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THE NATURE OF DEFORMATION WITHIN THE OUTER LIMITS OF THE CENTRAL APPALACHIAN FORELAND FOLD AND THRUST BELT IN NEW YORK STATE *

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ABSTRACT

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Residual strain, a self-equilibrating recoverable strain that remains in rocks even after external forces and moments are removed, is found NNW of the folded Appalachian plateau in the Devonian Onondaga limestone and the Silurian Lockport dolomite and Grimsby sandstone of western New York. This residual strain is manifest upon overcoring by a NNW directed maximum expansion of the limestone and sandstone and a random maximum expansion of the dolomite. Strains, recorded with strain gauge rosettes bonded to outcrops, are as high as 200 $\mu\epsilon$ (microstrain). Double overcoring of the sandstone and dolomite relieves smaller strains of the same orientation as the initial overcore. X-ray analysis of the Grimsby sandstone shows that the elastic residual strain locked in quartz grains is characterized by a NE principal extension of 60 $\mu\epsilon$ and a 10—30 $\mu\epsilon$ NW principal compression oriented 30° counterclockwise from the NNW compression indicated by overcoring. Sonic velocity tests on samples in the lab indicate that Grimsby sandstone is anisotropic with a NNW maximum P-wave velocity of 4.05 km/sec. This anisotropy correlates with the residual strain in Grimsby sandstone. Mechanical twinning of calcite within both the Onondaga limestone and Grimsby sandstone indicates that the rock contains a permanent compressive strain of less than 2% in the NNW direction. The permanent strain becomes progressively smaller in a series of samples from Syracuse to Buffalo, New York. The development of solution cleavage in Onondaga limestone also indicates a NNW compression. No evidence of permanent strain was found in the dolomite. The NNW compression of the limestone and sandstone is normal to the fold axes of the Appalachian foreland fold and thrust belt. This geometric relationship indicates that the residual strain well beyond the outermost Appalachian folds was caused by the same tectonic stresses responsible for folding the Appalachians during the late Paleozoic. Strain within the Appalachian plateau below the Silurian salt horizon suggests either the presence of a second décollement in, perhaps, Ordovician shales or a general NNW shortening of the crust under the Appalachian plateau.

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INTRODUCTION

How far were orogenic stresses transmitted away from the core of the Appalachian foreland fold and thrust belt? The answer is obtained by first recognizing how incipient deformation at the leading edge of a foreland fold and thrust belt is manifest and then tracing that deformation toward the craton from the fold and thrust belt.

Various geologic structures have been associated with the transmission of stresses toward the North American craton from the core of the Appalachians. Geologists in the first half of the 20th Century recognized folds within the Appalachian plateau which were arcuate about the same trend as the Valley and Ridge of central Pennsylvania (Wedel, 1932: Fettke, 1954). Likewise, vertical extension fractures change strike to conform to the arcuate tectonic pattern of the central Appalachian plateau (Parker, 1942; Ver Steeg, 1942, 1944; Nickelsen and Hough, 1967). Both of these observations along with the recognition of décollement tectonics under the Appalachian plateau (Rodgers, 1963; Gwinn, 1964) suggest the transmission of Paleozoic orogenic stresses throughout the folded portion of the Appalachian plateau.

In addition to the development of brittle fractures, décollement slip, and macroscopic folds, the transmission and dissipation of orogenic stress is also manifested by a penetrative strain of rocks within the Appalachian plateau (Nickelsen, 1966; Engelder and Engelder, 1977). Such strain is necessary to accommodate the geometry of plateau folds (Wiltschko and Chapple, 1977).

Fracture patterns northwest of the open fold zone of the Ouachita Mountains indicated to Melton (1929) that orogenic stresses are transmitted well beyond the last major structures of a foreland fold and thrust belt. Likewise, Ver Steeg's (1942, 1944) mapping indicated that transmission of orogenic stresses also occurred beyond the last open fold of the Appalachian plateau in Ohio. Friedman (1964) pointed out that the fracture pattern NW of the open folds of the Ouachita Mountains indicated that the maximum compressive stress (σ_1) was horizontal and normal to the fold axes, the intermediate principal stress (σ_2) was vertical, and the least compressive stress (σ_3) was horizontal and parallel to the fold axes. These fracture analyses leave little doubt that stress was transmitted beyond the last open fold of the Appalachian plateau but is it reasonable to expect that penetrative strain accompanied this transmission of stress?

Recent studies of folds indicate that major layer-parallel shortening occurs during the low-dip (early) stages of the development of buckle folds (Sherwin and Chapple, 1968). Strain recorded by calcite twinning in minor folds in central Pennsylvania resembles the patterns expected for the early stages of buckling (Groshong, 1975 a, b). Likewise, laboratory studies of buckle folds show early layer-parallel shortening prior to the buckle instability (Handin et al., 1972). If the major folds of the Appalachian plateau behave as these smaller folds, layer-parallel shortening may represent the initial stage in the development of folds preceding any change in limb dip. Thus it is not unreasonable to expect that incipient deformation by layer-parallel shortening may be found at the leading edge foreland fold belts where folds have not yet developed. In fact, Chinn and Konig (1973) show that favorably oriented calcite crystals twin under stress conditions insufficient to produce discernible fold forms northwest of the Ouachita fold belt.

In this paper I describe several measurements which show that the rocks of western New York were deformed by a NNW directed compression at locations beyond the outermost fold of the Appalachian plateau. The incipient deformation, at the leading edge of the central Appalachian foreland fold and thrust belt, is manifest in: (1) a residual elastic strain; (2) an elastic anisotropy; (3) a plastic deformation by mechanical twinning of calcite; and (4) a shortening by the development of solution cleavage.

GEOLOGY OF WESTERN NEW YORK

The outer limit of the central Appalachian foreland fold and thrust belt is found in western New York State where Paleozoic sediments (Silurian to Upper Devonian) were deposited on the North America craton (Fig. 1). Oldest rocks outcrop in the north as the result of a gentle dip to the south (3 m per km). Structures in the southern portion of western New York consist of extremely subdued and regularly spaced folds striking north of east (Wedel, 1932). Subsurface data show anticlines over imbricated high-angle faults (Bradley and Pepper, 1938). These structures are the outermost major folds of the Appalachian foreland fold and thrust belt. North of the folds the only major structure is the Clarendon—Lindin fault system which has been mapped by Van Tyne (1975) by correlation of well logs. This fault system appears at the surface as a gentle monocline striking N—S dipping at 3° to the west near Clarendon, New York (Fig. 1).

