2° from the normal to ϵ_f and four outcrops in which set II joints were not found.

Table 1 lists the mean for the direction of ϵ_f from three counties. These data come from the Canadaway Group and may be compared with the orientations of set Ib joints from the Canadaway Group. The direction of ϵ_f rotates counterclockwise from east to west by 13°. This compares to a 12° rotation of set Ib joints over the same area. For the Canadaway Group in three counties the angle between ϵ_f and the strike of joint set Ib is a consistent 8° to 14° over an east to west

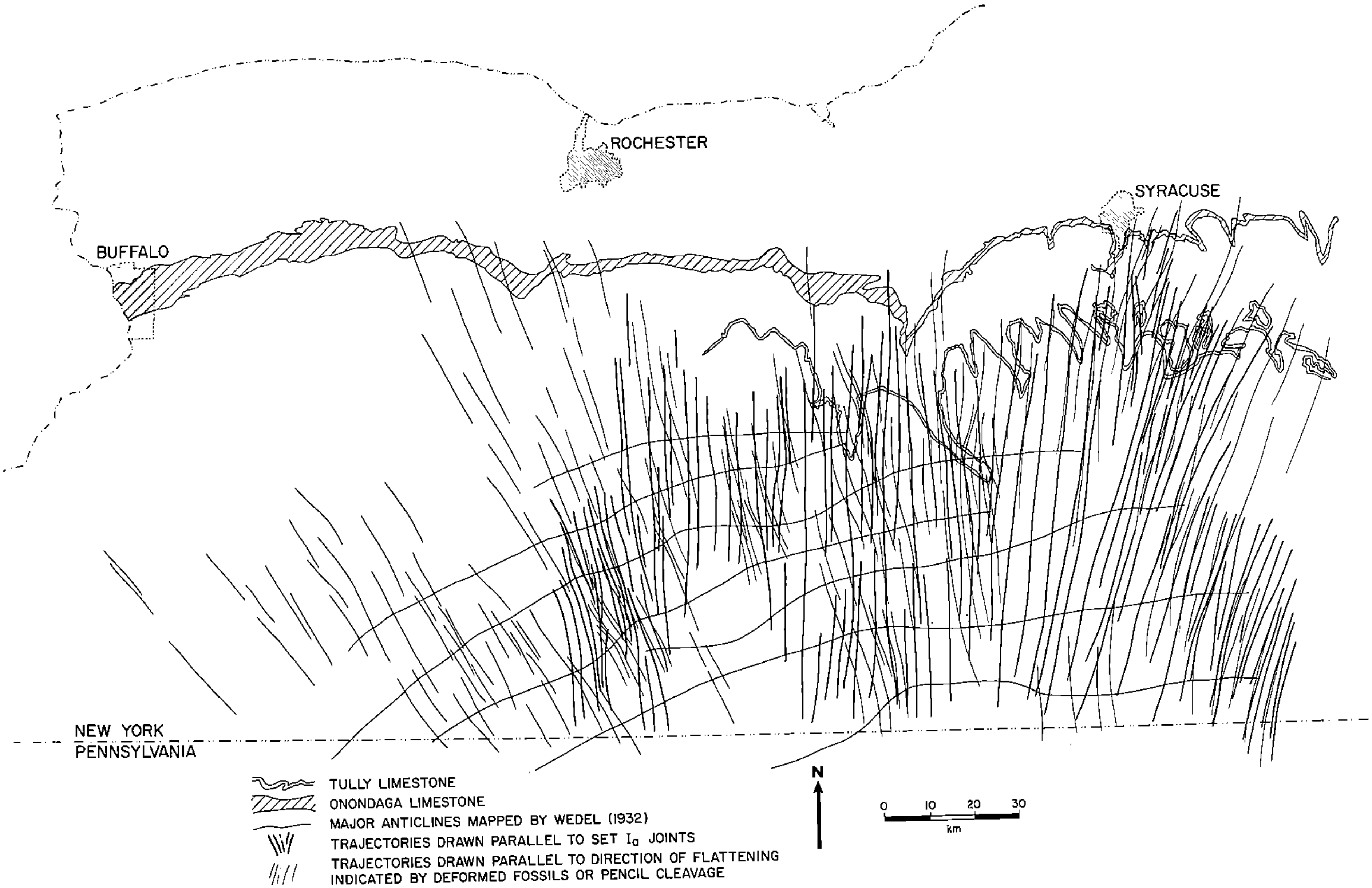


Fig. 4*a*. Plot of trajectories drawn parallel to the strike of set Ia joints (dark lines) and trajectories drawn parallel to the maximum horizontal compression (ϵ_f) indicated by deformed fossils and pencil cleavage (light lines) using data from *Engelder and Geiser* [1979].

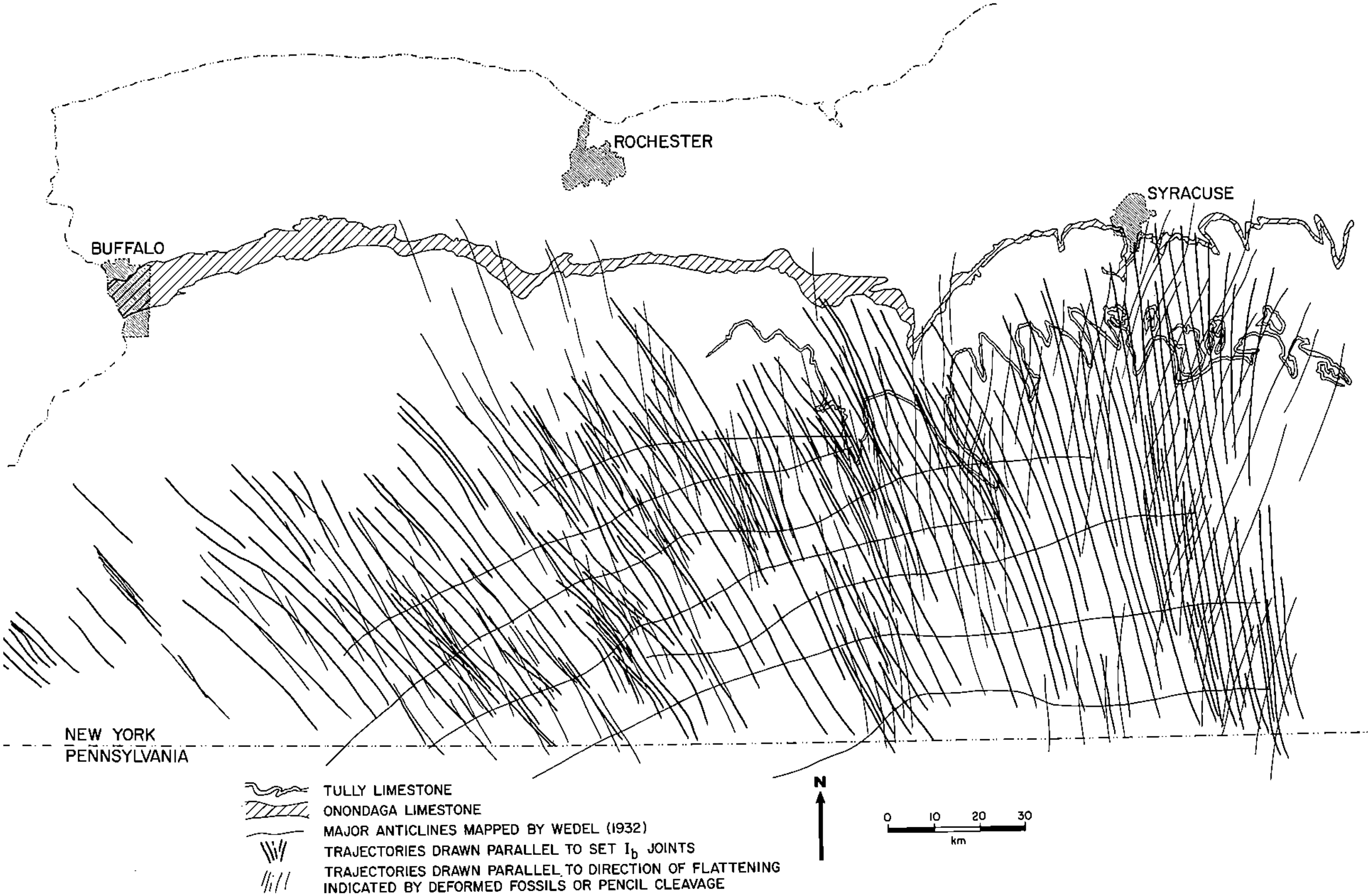


Fig. 4*b*. Plot of trajectories drawn parallel to the strike of set Ib joints (dark lines) and trajectories drawn parallel to the strike of set Ib joints (dark lines) and trajectories drawn parallel to the maximum horizontal compression (ϵ_f) indicated by deformed fossils and pencil cleavage (light lines) using data from *Engelder and Geiser* [1979].

TABLE 1. Orientation of Joints and Compression on the Appalachian Plateau, New York

		Set Ib			Set Ia					
Stratigraphic Group	County	Stations	Mean Strike	Standard Deviation	Stations	Mean Strike	Standard Deviation	Angle Ia to Ib	Angle Ib to II	Angle Ia to II
Conneaut	Cat.	8	316°	4°						92°
Conneaut	All.	12	319°	4°						98°
Canadaway	Cat.	9	312°	4°						104°
Canadaway	All.	12	317°	3°						93°
Canadaway	Ste.	23	323°	4°	9	350°	4°	27°	65°	94°
Java	Ste.	12	321°	5°	10	347°	5°	26°	71°	98°
West Falls	Liv.	10	322°	2°						104°
West Falls	Ste.	22	326°	4°	12	355°	5°	29°	76°	90°
West Falls	Ont.	13	326°	3°	10	005°	6°	39°	68°	98°
West Falls	Che.	14	332°	3°	5	006°	4°	34°	70°	104°
West Falls	Bro.	12	353°	2°	15	011°	2°	18°	88°	103°
Sonyea	Ont.-Ste.	7	333°	4°	7	353°	3°	20°	83°	104°
Sonyea	Sch.	7	336°	2°	4	356°	3°	20°	81°	101°
Sonyea	Cor.	7	354°	1°	4	012°	2°	18°	92°	107°
Sonyea	Bro.	23	356°	1°	20	013°	2°	17°	95°	110°
Genessee	Yat.	11	330°	5°	5	001°	3°	31°	72°	100°
Genessee	Sch.-Sen.	17	331°	6°	8	358°	5°	27°	72°	99°
Genessee	Tom.	11	341°	3°	2	007°	3°	26°	79°	99°
Genessee	Cay.	7	344°	4°						118°
Genessee	Cor.-Bro.	23	356°	1°	17	012°	3°	16°	94°	112°
Hamilton	Sen.	7	327°	4°	4	003°	7°	36°	75°	100°
Hamilton	Ono.	9	352°	3°	11	011°	2°	19°	91°	108°
					Hydraulic Fractures					
Canadaway	All.	8	063°	4°						
					Compression Indicated by Deformed Fossils					
Canadaway	Cat.	8	320°	3°						
Canadaway	All.	16	331°	4°						
Canadaway	Ste.	8	333°	3°						

Abbreviations: All., Allegany; Bro., Broome; Cat., Cattaraugus; Cay., Cayuga; Che., Chemong; Cor., Cortland; Liv., Livingston; Ono., Onondaga; Ont., Ontario; Sch., Schuyer; Sen., Seneca; Ste., Steuben; Tom., Tompkins; Yat., Yates.

change in orientation of 12° to 13° for both structural elements.

