

Appalachian Stress Study

2. Analysis of Devonian Shale Core: Some Implications for the Nature of Contemporary Stress Variations and Alleghanian Deformation in Devonian Rocks

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Detailed stress measurements in three boreholes penetrating horizontally bedded Devonian siltstones, sandstones, and limestones above a prominent salt decollement in the Appalachian Plateau of western New York have revealed variations in horizontal stress magnitudes which correlate with lithologic and stratigraphic units in all wells. High differential stress levels (up to 20 MPa) were found in shales of very high clay content, contrary to the proposition that such materials have negligible long-term strength. Elastic modulus data show that stiffer beds generally host higher stress levels and suggest that sand/shale stress contrasts result in large part from elastic shortening of the section in response to regional ENE compression. No correlation between stress and Poisson's ratio was found. However, a major systematic drop in stress level within the generally massive shales, which occurs across a group of sand beds near the base of the Rhinestreet formation, appears to be of different origin. The stress offset corresponds to the top of a section which we conclude, on the basis of local and regional total strain data derived from chlorite fabric measurements, once hosted abnormally high pore pressures. The total strain data also suggest the entire section above the salt has been uniformly shortened during Alleghanian compression. To explain the stress discontinuity, two kinematic patterns for Alleghanian deformation of the Devonian section are proposed, both involving abnormal pore pressure development in the sub-Rhinestreet section in response to limited drainage of fluid. Drainage of this paleo-overpressure is the best available explanation of the stress offset, although an additional remnant stress component must also be present to satisfy the stress data precisely.

INTRODUCTION

Price [1974] suggested that contemporary stresses in sedimentary basins might be quantitatively synthesized as a superposition of several constituent stress fields imparted during the burial and subsequent history of the strata. Some examples of potential sources of stress are gravitational loading or unloading [McGarr, 1988], contemporary tectonic compression [Voight, 1969; Sbar and Sykes, 1973], crustal flexure [Clark, 1982], and, of particular concern here, remnant stresses locked in from events in the past [Voight, 1974; Voight and St. Pierre, 1974]. The role of remnant stresses is particularly controversial since we know so little about the low-pressure rheological properties of common sedimentary rocks which might admit relaxation of shear stresses over comparatively short geological time scales [Warpinski, 1986]. An insight as to the respective importance and nature of these sources of stress could, in principle, be gained from examination of detailed stress data from regions in which the tectonic history of the sedimentary section is comparatively well known. For example, from such "high-resolution" stress data it might be possible to recognize the localized signature of a remnant stress component and, through

knowledge of tectonic background, to identify the mechanism and perhaps epoch of origin. In this paper we investigate the nature of stress variations observed in a sandstone/shale sequence in the Appalachian basin. These data are described by Evans *et al.* [this issue] (hereinafter referred to as paper 1) (or "P1" in figure citations). We assume that the reader has paper 1 available. The regional counterpart to this essentially local study is presented by Evans [1989].

The stress data show a clear systematic contrast in stress magnitude between shales and comparatively thin beds of sandstone and limestone. The profiles of stress in the shales also suggest the presence of an unusual component which is neither gravitational nor broad-scale "tectonic" in origin. The anomaly is stratigraphically correlated with a section of sediments which Engelder and Oertel [1985] have inferred, on the basis of regional compaction and joint propagation data, once hosted abnormally high pore pressures. To examine the association further, an X ray pole figure goniometer [Oertel, 1983] was used to measure total strains (or "finite strains" in British usage) resulting from compaction and tectonic deformation. The measurements were made on oriented core samples taken from throughout the sedimentary section in question. Static elastic moduli of the sands and shales were also measured. Aside from defining compaction throughout the stress-tested section, the results have

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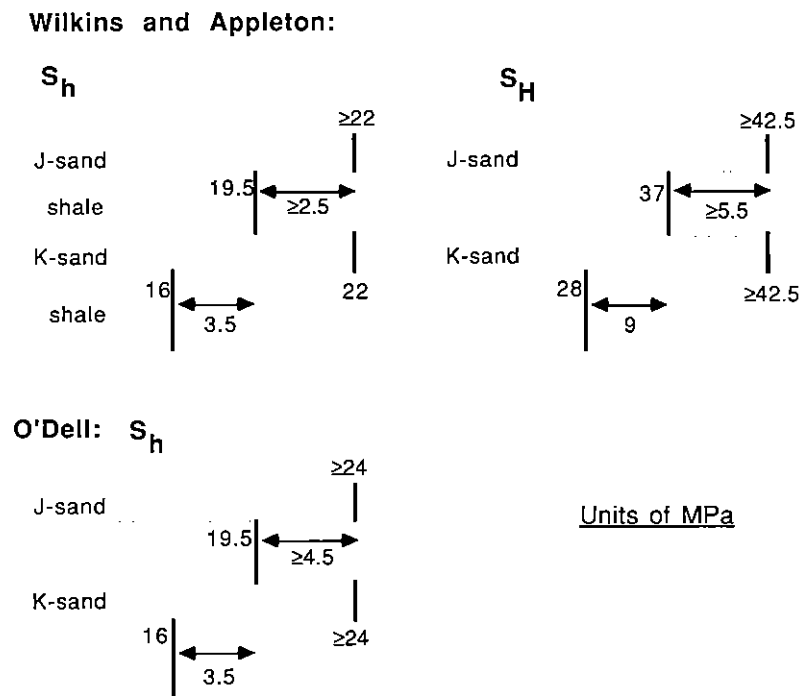


Fig. 1. Illustration of estimated stress magnitudes in sand and neighboring shale beds near the Rhinestreet base and the implied stress contrasts.

posed with common stratigraphic horizons aligned. For discussion of these data, we define three different stress regimes.

1. An upper regime lies above the H-sand horizon, in which the ISIP values define essentially linear trends that lie on or systematically above the lithostat for each well (see Figure 8 of P1). In paper 1 we show that these near-lithostat trends correspond to the predicted vertical stress profile for each well, and hence S_h is at least as great as S_v in this upper regime.

2. A transition zone lies between the H-sand and the K-sand horizons, in which S_h in the shales declines below S_v , the level of "clipping," whereas S_h in the sands remains at least as great as S_v . Major drops in shale stress occur across the J- and K-sands, and hence the contrast between sand and shale S_h levels becomes more pronounced with depth. This lithology-correlated stress contrast is seen most clearly in Figures 8 and 10 of P1.

3. A lower regime lies below the K-sand, in which S_h estimates obtained in shales at common stratigraphic levels are identical from well to well (Figures 9 and 10 of P1), despite varying overburden, and are significantly less than S_v . An ISIP measured in the Tully limestone, however, contrasts strongly with those in neighboring shales in that it lies on the anticipated S_v trend. Thus S_h is probably at least as great as S_v in that bed.

There are two aspects to the stress data that we seek to explain: the nature of the lithology-correlated stress contrasts, and the nature of the large negative offset in stress magnitude that occurs near the base of the Rhinestreet member of the West Falls group. The magnitudes of these features, which extend laterally over at least the 3 km sampled by the three boreholes, are illustrated in Figure 1. The principal drop in shale stress in the transition zone occurs across the K-sand bed, where S_h falls from approximately 19.5 MPa in the shale above to 16 MPa in the shale

immediately below. The corresponding drop in S_H is from 37 to 28 MPa. The offsets in S_h and S_H are thus 3.5 and 9.0 MPa, respectively. No systematic reorientation of principal axes accompanies this offset (Figure 9 of P1). The precise magnitude of the lithology-correlated contrasts in stress within the transition zone (and about the Tully limestone) is uncertain, owing to the likely underestimation of S_h in the sands and limestone. Data from the Wilkins and Appleton wells place lower bounds on the stress magnitudes in both K- and J-sands of 22 and 42.5 MPa for S_h and S_H , respectively. This suggests that the contrast in S_h and S_H between these sands and the intervening shale is at least 2.5 and 5.5 MPa, respectively. In fact, if the horizontal stresses in the sands are also laterally uniform, then the lower bound on S_h contrast between these sands and intervening shale is given by the O'Dell well ISIP contrast of 4.5 MPa. The S_h contrast between the K-sand and the underlying shale is at least 8.5 MPa.

Pursuant to an explanation of the stress features, we obtained oriented core samples from two neighboring wells for physical property measurements. The wells are two of a series of 31 Appalachian basin boreholes drilled as part of U.S. Department of Energy's Eastern Gas Shale Program (EGSP) and are denoted as NY1 and NY4 in Figure 1 of P1. Of greater importance is well NY1, located 20 km northwest of the South Canisteo wells, for it penetrates an almost identical stratigraphic section. The similarity in formation lithology is evident in Figure 2, which compares portions of the natural gamma, neutron porosity, and density logs obtained from the Wilkins and NY1 wells in the stratigraphic vicinity of the Rhinestreet base. The geophysical signature of respective formations is almost identical, although of the sands recognized in the Wilkins well, only the K-sand (alias Grimes sandstone) is sufficiently extensive to appear in NY1. Two types of studies were performed on the core samples. The first used a triaxial press to determine the static

implications for the kinematic history of the section during Alleghanian deformation.

STRUCTURAL, STRATIGRAPHIC, AND TECTONIC BACKGROUND

The stress study was conducted in Devonian shale rocks of the Appalachian Plateau near the town of South Canisteo in western New York (see Figures 1 and 2 of P1). The pertinent stratigraphic section is shown in Figure 3 of P1 which corresponds to the cross section A-A' of Figure 2 of P1. There is no vertical exaggeration. Bedding is essentially horizontal, dipping approximately 0.5° to the south. For a detailed description of local stratigraphy the reader is referred to the work of *Van Tyne* [1980, 1983]. The lowermost strata of relevance here are the Silurian salt beds of the Salina Group. When considered together, these beds define a continuous salt horizon that underlies most of the Appalachian Plateau (Figure 1 of P1). The salt exerted a profound influence on the development of structures in the plateau region during Alleghanian compression through its mechanical behavior as a decollement surface [*Frey*, 1973; *Davis and Engelder*, 1985]. Regional stress maps derived from commercial hydrofracturing operations show that the salt continues to influence the mechanical response of the basin sediments to contemporary tectonic stress [*Evans*, 1989]. In the Devonian section above the salt, inferred least horizontal stress magnitudes S_h generally approach or exceed the overburden, whereas in the section below the salt they are significantly less. Orientation of maximum horizontal stress in both sections parallels the structural trend of the basin and averages ENE. The discontinuity in S_h trend with depth suggests that the salt acts effectively to decouple the horizontal stress systems in the overlying and underlying sections, and hence we consider it to define the base of relevance to this study. Overlying the Salina are the carbonates and evaporites of the Bertie Group, the top of which is truncated by an unconformity zone containing the locally intermittent Oriskany sandstone. Above this lies the Onondaga limestone and the "Devonian shale" section.

