DEVONIAN FLUVIAL TO SHALLOW MARINE STRATA, SCHOHARIE VALLEY, NEW YORK

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Introduction

The Middle to Upper Devonian Catskill clastic wedge in New York State is part of a foreland-basin fill that has relatively thick non-marine deposits to the east and thinner marine deposits to the west. The overall succession of strata is indicative of marine regression: however, there is ample evidence in coastal and fluvial strata of periodic marine transgressions. There are several superimposed scales of cyclicity in these strata, at least three of which can be readily distinguished:

(1) 10- to 100-meter thick, asymmetrical coarsening-upward to fining-upward sequences, representing on the order of hundreds of thousands to millions of years, that have been interpreted as recording relative sea-level changes associated with eustasy, tectonically induced changes in sediment supply and subsidence rate, and possibly climate change (e.g., Dennison and Head, 1975; Dennison, 1985; House, 1983, 1985; Johnson et al., 1985; Ettensohn, 1985; Brett and Baird, 1986; Van Tassell, 1987; Willis and Bridge, 1988; Dennison and Ettensohn, 1994).

(2) Meter- to 10-meter thick fining-upward or coarsening-upward sequences of sandstone and shale (representing thousands to hundreds of thousands of years) that have been interpreted to have been formed by lateral migration, abandonment and filling of channels; progradation and abandonment of channel-mouth bars, crevasse splays and tidal flats; and filling of coastal bays (e.g., Johnson and Friedman, 1969; McCave, 1968, 1969, 1973; Miller and Woodrow, 1991; Bridge and Willis, 1994). Although these sequences could have been caused by autocyclic processes such as channel switching (avulsion), allocyclic causes such as eustatic sea-level changes must also be considered (Van Tassell, 1987; Bridge and Willis, 1994).

(3) Centimeter- to decimeter-thick, sharp-based sandstone stratasets capped by shale that are commonly interpreted as the deposits of individual floods in rivers and floodplains, or storm events at sea (e.g., Gordon and Bridge, 1987; Craft and Bridge, 1987; Halperin and Bridge, 1988; Willis and Bridge, 1988; Bridge and Willis, 1994; and many others).

The validity of these interpretations of the different scales of facies sequence hinges on describing them in detail, and being able to correlate them between different outcrops. Physical tracing of strata between outcrops is made difficult by the wide spacing of outcrops (and cores), variable dip of strata, lateral changes of facies, lack of distinct marker horizons, and the absence of seismic profiles. Biostratigraphic correlation is very crude because biozones extend for on the order of a million years (100 m of strata) and only recently has it been possible to correlate marine

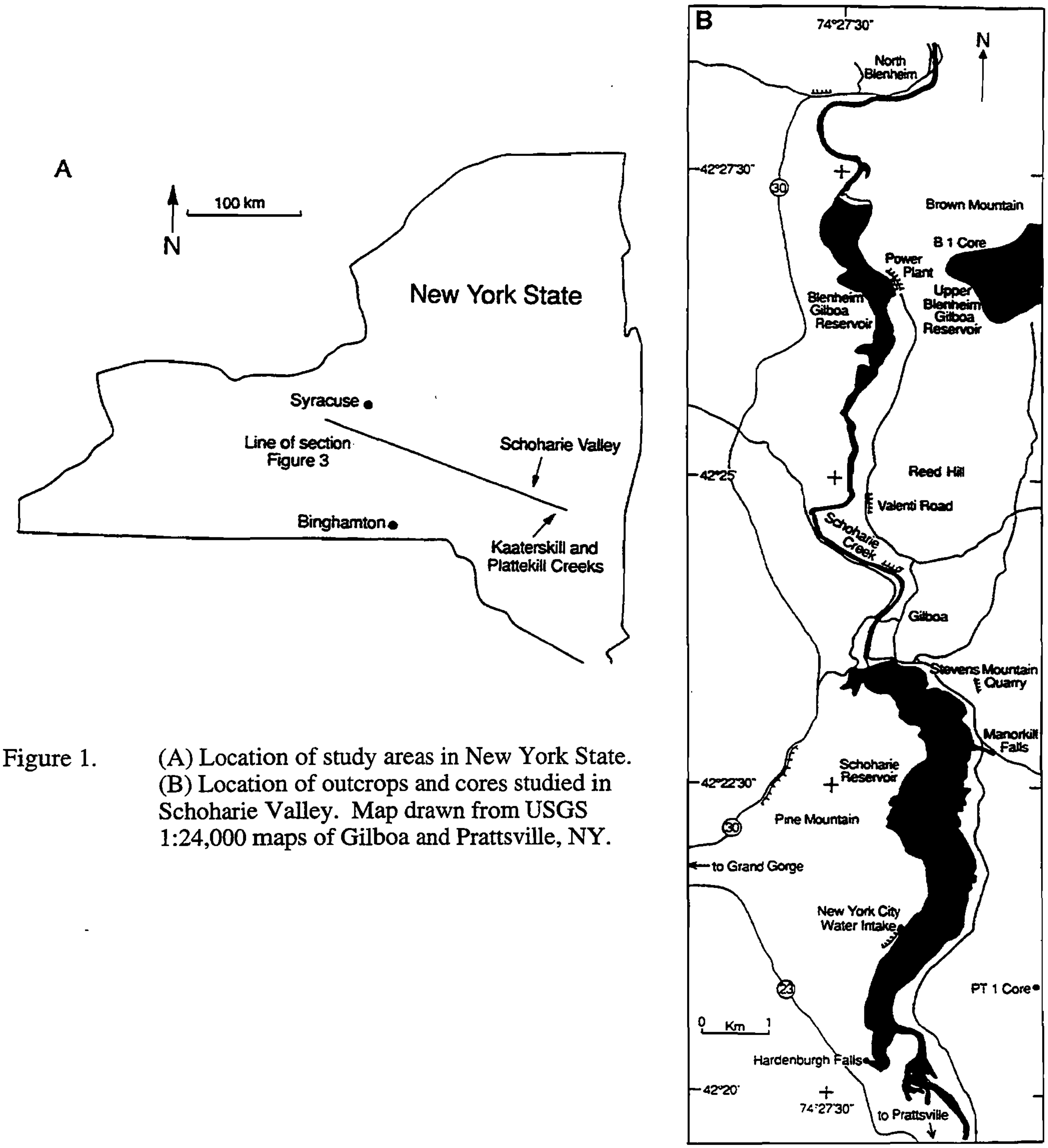

Figure 1. (A) Location of study areas in New York State. (B) Location of outcrops and cores studied in Schoharie Valley. Map drawn from USGS 1:24,000 maps of Gilboa and Prattsville, NY.