The joint pattern on the Appalachian plateau is further complicated by the presence of one joint set consistently oriented at about 060° and other joint sets unique to individual outcrops. The 060° joints are called set III joints over the region in which they can be distinguished from strike joints. A form line map was also drawn to show strike (set II) joints and to distinguish them from set III joints (Figure 6). For this map the strike lines of set II joints were started in the eastern portion of the map where angular differences between set II and set III joints were largest (see *Parker* [1942, Plate 4] for a map of set III joints). Also shown on Figure 6 are strike lines drawn parallel to solution cleavage, pencil cleavage, and the long axes of deformed fossils. Set II joints generally conform with changes in strike of the cleavage and folds but do not align perfectly with either. In the western portion of the map set II and set III joints are subparallel and cannot be distinguished. Once set II and set III joints merge to the same orientation, we arbitrarily identify the joints as set II.

The distribution of various joint sets is not uniform across the entire region. Sets Ib and II joints have the widest distribution as they are developed over the entire area. A western boundary to the distribution to set Ia joints is found at about the Allegany-Steuben County border; this distribution is not stratigraphically controlled as set Ia joints appear in all groups from the Conneaut down (Figure 1). West of the Steuben-Chemung County border, set II and set III joints cannot be distinguished.

CORRELATION OF IN SITU STRAIN AND JOINTING

Background

Further information on the process of jointing and hence the relationship between joint patterns and paleostress fields

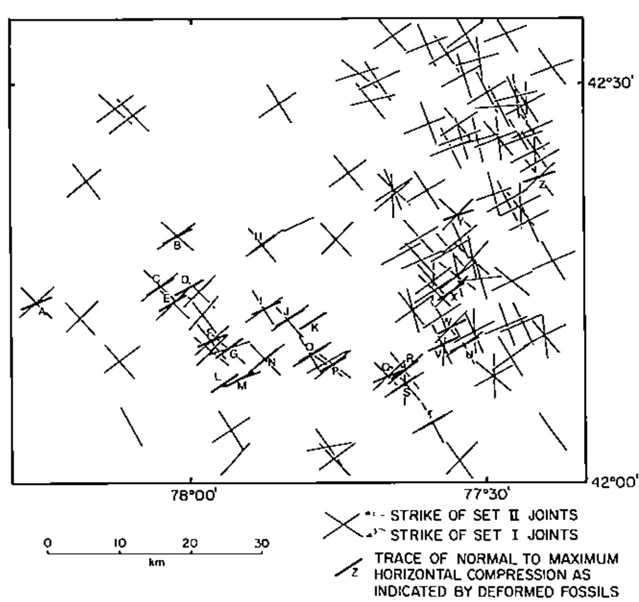


Fig. 5. A plot of the orientation of set Ia, set Ib, and set II joints in Allegany and Steuben counties. Also plotted are the normals to horizontal compression as indicated by deformed fossils (capital letters). The area of this figure is outlined in Figure 3.

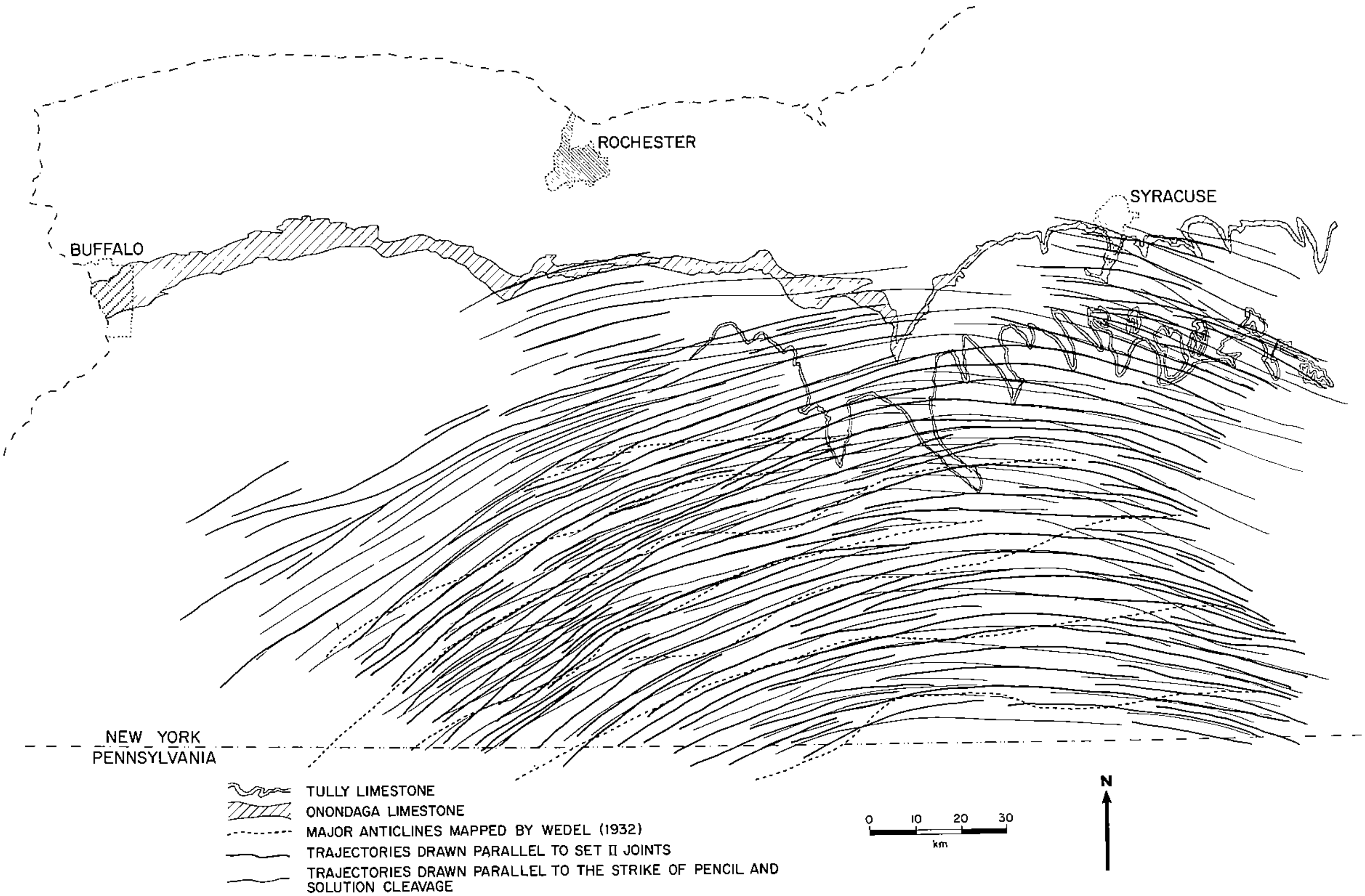


Fig. 6. Plot of trajectories drawn parallel to the strike of set II joints (dark lines) and trajectories drawn parallel to the strike of pencil cleavage, solution cleavage, and deformed fossils (light lines) using data from *Engelder and Geiser* [1979].

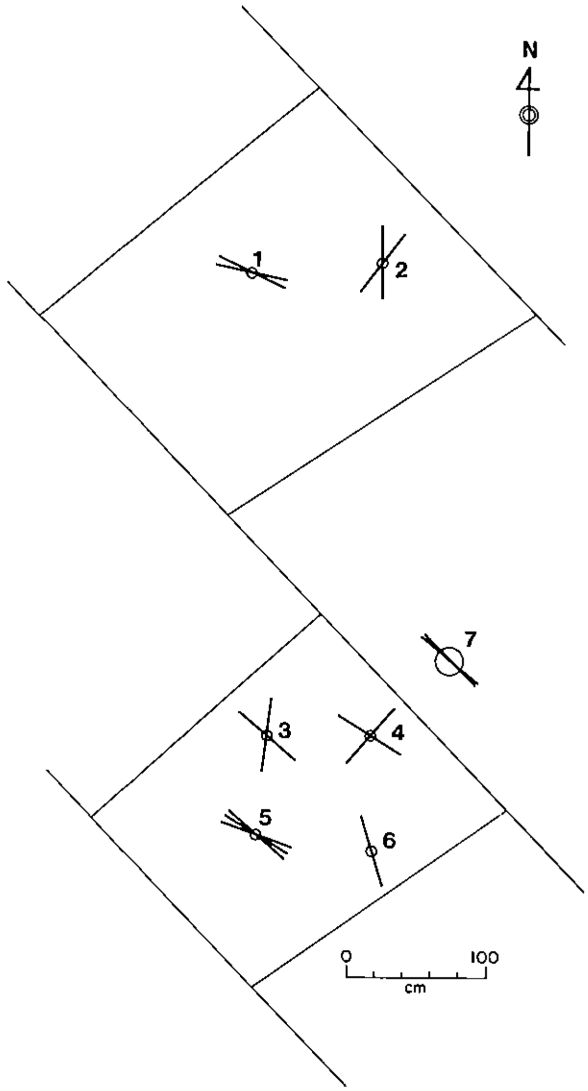


Fig. 7. Plot of the trace of joint zones and the orientation for maximum expansion (ϵ_{oc}) upon overcoring. The large circle represents the position of the U.S.B.M. borehole deformation gauge hole, whereas the small circles represent the position of six doorstopper holes.

can be gathered by measuring in situ strain using strain relaxation tests in jointed rock. Strain relaxation is any change in shape of a rock mass upon separation from a larger mass. This change in shape is referred to as a recoverable strain and is, in part, a measure of the elastic strain energy within the rock mass. Recoverable strain usually of the order of 10^{-3} to 10^{-4} is distinct from the nonrecoverable finite strain which may be on the order of 10^{-1} as indicated by various strain markers in rock. Relaxation may be either instantaneous or time dependent [*Nichols and Savage,* 1977]. Mechanisms for relaxation are poorly understood and can include (1) the recovery of elastic strains imposed by far field stresses [*Jaeger and Cook,* 1969], (2) the relaxation of residual elastic strains [*Friedman,* 1972], (3) the relaxation of thermal strains induced by near surface temperature gradients [*Hooker and Duvall,* 1971], and (4) the relaxation of strain induced during the chemical weathering of rock or even during the reaction of drilling fluids with the rock [*Savage,* 1979].