The Devonian section consists of alternating black and gray prodeltaic turbidite piles (mudstones and siltstones) the color of which reflects differences in organic content [*Van Tyne*, 1983]. The sequence has been interpreted by *Ettensohn* [1985] as denoting cycles of basin subsidence and quiescence in response to proposed collisional phases of the Acadian orogeny. Within this framework, the Onondaga and Tully limestones represent periods of tectonic quiescence between two major tectonic cycles, and the alternating black and gray "shales" represent subcycles reflective of episodic variations in water depth (Figure 3 of P1). Below the Sonyea Group, the lithology is dominantly calcareous, with the cyclical recurrence of calcareous siltstones grading upward into limestones [*Cliffs Minerals Inc.*, 1982]. Above the Sonyea, quartzitic clastics begin to appear with quartz-rich beds becoming increasingly common above the black shale base of the Rhinestreet shale. These quartz-rich beds, which range in thickness from 5 to 13 m, will be referred to as "sands," although in detail they are composed of intercalated beds of fine-grained sandstone and siltstone. They are identified by the prefixes B-K in Figure 3 of P1 and are widespread but rarely individually extensive on scales greater than tens of kilometers. One exception is the K-sand, known locally as the Grimes sandstone, which has been

mapped by *Bradley and Pepper* [1938] in adjacent counties to the north and west, and is also present in the core from well NY1, the location of which is shown in Figure 1 of P1.

The section below the sands is cut by a family of blind reverse faults which ramp up from the Salina salt detachment and show displacements of up to 15 m at the level of the Oriskany (A. Van Tyne, unpublished data, 1985). The larger of these northeast trending faults extend sufficiently high in section to cut the Tully limestone, but probably not the Grimes sandstone. The widespread nature of this fault population, which *Gwinn* [1964] and *Frey* [1973] have described as underlying plateau anticlinal structures, is evident in Figure 1 of P1 where recognized faults cutting the Onondaga limestone in the vicinity of western New York are shown. They most likely formed in response to compression during the Alleghanian (Permian) orogeny [*Engelder and Engelder*, 1977] and are a common feature of decollement thrusting over salt [*Davis and Engelder*, 1985]. However, both *Bradley and Pepper* [1938] and *Beardsley and Cable* [1983] report evidence of disturbance prior to this time, possibly through diapiric action of the salt. Thus some doubt remains about the genesis and history of slippage on these faults, although there is no published evidence for significant post-Alleghanian displacement.

The question of how deeply the Devonian shales of the study area have been buried in the past is uncertain. Regional stratigraphic thicknesses suggest that about 750 m of Upper Devonian section has been eroded, thereby implying for example that the lowermost Middle Devonian Genesee shale (Figure 3 of P1) has been buried to a depth of at least 2 km. The amount of Carboniferous and Permian section that has additionally been eroded from the plateau remains disputed [*Johnsson*, 1986; *Karig*, 1987]. On the basis of orogenic flexure models, *Beaumont et al.* [1988] suggest that it may be as much as 2 km in the study area. *Epstein et al.* [1977] report a conodont color alteration index value of about 2 for the stratigraphic level of the Genesee shale in the vicinity of South Canisteo which corresponds to burial depths in the range 2.4–3.6 km. However, mechanical compaction data from chlorite fabric analysis suggest depths closer to 2 km [*Engelder and Oertel*, 1985]. In the light of the above uncertainties we take 1–2 km as reasonable bounds on the maximum thickness of eroded section.

IN SITU STRESS DATA

A detailed three-dimensional description of in situ stress in the section above the Hamilton group was obtained by conducting a series of hydrofracture measurements in three uncased 200-mm-diameter boreholes. The three holes are ~1.5 km apart and are located on the flank of a 200-m-high hill (Figure 2 of P1). A total of 43 measurements were made in the Wilkins well, 24 in the Appleton well, and 10 in the O'Dell well. These data are reported in detail in paper 1 and are summarized here only briefly. An important aspect of interpretation is the recognition that where least horizontal stress S_h exceeds the vertical stress S_v , the instantaneous shut-in pressures (ISIP) is "clipped" at the level of S_v . This is most likely due to rapid fracture rotation to the horizontal plane during propagation from the wellbore [*Evans and Engelder*, 1989]. Thus ISIPs which fall on the anticipated vertical stress represent only lower bounds to the true value of S_h . Figures 9 and 10 of P1 present a summary of results in which the stress profiles from all three wells are super-

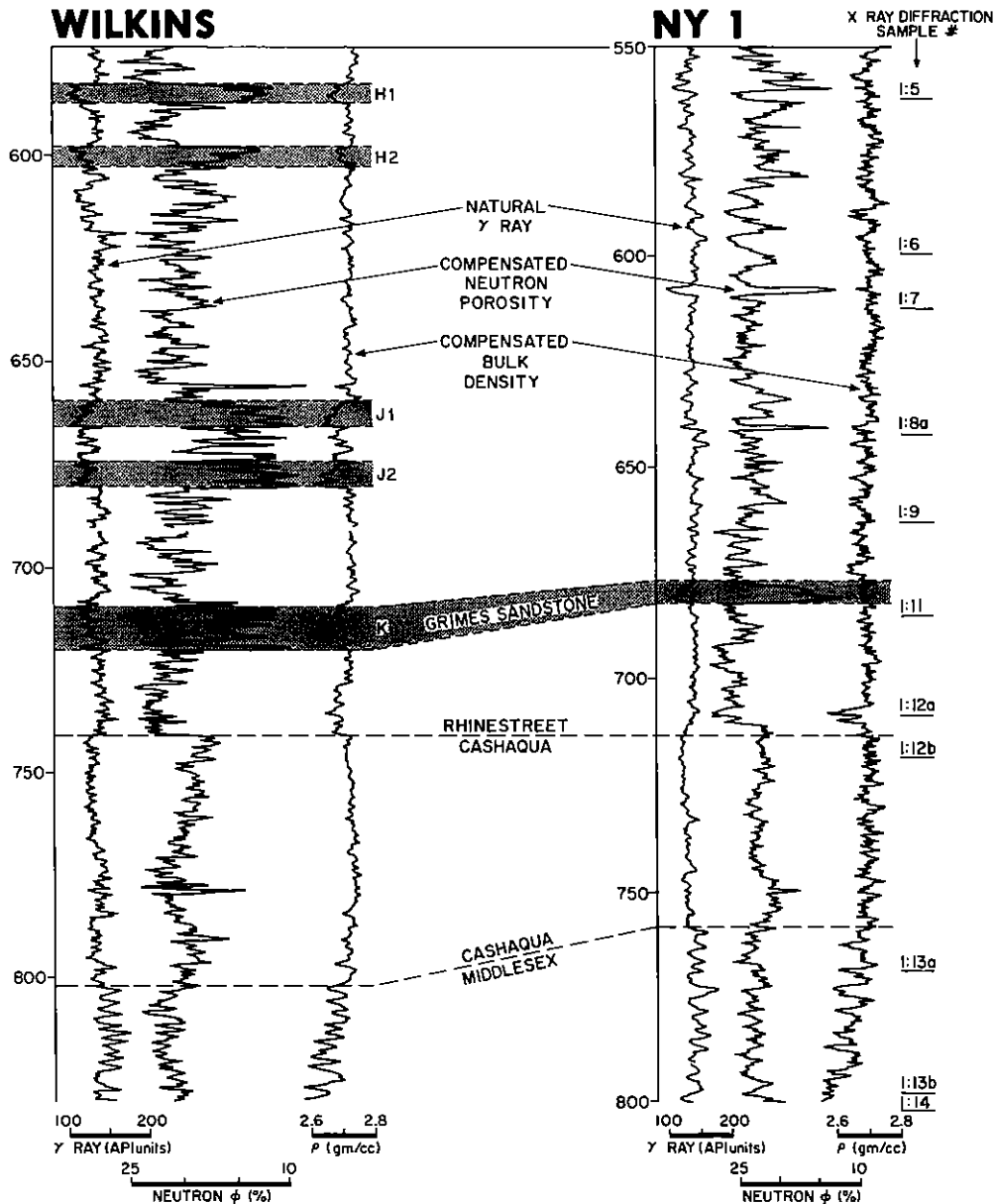


Fig. 2. Comparisons of geophysical log suites obtained in the same stratigraphic section of the Wilkins and the EGSP well NY1. The stratigraphic location of NY1 core samples used in the chlorite fabric analysis is shown on the right.

elastic constants of the various lithologies found in the section. The second used an X ray diffraction technique to estimate the total strains inherent in the samples, as recorded by the preferred orientation of chlorite grains.

STATIC ELASTIC CONSTANTS ON NY1 CORE AND THEIR IMPLICATIONS

A possible explanation of the stress features is that they result from contrasts in material elastic properties. A contrast in Poisson's ratio will produce a difference in the proportion of vertical stress coupled into the horizontal plane, and a contrast in Young's modulus will give rise to a proportionate difference in the stress developed in response to lateral compression. To investigate these possibilities, nine samples were prepared and tested to obtain estimates of uniaxial strain modulus E_{us} , Young's modulus E , and Poisson's ratio ν . All measurements were performed on 2.85-cm-diameter right circular cylinders, cut normal to bedding,

by steadily increasing axial stress between 20 and 45 MPa. Confining pressure for the measurements of E and ν was typically of the order of 15 MPa, and for E_{us} it averaged 25 MPa. Strains were measured with strain gages bonded either to a copper jacket or, in the case of the low-permeability shales, directly onto the sample underneath heat-shrink tubing. Difficulties were experienced in coring samples of the fissile shales which tended to part along bedding planes, and in several cases this limited sample lengths to less than two sample diameters, the point at which the sample end caps begin to corrupt the strains at the center of the sample. Consequently, the most relevant elastic constant that we could measure for all samples was the normal-to-bedding uniaxial strain modulus, a quantity which is unaffected by short sample length. We were unable to measure Young's modulus in the plane of the bedding. For the uniaxial strain measurements the output of the circumferential strain gage was maintained constant to within the equivalent of 3

TABLE 1. Results of Elastic Moduli Determinations

Depth, m	Sample	Formation	Description	E_{us} , GPa	E , GPa	Poisson's Ratio
310.3	1:2	Pipe Creek	light gray siltstone	31	30	0.11
409.1	1:3	Rhinestreet	dark gray silty mudstone	36	31	0.18
486.15	1:4	Rhinestreet	gray silty mudstone	22		
636.8	1:7	Rhinestreet	light gray siltstone	27		
682.6	1:10b	Rhinestreet (Grimes)	fine-grained sandstone	52	48	0.17
682.7	1:10h	Rhinestreet (Grimes)	fine-grained sandstone	52	46	0.17
684.0	1:10g	Rhinestreet	black silty mudstone	26	20	0.19
798.15	1:13b	Middlesex	black silty mudstone	23	17	0.13
801.95	1:14	West River	dark gray silty mudstone	32	26	0.17

μ strain. The results of the measurements are listed in Table 1. Young's modulus and Poisson's ratio estimates for samples shorter than two diameters are not included. Uniaxial strain modulus and Poisson's ratio are plotted against depth in Figures 3a and 3b. Formation tops are also indicated, although a better sense of the location from which the samples were removed is obtained from Figure 2, where all but the top three sample locations are indicated to the right of the geophysical logs.

