Correlation between abnormal pore pressure and tectonic jointing in the Devonian Catskill Delta

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ABSTRACT

Using the preferred orientation of chlorite, we measured vertical compaction in 53 samples from the Devonian Catskill Delta in central New York State. The marine part of this delta contains three levels based on different amounts of vertical compaction. These levels correspond roughly to (1) black shales at the base, (2) prodelta turbidites, and (3) a cap of shallow-water sediments including abundant storm-washed shell hashes deposited within the wave base. The cap is normally compacted, whereas the lower two levels are undercompacted. Tectonic (cross-fold) joints that propagated during the late Paleozoic Alleghanian orogeny are restricted to the deeper, undercompacted levels of the Catskill Delta, whereas unloading (cross-fold) joints pervade the cap. The correlation between undercompaction and the distribution of tectonic joints indicates that abnormal fluid pressure was a key mechanism during the propagation of these joints.

INTRODUCTION

Abnormal pore pressure develops during compaction when pore water cannot drain fast enough to stay at hydrostatic pressure. A number of mechanisms including poor collapse (Magara, 1978), clay dehydration (Powers, 1967), and aquathermal pressuring (Barker, 1972) may cause abnormal pressures. Undercompaction within a sedimentary rock is one piece of evidence suggesting that abnormal pore pressures occurred at some time during the burial history of the host. The purpose of this paper is to present evidence for undercompaction within the Catskill Delta and demonstrate a correlation between undercompaction (abnormal pore pressures) and the development of tectonic joints. The relationship of both undercompaction and tectonic jointing to the stratigraphy of the Catskill Delta is critical in establishing the correlation.

The Catskill Delta is a large clastic wedge that thickens from 2 to 7 km in a west to east section across New York State (Colton, 1970; Van Tyne and Foster, 1979; Friedman and Sanders, 1982). The delta was deposited during the Devonian Acadian orogeny with sediments derived from the uplifted highlands in western New England (Rodgers, 1970). In the western parts of the delta the sediments are exclusively marine, whereas in the eastern and upper parts fluvial sediments appear. The marine part of the Catskill Delta consists of three major stratigraphic levels (Fig. 1). At the base are black shales (Hamilton to Genesee Groups), showing little or no bottom-current sorting, which were deposited in 200 m of water (McIver, 1970). Interfingered siltstone and shale of prodelta turbidites (Genesee, Sonyea, and West Falls Groups) constitute the middle level. The division between deep marine shales and prodelta turbidites occurs above the Geneseo Shale of the Genesee Group. The cap (Canadaway and Conneaut Groups) includes interfingered siltstones, shales, and storm-washed shell hashes that were deposited above wave base.

The Catskill Delta was mildly deformed about 50–100 m.y. after deposition (Rodgers, 1970). During the Carboniferous Alleghanian orogeny the beds of the delta were pushed to the north-northwest on a decollement within Silurian Salina salts to form a narrow tapered wedge of deformed rock (Engelder and Engelder, 1977; Davis and Engelder, 1985). Compression of the delta is manifested by gentle folds, a disjunctive cleavage, twinning of calcite, and cross-fold joints (Geiser and Engelder, 1983).