Two general relationships between in situ strain and joints are the following: (1) The magnitude of recoverable strain is inversely proportional to the density of jointing in the outcrop. The larger the volume of unjointed rock the larger the strain relaxation [*Swolfs et al.,* 1974; *Engelder and Sbar,* 1977; *Lee et al.,* 1979]. (2) Strain relaxation tests within a rock mass containing a single predominant joint set shows a maximum expansion subparallel to the strike of that joint set [*Preston,* 1968; *Eisbacher and Bielenstein,* 1971; *Brown,* 1974; *Engelder and Sbar,* 1977]. Rocks containing residual elastic strain have the tendency to fracture parallel to the maximum compression associated with that residual strain if subject to point loading normal to bedding during laboratory tests [*Friedman and Logan,* 1970]. In the field, in situ strain is often coaxial with jointing, although it is not clear whether a rock fabric or residual strain controls the orientation of joint propagation and subsequently strain relaxation, or whether jointing process imposes an in situ strain which in turn controls strain relaxation. In the laboratory, joint orientation and strain relaxation are clearly controlled by a rock fabric or residual strain.

The Cambrian Potsdam sandstone and many other Lower Paleozoic outcrops on the west side of the Adirondacks and along the shore of Lake Ontario, contain joints with the same orientation (ENE) as set III joints in the southern tier of New York. Set I and II joints are not found in these Cambrian and Ordovician rocks. Post-glacial pop-ups in the Potsdam sandstone are oriented so that they relieved an ENE maximum compressive stress. Upon overcoring foil-resistance strain gauges bonded to outcrops, a strain relaxation with an ENE maximum expansion was recorded at outcrops where the ENE joint set was most prominent [*Engelder and Sbar,* 1977].

Two tests of the relationship between in situ strain and jointing were made in the allochthonous sheet of the Appalachian plateau. The sites for these tests were chosen so that behavior of a unit in the lower part of the allochthonous sheet (shortening primarily by splay faulting) could be compared with the behavior of a unit in the upper part (layer-parallel shortening). The Onondaga limestone [*Dunbar,* 1959] occurs in the lower part of the allochthonous sheet, whereas the Machias formation [*Woodruff,* 1942] occurs in the upper part.

TABLE 2. Overcoring Data From the Machias Formation, Belmont, New York

Hole	Depth, cm	ε_{max}, $\mu\varepsilon$	ε_{min}, $\mu\varepsilon$	$\theta_{\varepsilon_{max}}$
		*U.S.B.M. Borehole Deformation Gauge**		
7	5	141	−80	309°
7	8	321	−194	313°
		Doorstopper		
6	13	196	69	340°
2	15	356	36	358°
1	15	−48	−63	284°
5	17	112	81	289°
3	20	170	123	006°
1	30	−187	−466	300°
2	45	62	−25	038°
3	53	472	268	309°
4	71	295	15	300°
5	114	126	82	300°
4	114	166	59	041°
5	167	258	91	309°
Σ	Doorstopper	126	63	316°
Σ	U.S.B.M.	79	−47	312°

*The U.S.B.M. gauge measures the deformation of a borehole rather than true rock strain.

TABLE 3. Mechanical Tests of the Machias Formation

Pressure, MPa	ϵ_{max}, $\mu\epsilon$	ϵ_{min}, $\mu\epsilon$	$\theta_{\epsilon_{min}}$	Hole	Depth, cm
		U.S.B.M. Compression Test			
2	337	151	072°	7	8
		Doorstopper Compression Test			
2	112	100	015°	6	13
		Velocity Anisotropy			
	P_{max}, km/s	P_{min}	$\theta_{P_{max}}$	Hole	Depth, cm
	4.55	4.28	047°	5	17
	4.39	4.06	047°	3	53

Machias Formation

The Machias formation of the Canadaway Group is an Upper Devonian interbedded shale and siltstone with deformed fossils throughout and both cross-strike (set Ib) and strike (set II) joints. An outcrop of the Machias formation located in the Genessee riverbed at Belmont, New York, was selected because an area about 500 m^2 of relatively massive siltstone is exposed (Figure 3; site E in Figure 5). Three joint sets are evident with the most prominent being the set Ib joints oriented at 315°, spaced at about 2 m, and extending the full 50 m of outcrop (Figure 2*a*). A local joint set unrelated to set I, II, or III extends through the outcrop, striking 085° and spaced at about 6-m intervals. Set II cross joints strike about 045°–050° with a 2-m spacing. The massive siltstone contains fossils indicative of the Late Paleozoic compression oriented at 330° (ϵ_f) [*Engelder and Engelder,* 1977]. The particular advantage of this outcrop is that the direction of ϵ_f differs by 15° from the strike of the set Ib joints. If there is a relationship between in situ strain and the jointing process, that relationship might be distinguished from the relaxation of elastic residual strains parallel to ϵ_f.

Subsurface strain relaxation measurements were accomplished using both the doorstopper [*Sbar et al.,* 1979] and U.S.B.M. borehole deformation gauge techniques [*Hooker and Bickel,* 1974]. Six holes were drilled for the doorstopper and two for the U.S.B.M. gauge (Figure 7). The position of the holes was selected to measure strain relaxation at different positions relative to the joints that trace the boundaries of the jointed siltstone blocks.

A major difficulty was that siltstone of the Machias formation split parallel to bedding during drilling of 15.2-cm-diameter cores for the U.S.B.M. borehole gague measurements. Several attempts were made before two complete overcoring curves were recorded. Later use of the compression chamber to obtain the modulus of the rock was possible only by epoxying several pieces of core together. Splitting was not a problem during use of a 7.6-cm bit for overcoring the doorstoppers.

Table 2 shows that the data from the strain relaxation measurements have a large amount of scatter in both orientation and magnitude. The two U.S.B.M. borehole deformation gauge measurements showed consistent orientations but strain magnitudes that differed by 50%. The doorstopper data were far less consistent although some individual holes gave both repeatable orientations and signs of strain upon relaxations. For example, both cores from hole 1 contracted upon overcoring, and both showed a least contraction oriented within 16° of each other. Three cores from hole 5 all expanded with maximum expansion within 20° of each other. No obvious relationship was seen between either magnitude or orientation of strain relaxation and either position of the hole relative to the joints or depth of the measurement. The contraction of both samples in hole 1 indicates that hole 1 was located in a small volume of rock subject to small tensile stresses. The other contraction was measured in the same block in hole 2. These data may indicate some inherent difference in in situ strain between the block containing holes 1 and 2 and the block containing holes 3, 4, 5, and 6.

The tensor average of the doorstoppers gives a maximum expansion with a strike of 316°. In terms of a variation in orientation, the doorstopper data gave a mean strike of 314.6° and a standard deviation of 12.48°. The tensor average for the U.S.B.M. borehole deformation gauge gives a maximum expansion with a strike of 312°. Although there appears to be a great deal of scatter in the direction of ϵ_{max}, the data do not come from a uniform distribution. Assuming that the directional data fit a von Mises distribution, Rayleigh's test shows that the null hypothesis of a uniform distribution for orientation of ϵ_{max} is rejected at the 10% level [*Mardia,* 1972]. From this test we infer that there is no reason to reject data from holes where there was a great difference in orientation of ϵ_{max}.

The ϵ_{max} calculated using a tensor superposition is oriented within a degree of the strike of set Ib joints in the Machias formation. Because of the correlation in orientation of the tensor

TABLE 4. Overcoring Data From Onondaga Limestone, Le Roy, New York

Depth, m	ϵ_{max} $\mu\epsilon$	ϵ_{min}, $\mu\epsilon$	$\theta_{\epsilon_{max}}$
	Doorstopper		
0.61	258	65	301°
0.74	139	85	352°
0.79	96	−50	335°
0.93	426	59	024°*
	Surface Strain Gauges		
0.00	123	−7	355°
0.00	75	−9	332°
0.00	121	−3	338°
	U.S.B.M. Borehole Deformation Gauge†		
0.51	1295	466	294°
0.76	2495	1248	277°
1.02	1029	106	277°

*Strain gauge partially bonded to a chert nodule.

†The U.S.B.M. gauge measures the deformation of a borehole rather than true rock strain.

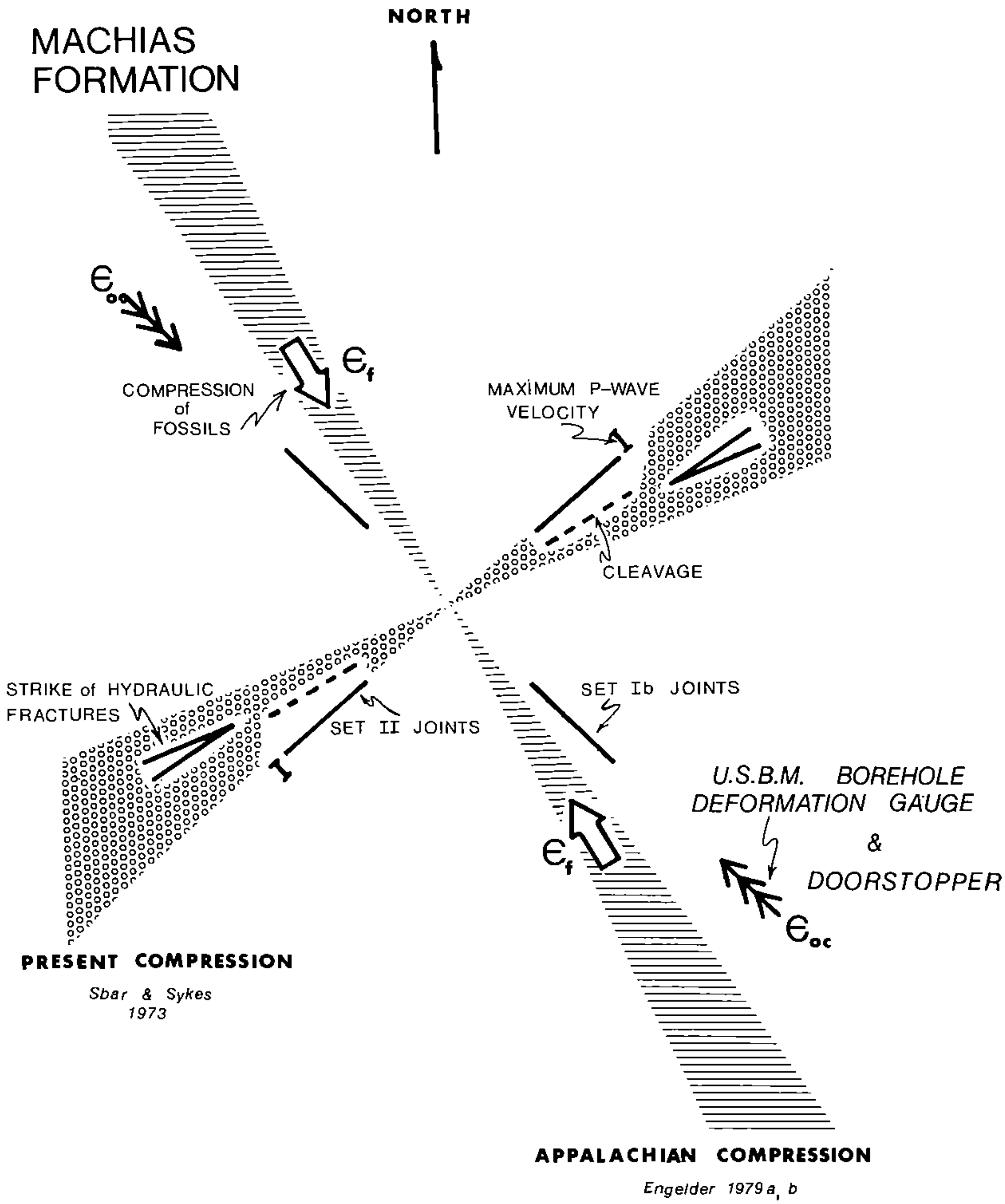


Fig. 8*a*. Summary plot showing orientations of joints, solution cleavage, maximum compressional strain by fossil distortion (ϵ_f), and maximum expansion on overcoring (ϵ_{max}) in the Machias formation at Belmont, New York, and hydraulic fracture orientation in Richburg sandstone of the Canadaway Group [*Overbey and Rough*, 1968]. Also shown is the general orientation of the Appalachian compression affecting Allegany County, New York, and the present compression affecting much of eastern North America.