Minor folds with an Appalachian trend (east—west fold axes) are found near the Onondaga limestone within 5 km of the outcrop *SYR* (Groshong, 1975 a, b) and 25 km *SE* of *OAK* (Fig. 1). E—W striking vertical stylolites occur as far west as *OAK* but these minor structures with an Appalachian trend die out farther to the west.

Fossil distortion and solution cleavage are associated with the folds of western New York (Groshong, 1975; Engelder and Engelder, 1977). The penetration strain is indicative of as much as 10% layer-parallel shortening normal to the Appalachian fold trend. The orientation of horizontal maximum compressional strain measured using the mechanical twins in calcite is shown in Fig. 1. The data on strain in calcite grains from localities *RAW*, *VAN*, *AND*, *CAM*, *ADI*, *SMB* and *TRI* are consistent with the compression of fossils as discussed in Engelder (1979).

Three stratigraphic units were sampled for this study: (1) the Grimsby formation, a Silurian unit of both marine and non-marine red sandstones with calcite cement (Fisher, 1954); (2) Lockport group, a Silurian dolomite

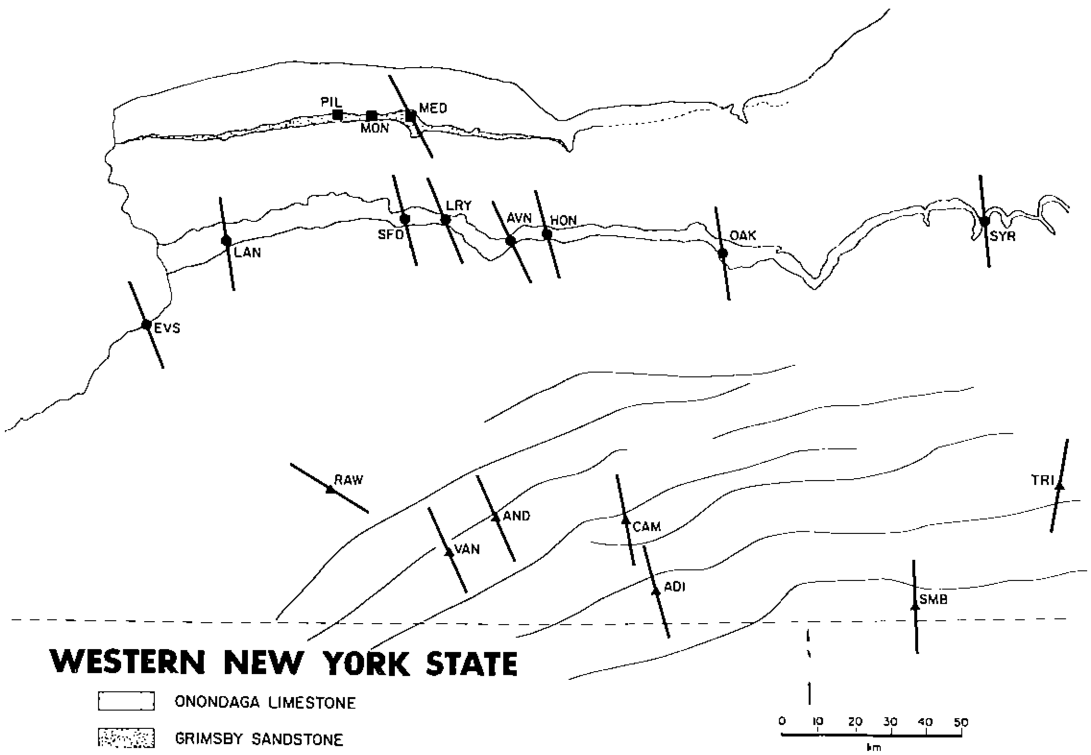

Fig. 1. The general tectonics of western New York State. Solid curved lines are anticlinal axes mapped by Wedel (1932). Dotted lines locate the Clarendon—Linden fault system mapped by Van Tyne (1975). Squares and circles with three letter labels are sample localities for this study and triangles for that of Engelder (1979). The orientation of horizontal maximum compressive strain measured using mechanical twins in calcite is shown for most localities.

which is thin-bedded and generally fine grained (Zenger, 1962); and (3) the Onondaga formation, a Devonian limestone with abundant corals and patch reefs (Dunbar, 1959). The Grimsby sandstone is equivalent to the Tuscarora sandstone in the Valley and Ridge region of central Pennsylvania. Because stratigraphic terminology in western New York is in a state of flux, in this paper these three units are called the Grimsby sandstone, Onondaga limestone and Lockport dolomite.

RESIDUAL STRAIN

Laboratory experiments of short duration, compared with the geological time scale, show that deformation of rocks includes a recoverable component which itself may be instantaneous (elastic) or time-dependent (c.f., Scholz and Kranz, 1974). It is a matter of debate whether these same recoverable deformations will remain locked in rocks from the time of the tectonic deformation as much as hundreds of millions of years before present. Field measurements show that many rocks contain recoverable strains which have apparently been locked in rocks since the Precambrian (Friedman, 1972). These potentially recoverable strains, called residual strains, are self-equilibrating and remain in the rock even after external forces and moments have been removed. Tectonic deformations are only one of many sources for residual strain. Others include thermal strains, burial loads, weathering, and chemical alteration. Often it is difficult to eliminate these latter mechanisms in favor of tectonic deformation as the cause of residual strains (Russell and Hoskins, 1973; Swolfs et al., 1974).

The mechanisms for locking in tectonic residual strains are not clear. All models for residual strain consist of two elements: those containing locking strains and those containing locked strains (Friedman, 1972). The corresponding stresses must balance one another in an equilibrium volume so that the volume average of stresses is zero (Varnes and Lee, 1973). On the scale of grain size Voight and St. Pierre (1974) and Gallagher et al. (1974) present a clear model for the tectonic compression of unconsolidated grains. While the grains are under load (strained) a cement is introduced which solidifies without strains. Once the load is removed, the grains partially relax, and thereby introduce strains into the cement. The grains (locked element) are constrained by the cement (locking element) and cannot completely relax. On a much larger scale Voight (1974) offers a mechanism for locking-in strain over an entire décollement sheet. If high fluid pressures or low viscosity in a fault zone were responsible for décollement motion, a conversion to low fluid pressure or high viscosity would increase the frictional strength of the décollement fault zone. "The high friction imposed by effective overburden pressure would prevent regional relaxation of the orogenic stress existing at that point in space and time." In effect, the fault zone and rocks below are the "locking" element and the décollement sheet is the "locked" element.