The results suggest that bedding-normal static moduli for the shales lie in the range 20–35 GPa. The sand is much stiffer, with a uniaxial strain modulus value of 52 GPa. Profiles of dynamic Young's modulus obtained from sonic and density logs run in the Wilkins well also suggest that the sands are systematically stiffer than the neighboring shales [Plumb et al., 1987; Evans and Engelder, 1987] as is the Tully limestone. No systematic contrast in Poisson's ratio between sand and shale is evident in Figure 3b, a result which is also supported by the log data. Thus it is reasonable to conclude that the systematic contrast in stress level between the sand/limestones and the shales is due to horizontal elastic straining of the beds in response to contemporary compression. That is, the section is responding as a "layer cake" in which the stiffer layers develop a greater

stress as the section is uniformly shortened. The hydrofracture orientations show this direction to be ENE, which is consistent with the maximum compressive stress direction that is found over large areas of the eastern United States [Plumb and Cox, 1987; Zoback and Zoback, 1989; Evans, 1989], and is thought to result from the push of the Mid-Atlantic Ridge transmitted across the continental margin [Voight, 1969; Sbar and Sykes, 1973]. Thus it is possible to explain the lithology-correlated stress variations within a large-scale context, although the precise mechanics of coupling are complex. Regional mapping of stresses at various stratigraphic depths in the Appalachian basin has shown that a discontinuity in the depth gradient of least horizontal stress across the Silurian salt horizon, with S_h stepping from generally superlithostatic values above the salt to markedly sublithostatic values below [Evans, 1989]. This implies that the salt decollement serves to mechanically decouple the Devonian "layer cake" from the underlying strata and suggests that the horizontal strains developed in the two sections in response to ENE compression may be slightly different. It is noteworthy that within this compression-dominated scenario, bed-to-bed variations in stress magnitude are determined largely by contrasts in elastic stiffness rather than Poisson's ratio, as is assumed in many commer-

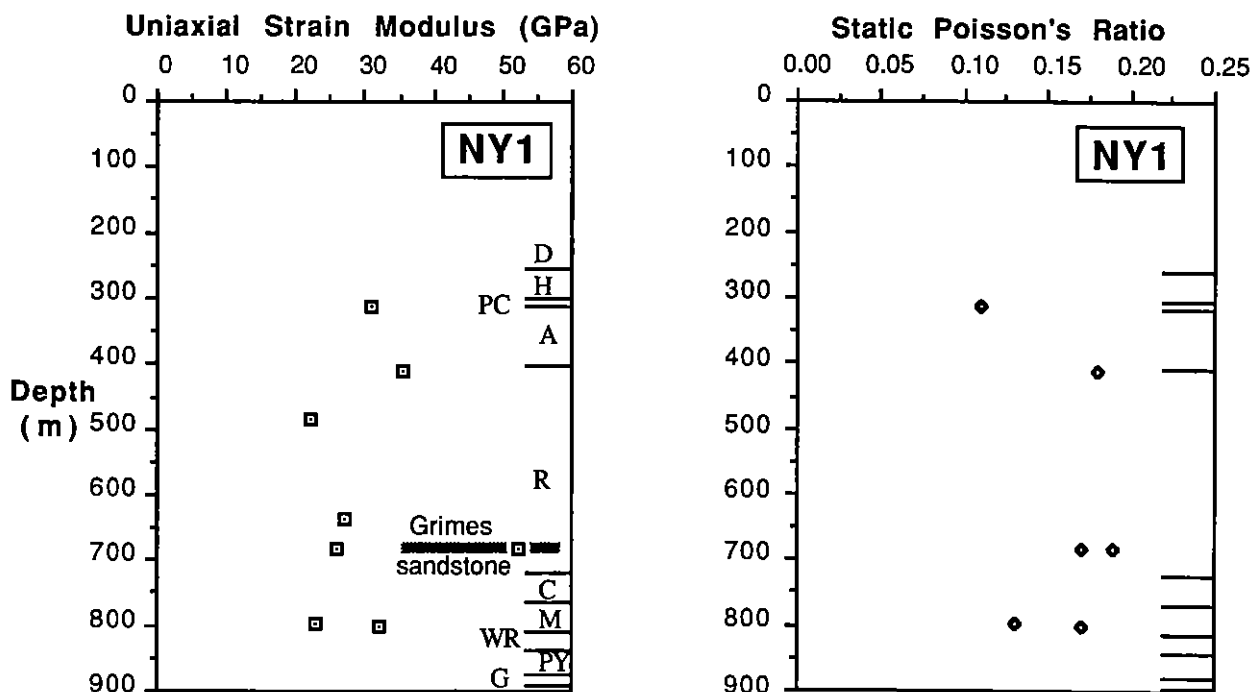


Fig. 3. Static elastic constants determined from cores samples recovered from well NY1. The legend to the stratigraphic index is given in Figure 3 of P1.

WELL NY1

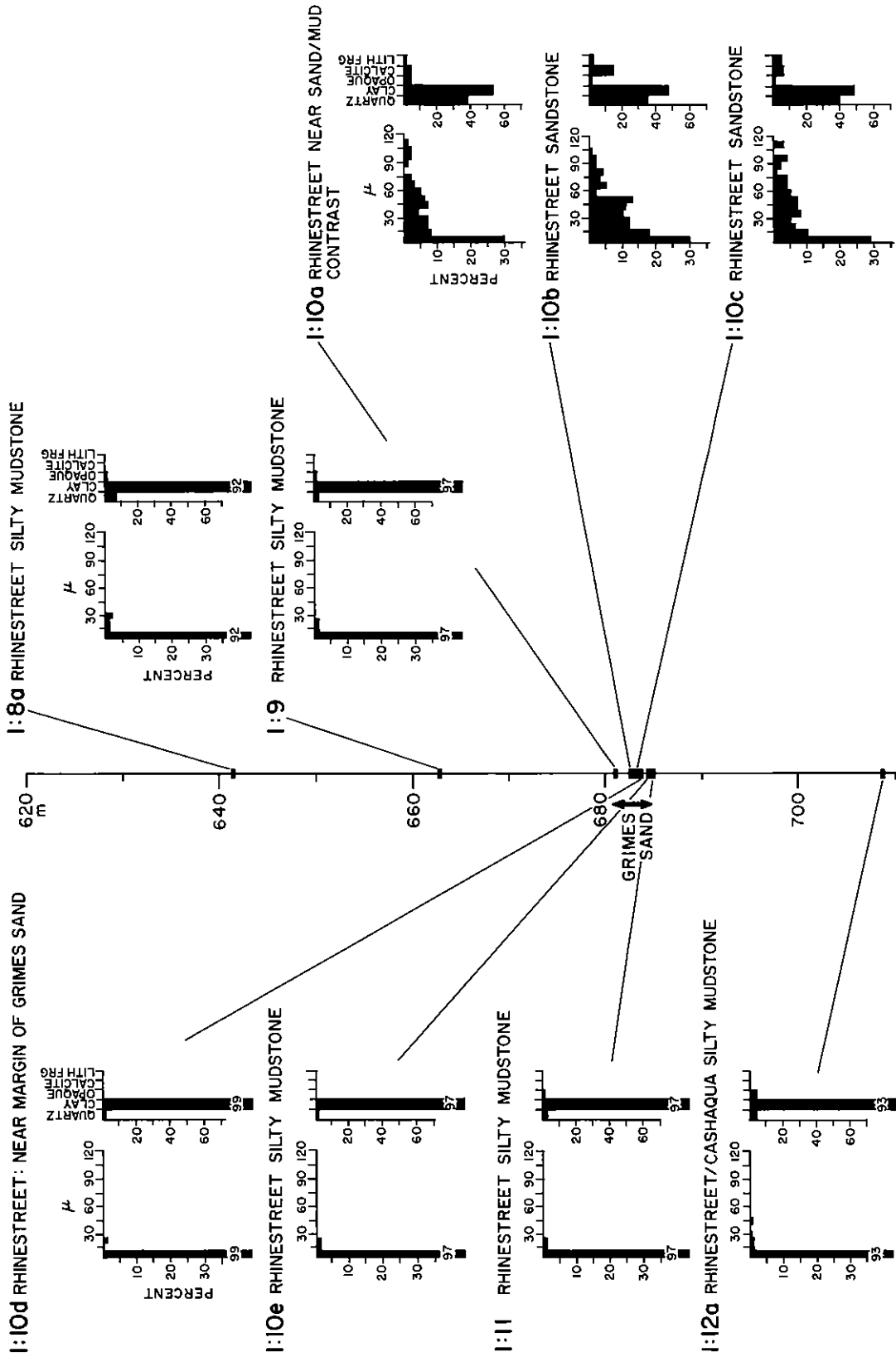


Fig. 4. Mineralogical composition of NY1 core samples determined from thin sections for sands and shales in the vicinity of the Grimes (K) sandstone.

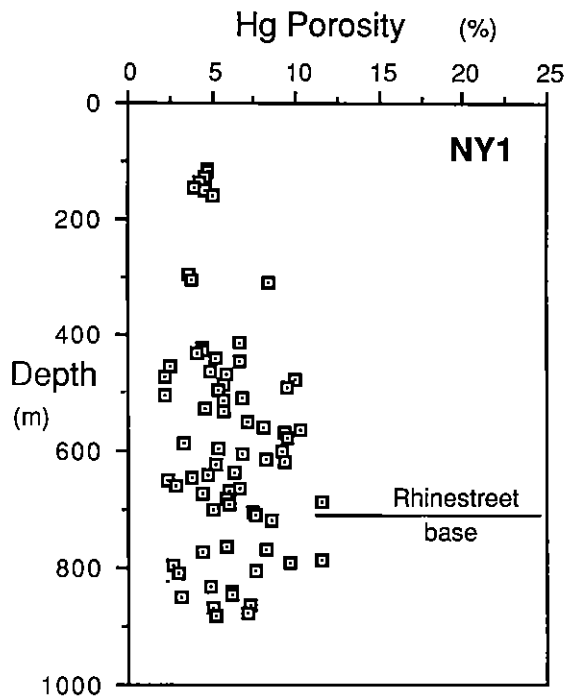


Fig. 5. Results of mercury porosimetry measurements conducted on NY1 core samples.

cial schemes for estimating stress contrasts [e.g., *Frisinger and Cooper*, 1985].

An important question is whether the elastic straining mechanism might also account for the drop in stress level below the K-sand. This would require that the bedding-parallel (horizontal) Young's modulus values of the shales below the K-sand be systematically less than those for the shales above. A sense of the magnitude of the contrast required can be obtained by noting that the lower bound on S_H contrast between the sands and the "upper" shales (above the K-sand) is ~ 5.5 MPa, whereas the contrast in S_H between the "upper" and "lower" shales is ~ 9 MPa. Thus, if the contrasts are to be ascribed entirely to ENE elastic straining and the lower bound on S_H in the sands is realized, then the requisite bedding-parallel modulus contrast between the "upper" and "lower" shales must be a factor of 9/5.5 greater than the modulus contrast between the sands and the "upper" shale. Although sand stress levels in excess of the measured lower bound will reduce this factor proportionally, for reasonable S_H stress levels the requisite modulus contrast between "upper" and "lower" shales remains considerable. The uniaxial strain modulus estimates shown in Figure 3 show no evidence of a contrast between "upper" and "lower" shales. However, these data describe modulus normal to bedding and do not necessarily discount the existence of a systematic contrast in bedding-parallel modulus. The dynamic modulus estimates derived from the sonic log measure a composite complicated property, although they too are biased toward measuring bedding-normal modulus. These estimates also show no evidence supporting systematically smaller moduli for the "lower" shales [*Plumb et al.*, 1987; *Evans and Engelder*, 1987]. The geophysical logs, presented in part in Figure 2, indicate that distinct physical changes occur between formations but show no features that clearly contrast between the "upper" and "lower" shales. A selection of results from thin section

analyses of the rocks that compose the study section are presented in Figure 4 and show that no pronounced change in mineralogical constituents occurs between shales above and below the Grimes (K) sandstone. Finally, in Figure 5 we show the results of mercury porosimetry measurements on core from NY1, reported by *Kalyoncu et al.* [1979]. Again, these data show no evidence of a systematic change in measured connected porosity near the Rhinestreet base. Thus despite strong evidence that the entire section is elastically strained due to ENE contemporary compression, resulting in stress contrasts between beds of significantly different stiffness, the material property data suggest that the major stress offset at the K-sand has a different origin. We shall return to the question of the nature of the stress offset after first discussing the X ray diffraction measurements and their implications for the deformational history of the section.