average with set Ib joints and despite the scatter in orientation of ϵ_{max}, we infer that set Ib joints and the in situ strain are genetically related. Likewise, we infer that the in situ strain is not genetically related to ϵ_f. Of 14 strain relaxation measurements, only one showed an ϵ_{max} within 15° of ϵ_f, whereas half showed an ϵ_{max} within 15° of the strike of the set Ib joints.

Compressional-wave velocities as measured on cores in the laboratory indicate a velocity anisotropy in the siltstone (Table 3). Tests on two different cores show that the core is most compliant parallel to the major outcrop joint set and the direction of maximum expansion as determined by tensor average (~315°). Compression tests on the 15.2-cm core with a U.S.B.M. borehole deformation gauge inserted show the compliant direction at 342°. This core was epoxied together because it split parallel to bedding. Compression tests on a doorstopper core show the compliant direction at 285°. In the case

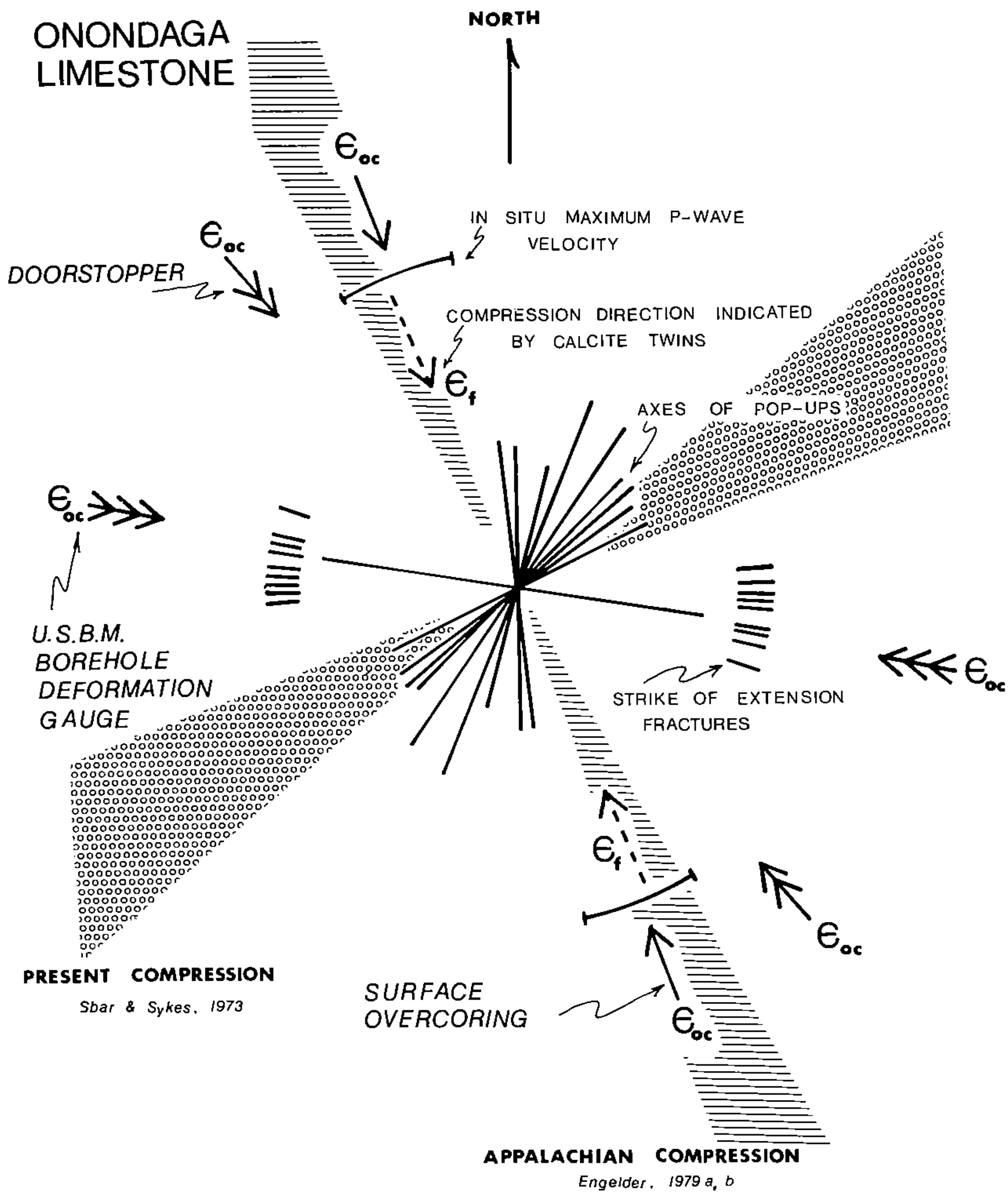


Fig. 8*b*. Summary plot of joint orientation, axes of pop-ups, and orientation of maximum expansion upon overcoring (ϵ_{max}) in the Onondaga limestone at LeRoy, New York. Also shown is the general orientation of the Appalachian compression affecting Genessee County, New York, and the present compression affecting much of eastern North America.

of the doorstopper core the ratio of ϵ_{max} to ϵ_{min} was so small that the orientation of the compliant direction is poorly constrained. The mechanism responsible for this strength anisotropy in the siltstone may be related to a mica fabric. G. Oertel (personal communication, 1979) has examined the preferred orientation of mica at three sites including C and F on Figure 5 and finds that in the horizontal plane the normal to the preferred orientation of mica is 8° to 18° counterclockwise from the direction of compression indicated by deformed fossils and penciling. This counterclockwise relationship is the same as the orientation of the velocity anisotropy relative to the finite strain indicated by fossil deformation at site E (Figure 5).

Maximum expansion on overcoring is parallel to the compliant direction of the Machias formation, yet data suggest that the recoverable strain was not a relaxation against a uniform radial load caused by mechanisms such as annual thermal stresses or drilling conditions. This is shown by mechanical tests in which a uniform radial load was reapplied to the samples. For both the U.S.B.M. core and the doorstopper cores the ratio $\epsilon_{max}/\epsilon_{min}$ was much different on overcoring than on application of a uniform radial load. In both cases the presence of a maximum principal stress parallel to the compliant direction was necessary to achieve the strain relaxation measured in situ.

The Onondaga Limestone

For our study of the relationship between jointing and in situ strain we chose to make surface strain gauge [*Engelder and Sbar,* 1976], doorstopper [*Sbar et al.,* 1979], and U.S.B.M. [*Hooker and Bickel,* 1974] overcoring measurements in the Onondaga limestone at the Le Roy Crushed Stone Quarry at Le Roy, New York (Figure 3). The quarry floor of Onondaga limestone contains many joints and post excavation pop-ups. However, the joints are related to neither set I nor II joints found in the Upper Devonian rocks to the south. The pop-up axes vary in orientation by as much as 30° with a general N-S trend. At the site chosen for strain relaxation measurements the pop-ups are oriented in response to a maximum principal stress striking about 280°, which is also the strike of the major joints in the floor of the quarry. The advantage of working this area of the quarry floor is that when compared with the outcrop of Machias formation the angle between joints and the direction of the Late Paleozoic compression is larger for the Onondaga limestone. Residual strain oriented in the direction of ϵ_f in the Onondaga limestone is most likely to be manifest by a ϵ_{max} upon relaxation striking at 340° whereas a measurement of stresses associated with the generation of the pop-up and jointing will be an ϵ_{max} striking at 280°.

Results from surface overcoring, doorstopper, and U.S.B.M. measurements are given in Table 4 and summarized in Figure 8*b*. Briefly, ϵ_{max} on overcoring surface strain gauges was oriented parallel to the direction of ϵ_f, whereas ϵ_{max} upon overcoring the U.S.B.M. borehole deformation gauge was parallel to the joints and to the compression causing the post-excavation pop-ups. Relaxation accompanying the doorstopper tests was ambiguous because of the scatter in orientation of ϵ_{max}. By excluding the data for the strain gauge partially bonded to the chert nodule, averaging the measurements gives an orientation of ϵ_{max} that falls within the acute angle between ϵ_{max} of the U.S.B.M. gauge and the surface strain gauges.

The point here is that jointing did not totally relieve subsurface stresses responsible for controlling the orientation of the joints. However, these stresses may be low enough so that a strain relaxation at the surface as measured by strain gauges bonded to the surface is controlled by a rock fabric or residual stress. The in situ velocity anisotropy as measured by a Bison seismograph over 30 m has a maximum velocity parallel to the rock fabric or residual stress [*Engelder and Sbar,* 1979]. The difference in strain relaxation between the rock surface and points within the rock has been observed elsewhere. The best documented example is Barre granite, where surface strain relaxation measurements by *Engelder et al.* [1977] and *Swolfs* [1977] show a maximum expansion normal to the rift plane, whereas subsurface measurements by *Nichols et al.* [1977] show a maximum expansion controlled by boundary loads.