In western New York and Ontario, Canada, the squeezing of underground openings at the Niagara Falls Hydropower Plants is attributed to the long-term recovery of a residual strain (Lee and Lo, 1976; Lee and Klym, 1976). Here the base of the Lockport group including the Gasport formation continues to expand against the underground workings more than 70 years after the initial excavations.

STRAIN RELAXATION

During the summer of 1975 Robert Kranz and I measured the recoverable in situ strain of some rocks in western New York (Fig. 2). This was accomplished by cutting 15 cm cores from bedrock and monitoring the strain change caused by freeing the core. Details of this technique are given in Engelder and Sbar (1976, 1977) and Engelder et al. (1977).

Eight sites in the vicinity of the Clarendon–Linden fault were selected for

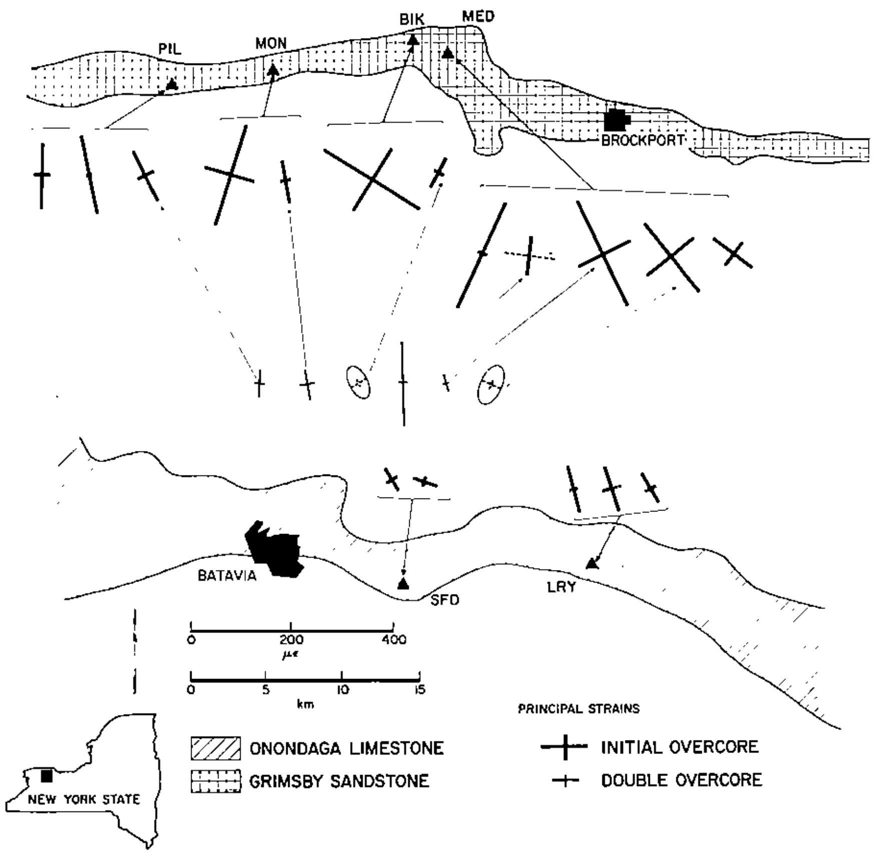

Fig. 2. Geology and in situ strain in the vicinity of Brockport and Batavia, New York. Six sample sites are labeled. The magnitude and orientation of the strain relieved by the initial overcore are indicated by dark lines. The magnitude and orientation of the strain relieved by a double overcore are indicated by light lines. Solid lines represent expansion and dashed lines represent contractions. The long axis of the ellipse is oriented in the direction of maximum expansion (minimum contraction) where contraction took place along both axes. The magnitude of relieved strain is represented by a scale in microstrain ($\mu\epsilon$).

strain relaxation measurements (Figs. 2 and 3). With the exception of a site labeled *MON* all were inside or on benches of stone quarries. In the formations described above we wished to sample outcrops between the intersections with the surface of the three branches of the Clarendon—Linden fault and at a distance from the fault system. Our experiment was designed to compare the variation in magnitude and orientation of strain relaxation among stratigraphic units and to measure the effect of the Clarendon—Linden fault on the in situ strain. Small earthquakes had been triggered along the fault by fluid injection near Dale, New York (Fletcher and Sykes, 1977).

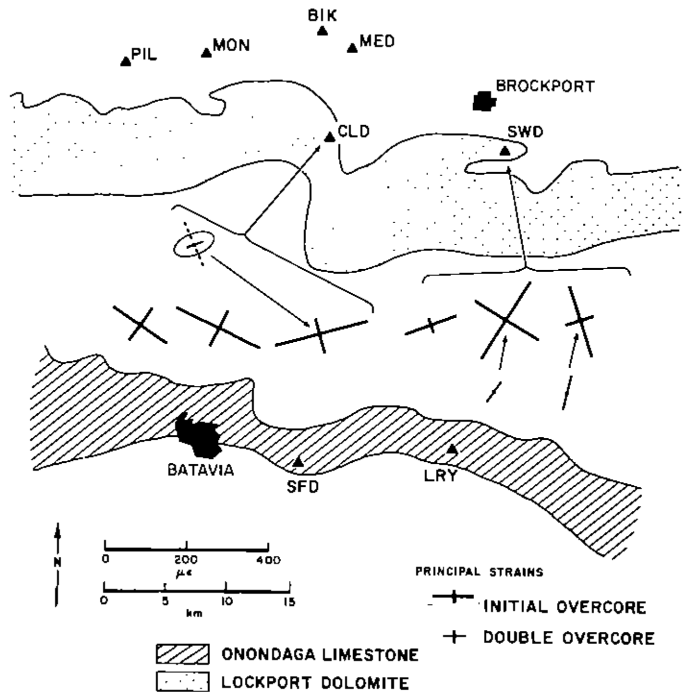

Fig. 3. In situ strain relaxation of the Lockport dolomite. The magnitude and orientation of the strain relieved by the initial overcore are indicated by dark lines. The magnitude and orientation of the strain relieved by a double overcore are indicated by light lines. Solid lines represent expansion and dashed lines represent contractions. The long axis of the ellipse is oriented in the direction of maximum expansion (minimum contraction) where contraction took place along both axes. The magnitude of relieved strain is represented by a scale in microstrain ($\mu\epsilon$).

A fault-plane solution from these earthquakes indicated a N80°W compressive stress and it was of interest to compare this stress orientation with that detected at the surface.