X RAY POLE FIGURE GONIOMETER MEASUREMENTS ON NY1 CORE

A total of 18 core specimens sampling the entire study section in the South Canisteo wells were selected for strain analysis using an X ray diffraction technique. Sample locations in the neighborhood of the Rhinestreet base are shown in Figure 2. An extensive review of the technique of total strain estimation from measurements of phyllosilicate preferred orientation has been given by *Oertel* [1983]. The objective of the technique is to estimate the distortion that a volume of mudstone or siltstone has undergone since it was deposited on the seafloor. We obtain a measure of the present preferred orientation of chlorite basal planes and assume that this describes the present preferred orientation of clay platelets initially deposited with their basal planes randomly oriented. As compaction proceeded and pore space was lost, the clay developed a preferred orientation with basal planes increasingly rotating toward parallelism with bedding. Compaction, or vertical strain, is correlated with the degree to which the clay platelets or their chlorite successors have become reoriented. Similarly, deformation in the horizontal plane, such as might result from shortening under the application of a tectonic compression, also results in a rotation toward parallelism with the plane normal to the shortening axis. The preferred orientation of the chlorite basal planes is measured by a pole figure goniometer [*Wenk*, 1985, pp. 39–40]. The theoretical framework for recovering quantitative estimates of strain from the measure of preferred orientation is that given by *March* [1932] (see also *Oertel* [1985]). In discussing total strains we adopt the convention that extension is positive.

It is important to appreciate that the output of the apparatus is an estimate of distortion only and contains no information about volume change. For convenience, the principal strains derived from the output are arbitrarily scaled such that the implied total deformation (compaction plus tectonic) would conserve volume. These strains, which we denote ϵ_1 , ϵ_2 , and ϵ_3 , are nominal values. To calculate the actual strain, a scaling factor is applied to the three principal stretches (strain augmented by unity). This factor is determined by making an assumption about the type of tectonic deformation that has occurred. For example, if we assume that areas in the bedding plane were unchanged by the tectonic deformation (i.e., "tectonic" horizontal shortening, ${}^b\epsilon_2$, is compensated by extension, ${}^b\epsilon_1$, along the orthogonal

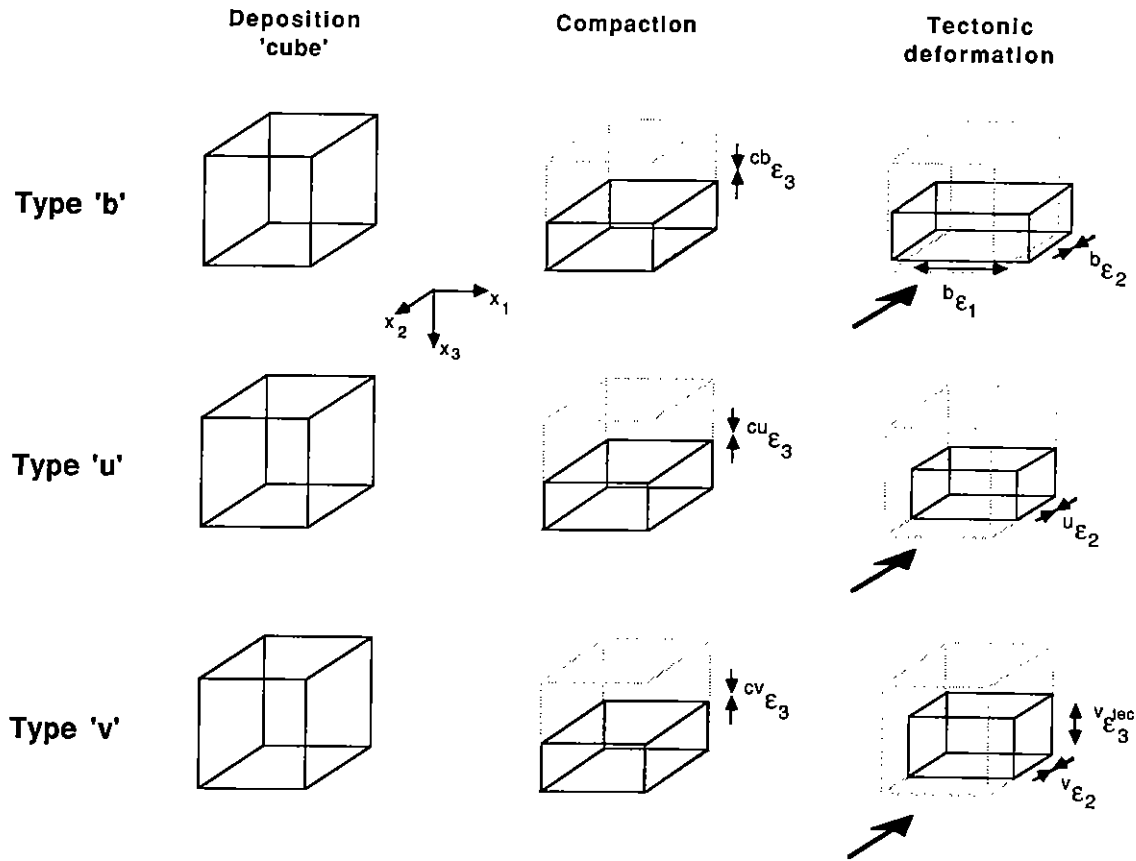


Fig. 6. The three types of tectonic deformation pattern considered to represent the kinematic response of the Devonian section to Alleghanian compression.

horizontal axis) and that no vertical tectonic strain occurred, then constancy of area gives

$$(1 + \epsilon_1)(1 + \epsilon_2) = 1 \tag{1}$$

The nominal, volume-preserving stretches, $(1 + \epsilon_1)$ and $(1 + \epsilon_2)$, can be scaled to satisfy (1) through multiplication by $1/\{(1 + \epsilon_1)(1 + \epsilon_2)\}^{1/2}$. Hence this factor transforms stretches obtained by assuming that volume was preserved by the total deformation to stretches which assume that bedding-plane area was preserved (distinguished by the superscript b). For this type of deformation, compaction, ${}^b\epsilon_3$, is given by

$${}^b\epsilon_3 = \left[\frac{(1 + \epsilon_3)}{\{(1 + \epsilon_1)(1 + \epsilon_2)\}^{1/2}} \right] - 1 \tag{2}$$

We consider three scalings which correspond to plausible tectonic deformation patterns for the northern Appalachian Plateau. These are illustrated in Figure 6. The first pattern (hereinafter referred to as type "b") is that described above. Under this assumption, volume is preserved by the tectonic deformation, and the estimate of compaction strain is a minimum. The second pattern, denoted type "u," considers that the tectonic deformation involved uniaxial shortening along one principal horizontal strain axis (x_2 axis), with no compensating extension along either the x_1 or vertical axis. In this case, the total deformation leaves lengths along the long axis (x_1 axis) unchanged, and the scaling factor is $1/(1 + \epsilon_1)$, where $\epsilon_1 > \epsilon_2$. For this case, compaction is given by

$${}^u\epsilon_3 = \left\{ \frac{(1 + \epsilon_3)}{(1 + \epsilon_1)} \right\} - 1 \tag{3}$$

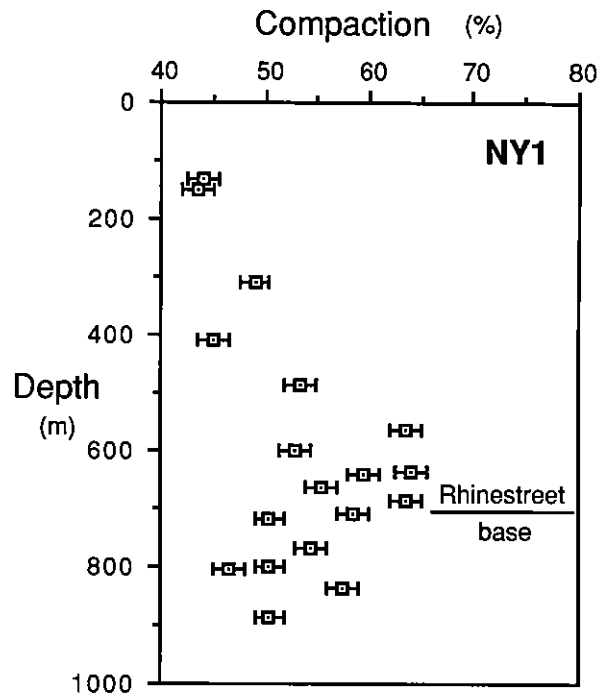


Fig. 7. Compaction estimates from chlorite fabric analysis. The "error bars" are not used in the conventional sense. Rather, they join compaction estimates derived assuming type "b" and type "u" deformation patterns. These correspond to minimum and maximum values of compaction strain under the assumption that vertical tectonic strain is zero.

TABLE 2. Results of Chlorite Fabric Analysis

Sample	Depth, m	Formation*	Compaction, %			Maximum Shortening, %			Azimuth of Maximum Shortening °E of North
			Type "b"	Type "u"	Type "v"	Type "b"	Type "u"	Type "v"	
<i>NY1</i>									
I:17	131	D	43	45	48	3	6	6	—
I:1	149.4	D	41	45	52	7	13	13	144
I:2	310.3	PC	48	50	53	3	6	6	135
I:3	409.0	A	44	46	50	4	8	8	160
I:4	486.2	A	52	55	61	7	13	13	158
I:5	562.7	R	62	65	69	7	13	13	169
I:6	599.3	R	52	54	58	4	8	8	163
I:7	636.7	R	63	65	70	7	13	13	209(151)
I:8a	641.6	R	58	61	66	7	13	13	162/198
I:9	662.6	R	54	57	63	7	13	13	134
I:11	685.2	R	62	65	69	7	13	13	169
I:12A	708.7	R/C	57	60	65	7	13	13	161
I:12B	718.7	C	49	52	56	4	9	9	151
I:13A	768.7	M	53	56	61	7	13	13	166
I:13B	798.3	M/WR	49	52	56	4	9	9	151
I:14	801.9	WR	45	48	54	6	11	11	151
I:15	836.1	PY	56	59	64	7	13	13	152
I:16	887.3	G	49	52	58	7	13	13	211(149)
Mean of 18 measurements			52	55	59	6	13	13	155
Error						+1; -1	+2; -2	+2; -2	+0.2; -7.6
<i>NY4</i>									
4:1A	927.8	G	57	60	65	7	13	13	229
4:1B	928.4	G	55	58	63	7	13	13	232
4:2B	1159.5	M	53	56	61	7	13	13	219
4:2B	1159.8	M	54	57	63	7	13	13	228

*D, Dunkirk; PC, Pipe Creek; A, Angola; R, Rhinestreet; M, Middlesex; WR, West River; PY, Pen Yan; G, Genesee black shale.