Comparison of Two Sites

Neither set Ib nor II joints propagated in the Onondaga limestone; yet, deformed calcite clearly indicates that the Late Paleozoic compression of the Appalachian plateau was transmitted as far NNW as LeRoy, New York. In the Machias formation at Belmont the propagation of set Ib and II joints appears to have been controlled by a stress field whose trajectories were 15° from the principal axes of ϵ_f. At both sites the maximum expansion upon overcoring the U.S.B.M. gauge was parallel to the most prominent joint set (Figure 8). In the Onondaga limestone the most prominent joint set was the only joint set, whereas in the Machias the set Ib joints were most prominent by virtue of being spaced at 2 m and extending the 50 m length of the outcrop. The joints in the Onondaga limestone most certainly propagated in the same stress field that was responsible for the pop-ups upon removal of overburden. The nature of this stress field is unknown but may be related to one of several possible sets of boundary conditions present during quarrying. Set Ib joints in the Machias formation occur throughout the Upper Devonian section of the Appalachian plateau in New York. Upon overcoring doorstoppers in the Machias formation, the tensor-average maximum expansion was parallel to set Ib joints, whereas maximum expansion recorded by doorstoppers bonded to Onondaga limestone was not parallel to the prominent joint set. Maximum expansion at the surface of the Onondaga limestone was parallel to the late Paleozoic compression.

DISCUSSION

Stress Trajectories and Timing of Jointing

The complicated structural patterns in the core zones of all mountain ranges represent irrefutable evidence that stress fields constantly change throughout the evolution of mountain ranges. In contrast, the simpler structural patterns of the forelands, of which the Appalachian plateau is an example, are the result of a less complicated stress history and yet may represent key events during which major pieces of crust collided, shifted or were consumed in the vicinity of the core zone of the mountain range. In other words, the foreland may contain a more useful record than the core zone for identifying such things as the direction of plate convergence during either crustal collision or subduction and the shape of plate boundaries during collision. In order to test this notion for the Appalachian plateau, our first step is to identify regional stress patterns. The next step, which is currently in progress, is to relate these stress patterns to processes driving deformation toward the foreland from the core zone of the Appalachian Mountains.

The relationship between finite inhomogeneous strain and a stress field is unclear, particularly if it is suspected that the orientation of the stress field was changing with time. Joints are more likely to indicate the orientation of a stress field at one instant in geologic time [*Bankwitz,* 1966]. However, the rate at which joints propagate is unknown; *Bahat* [1979] suggests that a joint surface with extensive hackling suggests rapid fracture, whereas plume structures represent slow crack propagation under either static or dynamic load. Our interpretation of the Appalachian plateau assumes that on the outcrop scale, even the slow propagation of cracks is fast enough to represent one instant in geologic time. However, before using joints to infer regional stress patterns, we must determine whether trajectories drawn parallel to the strike of joint sets represent a regional stress field or a rock fabric merely controlling the orientation of regional joint sets. That the latter is possible is suggested by *Nelson and Stearns* [1977] for regional joints on the Colorado plateau. Likewise, we must determine the sequence of development of the joints and finite strain. For this determination, much may be inferred from data on finite strain, joint orientation, and strain relaxation. The following are points for consideration.

1. We recognize two rock fabrics associated with the late Paleozoic compression. One is a spaced solution cleavage accompanying fossil deformation and the other is a penetrative fabric attributed to the preferred orientation of mica and manifest as a velocity anisotropy. Although we have documented this penetrative fabric at just one outcrop (site E, Figure 5), G. Oertel (personal communication, 1979) indicates that the mica fabric may be extensive in the western portion of Figure 1. Although direct evidence that the penetrative fabric formed prior to the spaced cleavage is lacking, the mechanical rotation of the mica during soft sediment deformation seems a likely mechanism. Such a deformation may not affect the fossils which acted as rigid inclusions in a deforming matrix. Once lithified the matrix could exert enough of a load on the fossils to promote deformation. If the mica fabric formed in a lithified sediment after fossil distortion, the event did not leave a visible trace on either the solution cleavage or the deformed fossils. The likelihood of this happening seems remote. The mica fabric is not coaxial with the compressional strain indicated by the spaced solution cleavage. The normal to the mica fabric correlates with the orientation of set Ib joints where documented in Figure 5 and strain relaxation in the Machias formation, whereas set Ia joints parallel the compression indicated by deformed fossils and cleavage development in the eastern portion of Figure 4*a*. The only evidence we have for the contemporaneity of jointing with the development of either rock fabric is that set Ia joints appear to be opening to permit precipitation of calcite at the same time that clay selvages were forming. This is seen in thin section where some calcite-filled joints cut clay selvages, whereas in other places in the same thin section clay selvages cut calcite-filled joints. It must be indicated, however, that the Ia joints may still predate the solution cleavage. The presence of pre-cleavage joint sets has been documented by *Dean and Kulander* [1977], *Nickelsen* [1979], and *Sansone* [1979].

2. Outcrops containing both set Ia and set Ib joints show no direct evidence for the sequence of formation because cutting relationships are ambiguous. Evidence supporting an early age for the Ia joints relative to Ib joints is the scattered calcite veins found at eight localities, seven in the Tully limestone and one in the Onondaga limestone. All the veins in the Tully parallel the Ia set. The single exception in the Onondaga is associated with a local shear zone. In thin section other units have calcite filled microjoints parallel to set Ia joints. Our idea is the same as *Charlesworth*'s [1968], who observed that when joints are associated with veins, the older joint set should have the higher percentage of veins. Although we recognize pitfalls with this association, we use this interpretation to suggest that set Ia joints formed prior to set Ib joints.

3. Dissolution along solution cleavage planes accounts for 4–18% layer-parallel shortening on the Appalachian plateau [*Engelder*, 1979*a*]. If strain of the entire Upper Devonian section were volume constant, the 4–18% of dissolved rock would be reprecipitated someplace in the section. Although there still remains the possibility that the dissolved rock could be deposited as a cement, we presently see no evidence to support this. Early-formed joints might be a major sink into which chemically charged fluid could migrate and reprecipitate dissolved rock. Early-formed joints may also be conduits [*Geiser and Sansone*, 1980] through which 4–18% of the rock could pass while in solution. In this latter case some reprecipitation may occur but the final volume of filled joints would be much less than the volume of rock removed in solution. Because calcite- and quartz-filled joints are rare in all units except the Tully limestone, we conclude that strain by layer-parallel shortening was not volume constant, and that early-formed joints served as conduits rather than sinks.

4. A mechanism for early or deep formation of joints is by hydraulic fracturing as the result of pore pressure buildup during compaction or lateral compression [*Secor*, 1965]. The only evidence that hydraulic fracturing may have had a role during the formation of set Ia joints is the rare occurrence of calcite-filled joints (set Ia) in the Tully limestone and microjoints in other units. The generation of two sets of joints at an angle of less than 30° by hydraulic fracture seems unlikely because the formation of the initial set Ia joints would have acted to drain off fluid, relieve excess pore pressures, and leave an open pore water system prior to rotation of the stress field. Set Ib and Ia joints are equally common in shales which therefore could not have acted as an impermeable layer for maintaining abnormal pore pressure necesary for the formation of a later joint set by hydraulic fracture.

5. Plumose structures, as opposed to slickensides, on both set Ib joints and set Ia joints leave little doubt that the joints formed normal to σ_3 as extension fractures rather than shear fractures [*Bahat*, 1979]. Therefore outcrops containing both set Ib and set Ia must be interpreted as evidence for the rotation of a stress field by as much as 30° in some cases. Such a rotation will result in a high shear stress parallel to the early formed set Ia joints. The lack of slickensides or other evidence for shear displacement on set Ia joints suggests that the shear stress following rotation was maintained at a low level. The presence of extension fracturing without shear fracturing suggests that net tensile stresses were present during jointing. Net tensile stresses may be achieved along several loading paths, all of which require that low deviatoric stresses be maintained. Loading paths to produce net tensile stresses include the addition of abnormal pore pressure [*Secor*, 1965], and thermal cooling, plus lateral contraction upon the removal of overburden [*Price*, 1966]. An effect of the latter is commonly observed in deep road cuts where there is an increase in jointing near the surface [*Wise et al.*, 1980]. Because of the lack of calcite filling, we infer that set Ib joints formed during contraction which accompanied removal of the overburden and cooling. Thus, although parallel to an early fabric, set Ib joints propagated very late in the history of Appalachian tectonics.

6. It is highly unlikely that set Ib joints propagated prior to the development of a penetrative fabric which may be the result of soft sediment deformation. Very delicate surface structures preserved on the surfaces of set Ib joints are associated with brittle fracture of indurated rock and, even if present after the rupture of a partially indurated sediment, would have been destroyed during the penetrative deformation causing the development of a fabric. The likely time for set Ib jointing is during contraction caused by removal of overburden and by thermal cooling. Because tectonic stresses may become small during contraction, joint propagation may be controlled by a rock fabric. The present fabric in the form of an in situ strain in the Machias formation favors the propagation of joints parallel to set Ib joint direction.

7. *Lachenbruch* [1961] calculated that in naturally occurring, rapidly evolving joint systems, the joint depth should be the same order of magnitude or less than the joint spacing.

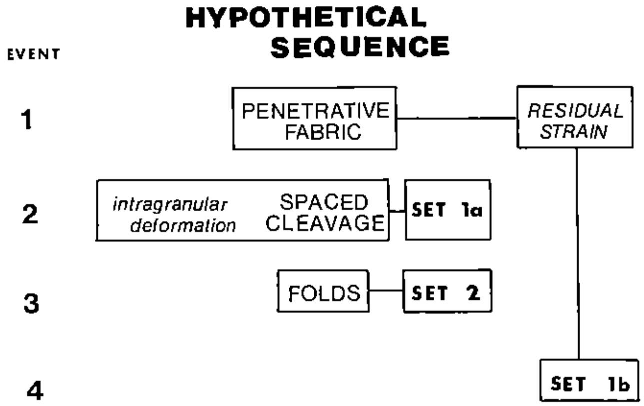


Fig. 9. A hypothetical deformation sequence for events affecting the Appalachian plateau.

Many joints on the Appalachian plateau seem to extend tens of meters vertically while having a spacing of a meter or less. Although this seems contrary to *Lachenbruch*'s [1961] concept of joint spacing, close examination reveals that many joints are really zones containing joints that extend a meter or less vertically before another joint starts almost in line with the first (Figure 2). This is also seen on joint faces were many plumose structures are restricted by shale layers to less than one meter in depth. However, there is no restriction on horizontal propagation of plumose structures indicative of single joints and some extend more than 5 meters horizontally. Many outcrop faces show plumes with length to depth ratios of 10:1 or greater. The control of bedding thickness on the vertical extent of joints is truly reflected by the joint spacing of less than a meter. This also illustrates the weak coupling across the shale interbeds [*Sowers*, 1972]. If there were no coupling joints would not form in zones but as isolated joints appearing without regard to joints in beds above and below.

8. The origin of set II joints presents a problem for which our observations provide no direct answer. Like *Muecke and Charlesworth* [1966] and many others who have dealt with multiple jointing, we must appeal to a rotation of σ_3 by 70° to 90° sometime during the history of the rock. The most likely time is during the development of folds while the upper beds are above a neutral fiber. This would place the propagation of set II joints between the early propagation of set Ia and late propagation of set Ib joints. In this scheme the abutting of one joint set against another should aid in sorting out the order of development. Younger joints abut against older ones. Unfortunately a preliminary analysis shows ambiguous relationships.