Maximum expansion following the initial overcore of both the Grimsby sandstone and Onondaga limestone was generally toward the NNW (Fig. 2). In strain-relief work the maximum expansion corresponds to the maximum compressive strain in the plane of the measurement prior to overcoring. The magnitude of the maximum expansion varied from the maximum of about 200 $\mu\epsilon$ in the Grimsby sandstone to a minimum of 50 $\mu\epsilon$ in the Onondaga limestone. Variation in orientation differed from outcrop to outcrop with *LRY* and *PIL* producing the most repeatable data. It may be of significance to note that both of these outcrops are farthest from the Clarendon—Linden fault system.

Strain relaxation of the Lockport dolomite shows a random orientation for maximum expansion following the initial overcore. In three cases the maximum expansion was more than 200 $\mu\epsilon$ (Fig. 3). The Gasport formation is exposed at site *SWD* whereas the Eramosa formation is exposed at *CLD*.

To test for residual strain a 7.6 cm overcore was placed inside some 15 cm cores (Friedman, 1972. Swolfs et al., 1974; Engelder and Sbar, 1976). Of six double overcores within Grimsby sandstone, the maximum expansion was oriented no more than 31° from a north—south trend with an average orientation slightly west of north (Fig. 2). This includes measurements for samples from both within the Clarendon—Linden fault zone and adjacent to it. For each of the three double overcores in Lockport dolomite the maximum expansion for the 7.6 cm core correlated with that of the initial overcore (Fig. 3). Measurement of relaxation following double overcoring of the Onondaga limestone failed because of strain gauge failure in one case and the breaking of a core in another case. In general the orientation of maximum expansion for the 7.6 cm overcore correlated with the orientation of the maximum expansion following the initial overcore.

Without further analysis two facts emerged from the strain relaxation data: (1) the maximum expansion of the Grimsby sandstone and the Onondaga limestone was compatible with neither the compression given by the earthquake fault-plane solution for the Clarendon—Linden fault (Fletcher and Sykes, 1977) not the regional stress field documented in Sbar and Sykes (1973); and (2) the similar orientation of strain relaxation in both initial and double overcores was independent of the orientation of the strain relaxation after the initial overcore. From these two facts two hypotheses emerge: (1) the orientation of strain relaxation following the initial overcore was the recovery of a residual strain not influenced by contemporary stresses; and (2) the NNW expansion of Grimsby sandstone and Onondaga limestone is the recovery from stresses imposed on those rocks during the late Paleozoic compression of the Appalachian foreland fold and thrust belt. The following analyses test these hypotheses.

X-RAY ANALYSIS FOR RESIDUAL STRAIN

An X-ray diffraction technique for measuring residual elastic strain in quartzose rocks was developed by Friedman (1967 a, b, 1972). He measures the elastic strain components parallel to a given direction in a 5-cm diameter disk of rock by comparing the measured *d*-spacing of $\{32\overline{5}4\}$ in quartz along that direction (d_{OBS}) with that of the strain-free material (d_{u}), i.e., $(d_{\mathrm{OBS}} - d_{\mathrm{u}})/d_{\mathrm{u}} = \epsilon$. The *d*-spacings are determined by computer analysis of fixed count data obtained by step scanning the diffraction profile. Orientations and magnitudes of the three principal strain axes are calculated via computer from five strain components inclined at 45° to the sample surface and the undistorted value of strain normal to the surface. By measuring more than five strain components inclined at 45°, multiple sets of principal strains are calculated to test for strain homogeneity as illustrated by the six sets of strain axes in Fig. 4.

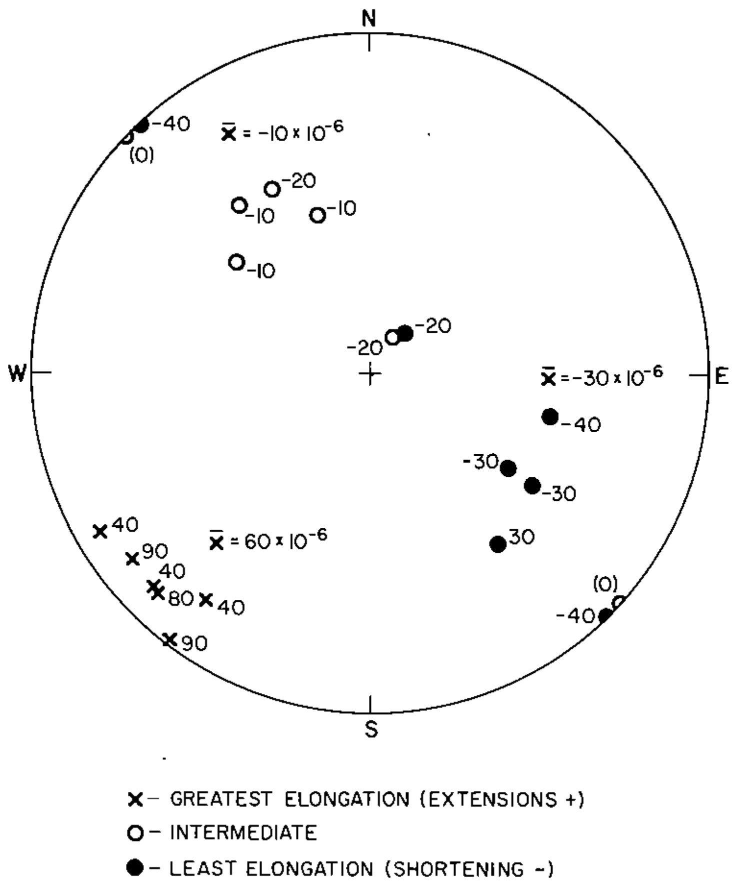

Fig. 4. Lower hemisphere projection of the state of residual elastic strain in Grimsby sandstone from *PIL*. Six sets of principal strain axes determined from elastic distortions in the quartz grains. Symbols cross, circle, and dot denote the greatest, intermediate, and least principal elongations, respectively, with compressive strains counted negatively. Average strain magnitudes are given.

Dr. Friedman of the Center for Tectonophysics, Texas A and M University, made an X-ray analysis of a sample of Grimsby sandstone from *PIL*. He observed an extension parallel to bedding and striking NE and a maximum shortening plunging less than 45° to the SE (Fig. 4). All the strains in the NW trending vertical plane are compressional. The overcoring measurements relate to the secondary principal strains in the circular section of the overcore while the X-ray determination is three-dimensional. Although not a perfect match to the orientation of the strain relaxation measured by overcoring, this measurement shows the presence of a grain by grain elastic residual strain. Likewise, this compression direct west of north is a theme that is repeated for each of the measurements reported here.