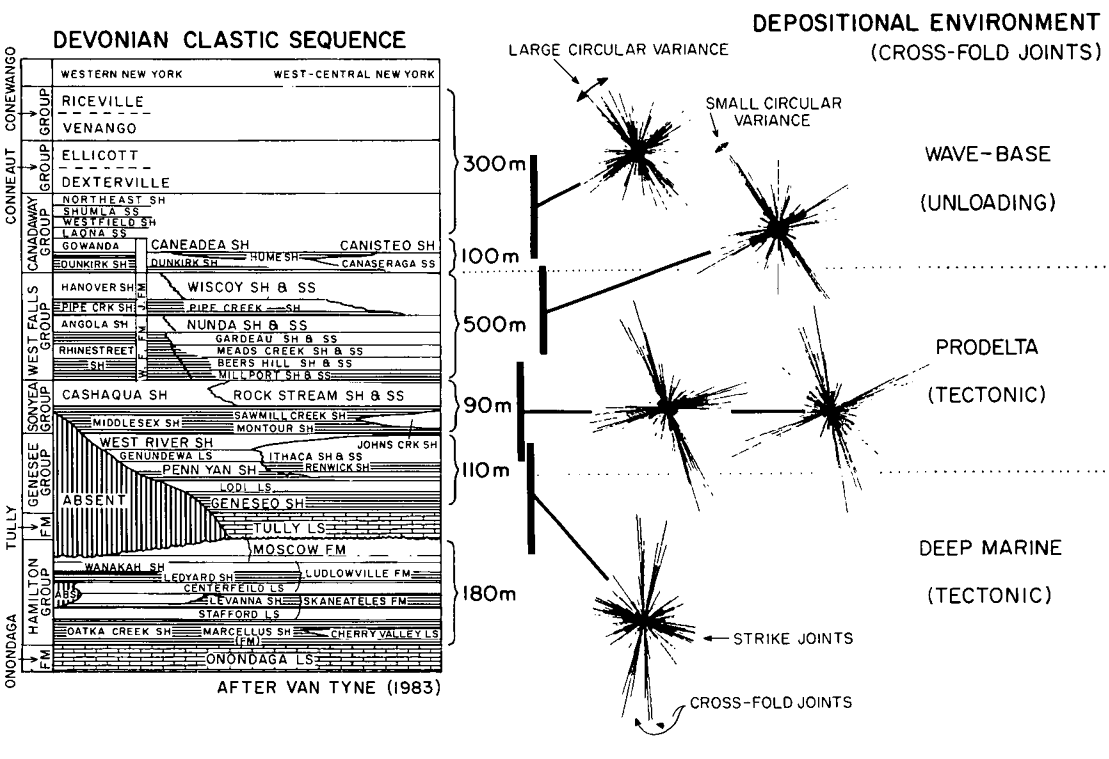

Figure 1. Stratigraphic column for Catskill Delta (after Van Tyne, 1983). Stratigraphic thicknesses taken from wells south-southwest of Rochester, New York. Rose diagrams consist of 500 or more joints at 30–50 outcrops of rocks from indicated parts of stratigraphic column. Outcrops cover area of six to ten 7½-minute quadrangles. Contained and uncontained tectonic joints refer to whether or not bedding interfaces stopped vertical growth of joints (see Engelder, 1985).

Two types of cross-fold joints (striking north-south to north-northwest) are found within the Catskill Delta: tectonic and unloading (Fig. 2) (Engelder, 1985). At least two sets of cross-fold joints can be dated as Alleghanian on the basis of their consistent orientation relative to the compressive pulses of that age and their cross-cutting relationship with the Alleghanian cleavage (Engelder and Geiser, 1980). These two sets of tectonic joints correlate with the two distinct phases of the Alleghanian orogeny, the Lackawanna and the Main phases (Geiser and Engelder, 1983). Cross-fold joints that postdate the Alleghanian orogeny are unloading joints. Unloading joints are most common within the cap rock of wave-base sediments. When viewed in the form of rose diagrams, one set of tectonic joints has smaller circular variance (<5°) than a set of unloading joints (>10°) (Fig. 1). Abutting relationships indicate that tectonic joints form before strike joints, whereas unloading joints form in conjunction with strike joints (Engelder, 1985). The boundary between the tectonic and unloading joints is found near the contact between the West Falls and Canadaway Groups; the tectonic joints are restricted to the deep marine shales and the prodelta turbidites.

Although both unloading and tectonic joints are mode I cracks showing no shear displacement, the loading path to achieve effective tension is different for the two types of joints (Engelder, 1985). Unloading joints formed near the surface in response to the thermal-elastic contraction accompanying erosion and uplift. Tectonic joints form during tectonic compression at any depth of burial where abnormal pore pressure exceeds the strength of the host rock.