9. On the Appalachian plateau slip on detachments was accommodated by shortening of the allochthonous sheet. There is a profound difference between the mechanism of shortening of the lower and upper parts of the sheet as represented by the massive Onondaga limestone and thinly bedded Machias formation, respectively. The limestone shortened 1–2%, as indicated by intragranular deformation of calcite [*Engelder*, 1979*b*] and a great deal more by duplication of section on splay faults [*Gwinn*, 1964]. The overlying shales and siltstones shortened 1–2% by folding over blind thrusts [*Rodgers*, 1963] and a great deal more by penetrative strain [*Engelder*, 1979*a*]. The cross-strike jointing that accompanied shortening in the upper part of the allochthonous sheet is not as common in the Onondaga limestone. This suggests that in the more deeply buried portion of the allochthonous sheet, the net σ_3 rarely dropped below the tensile strength of the limestone, whereas the net σ_3 commonly dropped below the tensile strength of the upper part of the allochthonous sheet.

In summary, four distinct events that affected the Appalachian plateau are listed in a hypothetical time sequence in Figure 9. Although we have no direct evidence indicating the timing of the formation of the penetrative fabric relative to the spaced cleavage, we speculate that the development of the penetrative fabric was event 1. Calcite-filled joints of event 2 show no evidence of a penetrative fabric. In addition, a penetrative fabric may form during lateral compaction of a semilithified sediment. A residual strain responsible for controlling the orientation of set Ib joints may also have been locked in the rock during event 1.

Outcrop and thin section evidence suggests that set Ia joints are early-formed joints. Whether or not the set Ia joints formed before cleavage developed is debatable, and thus the two are both listed as event 2. That cleavage formed before the folding is in part recognized through the significant rotation of the cleavage relative to that of the fold axes [*Engelder and Geiser*, 1979], as well as the presence of cleavage far in advance of the most external fold. The presence of calcite filling suggests that set Ia joints formed before set Ib and set II joints. That folding and set II joints occurred simultaneously (event 3) is suggested only because folding provides a mechanism to induce extension normal to fold axes. We infer that set Ib joints are late (event 4) and controlled by the penetrative fabric in the rock. By late we suggest they may have formed during removal of overburden and thus form after set II joints.

Appalachian Plateau Dynamics

With the present data on joints and finite strain we now suggest regional stress patterns and construct a model for the deformation of the Appalachian plateau of New York. Our model is designed to incorporate the following aspects of the structures found on the Appalachian plateau.

1. Our previous work on the New York plateau [*Engelder and Engelder*, 1977; *Engelder*, 1979*a*; *Engelder and Geiser*, 1979] strongly suggests that the Devonian cover has undergone a deformation over an area defined by the salts of the Salina Group. This deformation produced the present curvature in strike of the solution cleavage, penciling, and finite strain. The curvature suggests that the deformation may be thought of as a flow occurring in a trough defined by the base of the Salina Group. If we think of this flow as taking place in a trough, by analogy with other fluids, resistance at the walls of the trough will slow the flow. With flow restricted near the walls, material lines showing the progress of adjacent particles will have to bend and stretch. Yet, stretching parallel to fold axes did not occur, as is indicated by finite strain analysis [*Engelder*, 1979*a*]. Thus unlike flow in a trough, there is no evidence of bending of material lines despite the apparent curvature of the strike of solution cleavage, penciling, and finite strain.

2. Set Ia joints (Figure 4*a*), inferred to be of the same age as the cleavage in the eastern half of the plateau (from the longitude of Yates County to that of Onondaga County in Figure 1), show no evidence of the rotation shown by the cleavage. The strike of the joints maintains parallelism for over 100 km before abruptly rotating about 20° to the east (Figure 4*a*). If the cleavage strike lines are though to be a type of integrated velocity profile produced by the flow of an in-

compressible fluid, any set of joints which are at least as old as the cleavage should not rotate with the cleavage, and the set Ia joints do not.

3. *Jacoby and Dellwig* [1974] and *Murphy* [1980] have documented the existence of tear faults within the Salina Group. These tear faults have orientations similar to that of the set Ia joints, and are consistent with the 'keystone' deformation postulated by *Gwinn* [1964], *Rodgers* [1970], and *Engelder and Engelder* [1977], for the development of the orogenic curvature of the Appalachians. By keystone deformation we mean the process of enlarging the radius of curvature of a fold belt without internal extension parallel to fold axes. This is accomplished by slip of rectangular blocks separated by triangular blocks in a manner similar to *Laubscher*'s [1972] model for the Jura. If keystone deformation occurred on the Appalachian plateau, there should be a relationship between this deformation and the various joint systems.

4. The Appalachian plateau of New York has undergone a minimum of a 10–15% layer parallel shortening associated with cleavage and penciling. Yet the Ia joint set, which is synchronous with this shortening, shows no evidence of slip parallel to the joint surface.

5. If, as we believe, the joint sets represent stress trajectories, then the shape of the boundary associated with the generation of various joint patterns may be drawn.

The above aspects are included in the following kinematic and dynamic models for the deformation of the Appalachian plateau of New York. The lower boundary of deformation is placed within the Salina Group, where detachment of the overlying section occurs. The overlying section consists of two distinct lithotectonic units, a lower brittle unit and an upper ductile unit (Figure 10). Here brittle and ductile are used to note only the behavior of the lower unit relative to the upper unit. The lower unit includes the Onondaga limestone, Oriskany sandstone, and other units that have been imbricated by splay faults climbing up section to end as blind thrusts [*Gwinn*, 1964]. The ductile unit includes all of the Upper Devonian Group of western New York which drape over the imbricated rocks at depth.

For our kinematic model, Appalachian plateau deformation resembles that presented by *Wegmann* [1961] for the Jura (Figure 10). The difference is that basement rock under the Jura is equivalent to the brittle unit of the Appalachian plateau, but the plateau has detachment beneath the 'basement.' *Wegmann* [1961] envisions the basement of the Jura shuffling on a series of tear faults, whereas the brittle unit of the Appalachian plateau not only shuffles on tear faults [*Murphy*, 1980] but also shortens by imbricated thrust faults [*Gwinn*, 1964]. Although the deformation mechanism is different, the cover of the Jura resembles the ductile unit of the Appalachian plateau.

The dynamic model proposed for the Appalachian deformations depends in part on our interpretation of set Ia and set Ib joints. Strain relaxation and velocity anisotropy tests on the Machias formation show that the set Ib joints are coaxial with a penetrative fabric. The nature of the penetrative fabric is not clear, although it is likely to be related to a preferred orientation of phyllosilicates in the rock (G. Oertel, personal communication, 1979). Because set Ib joints are coaxial with in situ strain, it is reasonable to suggest that the in situ strain controlled the direction of propagation of set Ib joints in much the same manner as residual strain controls the fracture orientation in point load experiments [*Friedman and Logan*, 1970]. Likewise, the relationship between the penetrative fabric and strain relaxation measurements is obvious only to the extent that the two phenomena are coaxial. It is clear that the penetrative fabric is not coaxial with the spaced cleavage and finite strain by fossil distortion. Evidence suggests that set Ib joints may be joints forming on the removal of overburden. If this interpretation is correct and if the penetrative fabric is indeed widespread, then the orientation of set Ib joints actually parallels the compression direction during an event market by formation of the penetrative fabric. By this interpretation the trajectories of set Ib joints do not represent true stress trajectories but in fact follow a fabric representing finite strain (Figure 11*b*). Thus the boundary condition established on the basis of the set Ib joints are only an approximation. In Figure 11*b* the stress trajectories represent a NNW-directed push from a thrust sheet southeast of the New York–Pennsylvania border.

During the stage when set Ia joints propagated stress was transmitted to New York above the detachment by end loading (Figure 11*a*). A north-south directed end loading occurred along an east-west boundary running from the western margin of the set Ia joints to the area near Binghamton. The shortening strain set up in response to these stresses is taken up in a different manner by the brittle and ductile units. Pressure solution and intragranular strain plus some folding occur in the ductile superstructure, whereas splay faulting plus some ductile deformation occur in the brittle section.

The Ia joint set shows no rotation or sign of shear, and finite strain shows no indication of stretching parallel to fold axes. An explanation for these observations is indicated by the

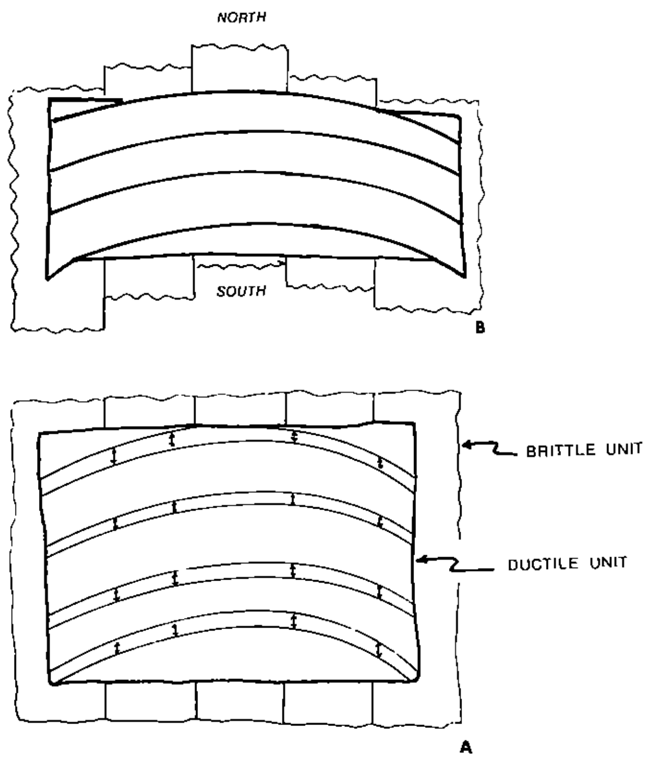


Fig. 10. (*a*) Plan view of dynamic model for the Appalachian plateau where the rocks above the detachment consist of a lower brittle unit and an upper ductile unit. (*b*) During deformation the lower brittle unit develops a series of parallel tear faults which do not propagate through the ductile unit. As slip occurs on the detachment an arcuate pattern is developed in the brittle unit. The same arcuate shape is developed in the ductile unit by solution along arcuate surfaces.