The mechanisms for the locking and relaxation of residual strain in rocks is still in doubt (Tullis, 1977). In western New York two lithologies, limestone and sandstone, relax with a NNW orientation for maximum expansion upon initial overcoring. Double overcoring of the Grimsby sandstone also triggers a NNW maximum expansion (Fig. 2). Both the Grimsby sandstone and Lockport dolomite show similar orientations for maximum expansion of initial and double overcores. To establish insight into the physical parameters associated with the recovery of residual strain by overcoring both static and dynamic mechanical properties, as well as visible fabric, are relevant. Often a mechanical anisotropy and fabric correlate with residual strains (Friedman and Logan, 1970; Friedman and Bur, 1974).

Dynamic mechanical characteristics include the elastic-wave transport properties of the samples. Pulse matching techniques described by Mattaboni and Schreiber (1967) were used to measure compressional wave velocities in samples recovered from double overcoring experiments. Flats were ground parallel to the cylindrical axis of 7.6 cm diameter cores in order to mate barium titanate transducers to the core. Travel times were measured in six directions at 30° intervals in a plane parallel to the strain gauge rosette and outcrop surface.

The compressional wave velocity within the Grimsby sandstone is between 3.3 and 4.0 km/sec in all samples whereas it is as much as 2 km/sec faster within the Onondaga limestone (Fig. 5). The sandstone has a velocity anisotropy of 3—9% whereas that in the limestone is less than 1%. The anisotropy varies from 4 to 9% for samples of sandstone taken within 2 m of each other at *PIL*. The maximum velocity within the sandstone is about parallel to the maximum expansion following initial relaxation. The 1% anisotropy in the limestone is not significant. Not shown in Fig. 5 are the data on the Lockport dolomite indicating a 1% anisotropy and a compressional wave velocity of 5.6 km/sec.

Static tests involved loading 5.5 cm diameter cores in a cylindrical test chamber described in Sbar et al. (1979). The core is subject to a uniform radial stress but it unconfined along its axis. This test simulates conditions on the outcrop surface where loads near the strain gauge bonded to the surface are radial but there is no vertical load. By loading 5.5 cm diameter cores with strain gauge rosettes attached, the shortening along three radii can be monitored as a function of the uniform radial stress. This relation between stress and strain is equivalent to a "linear compressibility" (Nye, 1957) for a two-dimensional measurement. This constant called "PLC" by Sbar et al. (1979) can be used to calculate the residual stress within an outcrop.

"PLC" for Grimsby sandstone from *PIL* and *MON* was measured (Table I). Both samples show a mechanical anisotropy to this static test with the stiffest (1/PLC) direction being NNW. This stiff direction correlates with the stiff direction indicated by the velocity anisotropy. The maximum stress

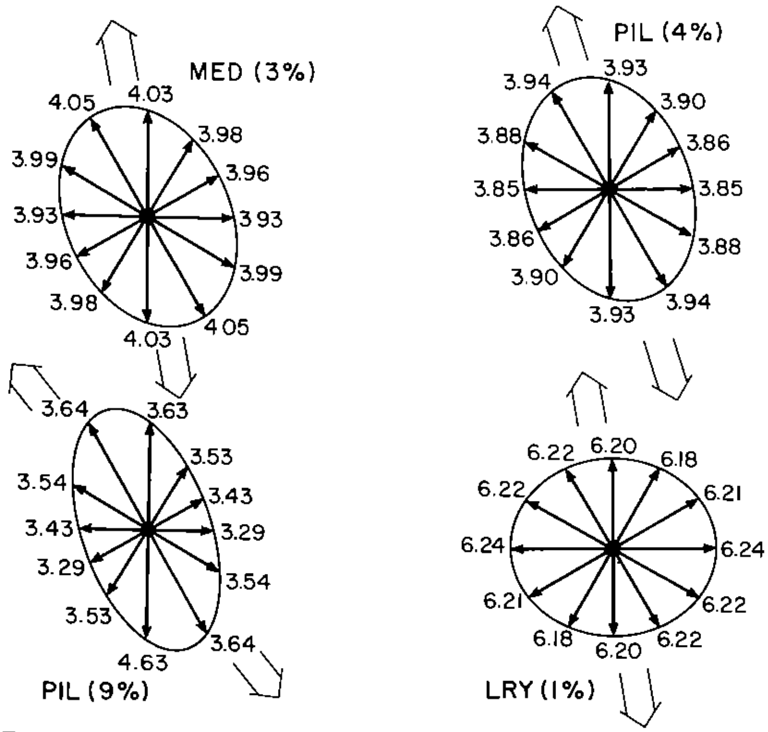

Fig. 5. Compressional wave velocities in the horizontal plane of samples. Units for velocity are km/sec. The orientation of the maximum expansion following initial overcoring is shown with arrows outside each velocity ellipse. Velocity anisotropy is listed next to the name of the sample site. Samples from *MED* and *PIL* are Grimsby sandstone. Sample from *LRY* is Onondaga limestone.

TABLE I

Strain relief to stress conversion

Site	Values parallel to gauge components			Principal values			Key
	N	SE	SW	max	min	strike of max	
MON	206	117	172	216	113	N18°E	$\mu\epsilon$
	3.1 mb^{-1}	3.1 mb^{-1}	3.8 mb^{-1}	3.8 mb^{-1}	2.9 mb^{-1}	N60°E	PLC
	66 b	37 b	45 b	67 b	32 b	N08°E	σ
BIK	173	229	165	229	148	N56°W	$\mu\epsilon$
	3.4 mb^{-1}	3.7 mb^{-1}	3.8 mb^{-1}	3.9 mb^{-1}	3.4 mb^{-1}	N83°E	PLC
	50 b	62 b	42 b	63 b	40 b	N48°W	σ

Key: $\mu\epsilon$ = strain relief upon overcoring in microstrain (strain $\cdot\ 10^{-6}$); PLC = two-dimensional linear compressibility (mb = b $\cdot\ 10^6$); σ = calculated stress and stress difference calculated for both samples is nearly the same. The difference in orientation of principal stresses between the samples reflects the choice of samples with divergent strain relaxation orientations after initial overcoring.

Petrographic observations included the measurements of the orientation of: (1) microfractures and fluid inclusions; and (2) the long axes of quartz grains; and (3) strain associated with mechanical twinning. Any microfractures and fluid inclusions in the Grimsby sandstone, Lockport dolomite, and Onondaga limestone assume no preferred orientation. The long axes of quartz grains in the Grimsby sandstone are aligned in a north-south to a northeast—southwest direction as indicated by samples from *PIL* and *MED* (Fig. 6). Paleocurrents during alluvial deposition of the Grimsby were generally toward the north (Hunter, 1960). The relaxation of strain and mechanical anisotropy of the samples from western New York cannot be correlated with either a preferred orientation of microfractures or grain boundaries. The correlation of stiff sample direction with the direction of maximum compression for residual stresses is common (Friedman and Bur, 1974). These static and dynamic tests leave little doubt that the Grimsby sandstone was once subject to a NNW-directed compression.