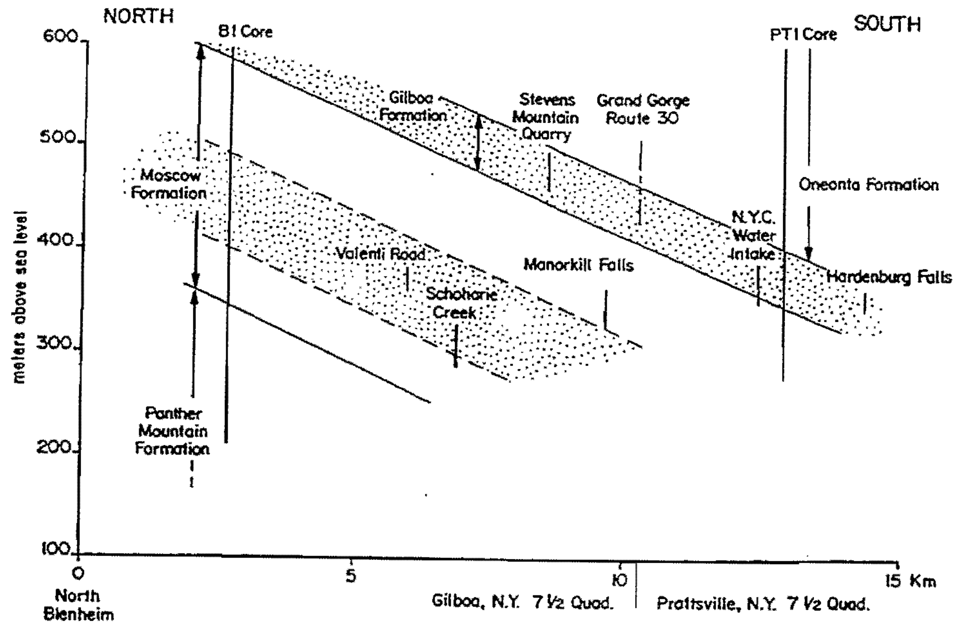

Figure 2. Stratigraphic section oriented N-S showing positions of outcrops and cores, and correlation of formations and other distinctive lithostratigraphic units within Schoharie Valley. Line of section approximately follows Blenheim-Gilboa and Schoharie Reservoirs. (Fig. 1). Correlations were made based on a regional dip to the S. of approximately 1.5° and matching of distinctive sequences which are on the order of tens of meters thick. Sandstone-dominant parts are stippled.

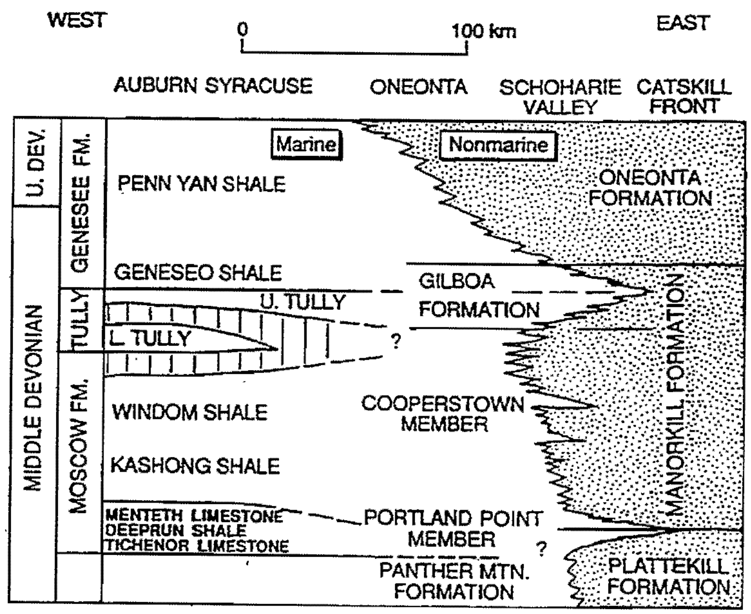

Figure 3. Stratigraphic section oriented approximately ENE-WSW (see Fig. 1 for location) showing correlation of formations in Schoharie Valley with marine formations to the west and fluvial formations to the east. Revised from work of Cooper (1933, 1934), Fletcher (1963, 1967), McCave (1968,

and non-marine rocks using miospores. Stratigraphic correlation of 10 meter thick to 100 meter thick sequences in marine strata is accomplished using laterally extensive black shales and limestones (Sutton et al., 1962; Sutton, 1963; Woodrow and Nugent, 1963; McCave, 1969, 1973; Rickard, 1975, 1989; Brett and Baird, 1985, 1986, 1990). However, their correlation with coastal and non-marine rocks remains uncertain (Halperin and Bridge, 1988; Bridge and Willis, 1994). In coastal and fluvial strata, it is difficult to correlate 10 meter thick and thinner stratasets for more than on the order of kilometers. This inability to correlate any but the thickest sequences across the basin makes it difficult to interpret these deposits in a sequence stratigraphic context, and difficult to assess whether the controls on their formation are local, regional or global.

Examples of the different scales of sequence in fluvial, coastal and shallow-marine deposits will be examined in the Schoharie Valley (in the vicinity of Gilboa, Grand Gorge and Prattsville). Bridge and Willis (1994) studied selected outcrops and cores through the interleaved fluvial and marine deposits of the Schoharie Valley (Figure 1), in order to expand upon previous descriptions and interpretations of these strata (e.g., Johnson and Friedman, 1969; McCave, 1968, 1969, 1973; Miller and Woodrow, 1991). Attention was focussed on the influence of wave, tidal and fluvial currents in shaping the coastal region, and on how these currents and coastal morphology changed during marine transgressions and regressions. Basin-scale controls on deposition were evaluated by considering coeval fluvial deposits to the east (Willis and Bridge, 1988) and offshore marine deposits to the west (Brett and Baird, 1985, 1986, 1990).