EVIDENCE FOR UNDERCOMPACTION

The preferred orientation of chlorite basal planes in shales and siltstones was used to measure overburden compaction. A key assumption with this technique is that the clay grains were originally deposited so that their basal planes were oriented at random, either as flocs or in fecal pellets. The resulting uniformly distributed orientations were shown to exist by measurements of the compaction in and around an early ironstone concretion (Oertel and Curtis, 1972). During collapse of pore space, accompanying loss of porosity, and compaction, the clay develops a preferred orientation with basal planes concentrated subparallel to bedding. Compaction correlates with the degree to which the clay becomes reoriented toward parallelism with the bedding plane.

The instrument measuring the preferred orientation of chlorite basal planes is a pole-figure goniometer attached to an X-ray diffractometer. The goniometer permits the systematic rotation of a sample to all orientations. The intensity of X-rays diffracted from the basal planes of chlorite grains varies with the volume of chlorite grains in the sample having a particular orientation. Principal normalized pole densities were found, and the three-dimensional strain at constant volume according to March (1932) was calculated as described by Wood et al. (1976).

By itself, the preferred orientation of crystals does not yield any information about volume change; statements about vertical compaction are, however, possible if additional assumptions can be made about the processes that caused the strain. Helpful for the formulation of such assumptions are certain regularities in the strain data calculated according to March (1932). The least strain is invariably normal to bedding (and thus vertical) within the margin of the orientation of error of sampling. This implies that the principal directions for the maximum and the intermediate strain lie, within the margin of error, in the bedding and the horizontal plane. We assume, then, that the vertical principal strain is, up to an insignificant error, the bedding-normal shortening due to overburden compaction. During pure, nontectonic, compaction there is no strain, and thus no area change, in the horizontal and bedding plane. Within the Catskill Delta the tectonic strains in the bedding plane are small by comparison with the vertical compaction strain; we assume, for the purpose of the calculation of overburden compaction, that the tectonic deformation did not change areas in the bedding plane. Although we do not in fact accept this assumption ourselves, it is a useful approximation of the minimum possible overburden compaction.

If the two bedding-parallel principal strains are ${}^{b}\epsilon_1$ and ${}^{b}\epsilon_2$, then the condition of constancy of area in the bedding plane is

$$(1 + {}^{b}\epsilon_1)\,(1 + {}^{b}\epsilon_2) = 1. \tag{1}$$

This fact is used to scale the constant-volume strain data in the following way:

$${}^{b}\epsilon_i = (1 + \epsilon_i) \,/\, \sqrt{(1 + \epsilon_1)\,(1 + \epsilon_2)} - 1 \qquad (i = 1, 2), \tag{2}$$