A striking contrast between the Lockport dolomite and the Onondaga limestone is the presence of mechanical twins in the latter and their complete absence in the former. From this observation it is clear that the Onondaga limestone was subject to a high enough deviatoric stress to deform

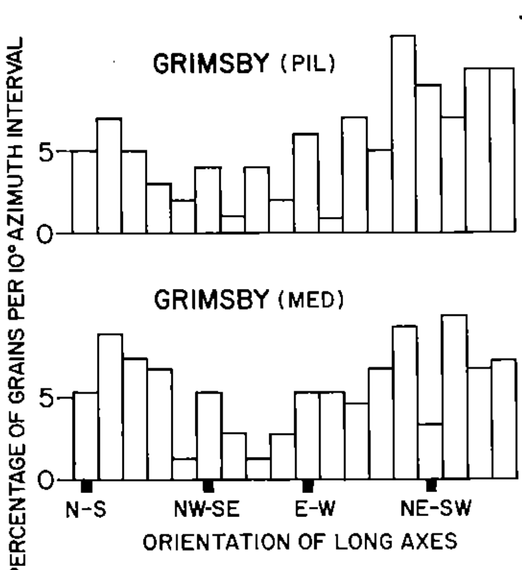

Fig. 6. Orientation of long axes of quartz grains in sandstone. Long axes of 150 grains have been measured from each thin section cut parallel to the horizontal plane of the outcrop. Azimuths have been divided into 10° intervals.

plastically sometime during its history. Such cannot be said of the Lockport dolomite. Other calcite-bearing units in western New York such as the Tichenor formation (*EVS*) and Grimsby formation (*MED*) also show mechanical twins (Fig. 1). The sandstones of western New York contain no quartz deformation lamellae.

STRAIN RECORDED BY CALCITE TWINNING

Residual strain was measured in a relatively small area of the Appalachian plateau more than 50 km from the nearest Appalachian fold. Based on data from this area alone the tie between the NNW compression indicated by the residual strain and the late Paleozoic Appalachian compression is just the correlation in orientation. A stronger tie can be established by sampling the Onondaga limestone eastward in New York State to Syracuse where folds of the Appalachian trend intersect the Onondaga limestone. Outcrops used for this sampling are shown in Fig. 1. One sample of middle Devonian Tichenor limestone of the Hamilton group (*EVS*) was also taken.

The mechanism of plastic deformation of the calcite is by the growth of twins parallel to the e-planes of the calcite (Turner, 1953). Calcite twins may be used for strain analysis in lightly deformed rocks such as those of the rocks of western New York (Groshong, 1972). Groshong's (1972) technique for measuring strain relies on Conel's (1962) observation that tensor shear strains (Γ) for individual grains of calcite is a function of the thickness of the mechanically twinned portion of the grain (t_i) by the equation:

$$\Gamma = \tfrac{1}{2} \tan \Psi = \frac{0.347}{t} t_i$$

where Ψ is the angular shear strain which is fixed at $38°17'$ for calcite twin gliding and t is the thickness of the grain normal to the twin set. The key assumptions of this technique are that the strain is homogeneous and that the amount of shear strain in a given twin set is a function of its orientation relative to the principal strain axis. Strain from five twin sets is used to solve a system of five simultaneous equations for the five unknown components of the strain tensor. Because simple shear of calcite by twinning is volume constant, the sixth component of the strain tensor, $\epsilon_z = -(\epsilon_x + \epsilon_y)$. More than five twin sets are used because the strain is rarely perfectly homogeneous and a best fit strain tensor must be computed using the method of least squares (Groshong, 1972).

In the area of the strain relaxation experiments calcite in the Onondaga limestone (*LRY* and *SFD*) and cement of the Grimsby sandstone (MED) contain mechanical twin lamellae. The orientation of the compression axes determined by the Turner method (Turner, 1953) for each of about 40 twin planes is plotted in Fig. 7. There is a diffuse concentration of compression axes with an azimuth slightly west of north for both the limestone and sandstone. This pattern is similar in orientation to that from twinned calcite sampled on the folded portion of the Appalachian Plateau (Engelder, 1979).

Strain associated with the mechanical twinning of calcite shows a N to NNW maximum compression of the Onondaga limestone across the entire western New York State (Fig. 1). In general the calculated three-dimensional strain tensor indicates that compressive strain occurred subparallel to the horizontal plane whereas extension dipped at a steeper angle (Table II). With one exception the maximum compression plunges less than 40° and in only two cases is the plunge of the maximum extension less than 40°C. This strongly suggests that the orientation for deviatoric strain was consistent regardless of the distance of the outcrop from the nearest folds on the Appalachian plateau.

Strain was higher in the samples of Onondaga limestone from the eastern end of the outcrops (Table II). The decrease in strain from east to west is reasonable in light of the tectonics of western New York. The outermost major fold mapped by Wedel (1932) on the Appalachian Plateau is the Lodi anticline which strikes east—west south of *SYR* and *OAK* and then assumes an ENE strike farther west (Fig 1). Thus Onondaga limestone west of *OAK* outcrops progressively farther from the northwestern fold in the Appalachian foreland fold and thrust belt. Local folds are found 5 km south of *SYR* and 25 km to the southeast of OAK. A plot of compressive strain versus distance from the nearest Appalachian fold shows a systematic