Ten-meter thick sequences were correlated between cores and outcrops over distances of up to 10 km (Figure 2; Bridge and Willis, 1994). Stratigraphic subdivision of these rocks (Figure 3; modified from Rickard, 1975, 1989) is due mainly to the work of Cooper (1930, 1933, 1934) and Cooper and Williams (1935). Some formations were defined based on fossils in fully marine deposits, but key zone fossils are absent in coastal and fluvial deposits. Nevertheless, the formations defined by Cooper can generally be recognized in Schoharie Valley. Outcrops to be visited span the upper Moscow Formation (Manorkill Falls), the Gilboa Formation (Route 30 near Grand Gorge, Hardenburgh Falls; Stevens Mountain Quarry), and the lower Oneonta Formation (Route 30 near Grand Gorge). These Formations are mid to late Givetian in age. This stratigraphic interval includes a major transgression and regression, upon which are superimposed smaller scale changes in relative sea level.

Hardenburgh Falls

Description

The Hardenburgh Falls outcrop (Figure 4) is assigned to the middle-upper part of the Gilboa Formation, and is equivalent to the middle part of the Stevens Mountain Quarry section (Figure 2). The lowest 6 meters of this exposure (Figure 4) comprise gray sandstones, dark gray siltstones and shales. The mudstones are sparsely fossiliferous and typically bioturbated. The fine to very-fine grained sandstones are sharp-based, cm- to dm-thick sheets and lenses. Thinner ones contain wave-ripple cross strata and are capped by wave ripple marks (2-D, 3-D, and interfering types). Thicker, coarser grained sandstones have hummocky-swaley cross strata and/or planar strata in their lower parts, and abundant intraformational shale fragments and disarticulated shelly

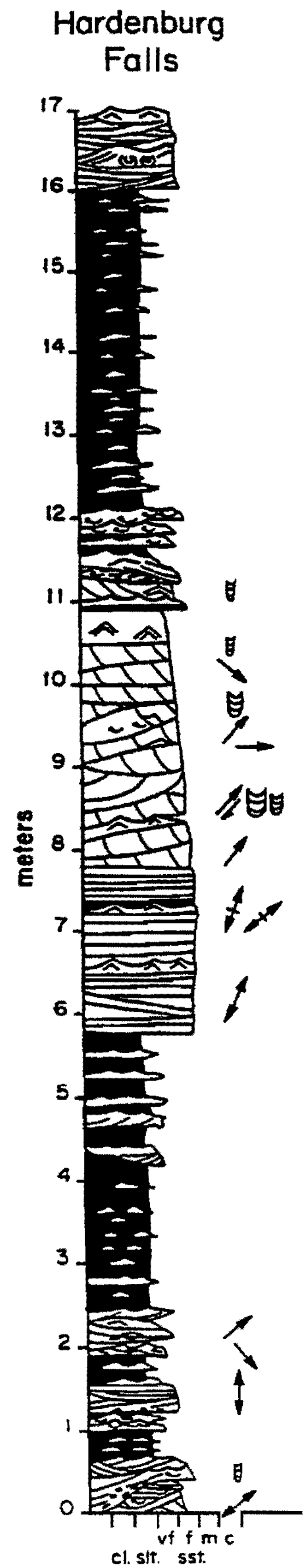

Figure 4 (A). Sedimentological log from Hardenburgh Falls. The sandstone body in this outcrop is equivalent to sandstone body (b) of Figure 5. Symbols explained in Figure 4 (B).