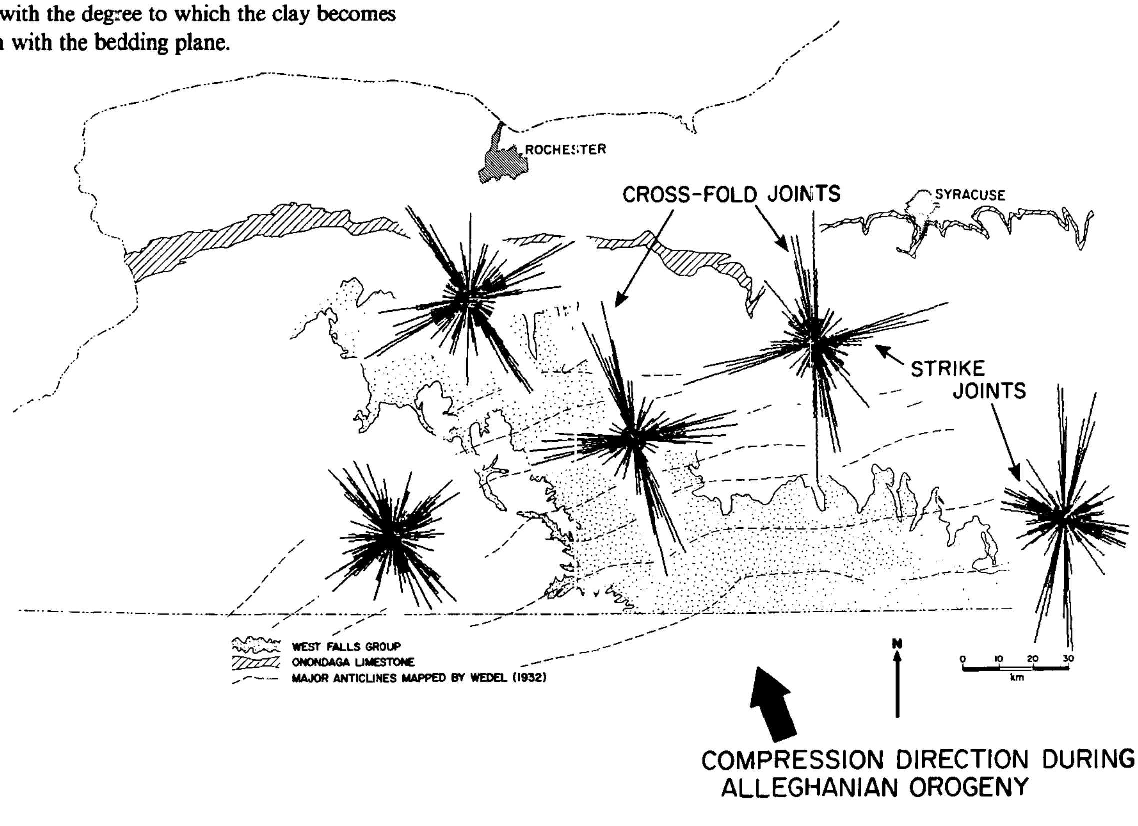

Figure 2. Regional joint pattern within Catskill Delta of central New York. Rose diagrams, which are same as in Figure 1, cover area in which data were collected.

where ϵ_1 and ϵ_2 are the two horizontal principal strains in the constant-volume form. The same scale factor can also be applied to the vertical least strain to yield the minimum overburden compaction strain where

$$^{c}\epsilon_{min} = (1 + \epsilon_3) \,/\, \sqrt{(1 + \epsilon_1)(1 + \epsilon_2)} - 1. \tag{3}$$

If tectonic (lateral) compaction has occurred, the assumption of constancy of area in the bedding plane is no longer valid. In the extreme case of all the tectonic strain being due to horizontal shortening without compensating extension, we could make the assumption that the long axis never changed length and use that as the basis for an estimate of maximum overburden compaction. The formula for the equal-longest-length compaction $^{c}\epsilon_{max}$ is as follows:

$$^{c}\epsilon_{max} = (1 + \epsilon_3) \,/\, (1 + \epsilon_1) - 1. \tag{4}$$

If we express porosities in volume percent and if we assume that all the volume loss of compaction is due to vertical shortening, then the following relationship exists between the original and present porosities P_o and P and the vertical strain due to compaction $^{c}\epsilon$:

$$^{c}\epsilon = (P_o - P)/(P - 100). \tag{5}$$

Oriented samples were taken from 53 outcrops representing most units within the Catskill Delta of central and western New York State. Most of our samples from the Catskill Delta contain, apart from chlorite and muscovite, enough fine quartz grains to indicate that their protolith was a rock somewhere between a shale and a very fine siltstone. These samples were buried between at least 250 and 2000 m, where the depth of burial is taken from the estimates of Colton (1970) and checked using well logs such as those of Van Tyne and Foster (1979). Because the thickness of the Carboniferous section in western New York is unknown, these depths of burial are minima based only on the thickness of the Devonian section. A small amount of data based on conodont color alteration suggests that the Carboniferous section in western New York was less than 1 km thick (Epstein et al., 1975).

The overburden compaction data are divided roughly according to the stratigraphic levels of the Catskill Delta and plotted against two compaction curves from Bond and Kominz (1984). The compaction curves represent the mean porosity change for normally consolidated shales and siltstones.