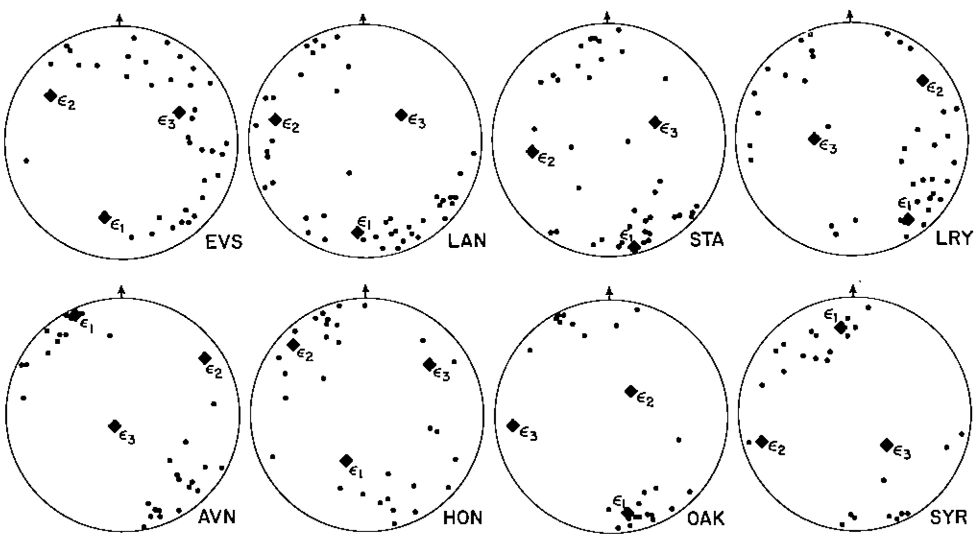

Fig. 7. Dynamic analysis of the mechanical twins in calcite for given samples of Onondaga limestone and one sample of Tichenor limestone (*EVS*). Compression axes for individual grains are shown in lower hemisphere, equal-area projection. The principal axes of strain determined by the Groshong method (Groshong, 1972) are indicated by diamonds with ϵ_1 the maximum principal compressive strain and ϵ_3 the maximum principal extension strain.

TABLE II

Three-dimensional strain tensors indicated by the mechanical twinning of calcite. Strain calculated using the method of Groshong, 1972 (compressive strain is negative)

Section	No. of thin sets	Negative expected values (%)	Principal strains (%)	Extimated error ±	Bearing (°)	Plung (°)	Formation
SYR	25	32	1.48	0.42	135	58	Onondaga
			0.68		254	16	
			−2.16		352	26	
OAK	24	25	1.80	0.57	264	17	Onondaga
			−0.45		44	67	
			−1.35		170	13	
HON *	30	40	4.16	1.81	51	33	Onondaga
			1.18		314	11	
			−5.34		208	54	
AVN	38	43	0.82	0.24	229	75	Onondaga
			−0.15		53	15	
			−0.66		323	1	
LRY	39	45	0.25	0.06	272	63	Onondaga
			0.03		51	21	
			−0.22		147	16	
SFD	51	37	0.70	0.08	70	55	Onondaga
			0.11		262	34	
			−0.59		169	5	
LAN	52	20	0.20	0.04	56	58	Onondaga
			0.03		284	23	
			−0.18		184	22	
EVS	36	29	0.19	0.05	64	42	Tichenor
			−0.08		305	28	
			−0.11		193	34	
MED	28	40	1.40	0.25	241	52	Grimsby
			−0.29		142	7	
			−1.10		47	37	

* Values for principal strains are high because of grains with thick twins and negative expected values.

decrease with increasing distance (Fig. 8). Data from *HON* are omitted from this plot because of the influence of thick twins with negative expected values. This example of the gradual dissipation of orogenic strain with distance away from the orogenic belt ties the nonrecoverable strain at *LRY* and *SFD* with the Appalachian compression. The magnitude of compressive strain at *SYR*, *OAK* and *HON* is similar to that reported for the folded Appalachian plateau (Engelder, 1979).

Calcite cemented Grimsby sandstone from *MED* also shows a compres-

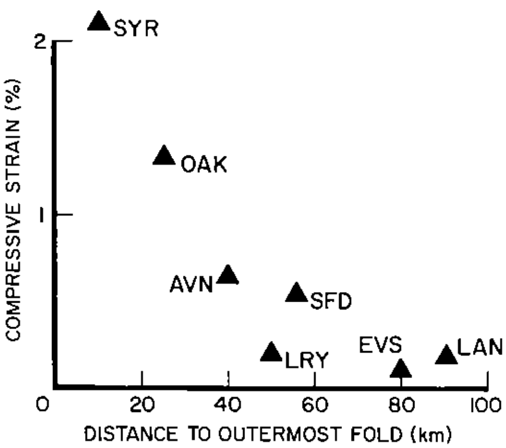

Fig. 8. Strain associated with the mechanical twinning of calcite vs. distance from perceptible folding on the Appalachian plateau.

sion oriented NNW although the maximum compressive strain dips at 37° (Table II). A compressive strain of equal magnitude affected both the Onondaga limestone and Grimsby sandstone.

SOLUTION CLEAVAGE

Another deformation structure which decreases in prominence from east (*SYR*) to west are solution cleavage planes. Groshong (1975a) describes these features near *SYR* and they are prominent at *OAK* as E—W vertical planes which are classified as weak cleavage according to Alvarez et al. (1978). Solution cleavage becomes much less obvious west of *OAK*. Solution cleavage is a common structure of the folded Appalachian plateau (Engelder, 1979).

DISCUSSION

Sbar and Sykes (1973) present data from hydrofracture measurements and earthquake fault-plane solutions from parts of New York State which indicate that the current tectonic stress has an ENE maximum compression. The fault-plane solution of Fletcher and Sykes (1977) for the Clarendon—Linden fault deviates from this ENE trend because of the strike of the fault. These data suggest that the relaxation of an applied component of strain during overcoring should have yielded an ENE expansion. Because very little strain was relieved in the ENE direction and because there is no evidence suggesting a local deviation of the applied stress field to NNW, we conclude that little, if any, applied stress is transmitted to the surface rocks of western New York. The similar orientation for maximum expansion of the initial and double overcore into Grimsby sandstone and Lockport dolomite does suggest that the strain relieved by the initial overcore was purely residual. Apparently a reorientation of residual strains does not occur upon partial relief of residual strains provided the introduced free surface is cylindrical. This is consistent with Engelder et al.'s (1977) experiments on Barre granite in which microcracks controlled the orientation of strain relaxation following both initial and double overcores. The residual strain in the sandstone and limestone does differ from that in the dolomite. The compression for the residual strain in the limestone and sandstone has the orientation of the late Paleozoic compression for the central Appalachian foreland fold and thrust belt. The dolomite shows a random orientation for residual strain.

Models for the locking and relaxation of residual elastic strain include the two-element model of Friedman (1972) where a "locking" element prevents the "locked" element from relaxing. Swolfs et al. (1974) and Tullis (1977) state that these two-element models must have equilibrium volumes much larger than sample size (e.g., Voight, 1974) or much smaller than sample size (e.g., Voight and St. Pierre, 1974). However, Tullis (1977) concludes that some overcoding data do not appear to reflect a residual strain with either very large or very small equilibrium domains; with this conclusion he suggests that non-elastic behavior must have occurred during overcoring. Nichols and Savage (1978) present data showing time-dependent relaxation upon overcoring. Rocks at the Niagara hydropower plants certainly show this characteristic (Lee and Klym, 1976).