This yields the largest estimate of compaction for deformation schemes which do not admit the presence of a vertical tectonic strain component. The third scenario, denoted type "v," is the same as type "u" with the exception that shortening is wholly compensated by vertical extension. Thus volume is conserved under this assumption, unlike the second case. The scaling factor again reflects preservation of "long" axis and is the same as in the second case giving a vertical strain estimate of

$${}^v\varepsilon_3 = \{(1 + \varepsilon_3)/(1 + \varepsilon_1)\} - 1 \quad (4)$$

This vertical strain, however, must be interpreted as the cumulative effect of a negative compaction strain ${}^v\varepsilon_3$ and a positive tectonic strain ${}^v\varepsilon_3^{tec}$, compensating for the horizontal shortening $(1 + {}^v\varepsilon_2)$; that is, the tectonic component partly "undoes" the compaction strain which supposedly preceded it. The initial compaction strain is thus given by

$${}^{cv}\varepsilon_3 = \{(1 + {}^v\varepsilon_3)(1 + {}^v\varepsilon_3^{tec})\} - 1 \\ = \{(1 + \varepsilon_3)(1 + \varepsilon_2)/(1 + \varepsilon_1)^2\} - 1 \quad (5)$$

This assumption leads to a strict maximum estimate of the compaction strain. A more detailed discussion of the assumptions underlying equations (1)–(5) is presented by *Oertel et al.* [1989].

The resulting determinations of compaction are listed in Table 2 and plotted as a function of depth in Figure 7. The horizontal bars link compaction estimates derived for type "b" and "u" deformation patterns and hence span the range of permissible compaction estimates under the assumption of no vertical tectonic strain. They are not error bars in the usual sense. We find that compaction increases approxi-

mately linearly with depth, from 40% at the present surface to ~65% at the base of the Rhinestreet. Thereafter a marked drop is observed to ~50% in the underlying Cashaqua Formation. Examination of thin sections from shales taken from above and below the drop in compaction showed no evidence of systematic lithological differences which might explain it (Figure 4). This drop in compaction coincides stratigraphically with the drop in horizontal stress observed in the Canisteo wells.

The corresponding estimates of principal horizontal shortening amount and direction are also listed in Table 2 and are plotted for the case where the deformation preserved horizontal "long" axes (types "u" and "v") in Figure 8. The case where the deformation preserved areas in the horizontal plane (type "b") yields estimates of shortening which are one half those shown in Figure 8a. Data from both NY1 and NY4 are shown so as to extend the sampling throughout the whole of the section above the salt. The results indicate that the entire section has been uniformly shortened. Some scatter is evident, but it is not systematic. The amount of shortening is between 6% and 13% ($\pm 1\%$) depending upon the degree to which the tectonic deformation was area preserving or "long-axis" preserving. The direction of maximum shortening inferred from the NY1 data is N24.4°W (+0.2°; -7.6°). The uncertainty values in parentheses represent one standard deviation about the stated mean when the Gaussian averaging is performed. Prior to averaging, we apply the appropriate tensor rotation to each natural strain (natural logarithm of the principal stretch $(1 + \varepsilon)$) so that all data points are expressed in terms of a common coordinate system. For three core samples, orientation was either ambiguous or suspect. Both results are given in Table 2 for

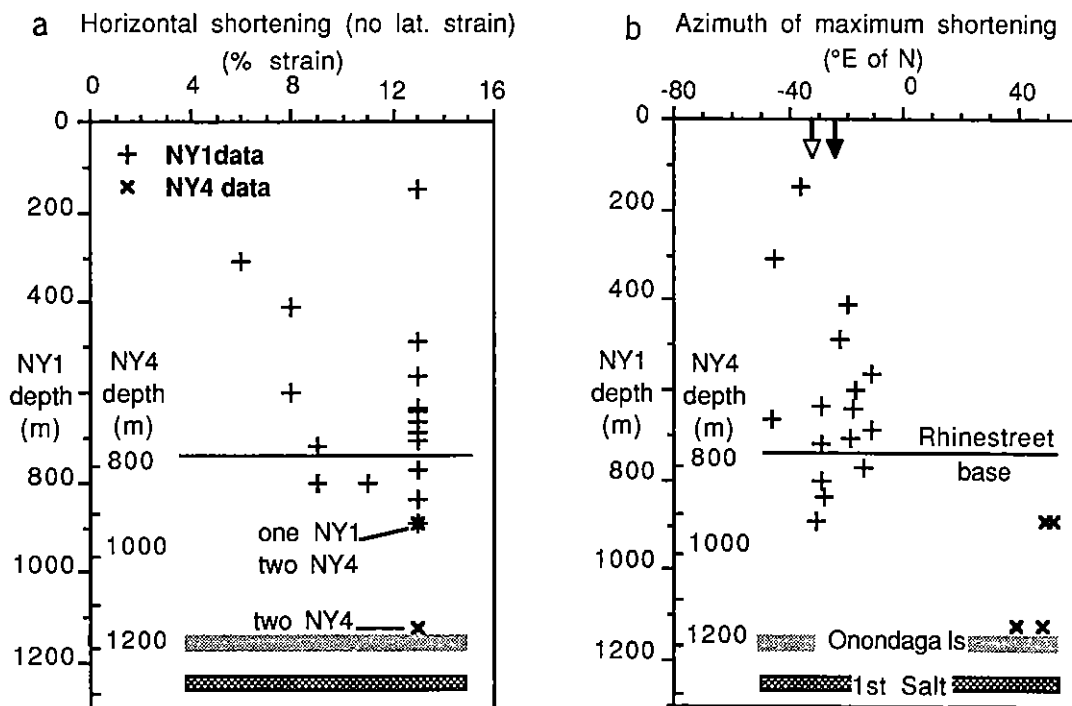


Fig. 8. Horizontal "tectonic" strain as determined from the chlorite fabric analysis of NY1 and NY4 core samples. (a) Estimates of maximum shortening strain, in percent, which apply for deformation types "u" and "v". (b) Corresponding orientation of the maximum shortening strains: the open arrow shows the direction of maximum shortening of deformed crinoids in the locality; the solid arrow indicates the normal-to-strike of anticlines and their underlying reverse faults in the study area. The NY4 orientation data are believed to be systematically in error due to some core recovery error.

the ambiguous case, and our suggested correction for misorientation is given in parentheses. These three samples were not used for the calculation of mean orientation. The orientation data from NY4 are suspect since velocity anisotropy data obtained from the same samples show an orientation which is $\sim 90^\circ$ rotated from a structure-parallel trend that is well established by other wells in the basin (Figure 10). Since there is no evidence to suggest that the well is drilled in a rotated block, we believe that the anomaly is due to error in core recovery and that the shortening direction at NY4 coincides with that at NY1. The results are in agreement with the observed deformation of crinoids near Andover, some 15 km SW of the NY1 site which, if interpreted as having resulted from a "long-axis" preserving strain, suggest a 14% shortening in a direction N33°W [Engelder, 1979]. Furthermore, the direction of shortening is perpendicular to the strike of anticlines and Alleghanian age reverse faults transecting the Canisteo region (Figures 1 and 2 of P1) and of mapped pencil cleavage in the area [Engelder and Geiser, 1979].

The question of which type of strain, either "bedding-area" or "long-axis" preserving, best describes the Alleghanian deformation remains uncertain, although there is some evidence that favors the latter. From an examination of the mechanisms responsible for crinoid deformation, Engelder [1979] concluded that the regional kinematic deformation pattern was better described as types "u" or "v" rather than type "b." The dominant mechanism responsible for crinoid deformation is stress solution. If shortening were compensated by orthogonal horizontal extension (i.e., "bedding-area" preserving), calcite redeposition around portions of the crinoid rim facing the extension direction might be anticipated, giving rise to "beards." These features are not

observed in crinoids found in the Appalachian Plateau. Additional evidence favoring "long-axis" preserving strain is given by consideration of the structures developed during shortening. The preferred-orientation data strongly suggest that the entire section above the salt strained uniformly (in plan view) during Alleghanian compression. The family of reverse faults ramping upward from the Salina salt beds suggest that at least some of the shortening of the Onondaga and Oriskany beds was accomplished by brittle failure. These faults die out upon reaching the Devonian shale beds which accommodated the strain in a ductile manner. If it is accepted that slippage on these faults was occurring during Alleghanian compression, then the absolute state of stress in these beds at the time of shortening must have conformed to "thrust" regime conditions, rather than "strike-slip." Hence, if stress solution was also occurring, the stress regime would not have favored the redeposition necessary to accommodate lateral extension. Rather, the pattern of strain would most likely mimic and enhance that accomplished by slippage on the faults, which implies type "u" or "v" deformation. The faults cannot accommodate lateral extension in compensation of the NNE directed shortening. Thus we conclude that the tectonic strain affecting the NY1 section is the result of Alleghanian compression and adopt the view that is better described by assuming preservation of "long" axis than of bedding-plane area.

The estimates of the amount and direction of shortening shown in Figure 8 are remarkably consistent throughout the part of the section where a decline in compaction is inferred, and demonstrate that the feature is not an error of measurement. Additionally, Engelder and Oertel [1985] performed an identical strain analysis on samples taken from broadly separated outcrops which expose the same Devonian sec-

tion. They also found evidence of undercompaction of the section below the West Falls group (of which the Rhinestreet is the lowermost member), although their results show much greater scatter than the NY1 data (which might perhaps be attributed to the geographical variation of their outcrop sources). Nonetheless, their results provide a basis to assert that the onset of undercompaction which we infer occurs at the Rhinestreet base near the NY1 site is a regional phenomenon and that it also occurs at the same stratigraphic level in the neighboring Canisteo wells. It is possibly related to the reduction in horizontal stress that occurs at the same level.

MODELS OF ALLEGHANIAN DEFORMATION

It is well known that undercompaction and pore fluid pressure in excess of hydrostatic are often correlated [Magara, 1978]. This correlation led Engelder and Oertel [1985] to interpret their results as the consequence of high pore fluid pressures in the section below the Rhinestreet which acted to limit compaction during burial and subsequent tectonism. Further evidence that this section was overpressured during Alleghanian times derives from a family of vertical joints which strike parallel to Alleghanian paleostress trajectories and were presumably driven by abnormally high fluid pressures [Engelder, 1985]. This family is largely confined to the section below the Rhinestreet. There is some evidence to suggest that these joints propagated prior to the period of major tectonic compaction on the plateau [Engelder, 1989], consistent with Nickelsen's [1979] observation that tectonic joint propagation preceded folding in the Valley and Ridge province. An implication is that near-lithostatic fluid pressures developed in the lower section early in the orogeny (i.e., prior to major shortening). At this time of joint propagation, the least principal stress was horizontal, although deviatoric stress was certainly very low, since the shear strength of the rock under the implied conditions of zero effective stress is essentially limited to the Coulomb cohesive strength [Etheridge, 1983]. As the reverse faults cutting the Onondaga suggest that a thrust regime prevailed during the episode of major shortening, the joint and shear fracture observations imply that a change in least principal stress direction from horizontal to vertical occurred, presumably after a small amount of shortening had been accommodated.