When plotted against the compaction curves for shale and siltstone, the samples from the top of the delta (the wave-base sediments of the Canadaway Group and above) show a tendency to be overcompacted, if Devonian thicknesses are used (Fig. 3). However, if the compaction curves are adjusted upward to account for about 300 m of Carboniferous sediments, then the fine siltstones show a tendency to be normally compacted. Samples showing a compaction of more than 0.55 are coarser grained siltstones containing many large detrital chlorite grains. These grains lie parallel to bedding and contribute to the higher pole densities for bedding-parallel orientation. These samples cannot be used as a measure of overburden compaction.

Samples from the upper part of the prodelta turbidites (the West Falls Group) follow a normal compaction curve for a very fine siltstone (Fig. 4). Moving the compaction curves for siltstone and shale up 300 m to account for the addition of Carboniferous rocks on top of the Devonian section indicates that the West Falls Group is undercompacted. The lower part of the prodelta turbidites shows a marked tendency toward undercompaction.

Although most of the marine shales of the Genesee and Hamilton Groups are finer grained and were more deeply buried than those taken from the prodelta turbidites, samples from these units show less compaction (Fig. 5). This tendency toward marked undercompaction begins to be observable at depths of 1400 m, and the bottom of the slightly undercompacted section occurs at about the 1500-m level of the Genesee Group.

DISCUSSION

During burial under 2 km or more of overburden, clay deposited on the sea bed with an initial porosity of about 70% normally compacts to less than 20% porosity (Rieke and Chilingarian, 1974; Magara, 1978). Shale and siltstone that retain a porosity significantly greater than 20% are undercompacted. One important mechanism for undercompaction is the support of pore space by pore fluid at a pressure above hydrostatic (abnormal pore pressure). Sediments of the Gulf Coast of the United States are well known for abnormally pressured shales; plots of shale density or porosity against pore pressure show that undercompaction correlates with abnormal pore pressures (Dickinson, 1953; Magara, 1978).

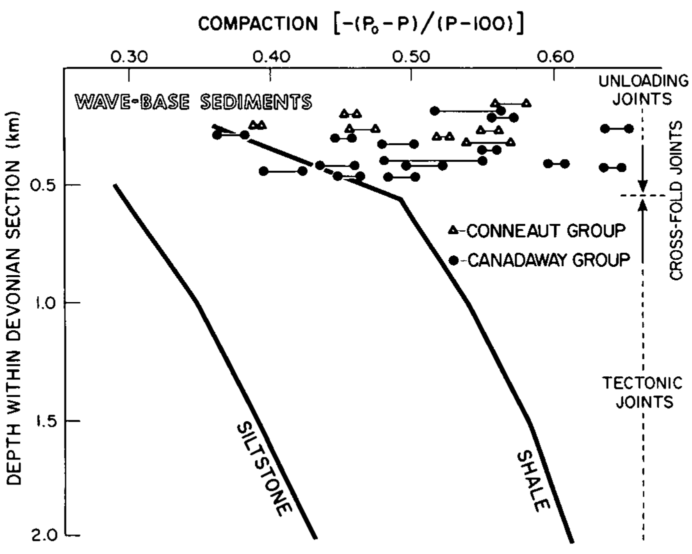

Figure 3. Compaction of samples taken from Canadaway and Conneaut Groups. Error bars indicate minimum and maximum compaction determined from formula assuming constancy of area in horizontal plane, or loss of area with long axis constant, respectively. Normal compaction curves are taken from Bond and Kominz (1984).