An easily visualized mechanism for non-elastic behavior upon overcoring is the opening of microcracks (Tullis, 1977). Engelder et al. (1977) present evidence for the opening of microcracks during the overcoring of Barre granite where the normal to the microcracks is parallel to the direction of maximum expansion and the core's most compliant direction. However, microcracking is evident in neither Grimsby sandstone or Lockport dolomite. In addition, maximum expansion is parallel to the stiff direction of the sandstone.

Our experiments suggest that residual strain equilibrium volumes in the rocks of western New York must at least be small relative to the volume of the core. This has to be true for X-ray analysis to show a residual strain, but this must also be the case for a double overcore to expand with a consistent orientation relative to the initial (Tullis, 1977). However, these data do not preclude the possibility of a residual strain with an equilibrium volume much larger than the overcore. This residual strain may have been locked in a manner such as proposed by Voight (1974). Regardless, the mechanism for locking and relaxation of residual strain remains unclear. Likewise, it is unclear why the sandstone and limestone strains should show an Appalachian trend whereas the dolomite does not.

It is of interest to compare crude estimate of stress difference ($\sigma_1 - \sigma_3$) calculated from overcoring, X-ray analysis and calcite twinning. Using a value for linear compressibility of 3.1 mb^{-1} for Grimsby sandstone and the initial strain relaxation for *PIL* ($\epsilon_1 - \epsilon_2 \approx 2 \cdot 10^{-4}$), a stress difference of about

65 b is estimated. Other overcoring data yield smaller stress differences. For X-ray analysis of residual strain, Friedman (1972) uses a value of $0.89 \cdot 10^6$ b for Young's modulus perpendicular to the $\{32\bar{5}4\}$ planes of quartz. Changes in the *d*-spacing of $\{32\bar{5}4\}$ in quartz from *PIL* multiplied by the Young's modulus gives an estimated stress difference of 81 b. Jamison and Spang (1976) use calcite twin lamellae to infer a stress difference at the time of twinning. Their estimate is based on a critical resolved shear stress of 100 b for twin gliding in calcite (Heard, in Friedman, 1967). Calcite grains from *LRY* and *SFD* rarely contain more than one twin set. This places the differential at the time of late Paleozoic deformation between 200 and 333 b. This is probably a high estimate because of the unknown effect of duration of load for which the number of twinned grains increases with time (Friedman and Heard, 1974). It is certainly conceivable that a 65—80 b differential stress maintained from late Paleozoic to present is high enough to twin calcite.

Using a modulus of $6 \cdot 10^5$ b for the Lockport dolomite (Lee and Lo, 1976), overcoring of the Lockport dolomite in western New York indicates the relief of about 80 b stress difference. Apparently such a stress difference was not large enough to mechanically twin dolomite whereas it was more than enough to twin calcite. This relationship is consistent with laboratory data on the relative strengths of calcite and dolomite (Higgs and Handin, 1959).

The fracture patterns observed beyond the last open folds of the Ouachita Mountains in Oklahoma (Melton, 1929) and the Appalachian plateau in Ohio (Ver Steeg, 1942, 1944) are not obvious in the same structural position in western New York (Isachsen and McKendree, 1977). This change in structural style suggests a change in state of stress along strike of the Appalachian trend provided the increased prominence of fracturing in Ohio is not lithologically or stratigraphically controlled. It is not unknown why a prominent NNW extension fracture is not seen in the Onondaga limestone and Grimsby sandstone particularly because: (1) the NNW extension fracture is most prominent in the Devonian rocks within the folded portion of western New York (Parker, 1942); (2) Friedman and Logan (1970) show that extension fractures are more likely to propagate parallel to the residual strain direction of maximum shortening (i.e., NNW to NW in the Grimsby sandstone).

Voight (1974) presented a model for the locking-in of orogenic stresses by changing the frictional properties of the décollement fault zone while the décollement sheet is under load. For this model to be applicable to western New York the present concept of Appalachian plateau décollement tectonics must be modified. The residual strain in the Lockport dolomite and the residual and permanent strain in the Grimsby sandstone suggest that a décollement fault zone must be below the Grimsby. Current concepts of plateau décollement are expressed by Gwinn (1964) who points out that most plateau folds die out at the level of the thick salts of the Upper Silurian Salina group. In this view the level of the décollement is above the Grimsby sandstone and Lockport dolomite but below the Onondaga limestone. Furthermore, Rodgers (1970) notes, "The limit of more than very mild folding corresponds quite well with the limit of relatively thick salt (75 m) in the upper part of the Upper Silurian Syracuse formation (Rickard, 1969)". If Voight's (1974) model for locking of residual strain is applicable to western New York, our data require a second décollement below the Grimsby formation. Possible candidates are the Utica and Canajoharie shales which are 113 m thick in central New York (Colton, 1970).

A second and perhaps more radical interpretation of the residual strain data from the Grimsby sandstone is that both the Paleozoic section and the basement were subject to a NNW compression during the late Paleozoic. Strain data do not require more than about 1% compressive strain. Maybe a "rigid" basement can absorb such a small compressional strain.

CONCLUSIONS

Using data from X-ray analyses, mechanical tests and petrographic observations, I infer that the rocks of western New York contain both a residual and permanent strain imposed during transmission of tectonic stresses during the development of the Appalachian foreland fold and thrust belt to the southeast. Incipient deformation at the leading edge of the central Appalachian foreland fold and thrust belt includes: (1) residual strain which is recoverable upon overcovering; (2) an elastic distortion of quartz grains which is detected by X-ray techniques; (3) a plastic deformation by mechanical twinning of calcite; and (4) a N to NNW shortening by solution cleavage. This very early stage of deformation imprints a small (~5%) velocity anisotropy on the mechanical properties of the Grimsby sandstone. A gradual decrease in NNW-directed compressional strain recorded in the mechanical twinning of calcite in Onondaga links the incipient deformation to more strongly deformed portions of the Appalachian plateau.

The mechanical behavior of the rocks during incipient deformation in the field is consistent with the laboratory behavior of the same rocks under light loads in the following respects: (1) all rocks show a component of elastic strain; (2) the critical resolved shear stress for the mechanical twinning of calcite is lower than that for the mechanical twinning of dolomite.

Residual strain data from western New York require some mechanism for NNW shortening under the classical décollement surface in the Upper Silurian Salina salts. Two possibilities are a décollement below the Grimsby in Ordovician shales or a general crustal shortening during the late Paleozoic.

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