LEGEND

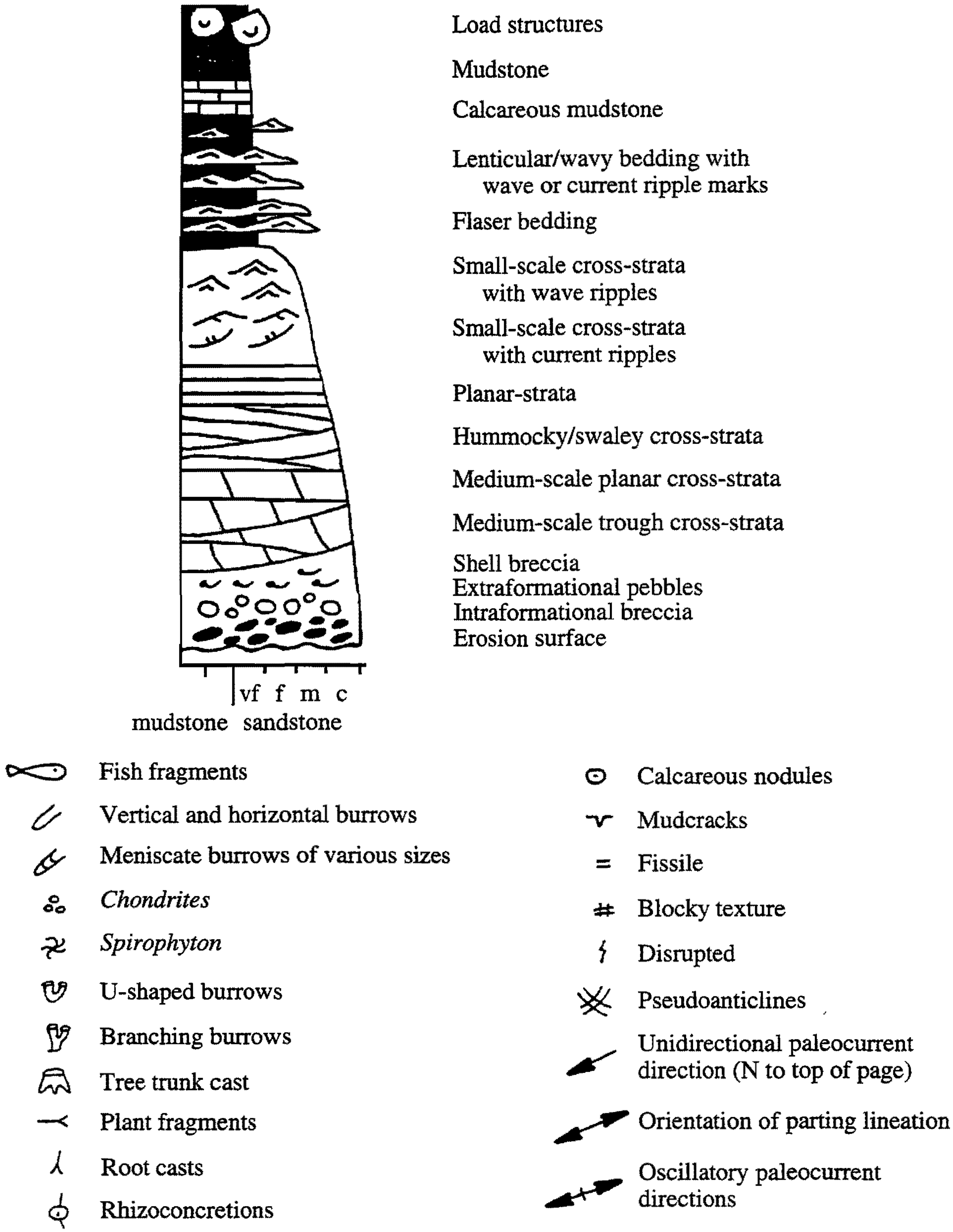

Figure 4 (B). Legend for Figures 4-7. Set thicknesses of medium-scale cross strata are cm to dm thick, and the cross strata are normally at the angle of repose; however, they are locally lower angle and transitional to hummocky-swaley cross strata.

fossils. Near the base of the section (e.g. Figure 4, meter 2) relatively thick sandstone stratasets have unidirectionally dipping low-angle to high-angle, medium-scale cross strata with easterly paleocurrents, and extraformational pebbles. At the base of the section, amalgamated sandstone stratasets fill N-S oriented channels that are up to decimeters deep and meters wide. These relatively coarse grained strata contain rounded extraformational chert and quartzite pebbles.

Shell accumulations include the brachiopods *Spinocyrtia, Mucrospirifer, Mediospirifer, Orthospirifer, Cupularostrum,* the bivalves *Goniophora, Actinopteria, Palaeoneilo,* and crinoid ossicles. Most of the disarticulated shell fragments lie concave down along stratal surfaces and show little evidence of breakage or abrasion of fine detail. Some shell concentrations are markedly monospecific (e.g *Cupularostrum).* Small, centimeter-diameter vertical burrows occur in the tops of sandstone strata.

The 5.5 meter thick, medium- to fine-grained sandstone body in the middle of this section has a sharp base that is overlain by planar strata and swaley cross strata with horizons of wave ripples draped by shale. Asymmetrical wave ripples indicate a NE-SW oscillation with a preferred migration to the SW. The overlying medium-scale trough cross strata dip to the NE mainly (with a rare, oppositely directed set) and contain superimposed wave- and current-ripple marks. Medium-scale planar cross strata with a SE dip, occurring above the trough cross strata, climb up low-angle surfaces that dip to the NW. Vertical, meniscate burrows of varying diameter occur throughout the sandstone body. Several stratasets of shell-rich sandstone with hummocky cross strata and wave ripples occur immediately above the sandstone body.

Interpretation

Interbedded mudstones and sandstones with hummocky cross strata, planar strata, wave-ripple marks and associated cross strata, and abundant shelly-fossil concentrations are interpreted as nearshore marine deposits formed below fair-weather wave base (e.g. Dott and Bourgeois, 1982; Hunter and Clifton 1982; Swift et al., 1983). Sandstones were formed during storms by either dominantly wave currents or by combined wave and unidirectional currents (Craft and Bridge, 1987; Southard et al., 1990, Duke et al., 1991), whereas mudstones represent fair-weather deposits or those accumulated below storm-wave base. Amalgamated sandstone stratasets indicate deposition mainly above fair-weather wave base. The preferred easterly paleocurrent orientation of some low- and high-angle cross strata, and the association of increasing grain size with increasing inclination of these cross strata, suggest onshore directed currents (asymmetrical wave, and/or flood tidal) that in places were strong enough to change hummocky bedforms to dunes (e.g. Nottvedt and Kreisa, 1987; Arnott and Southard, 1990; Myrow and Southard, 1991; Cheel and Leckie, 1992). The shallow channels may be cut by either storm-induced or tidal currents (e.g. Cacchione et al., 1984; Craft and Bridge, 1987).

The sandstone body in the middle of this exposure is interpreted as the deposits of a tidal-channel mouth bar that prograded into a storm-wave dominated sea. The vertical variation in sedimentary structures indicates wave currents or combined wave and unidirectional currents lower down, but dominantly unidirectional currents higher up. Paleocurrents associated with wave ripples are generally alongshore (NE-SW). The alongshore to landward-directed trough cross strata (NE-E) are probably associated with sinuous-crested dunes formed by tidal flood currents, and ebb tidal currents were relatively unimportant. Planar cross strata migrating to the SE may represent wave-formed swash bars or tide-formed straight-crested dunes (e.g. Hayes and Kana, 1976; Boothroyd, 1978; Fitzgerald, 1984; Sha and De Boer, 1991). Superimposed wave- and current-ripples indicate periods of reduced wave- and tidal-current strength. The sparse fauna and low degree of bioturbation, but dominance of the *Skolithos* ichnofacies, is typical of sandy deposits in coastal areas (Howard and Reineck, 1981).

The relatively sharp base and top of this sandstone body could be interpreted as due to diversion of a channel into this area followed by abandonment, reduction in sand supply, and marine transgression. However, an alternative explanation could be one of progradation of a sandy shoreface during a relative sea-level fall (a forced regression), followed by modest erosion of the top of the sandstone body during marine transgression. These alternatives are discussed below.