The question of interest here is how might a stratigraphically localized overpressure develop given the constraints imposed by the strain data and the fracture observations. The simplest explanation (model 1) is that the overpressuring developed in large part during burial, as is commonly observed in low-permeability clay-rich rocks where diagenesis, aquathermal expansion, and pore volume reduction generate superhydrostatic fluid pressure faster than the permeability of the material allows to dissipate. The depth of burial of the Rhinestreet base prior to the onset of major shortening is uncertain, although stratigraphic data suggest a depth of at least 1.7 km. Available worldwide evidence suggests this is sufficient to initiate overpressuring in shales [e.g., Rhem, 1972]. Topographically driven flow from orogenic highlands to the east could also have contributed to early overpressuring. Modest shortening might then have driven fluid pressures to near-lithostatic levels. That an overpressure did not develop in the upper section during burial, despite a similarly high clay content, might be attributed to the sand-rich beds serving as escape conduits for the excess fluid generated in the shales. In this scheme, horizontal shortening of the

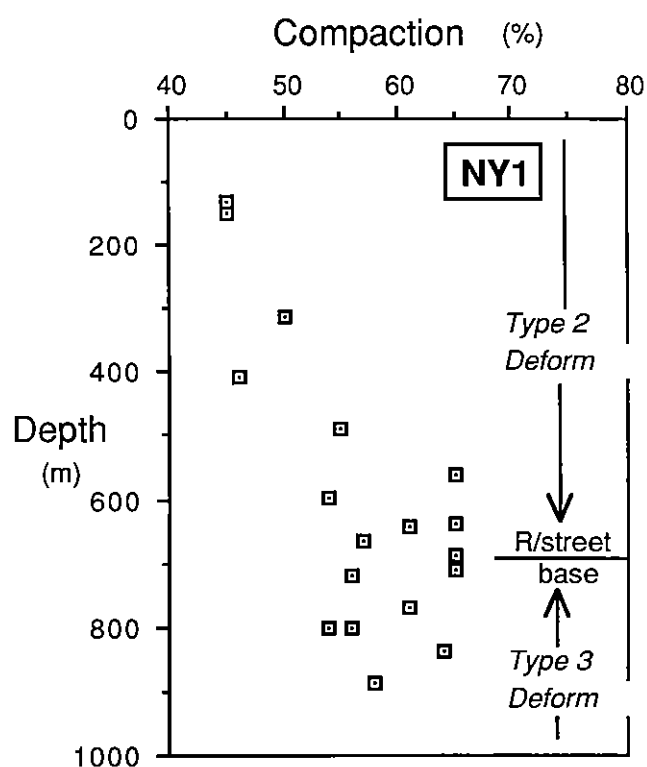


Fig. 9. Compaction developed before the advent of Alleghanian shortening as given by model 2. Deformation types "u" and "v" both assume zero strain normal to the principal shortening direction and hence are identical in plan. They differ in that the latter is volume constant and imparts vertical strain, whereas the former does not.

Devonian section in response to Alleghanian compression would have proceeded as uniaxial strain (deformation type "u") and left the burial-induced vertical strain unaltered. An implication of this model is that the entire section suffered a tectonically induced 12% volume reduction. How this could have been accomplished in the already overpressured section below the Rhinestreet base where fluid circulation was restricted is problematical.

An alternative model which does not suffer from this problem holds that the lower section (from the Salina salt decollement to the Rhinestreet base) deformed in a volume-constant manner resulting in a 13.6% vertical strain (deformation type "v"), whereas the upper section underwent uniaxial strain and suffered a 12% volume loss (deformation type "u"). Thus, in plan, the deformation patterns are identical, although the lower section has undergone a 13.6% vertical extension which accumulates with the preceding compaction strain to produce the observed anomaly (Table 2). It is interesting to examine whether the vertical strain anomaly can be explained entirely by this model without preshortening overpressure development. In Figure 9 we show the predicted profile of compaction prior to the advent of Alleghanian shortening calculated for this model. Compaction of the section above the base of the Rhinestreet is calculated according to equation (3), whereas equation (5) was applied for the lower section. We find that compaction of the lower section is still somewhat low, although only slightly so. This demonstrates that undercompaction of the lower section cannot wholly be explained as a volume-constant response to shortening, and hence we infer that overpressuring had developed to some degree prior to tectonic shortening. Thus our favored model (model 2) consists

of overpressure development in the lower section during burial, with subsequent elevation to the requisite levels for joint propagation during the earliest phases of Alleghanian shortening. Subsequently, shortening proceeded as type "v" deformation in the lower section and type "u" above, with further joint propagation suppressed by least horizontal stress levels in excess of the overburden. The two models we have discussed are end-members to a family of hybrid models which are equally consistent with the observations. Some volume loss must have occurred in the lower section during early shortening to propagate the joints.

SPECULATIONS ON THE NATURE OF THE SUB-RHINESTREET STRESS ANOMALY

Paleo-overpressure drainage. The association between undercompaction and overpressure offers, in principle, an explanation of the spatially coincident stress anomaly, provided it is assumed that cementation of the Devonian section occurred before this overpressure could leak down to normal hydrostatic levels. Evidence that these shales are now cemented is given by the high differential stress levels of up to 20 MPa supported by them (Figures 9 and 17 of P1). Indeed, the stress data demonstrate that a "lithostatic" stress state in shales cannot be presumed a priori, even given very high clay content (Figure 4). If it is accepted that at the close of Alleghanian tectonism the sub-Rhinestreet section was overpressured, whereas the overlying section was normally pressured, and that the material lithified while this difference in pore pressure lasted, then subsequent drainage of the overpressure to normal levels will result in a tendency for the rock to contract elastically in proportion to the difference between the bulk compressibility C_b and the intrinsic compressibility of the solid constituent C_i [Nur and Byerlee, 1971]. However, because the rock is constrained in situ, it can contract only in the vertical direction. The response in the horizontal plane is to develop a tensile stress component which reduces the preexisting total horizontal stress. The magnitude of this poroelastic component pS_h is given by

$${}^pS_h = \frac{(1 - 2\nu_d)}{(1 - \nu_d)} \alpha dP_p \quad (6)$$

where α is the Biot constant defined by $\{1 - C_i/C_b\}$, ν_d is Poisson's ratio measured under drained conditions, and dP_p is the change in pore pressure. We measured the bulk compressibility of sample 13b (Table 1) as $5.7 \times 10^{-11} \text{ Pa}^{-1}$. Taking a value for C_i of $1.6 \times 10^{-11} \text{ Pa}^{-1}$, as estimated from the bulk compressibility data for Devonian shales at 4 GPa confining pressure reported by Heard and Lin [1986], yields an estimate for the current value of α for the siltstones of about 0.7. This value was undoubtedly even larger in the past when the hypothetical overpressure gradually dissipated. Assuming a reasonable value for ν_d of 0.17, the reduction in horizontal total stress effected by this mechanism is at least 55% of the pore pressure reduction. Pore pressure is generally constrained to be less than the overburden by the criterion for the development of horizontal hydrofractures. Thus the maximum possible drop in P_p that can be admitted increases with depth as the difference between the vertical stress and the hydrostat. Assuming a depth of burial for the Rhinestreet base of 1.7 km and an overburden density of 2650 kg/m^3 , the corresponding maxi-

mum drop in pore pressure that can be admitted at the level of the K-sand is 27.5 MPa. If such an extreme drop occurred after cementation, we would expect a concomitant reduction in horizontal stress of at least 55%, that is, 15.1 MPa. The observed offsets in S_H and S_h of 9.0 and 3.5 MPa, respectively, are well within this limit.

A problem with the paleopressure drainage hypothesis is that it predicts a stress offset which is independent of azimuth; that is, the reduction in S_H would be the same as that in S_h . The stress data, however, suggest that S_h is reduced by 3.5 MPa and S_H by 9 MPa. The chlorite fabric analysis demonstrates that these rocks are strongly anisotropic, and hence the assumption of elastic isotropy inherent in equation (6) is not valid. This may partly explain the disparity. However, we feel it is too large to accredit entirely to anisotropy effects. Accepting the estimates of stress on face value, we seek an explanation for the difference in the offsets of S_H and S_h .

Horizontal detachment at the K-sand horizon. That no change in stress orientation accompanies the offset implies that the principle axes of the causative stress component and that of contemporary tectonic compression are coaxial. This might be taken as suggesting that the offset simply represents a subhorizontal detachment on which bedding plane slippage has taken place in response to contemporary compression: that is, rocks of the "upper" regime have suffered greater ENE-WSW elastic shortening than those of the "lower" regime. In paper 1 we show that stress magnitudes in the "upper" regime suggest a stress state which is close to the threshold for slippage on favourably oriented frictional interfaces, the sense of failure being "thrust." If bedding-plane slippage had occurred on scales at least as great as several kilometers (the minimum extent of the stress anomaly), then we might anticipate the development of NNW-SSE trending structures and faults within the section above the salt, since Alleghanian faults are incorrectly oriented to accommodate slip in response to the contemporary stress field. Despite abundant data from boreholes, there is no evidence for this, either locally or regionally. Furthermore, we consider it unlikely that an extensive horizontal detachment would develop in stratigraphic proximity to the major salt detachment. This leads us to search for another mechanism which might augment or replace the paleo-overpressure drainage hypothesis.

Remnant stresses from the Alleghanian orogeny. The orientation of horizontal principal stress axes during Alleghanian times was essentially orthogonal to the contemporary stress axes. Consequently, the presence of a stratigraphically localized stress component that is remnant from the Alleghanian compression would not affect the orientation of the contemporary principal axes: rather its presence would be manifest only in the magnitude of the contemporary stresses. It is worth examining the possibility that the stress anomaly, or a part of it, constitutes an Alleghanian remnant stress. Our usage of the term "remnant stress" is similar to that first employed by Voight [1974] and later clarified by Ranalli and Chandler [1975] and applies to any localized paleostress component arising from elastic strains locked into the rock through antagonistic force balance. It is analogous to the concept of "residual stress" defined by Friedman [1972], which is a specific form of remnant stress, defined at the grain scale, in which the "locked-in" and "locking-in" stresses reside in constituent grains and ce-