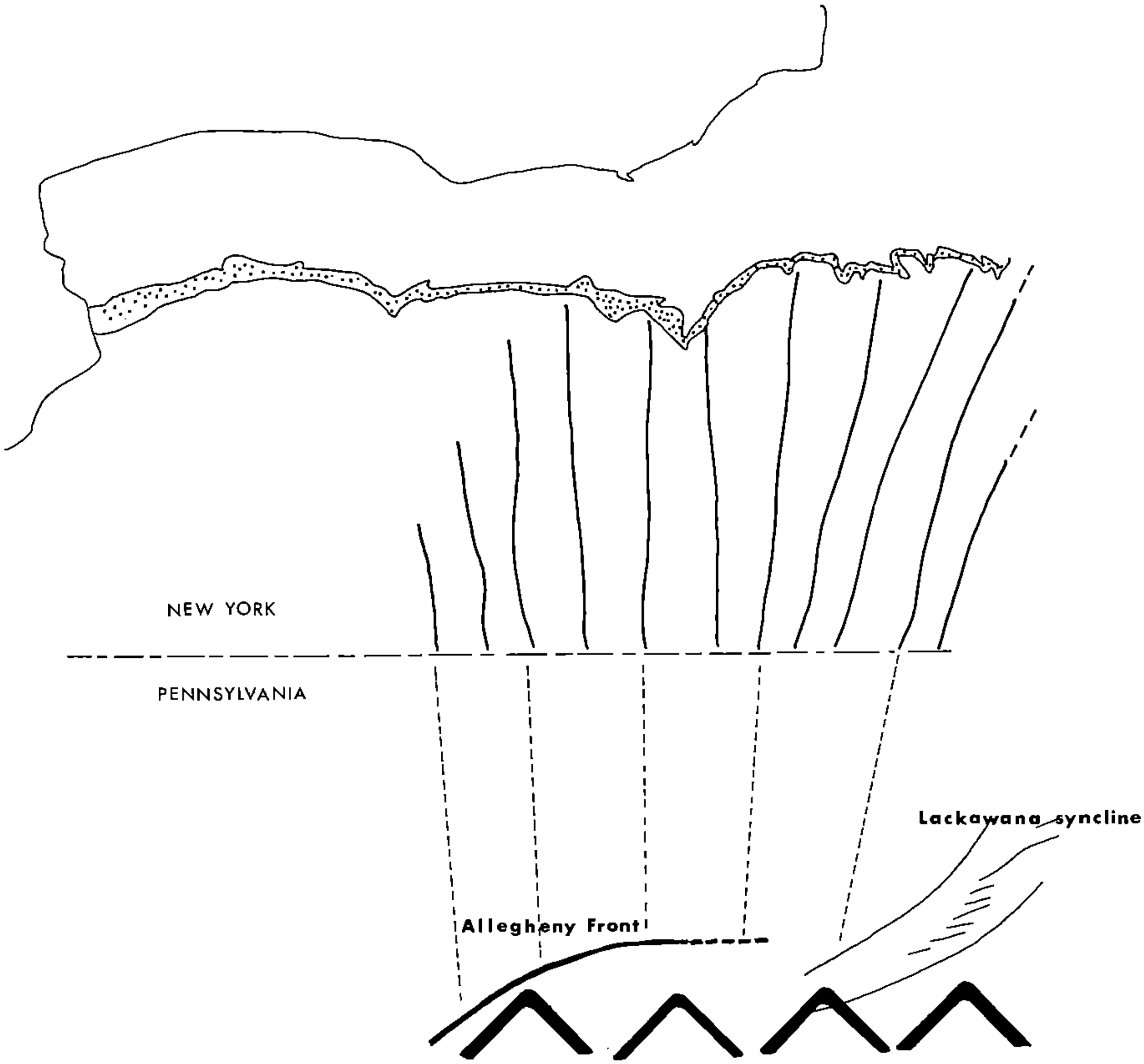


Fig. 11*a*. Stress trajectories drawn parallel to set Ia joints (solid lines). The stress trajectories have been extrapolated to the Allegheny front and Lackawana syncline (dashed lines). The outcrop pattern of the Onondaga limestone shown as the dotted area. The dark arrows represent a NNW-directed push from rocks to the southeast. Folds of the Lackawana syncline after *Rodgers* [1970].

nature of the solution process by which the ductile horizon is deforming. A critical aspect of this process that pressure solution shortening does occur by volume-decrease strain [*Wright et al.*, 1979]. This observation allows us to hypothesize the deformation model shown in Figure 10. Here as the ductile superstructure deforms, a curving family of surfaces is set up within the body across which dissolution is maximized. In essence the rectangle in Figure 10*a* is deformed into that shown in Figure 10*b* by removal of a nested set of curved slices. The mechanism for initiation of dissolution along curved surfaces is not understood. The curvature is apparently a function of the boundary conditions but not due to the rotation and stretching of material lines; consequently, the Ia joint set does not rotate.

Based on the stress trajectories drawn across the Appalachian plateau, we suggest that two events are recorded that reflect major deformation within the core zone of the Appalachian Mountains. The first event reflects a NNW-directed compression and may correlate with the decollement associated with the Lackawanna syncline [*Geiser*, 1980]. This event is represented by the SSE extrapolation of stress trajectories drawn parallel to set Ib joints (Figure 11*b*). The second event reflects a northward transmission of stress, and may correlate with the refolding of the edges of the Lackawanna syncline [*Rodgers*, 1970; *Geiser*, 1980; *Washington*, 1980]. Likewise, this event is represented by the southward extrapolation of stress trajectories drawn parallel to set Ia joints [*Washington*, 1980]. The second event is also recorded by fossil distortion and, hence, major decollement slip. Here it is not clear why fossil distortion occurs further to the west than set Ia joints.

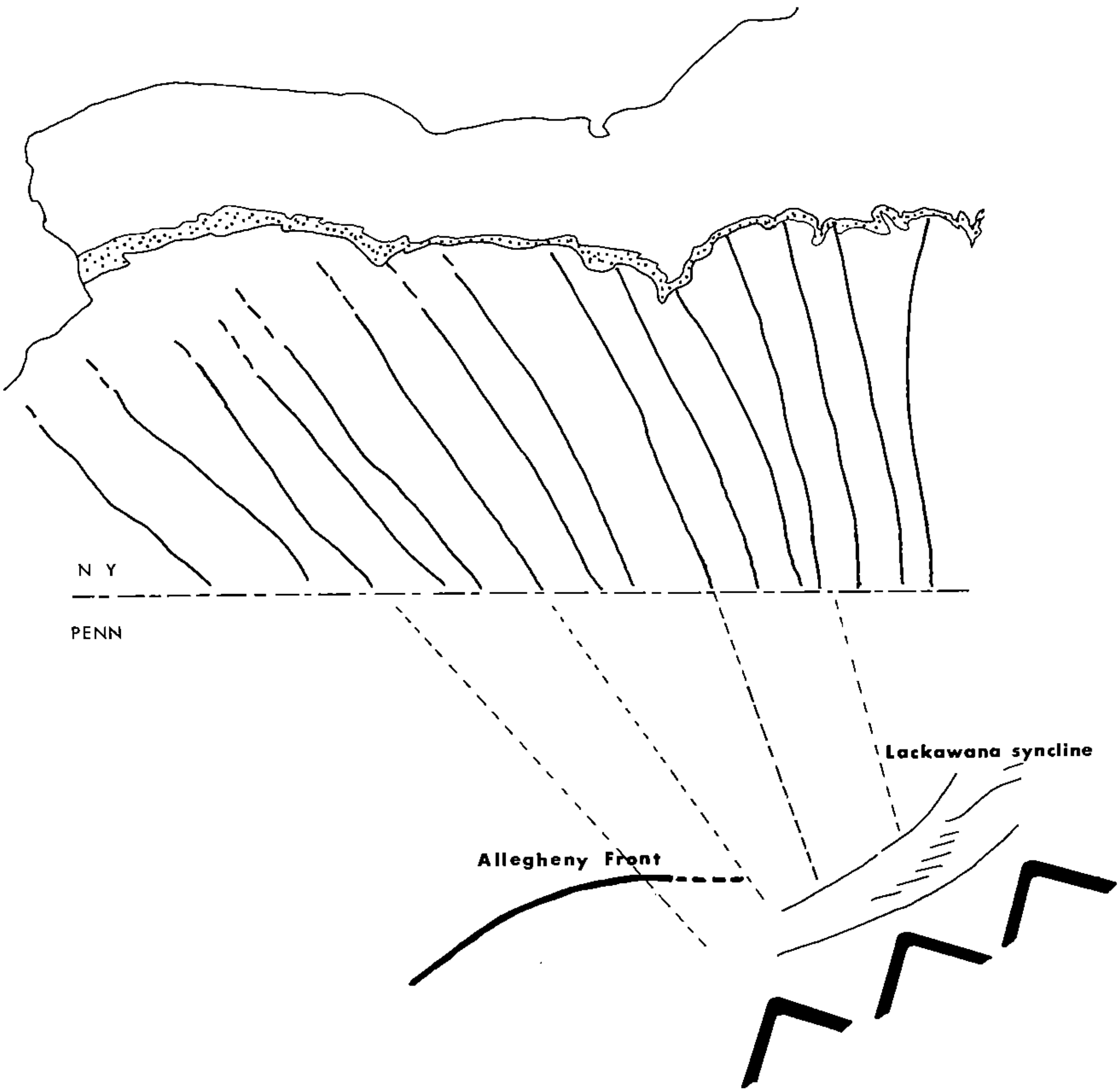


Fig. 11*b*. Stress trajectories drawn parallel to set Ib joints (solid lines). The stress trajectories have been extrapolated to the Allegheny front and Lackawana syncline (dashed lines). The outcrop pattern of the Onondaga limestone shown as the dotted area. The dark arrows represent a north-directed push from rocks to the southeast. Folds of the Lackawana syncline after *Rodgers* [1970].

Our model for Appalachian plateau dynamics differs from *Laubscher*'s [1972] model for Jura dynamics. In the Jura, tear faults form a conjugate pattern for which Laubscher draws stress trajectories bisecting the acute angle of the conjugate faults. This pattern permits extension parallel to fold axes throughout the 200 km long Jura Mountain chain. Our model for Appalachian plateau deformation, at least during the event represented by set Ia joint trends, shows stress trajectories subparallel to tear faults. Where stress trajectories are parallel to tear faults, decollement tectonics does not require extension parallel to fold axes. This is consistent with fossil distortion which shows no extension parallel to fold axes [*Engelder*, 1979*a*]. However, conjugate shears are found in central Pennsylvania, where *Nickelsen* [1966] reports extension parallel to fold axes (Figure 12).

During deformation of the Appalachian plateau we hypothesize that detachment and segmentation of the post-Salina cover developed along the lines shown in Figures 10 and 12. In this model, translation is accompanied by layer-parallel shortening in northeastern Pennsylvania and New York. Extension parallel to fold axes is restricted to the keystone region of Pennsylvania, where translation and a small amount of bending produced the 4–5% extension found by *Nickelsen* [1966].