Stevens Mountain Quarry

Description

Stevens Mountain Quarry exposes a large proportion of the Gilboa Formation (Figure 2). Attention will be directed to the 2 to 6 meter thick sandstone bodies that occur in the upper 30 meters of the quarry (Figure 5). These fine-to medium-grained sandstone bodies have sharp, erosional bases with tool and flute marks, overlain by intraformational breccia comprising shale clasts, plant fragments and shelly fossils (bivalves, brachiopods, *Tentaculites* and crinoid fragments). Relatively thin (2 to 3 meters) sandstone bodies are composed of amalgamated, dm- to m-thick stratasets containing swaley cross strata and planar strata. The erosional bases of individual stratasets are commonly overlain by transported shell fragments. These sandstone bodies vary little in grain size throughout their thickness: however, they tend to be capped with concentrations of shells, intraformational shale fragments and rounded extraformational pebbles (e.g. Figure 5A, 15-17m, Figure 5B, 7-9m and 12-15m). As the thickness of these sandstone bodies increases, stratasets of medium-scale cross-stratified sandstone begin to dominate their upper parts (e.g. Figure 5B, 19-23m). This kind of sandstone body may fine upwards, coarsen upwards or show little vertical variation in grain size. Like the thinner examples, they are capped by concentrations of shells, intraformational breccia and extraformational pebbles. The thickest sandstone bodies are medium grained and composed almost entirely of medium-scale cross strata, with lenses of planar strata. Erosional reactivation surfaces in cross sets are common. In places, asymmetrical ripple marks on reactivation surfaces and cross stratal surfaces indicate a paleocurrent direction opposite to the medium-scale cross strata (e.g. Figure 5A, 33-39m, Figure 5B, 29-34m). Stratasets in these sandstone bodies may be inclined at several degrees relative to the basal erosion surface (i.e. large-scale inclined strata) and sandstone-filled channels may occur in their upper parts. Immediately above some sandstone bodies, sandstone stratasets may have unidirectionally dipping low-angle and angle-of-repose cross strata, abundant shell accumulations and intraformational and extraformational pebbles. Such stratasets may also be markedly lenticular and fill channels up to dm deep and meters wide (as at the base of Hardenburgh Falls). Laterally extensive horizons of sandstone load casts occur immediately beneath some sandstone bodies.

Stevens Mountain Quarry North

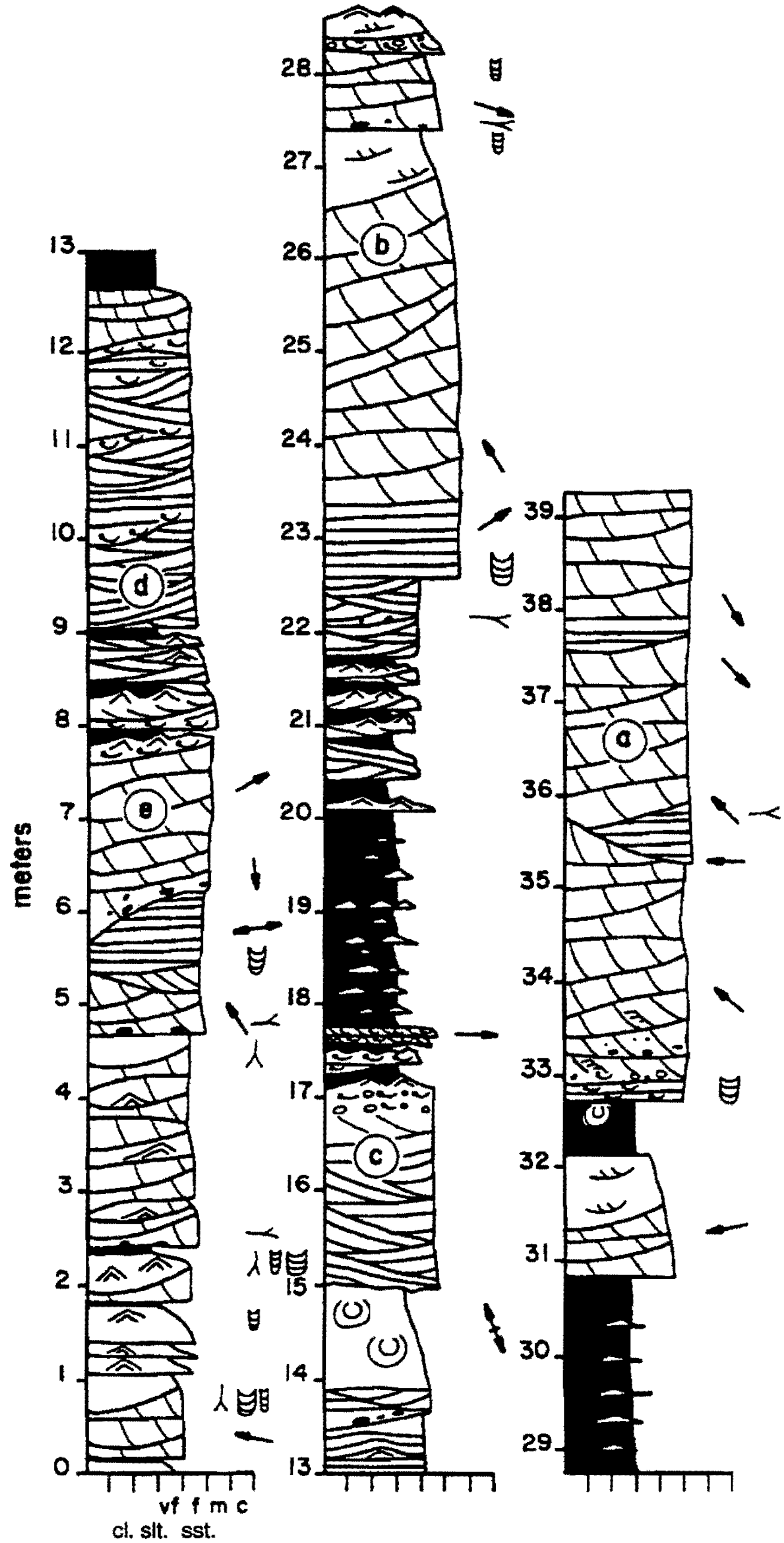

Figure 5 (A). Sedimentological log from the north end of Stevens Mountain Quarry. Sandstone bodies designated (a) through (e) are identified on both logs in order to illustrate lateral variations.