TABLE 3. Velocity Anisotropy Measured on EGSP Well Core Samples

Well	County	Mean Directional <i>P</i> Velocity, km/s and °E of North						Maximum Variation, %	Orientation of V_{\max}	Sample Depths, m	Number of Samples Tested
		0°	30°	60°	90°	120°	150°				
KY1	Perry	4.505	4.51	4.485	4.47	4.475	4.487	<1		725-821	45
KY2	Christian	4.69	4.74	4.74	4.73	4.74	4.72	<1		664-706	18
KY3	Martin	4.595	4.655	4.655	4.60	4.57	4.545	1.9	30-60°	761-1084	30
KY4	Johnson	4.775	4.785	4.79	4.795	4.78	4.78	<1		299-460	49
KY5	Rowan	4.085	4.09	4.09	4.09	4.075	4.075	<1		139-193	87
NY1	Allegany	5.035	5.135	5.211	5.187	5.09	5.02	3.8	60°	135-878	139
NY4	Steuben	4.915	4.81	4.79	4.88	4.96	4.975	3.9	150°	933-1169	11
OH1	Carrol	4.29	4.27	4.26	4.265	4.28	4.285	<1		601-1318	33
OH2	Washington	4.575	4.595	4.595	4.55	4.56	4.545	<1		1071-1100	9
OH3	Knox	4.625	4.61	4.605	4.62	4.63	4.635	<1		187-383	37
OH4	Ashtabula	4.87	4.875	4.89	4.885	4.88	4.87	<1		240-421	33
OH6/4	Gallia	4.735	4.735	4.745	4.745	4.745	4.74	<1		607-835	36
OH7	Trumbull	4.835	4.87	4.875	4.84	4.81	4.80	<1		468-834	99
OH8	Noble	5.075	4.98	4.875	4.86	4.93	5.045	4.4	0°	534-1265	113
PA1	McKean	4.80	4.87	4.94	4.93	4.84	4.77	3.6	60°	1059-1579	46
PA2	Alleghany	5.04	5.075	5.06	5.01	4.99	5.00	1.7	30°	2124-2231	12
PA3	Erie	4.81	4.775	4.795	4.795	4.81	4.81	<1		128-379	31
PA4	Indiana	5.31	5.365	5.34	5.255	5.195	5.22	3.3	30°	2164-2440	53
PA5	Lawrence	4.93	4.98	4.975	4.90	4.845	4.855	2.8	30°	1074-1256	58
T9	Grainger	5.16	5.245	5.215	5.165	5.12	5.11	2.4	30°	361-526	11
V1	Wise	4.715	4.82	4.84	4.775	4.67	4.63	3.6	60°	1484-1750	33
WV1	Jackson	4.64	4.70	4.645	4.59	4.58	4.63	2.6	30°	1049-1139	6
WV2	Jackson	4.485	4.55	4.555	4.475	4.415	4.425	3.1	30°	983-1125	26
WV4	Lincoln	22.03*	21.76*	21.80*	22.08*	22.31*	22.22*	2.5	30°	842-1209	15
WV5	Mason	4.955	4.98	4.98	4.975	4.96	4.945	<1		831-1023	40
WV6	Monongalia	5.355	5.38	5.365	5.315	5.28	5.285	1.9	30°	2191-2236	5
WV7	Wetzel	5.215	5.22	5.175	5.115	5.115	5.16	2.0	30°	1875-2017	36
20403†	Lincoln	5.12	5.20	5.275	5.26	5.20	5.12	3.0	60°	831-1228	126

**P* wave travel times; anisotropy estimate assumes equal path lengths.

†Data from Peng and Okubo [1978].

ment, respectively. (Note that residual stress as referred to by Holzhausen and Johnson [1979] corresponds to remnant stress in the present usage.) Although residual stresses may be present in Devonian rocks, they would be detected only by stress measurement techniques which average over small areas, such as overcoring. To first order, hydrofracture measurements should not be affected by the presence of grain scale residual stresses.

To explain the observed stress offset precisely would require that Alleghanian shortening leave a higher NNW directed horizontal compressive paleostress in the lower section than in the upper section. When superposed with a dominant, biaxially tensile, paleopressure drainage component, the effect will be to reduce the net offset in the NNW oriented principal stress below the Rhinestreet base, as required. The hypothesis implies that during shortening, the NNW oriented component of horizontal stress in the lower section exceeded the overburden by a greater margin than was the case in the upper section and that a remnant of this horizontal stress difference persists. That is, a higher horizontal shortening stress in the lower section has been locked-in by elastic coupling to the upper (locking-in) section. This is not compatible with our preferred model of Alleghanian deformation. We have argued that "thrust" regime conditions prevailed throughout the section during all but the earliest phases of shortening and that near-lithostatic fluid pressures had developed in the lower section at early times, when the stress state was marginally "strike-slip" and vertical

joints propagated. Under such conditions the maximum deviatoric stress in the lower section at the time of joint propagation is limited to several megapascals [Etheridge, 1983]. As there is no reason to suppose that near-lithostatic fluid pressures were not maintained during the major phase of shortening, this constraint appears to rule out the possibility that deviatoric stresses in the lower section were significantly greater than in the upper section. Indeed, the observation of abundant deformed crinoids in the upper section suggests that significant horizontal deviatoric stresses were present, although too little is known about the stress solution process to interpret this in terms of predicted stress levels.

On balance, the evidence does not favor higher paleo-compressive deviatoric stress levels in the lower section, as required by the most simple remnant stress explanation of the nonbiaxial component of the stress anomaly. Thus the precise nature of the sub-Rhinestreet stress anomaly remains somewhat of an enigma, although the paleopressure drainage hypothesis fits the observations best.

CHLORITE FABRIC AND THE MECHANICAL ANISOTROPY IN DEVONIAN SHALES

The Devonian shales of the Appalachian basin possess a strong mechanical anisotropy in the horizontal plane. Specifically, the rock exhibits a preference for tensile failure in a direction which generally parallels the structural trend. This direction also generally coincides with the fast direction for

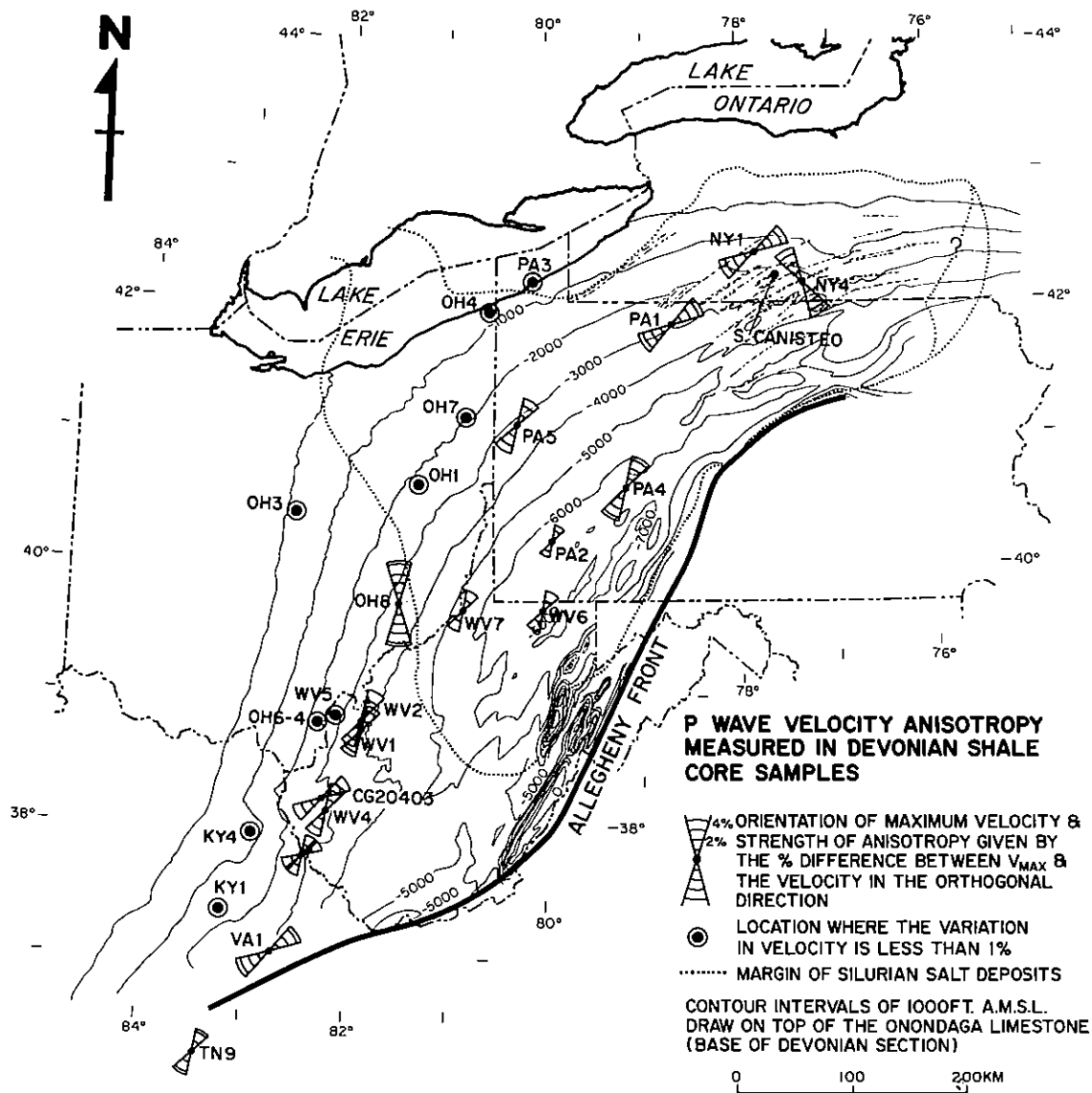


Fig. 10. P wave velocity anisotropy magnitude and fast direction observed in Devonian shale core samples from DOE's Eastern Gas Shale Project wells of the Appalachian basin.

horizontally propagating p waves measured in core samples under atmospheric pressure conditions. The data defining the velocity anisotropy are derived from the testing of core samples recovered from the 31 EGSP boreholes and are listed in Table 3 and plotted in Figure 10. (These data were culled from the phase 3 well completion reports compiled by Cliffs Minerals Inc. for Morgantown Energy Technology Center, Morgantown, West Virginia.) Note the anomalous orientation of the NY4 data point, also found in the chlorite fabric analysis, which we ascribe to core recovery error in the absence of evidence of block rotation. Both strength and velocity anisotropies appear to become more pronounced closer to the Allegheny front [Gregg, 1986]. The nature of this anisotropy is undetermined. Indeed, it has not been unequivocally demonstrated that it exists in situ and is not an artifact of removal of the core from confinement. Peng and Okubo [1978] proposed that the anisotropy results from a population of in situ microcracks. The sense of the anisotropy is such that the microcrack population would be aligned with what is apparently the maximum principal horizontal stress direction in the Devonian shales (Evans, 1989) and hence could represent "extensive

dilatancy anisotropy" which according to Crampin *et al.* [1984] is a state that characterizes rocks close to shear failure. An alternative explanation is that the chlorite fabric that we have used in this study imparts an anisotropy in both tensile strength and velocity which strengthens with deformation (i.e., as the Allegheny front is approached). Alexandrov and Ryzhova [1961] have measured directional wave velocities in layered silicates crystallographically similar to chlorite and report much faster p wave propagation in the basal plane (001) than in the normal direction, as required by the hypothesis. If the observed anisotropy is entirely due to the preferred orientation of platy minerals, then the fast p wave data shown in Figure 10 would provide some measure of the variation in direction and amplitude of local Alleghanian shortening in Devonian rocks about the basin.

CONCLUSION

Static and dynamic modulus data indicate that lithology-correlated variations in horizontal stress observed in a sandstone/shale/limestone sequence above a prominent salt detachment result in large part from elastic shortening of the

section in response to regional ENE-WSW compression. No correlation between stress and Poisson's ratio was found, suggesting that in compressional situations, bed-to-bed variations in elastic stiffness may be more important in generating bed-to-bed stress contrasts than variations in Poisson's ratio. Shales of very high clay content were found to be supporting high differential stress (at least 20 MPa), contrary to the belief that the state-of-stress in such materials is "lithostatic." Although bed-to-bed variations in stress are well explained by stiffness contrasts, the physical property data suggest that a major systematic drop in stress level that occurs in the generally massive shales across a group of sand beds near the base of the Rhinestreet is not of this origin. Rather, the discontinuity corresponds to the top of a section which we conclude, on the basis of local and regional total strain data, once hosted abnormally high pore pressures.