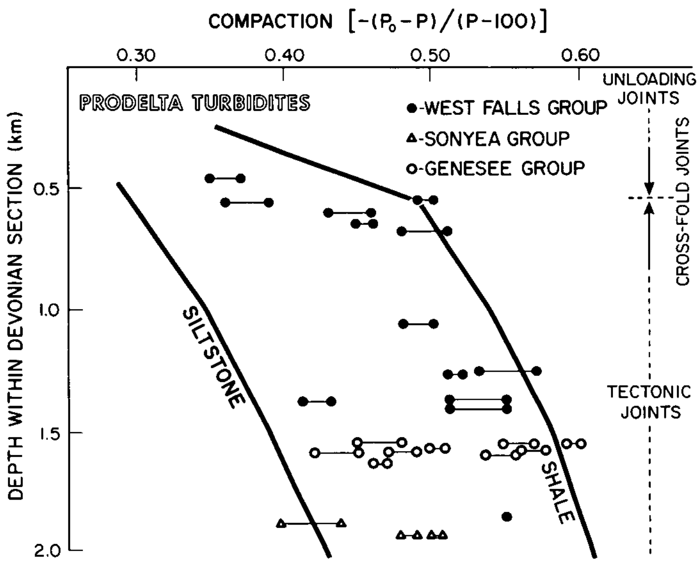

Figure 4. Compaction of samples taken from prodelta turbidites of West Falls, Sonyea, and Genesee Groups. Error bars as in Figure 3.

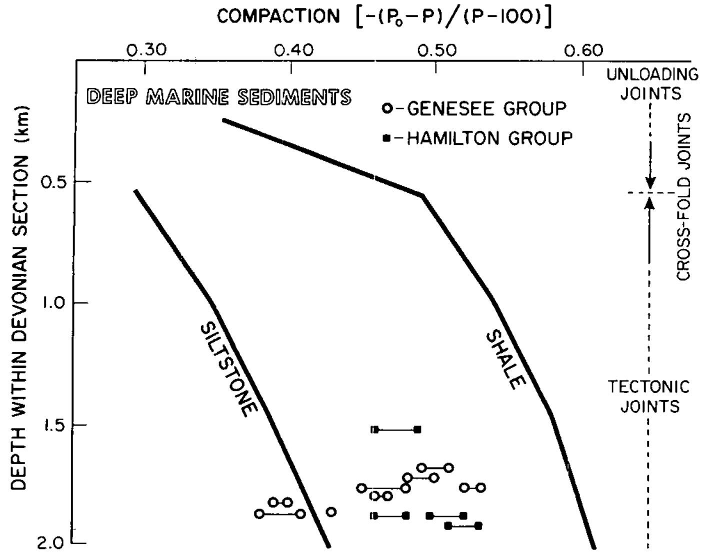

Figure 5. Compaction of samples taken from deep marine shales of Genesee and Hamilton Groups. Error bars as in Figure 3.

In wells from the Gulf Coast, the top of the abnormally pressured zone corresponds to a marked increase in the ratio of shale to sandstone (Schmidt, 1973). A similar trend is seen in moving downsection from the cap rock of wave-base sediments and into the prodelta turbidites of the Catskill Delta.

If abnormal pore pressures are high enough, the tensile strength of the host rock is exceeded, and mode I cracks will nucleate and propagate (Secor, 1965). Engelder (1985) distinguished two types of joints that form under abnormal pore pressures: tectonic and hydraulic. Their distinction is artificial but made necessary by the fact that the two types of compaction can lead to high pore pressures in a sedimentary basin. Hydraulic joints form under the influence of abnormal pressures developed solely during burial and the accompanying overburden compaction. A major tectonic compression is not involved. In contrast, tectonic joints develop during orogenic events which cause lateral shortening with volume loss inducing an increment of abnormal pressure beyond that due to compaction by burial alone; this increment of fluid pressure drives the effective maximum tensile stress to the point of joint propagation and the formation of cross-fold joints.

The correlation between abnormal pore pressures and the propagation of cross-fold joints in the Catskill Delta of central New York State is evident. Our thesis, however, is that abnormal pore pressure, which in part was caused by some combination of overburden compaction (as measured using the chlorite fabric), aquathermal pressuring, and clay dewatering during burial, was further enhanced by lateral (tectonic) compaction. The combination of this enhanced overpressure and tectonic stress led to the nucleation and propagation of tectonic joints some 50–100 m.y. after the initial deposition of the Catskill Delta. If overburden compaction alone was responsible for the cross-fold joints of the Catskill Delta, there would have been no correlation in time and space between the initial cross-fold joint sets of the Catskill Delta and the Alleghanian orogeny. This association of lateral compaction and enhanced abnormal pore pressure has been inferred (e.g., Berry, 1973) but never before documented.

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