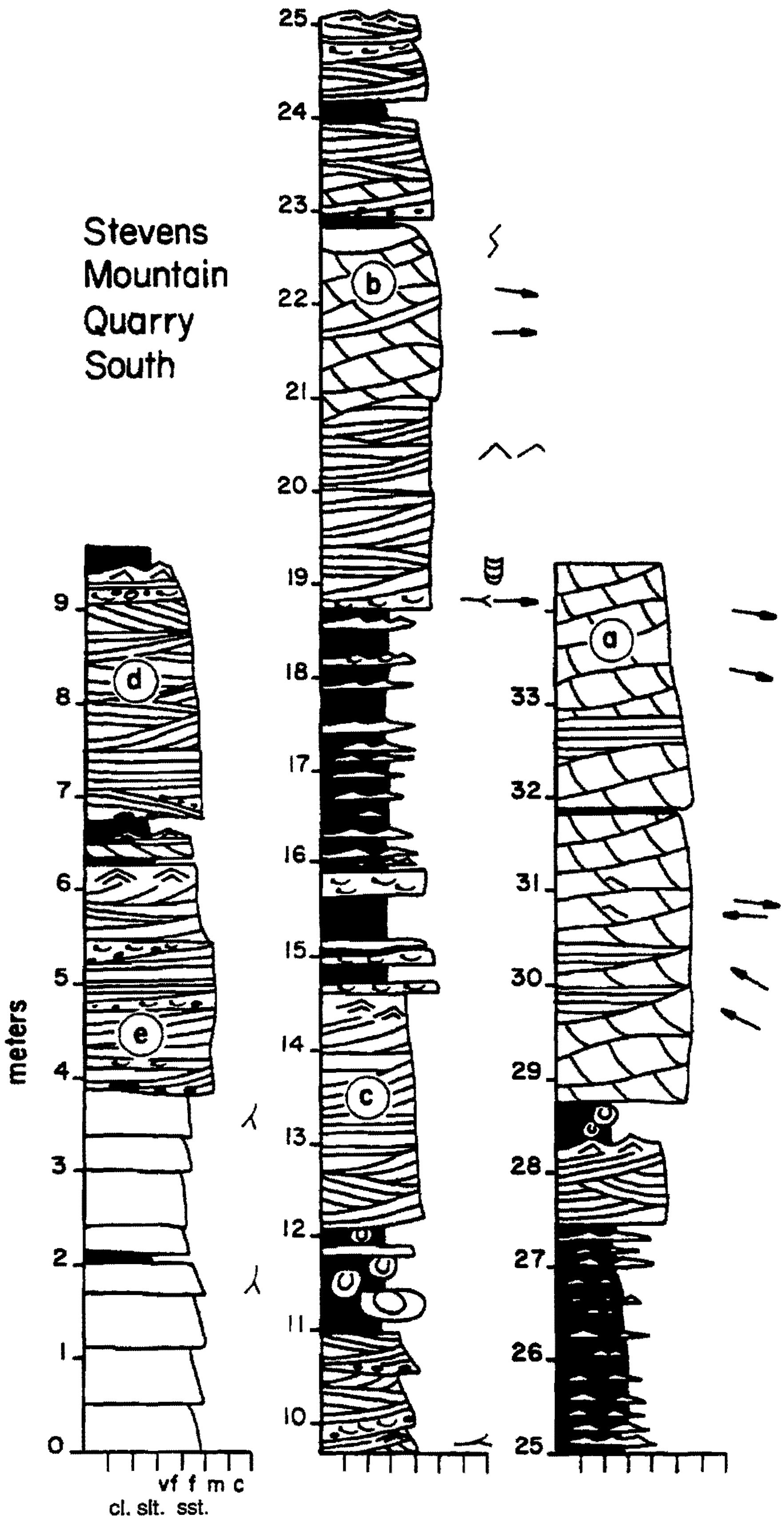

Figure 5 (B). Sedimentological log from the south end of Stevens Mountain Quarry. Sandstone bodies designated (a) through (e) are identified on both logs in order to illustrate lateral variations.

Paleocurrent directions from medium-scale cross strata in and above the upper parts of all sandstone bodies are generally in an easterly direction (range from NNE to SE). Medium-scale cross strata in the lower parts of the thickest sandstone bodies are generally oriented approximately 180 degrees to those higher up. Tool marks at the base of sandstone bodies give paleocurrents to the W and NW or to the E and SE.

Skolithos burrows (mm to cm diameter) are common in tops of sandstone stratasets. *Arenicolites* and *Chondrites* also occur. Large (up to 0.lm diameter) vertical, sand-filled meniscate burrows are characteristic of thicker sandstone bodies. Some sandstones are so bioturbated that primary sedimentary structures are difficult to discern (e.g. base of quarry). Fossils in the shell accumulations include brachiopods (*Spinocyrtia, Mucrospirifer, Tropidoleptus carinatus, Cupularostrum (Camarotoechia), Mediospirifer, Orthospirifer mesastrialis),* bivalves *(Goniophora, Grammysia, Actinopteria, Cypricardella, Palaeoneilo), Tentaculites* and crinoid ossicles (Cooper and Williams, 1935).

The thickness and sedimentary characteristics of sandstone bodies and mudstone intervals change laterally over hundreds of meters (compare sandstone bodies e and b in Figures 20A and B). Also, the thickest sandstone bodies dominated by angle-of-repose cross strata occur at the top of the quarry, whereas relatively thinner ones with mainly swaley cross strata and abundant shelly fossil accumulations occur lower down.

Interpretation

The thickest cross-stratified sandstone bodies with erosional bases overlain by intraformational breccia with shelly fossils, large-scale inclined strata and channel-fills, and opposing paleocurrent directions, are interpreted as deposits of laterally migrating tidal channels (see also Johnson and Friedman, 1969). Dominance of ebb-directed paleocurrents lower in sandstone bodies and flood dominance in the upper parts (in places associated with channel fills) is typical of strongly asymmetrical tidal currents in tidal inlets, estuaries, and tide-influenced deltaic distributaries. Variable dip angles of medium-scale cross-strata, erosional reactivation surfaces and superimposed current ripples are further evidence of strong tidal-current asymmetry. They indicate growth of sinuous-crested dunes to 'full vortex' stage during the dominant tidal current, slackening of the current and modification of dune geometry, then erosion of the dune by the subordinate tide. Such features are commonly reported from mesotidal and macrotidal settings (e.g. Terwindt, 1981; Boersma and Terwindt, 1981; Dalrymple, 1984; DeMowbray and Visser, 1984). General absence of tidal-bundle sequences and rarity of current ripples on reactivation surfaces are due to an erosional or nondepositional subordinate flood-tidal current and a dominant ebb-tidal current that may have been reinforced by a fluvial current.