The total strain data also show that the entire section above the salt has been uniformly shortened during Alleghanian compression. Two kinematic patterns for Alleghanian deformation of the Devonian section are proposed, both involving abnormal pore pressure development in the sub-Rhinestreet section in response to limited drainage of fluid. Dissipation of the abnormal pressure following lithification is shown to be qualitatively but not quantitatively consistent with the observed stress discontinuity. The model predicts that the resultant perturbation to the stress field should be biaxial, whereas the data suggest that the offset mimics the current dominant stress field component in terms of the ratio of the two principal horizontal stresses and their orientation. Postulation of an additional remnant stress component arising from "locked-in" Alleghanian compressive stress requires implausible stress conditions to develop in the upper and lower (overpressured) section in response to Alleghanian compression. Paleopressure drainage is the best available explanation of the stress offset. The fabric imparted to the shales during Alleghanian compression may explain their mechanical anisotropy.

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REFERENCES

- Alexandrov, K. S., and T. V. Ryzhova, Elastic properties of layered silicates, *Izv. Acad. Sci. USSR Phys. Solid Earth*, Eng. Transl., 9, 1165-1168, 1961.
- Beardsley, R. W., and M. S. Cable, Overview of the evolution of the Appalachian basin, *Northeast. Geol.*, 5, 137-145, 1983.
- Beaumont, C., G. Quinlan, and J. Hamilton, Orogeny and stratigraphy, numerical models of the Paleozoic in the eastern interior of North America, *Tectonics*, 7, 389-416, 1988.
- Bradley, W. H., and J. F. Pepper, Structure and gas possibilities of the Oriskany sandstone in Steuben, Yates and parts of the adjacent counties, *U.S. Geol. Surv. Bull.*, 899-A, 1938.
- Clark, J. A., Glacial loading: A cause of natural fracturing and a control of the present stress state in regions of high Devonian shale gas, paper presented at Unconventional Gas Recovery Symposium, Soc. of Pet. Eng., Pittsburgh, Pa., May 16-18, 1982.
- Cliffs Minerals Inc., Analysis of the Devonian shales in the Appalachian basin, final report to U.S. Dep. of Energy under contract DE-AC21-80MC14693, vol. 1 and 2, Morgantown Energy Technol. Cent., Morgantown, W. Va., 1982.
- Crampin, S., R. Evans, and B. Atkinson, Earthquake prediction: A new physical basis, *Geophys. J. R. Astron. Soc.*, 76, 147-156, 1984.
- Davis, D. M., and T. Engelder, The role of salt in fold and thrust belts, *Tectonophysics*, 119, 67-88, 1985.
- Engelder, T., Mechanisms for strain within the upper Devonian clastic sequence of the Appalachian Plateau, western New York, *Am. J. Sci.*, 279, 527-542, 1979.
- Engelder, T., The role of pore water circulation during the deformation of foreland fold and thrust belts, *J. Geophys. Res.*, 89, 4319-4325, 1984.
- Engelder, T., Loading paths to joint propagation during a tectonic cycle: An example from the Appalachian Plateau, *J. Struct. Geol.*, 7, 459-476, 1985.
- Engelder, T., Smoluchowski's dilemma revisited: A note on the fluid pressure history of the Central Appalachian fold-and-thrust belt, Report on Fluids in the Earth, edited by T. Usselman, Natl. Res. Council, Washington, D. C., in press, 1989.
- Engelder, T., and R. Engelder, Fossil distribution and decollement tectonics on the Appalachian plateau, *Geology*, 5, 457-460, 1977.
- Engelder, T., and P. Geiser, The relationship between pencil cleavage and lateral shortening within the Devonian section of the Appalachian Plateau, New York, *Geology*, 7, 460-464, 1979.
- Engelder, T., and G. Oertel, Correlation between abnormal pore fluid pressure and tectonic jointing in the Devonian Catskill delta, *Geology*, 13, 863-866, 1985.
- Epstein, A. G., J. B. Epstein, and L. D. Harris, Conodont color alteration—An index to organic metamorphism, *U.S. Geol. Surv. Prof. Pap.*, 995, 1977.
- Etheridge, M. A., Differential stress magnitudes during regional metamorphism, *Geology*, 11, 231-234, 1983.
- Eitensohn, F. R., The Catskill delta complex and the Acadian orogeny: A model, *Spec. Pap. Geol. Soc. Am.*, 201, 39-49, 1985.
- Evans, K. F., Appalachian stress study, 3, Regional scale stress variations and their relation to structure and contemporary tectonics, *J. Geophys. Res.*, in press, 1989.
- Evans, K. F., and T. Engelder, A study of stress in Devonian shales of the Appalachian Plateau, final report to U.S. Dep. of Energy under contract DE-AC21-83MC20337, Morgantown Energy Technol. Cent., Morgantown, W. Va., 1987.
- Evans, K. F., and T. Engelder, Some problems in estimating stress magnitude in thrust stress regimes, *Int. J. Rock Mech. Min. Sci. Geomech. Abstr.*, in press, 1989.
- Evans, K. F., T. Engelder, and R. A. Plumb, Appalachian stress study, 1, A detailed description of in situ stress variations in Devonian shales of the Appalachian Plateau, *J. Geophys. Res.*, this issue.
- Frey, M. G., Influence of salina salt on structure in the New York-Pennsylvania part of Appalachian Plateau, *Am. Assoc. Pet. Geol. Bull.*, 57, 1027-1037, 1973.
- Friedman, M., Residual elastic strain in rocks, *Tectonophysics*, 15, 297-330, 1972.
- Frisinger, M., and R. E. Cooper, Techniques for quantifying formation stress and fracture fluid response for optimum hydraulic fracture design, paper SPE 14015 presented at Offshore Europe 85 Conference, Soc. of Pet. Eng., Aberdeen, Scotland, Sept. 10-13, 1985.
- Gregg, W. J., Mechanical fabric and in-situ stress orientations in the Devonian gas shales of the Appalachian Basin, *Proc. U.S. Symp. Rock Mech.*, 27th, 709-715, 1986.
- Gwinn, V. E., Thin-skinned tectonics in the plateau and northwestern Valley and Ridge Province of the central Appalachians, *Geol. Soc. Am. Bull.*, 75, 863-900, 1964.
- Heard, H. C., and W. Lin, High pressure mechanical and sonic properties of a Devonian shale from West Virginia, *Rep. UCID-20612*, Lawrence Livermore Natl. Lab., Livermore, Calif., 1986.
- Holzhausen, G. R., and A. M. Johnson, The concept of residual stress in rock, *Tectonophysics*, 58, 237-267, 1979.

- Johnsson, M. J., Distribution of maximum burial temperatures across northern Appalachian Basin and implications for Carboniferous sedimentation patterns, *Geology*, *14*, 384-387, 1986.
- Kalyoncu, R. S., J. P. Boyer, and M. J. Snyder, Devonian shales: An in-depth analysis of well NY#1 with respect to shale characterization, hydrocarbon content and wire-log data, paper presented at 3rd Eastern Gas Shales Project Symposium, Dep. of Energy, Morgantown, W. Va., Oct. 1-3, 1979.
- Karig, D. E., Comment to 'Distribution of maximum burial temperatures across northern Appalachian Basin and implications for Carboniferous sedimentation patterns' by M. J. Johnsson, *Geology*, *15*, 278-279, 1987.
- Magara, K., *Compaction and Fluid Migration*, 319 pp., Elsevier, New York, 1978.
- March, A., Mathematische Theorie de Regelung nach der Korngestalt bei affiner Deformation, *Z. Kristallogr.*, *81*, 285-297, 1932.
- McGarr, A., On the state of lithospheric stress in the absence of applied tectonic forces, *J. Geophys. Res.*, *93*, 13,609-13,617, 1988.
- Nickelsen, R. P., Sequence of structural stages of the Alleghany orogeny at the Bear Valley strip mine, Shamokin, PA, *Am. J. Sci.*, *279*, 225-271, 1979.
- Nur, A., and J. D. Byerlee, An exact effective stress law for elastic deformation of rocks with fluids, *J. Geophys. Res.*, *76*, 6414-6419, 1971.
- Oertel, G., The relationship of strain and preferred orientation of phyllosilicate grains in rocks-A review, *Tectonophysics*, *100*, 413-447, 1983.
- Oertel, G., Reorientation due to grain shape, in *Preferred Orientation in Deformed Metals and Rocks: An Introduction to Modern Texture Analysis*, edited by H.-R. Wenk, pp. 259-265, Academic, San Diego, Calif., 1985.
- Oertel, G., T. Engelder, and K. F. Evans, A comparison of the strain of Crinoid columnals with that of their enclosing silty and shaly matrix on the Appalachian Plateau, *J. Struct. Geol.*, in press, 1989.
- Peng, S. S., and S. Okubo, Prediction of in-situ stresses from directional properties of rock cores for field development of Devonian shales, final report to U.S. Dep. of Energy under contract EY-77-C-21-8087, Morgantown Energy Technol. Cent., Morgantown, W. Va., 1978.
- Plumb, R. A., and J. Cox, Stress directions in eastern North America determined to 4.5 km from borehole elongation measurements, *J. Geophys. Res.*, *92*, 4805-4816, 1987.
- Plumb, R. A., K. F. Evans, and T. Engelder, Correlation of mechanical properties and in-situ stress contrasts (abstract), *Eos Trans. AGU*, *68*, 627, 1987.
- Price, N. J., The development of stress systems and fracture patterns in undeformed sediments, *Proc. Rock Mech. Congr.*, *3rd*, 487-496, 1974.
- Ranalli, G., and T. E. Chandler, The stress field in the upper crust as determined from in-situ measurements, *Geol. Rundsch.*, *42*, 653-674, 1975.
- Rhem, W., Worldwide occurrence of abnormal pressures, part 2, paper SPE 3845 presented at Abnormal Subsurface Pressure Symposium, Soc. of Pet. Eng., Baton Rouge, La., May 15-16, 1972.
- Sbar, M. L., and L. R. Sykes, Contemporary compressive stress and seismicity in eastern North America: An example of intraplate tectonics, *Geol. Soc. Am. Bull.*, *84*, 1861-1882, 1973.
- Van Tyne, A., Inventory and analysis of the oil and gas resources of Allegheny and Cattaraugus counties, New York, parts I and II, Southern Tier West, Reg. Plann. and Dev. Board, Salamanca, N. Y., 1980.
- Van Tyne, A., Natural gas potential of the Devonian black shales of New York, *Northeast. Geol.*, *5*, 209-216, 1983.
- Voight, B., Evolution of the North Atlantic Ocean: Relevance of rock pressure measurements, in North Atlantic—Geology and continental drift, *Mem. Am. Assoc. Petrol. Geol.*, *12*, 955-962, 1969.
- Voight, B., A mechanism for 'locking-in' orogenic stress, *Am. J. Sci.*, *274*, 662-665, 1974.
- Voight, B., and B. H. P. St. Pierre, Stress history and rock stress, *Proc. Rock Mech. Congr.*, *3rd*, 580-582, 1974.
- Warpinski, N. R., Elastic and viscoelastic calculations of stresses in sedimentary basins, paper SPE 15243 presented at Unconventional Gas Technology Symposium, Soc. of Pet. Eng., Louisville, Ky., May 18-21, 1986.
- Wenk, H.-R., Measurement of pole figures, in *Preferred Orientation in Deformed Metals and Rocks: An Introduction to Modern Texture Analysis*, edited by H.-R. Wenk, pp. 11-47, Academic, San Diego, Calif., 1985.
- Zoback, M. L., and M. D. Zoback, Tectonic stress field of the continental U.S., Geophysical Framework of the United States, edited by L. C. Pakiser and W. D. Mooney, *Mem. Geol. Soc. Am.*, in press, 1989.

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