Nickelsen and Hough [1967] suggest that the acute intersection between cross-strike joints are due to overlapping joint domains. Their idea was that joint sets attempt to remain perpendicular to fold axes, but as the fold axes bend, the joints do not change orientation but rather one set is replaced by another set. Present data do not support this but rather suggest

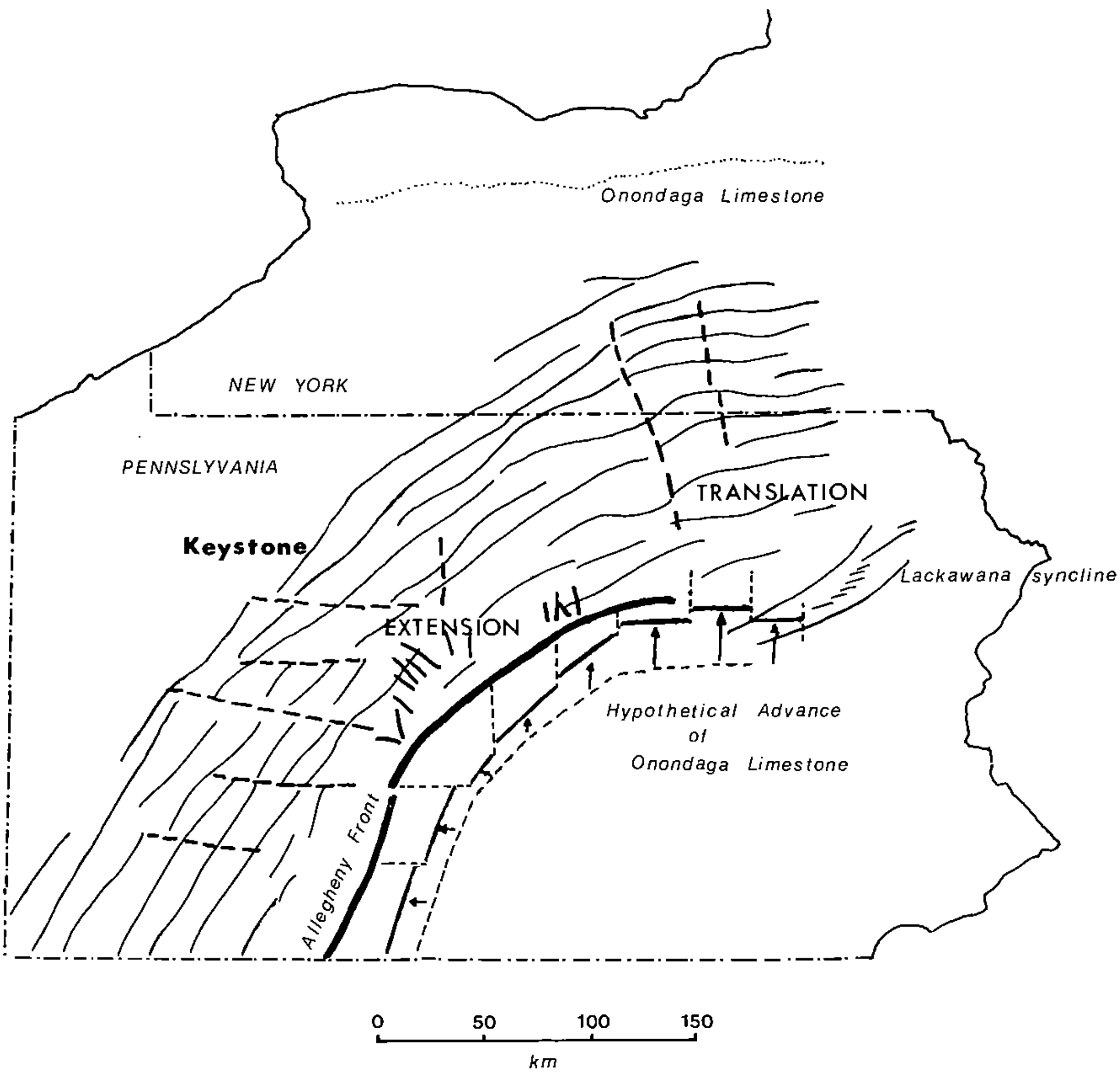


Fig. 12. Schematic map of the hypothesized kinematics of the central Appalachian plateau [after *Rodgers*, 1970]. Fold axes are the solid lines striking northeast to east-west. Heavy dark line is the Allegheny front. Dark dashed cross-strike lines are zones of potential strike-slip movement [after *Rodgers*, 1970; *Murphy*, 1980]. Segmentation of an original marker line on the Onondaga-Oriskany horizon is hypothesized (light dashed lines).

the intersecting joint sets reflect either two or more events during which decollement slip was redirected or the development of two different decollements.

Western New York In Situ Stress Measurements

Western New York has not been an easy area in which to measure and understand the state of stress in the upper crust. Various experiments in a variety of rock types give different answers concerning the state of stress. In the 1950's, hydraulic fracturing to stimulate recovery of petroleum in Allegany County, New York, and McKean County, Pennsylvania, showed a tendency to increase communication between wells aligned ENE to WSW. *Heck* [1960] demonstrated that fractures propagated from the borehold in this direction. *Sbar and Sykes* [1973] later included these measurements in their compilation showing a large region west of the Appalachian Valley and Ridge subject to a ENE horizontal compression. However, the strain relaxation measurements in the Machias formation show no relationship to the hydraulic fracture measurements in western New York (Figure 8*b*). This lesson serves as a good illustration of the care that must be exercised in using near surface strain relaxation measurements to predict the state of stress at depth.

The interpretation of hydraulic fracture measurements in Allegany County, New York, should also be treated with care. *Heck*'s [1955] interpretation of the control of hydraulic fracture orientation was that the rock contained planes of weakness along which fractures propagated. He correlated the hydraulic fractures with surface jointing, although there is a 10°–15° difference between set II joints in Allegany County and the induced hydraulic fractures (Figure 5). However, pencil cleavage in Allegany County is within a degree of parallelism with the mean orientation of the hydraulic fractures (Figure 8*b*). In a low stress situation a fabric such as a spaced cleavage could very well be the deciding factor in controlling the orientation of fracture propagation. This interpretation is more appealing that Heck's because joints are spaced more than a meter apart, whereas spaced cleavage develops with a much closer spacing.

The Onondaga limestone shows that jointing at very best partially relieves the stress field in which the joint propagates (Figure 7). Likewise, a residual stress which is most readily detected by overcoring strain gauges bonded to surfaces is less apparent when strain relaxation is measured below the surface.

Strain relaxation of the Machias formation appears to differ from the Onondaga limestone by virtue of the mechanism controlling the orientation of jointing. In the Onondaga quarry no rock fabric is coaxial with the joints and compression causing the quarry floor to pop up. The joints are more likely to have propagated parallel to true stress trajectories present during quarrying rather than parallel to a penetrative fabric as is thought to be the case for the set Ib joints in the Machias formation. We conclude that subsurface strain relaxation in the Machias formation is controlled by a residual stress, although the mechanisms for locking and relaxation of this residual stress are not understood.

Reik and Currie [1974] correlated joints and rock fabric in the Cardium sandstone of Alberta, Canada, in much the same manner as was done for the Machias formation at Belmont, New York. The Cardium sandstone exhibits a maximum stiffness parallel to both the local fold axes and joints equivalent to set II joints. Likewise, the Machias formation also exhibits a stiff direction parallel to set II joints. The difference between the Machias formation and the Cardium sandstone is that the former shows a tendency to expand parallel to set Ib joints, whereas the latter contains a compressive residual strain [*Friedman,* 1972] parallel to set II joints. However, in both cases the last joint set to form is parallel to the maximum recoverable strain.

CONCLUSIONS

1. The orientation of jointing is independent of stratigraphic units on the Appalachian plateau.
2. The orientations of set Ia and set Ib joints vary independently of each other.
3. The trace of set Ib joints corresponds with the compression direction indicated by a penetrative fabric in outcrops in Allegany County, whereas set Ia joints parallel the compression direction indicated by deformed fossils in many outcrops in the eastern portion of the New York plateau.
4. The pattern of set Ia joints is more likely to represent a paleostress field than Ib joints.
5. Subsurface samples of the Machias formation show a maximum expansion parallel to set Ib joints and a penetrative fabric rather than the direction of compression indicated by deformed fossils.
6. At the outcrop surface of Onondaga limestone a residual strain associated with an event other than jointing controls the direction of maximum expansion.
7. Surface and shallow near-surface stress measurements have been shown to be related to past strain history recorded in the rocks and unrelated to the modern, tectonic stress field.

APPENDIX 1: DEFINITIONS

Joint. A crack in rock, generally transverse to bedding, along which no appreciable shear displacement has occurred. A joint is an extension fracture.

Joint set. A group of more than two joints with a specific attribute; usually a common orientation.

Systematic joints. Subparallel joints generally having a regular spacing that depends on several parameters including bedding thickness and lithologic character [after *Hodgson,* 1961]. These are usually planar.

Nonsystematic joints. Nonparallel joints that are generally nonplanar [after *Hodgson,* 1961].

Cross joints. Those joints that are subnormal to systematic joints without cutting the systematic joints [after *Hodgson,* 1961].

Zonal joints. Joints that form narrow zones of close spacing compared to other joint sets in the outcrop in question. Zonal joints are often arranged edge to edge so that individual joints are only slightly offset from each other [after *Hodgson,* 1961].

Cross-strike joints. Vertical joints that cut fold axes at high angles. These are not to be confused with cross joints.

Strike joints. Vertical joints that are subparallel to fold axes.

APPENDIX 2: NOMENCLATURE

Strain ϵ_f. Maximum compressive strain as recorded by fossil distortion oriented parallel to bedding on the Appalachian plateau. These deformed fossils do not necessarily record the total finite strain on the Appalachian plateau as is discussed within this paper.

Strain ϵ_{oc}. Maximum in situ compressive strain as measured parallel to bedding by overcoring techniques.

Expansion ϵ_{max}. Maximum expansion measured during overcoring; $\epsilon_{oc} = \epsilon_{max}$.

Expansion ϵ_{min}. Minimum expansion measured during overcoring.

Joint set Ia. Cross-strike joints that generally strike parallel to ϵ_f and are sometimes calcite filled. Included as part of joint set I by *Parker* [1942]. These joints were called dip joints by *Sheldon* [1912] and are equivalent to the systematic joint set E described by *Nickelsen and Hough* [1967].

Joint set Ib. Cross-strike joints that strike at 16°–34° counterclockwise from ϵ_f. This joint set strikes parallel to ϵ_{oc} in the Machias formation. Included as part of joint set I by *Parker* [1942]. These joints are called dip joints by *Sheldon* [1912] and are equivalent to the systematic joint set A described by *Nickelsen and Hough* [1967].

Joint set II. Joints that strike subparallel to fold axes. Joint set II of *Parker* [1942]. These joints were called strike joints by *Sheldon* [1912] and have the same orientation relative to fold axes as the nonsystematic joints described by *Nickelsen and Hough* [1967].

Joint set III. A joint set not obviously related to either ϵ_f, ϵ_{oc}, or fold axes. Joint set III of *Parker* [1942].

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