Sandstone bodies in which lower parts have swaley cross strata, planar strata and wave ripples but upper parts have angle-of-repose cross strata, suggest wave currents or combined wave and unidirectional cants lower down but dominantly unidirectional currents higher up. Interpretation of the depositional environment of these sandstone bodies hinges critically on their lateral transition to thicker sandstone bodies dominated by angle-of-repose cross strata and thinner sandstone bodies dominated by amalgamated swaley cross strata. In some locations, sandstone bodies have sharp erosional bases, but in others they occur at the top of coarsening-upward sequences with interbedded hummocky cross-stratified sandstones and shales immediately beneath. These features suggest deposition on channel mouth bars that were prograding into a marine, storm-wave dominated area. Such bars may have been on the seaward side of tidal inlets (i.e. ebb-tidal deltas) associated with estuaries or barrier-beach shorelines, or associated with tide-influenced deltaic distributaries. Landward-directed (NE to SE) trough cross-strata are most likely associated with sinuous-crested dunes formed by tidal flood currents, but also perhaps with asymmetrical shoaling wave currents.

The relatively thin sandstone bodies dominated by amalgamated swaley cross- stratified strata are interpreted as more distal parts of channel mouth bars, that were completely dominated by storm waves. The lack of mud suggests deposition above fair weather wave base. The common occurrence of load casts near the base of these sandstone bodies suggests rapid deposition of sand on mud. Such soft-sediment deformation features are common offshore from Mississippi delta distributaries (e.g. Coleman and Prior, 1982). Paleocurrent indicators in the bases of sandstone bodies suggest a dominantly offshore directed unidirectional current, as is common in similar Frasnian rocks in the Catskill Region (Craft and Bridge, 1987; Halperin and Bridge, 1988; Bishuk et al., 1991) and in many other ancient storm-dominated shelf deposits (Leckie and Krystinik, 1989). Offshore-directed currents may be associated with coastal set-up during storms (e.g. Craft and Bridge, 1987, Halperin and Bridge, 1988; Leckie and Krystinik, 1989; Duke, 1990; Duke et al., 1991). However, fluvial and tidal currents may also have had an influence. Lenticular, pebbly and shelly sandstone strata, locally in channels, at the top of these sandstone bodies are evidence for relatively strong shoreward directed currents, possibly due to tidal flood, "storm-tides", or asymmetrical shoaling-wave currents (e.g. Cheel and Leckie, 1992). Thus, these coarse, shelly strata on top of sandstone bodies are associated with rapid abandonment of the channel-mouth bar, reduction in sand supply, and marine transgression (see, for example, Boyd and Penland, 1988; Boyd et al., 1989; Sha and DeBoer, 1991; Penland et al., 1988).

Bioturbation is characteristically rare in coastal sand bodies such as these (Howard and others, 1975; Howard and Reineck, 1981). Large, vertical meniscate burrows are similar to those reported in fluvial and coastal channel deposits elsewhere (Thoms and Berg, 1985; Miller, 1979; Bridge, Gordon and Titus, 1986), and may be due to upward escape of bivalves. The brachiopod-bivalve fauna is typical of Devonian nearshore communities (McGhee and Sutton, 1985; Sutton and McGhee, 1985).

The vertical variations in sandstone body characteristics within the upper 30 meters of Stevens Mountain quarry suggest a decreasing marine influence, with progressively more proximal channel bar deposits higher up. This trend is continued upwards into the non-marine Oneonta Formation.

Manorkill Falls

Description

The Manorkill Falls section (Figure 6) exposes the upper part of the Moscow Formation where a pronounced grain-size fining occurs (Figure 2). When the water level in Schoharie Reservoir is unusually low, up to 10 meters of sandstone is exposed (below the log in Figure 6). Many sandstone casts of tree trunks (part of the famous Gilboa fossil forest) are seated in the gray-green disrupted mudstones immediately above these sandstones. The casts stand upright within the lowest, decimeter-thick sandstone strataset of a 2-m thick sandstone body (Figure 6, 3.5-5.5 m). Within this upward-fining sandstone body, sheet-like stratasets contain mainly medium-scale cross strata with easterly paleocurrents. Reactivation surfaces with ripple marks are superimposed on these cross strata. Sandstone stratasets between meters 6 and 7 have low-angle (swaley?) cross strata, wave-rippled upper surfaces, and rare shelly fossils (bivalves, brachiopods, crinoid ossicles, bryozoa). The distinctive wedge-shaped sandstone body between meters 7 and 8 has low-angle bedding surfaces that dip in the direction the sandstone body thins. Upper surfaces of these large-scale inclined strata are covered with symmetrical and interfering wave-ripple marks. Ichnofossils are abundant and diverse throughout these sandstone bodies, and include *Skolithos, Meunsteria, Arenicolites and Diplocraterion*.

The overlying mudstone-dominated sequence (meters 8-24, Figure 6) contains relatively undisrupted and fissile mudstones that grade up into intensely disrupted mudstones with desiccation cracks, root casts, burrows, and calcareous rhizoconcretions. Interbedded sandstone stratasets are mainly wave-ripple cross stratified, but may have planar or hummocky-swaley strata at the base. Between meters 18 and 21 (Figure 6) such stratasets occur in E-W oriented, fine-grained channel fills. Some upward-fining sandstone stratasets have medium-scale and small-scale stratasets with easterly paleocurrents (e.g., 16-18 m; Figure 6) Rare transported shelly fossils and tree trunk casts occur again near the top of the section. At the top of the measured section, a 6 meter thick sandstone body has stratasets inclined at up to 5 degrees relative to its basal erosion surface (i.e. large-scale inclined strata). Stratasets lower down in the sandstone body have planar and swaley stratification with superimposed ripple marks. Higher up, stratasets of medium-scale cross strata fill channel forms.

Interpretation

The mudstones at the base of the measured section record floodbasins with soils that supported a forest of trees. The sandstones that overlie these deposits, including those that buried and preserved the bases of tree trunks, show evidence for landward progradation of wedges of sand into standing water. This and the fossil content indicate a back-shoal washover origin. The overlying fissile mudstones with wave-rippled sandstone indicate deposition in quiet standing water that was periodically agitated by waves, possibly an interdistributary bay or back-barrier lagoon. A return to subaerial conditions is signalled by the occurrence of desiccation cracks, root casts, and calcareous paleosol nodules. Overlying mudstones and sandstones with evidence of landward-directed unidirectional paleocurrents, wave currents, periodic exposure and fine-grained channel fills may have been deposited on sandy to muddy tidal flats that changed upwards to

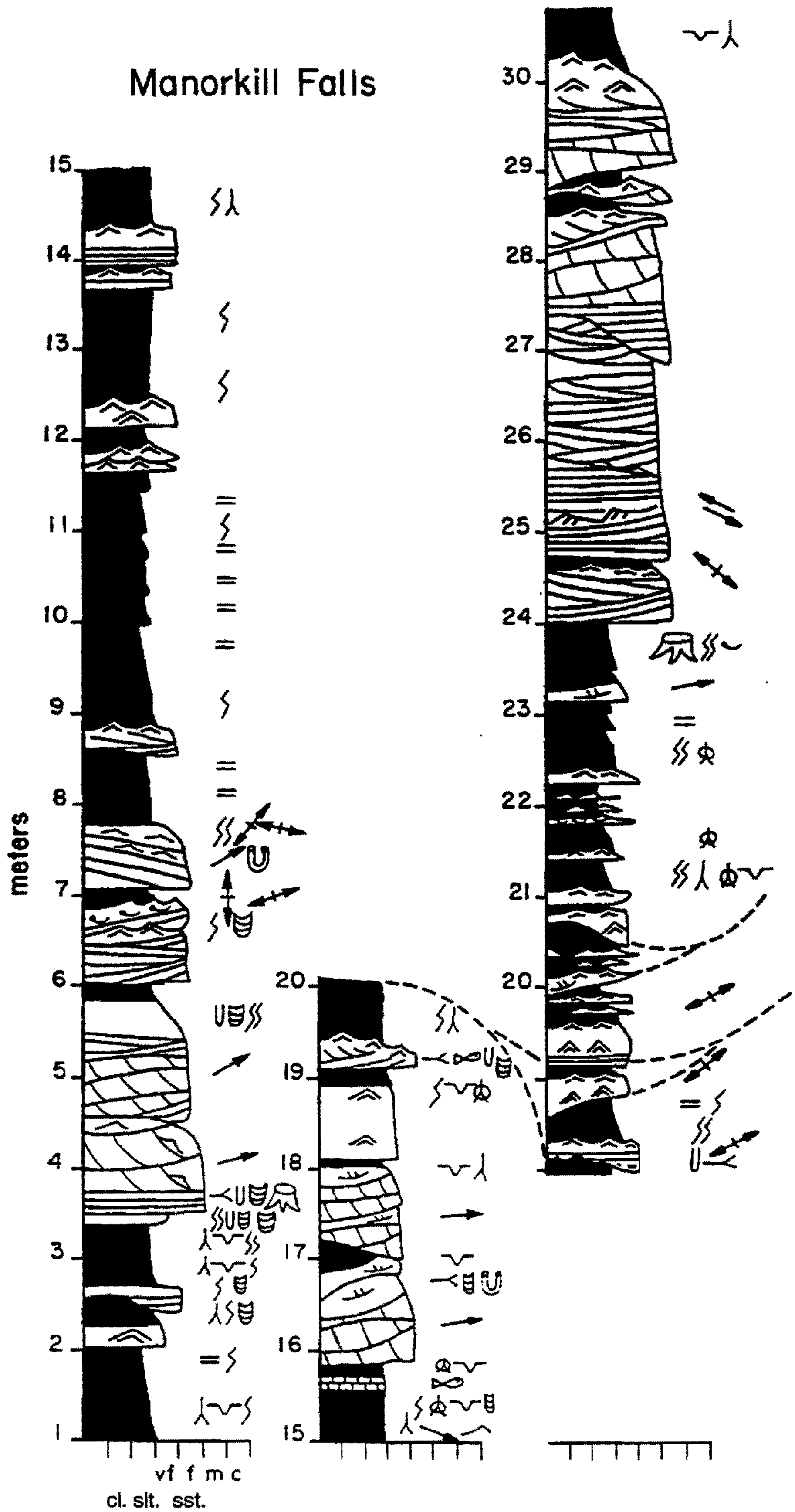

Figure 6. Sedimentological log of Manorkill Falls Section, east side of Schoharie Reservoir.

subaerial plains. The landward-directed paleocurrents may be associated with tidal floods and/or storm overwash. The sandstone body at the top of this section is interpreted as a tidal-channel mouth-bar deposit.

Gilboa

At the side of the road at the northern tip of Schoharie Reservoir, immediately to the west of Schoharie Creek, there is an exhibit of sandstone casts of tree trunk casts that were recovered during excavations associated with the construction of the dam. There are a number of places in this area where such tree trunk casts can be seen in situ. One of these is Manorkill Falls.

Route 30 near Grand Gorge

Description of lower part

This road cut just north of Grand Gorge (Figure 1) exposes the upper Gilboa and lower Oneonta Formations, the overall sequence recording marine regression. The lowest 28 m of the section (Figure 7) comprises the Gilboa Formation. This Formation is composed of 2 to 6 m thick sandstone bodies separated by mudstone-dominated intervals.

The sandstone bodies have sharp bases with erosional sole marks, overlain in places by intraformational breccia composed of shale clasts, plant fragments, and shelly fossils (bivalves, brachiopods, crinoid fragments, and *Tentaculites*). The thinner sandstone bodies (2-3 m) are composed of amalgamated, decimeter to meter thick stratasets of swaley cross strata and planar strata. Erosional bases of individual stratasets are commonly overlain by transported shell fragments, and wave-ripples are rarely preserved beneath the thin shales draping the stratasets. Grain size of these sandstone bodies varies little through their thickness, except at the top where concentrations of shells, intraformational and extraformational pebbles occur.

As the thickness of these sandstone bodies increases, medium-scale cross strata begin to dominate their upper parts (e.g. meters 14-28). The thickest sandstone bodies are almost entirely composed of medium-scale cross strata. Medium-scale cross strata show opposing paleocurrent directions, especially in the upper parts of sandstone bodies. Symmetrical and asymmetrical ripple marks (draped with shale in places) superimposed on medium-scale cross strata also indicate divergent paleocurrents. These sandstone bodies may fine upwards, coarsen upwards, or show little vertical variation in mean grain size. They may also be capped with pebbly and shelly beds.

The dark gray, sparsely fossiliferous shales are typically bioturbated and interbedded with gray, centimeter to decimeter thick sandstone sheets and lenses. The thinner sandstones have wave-ripples and associated cross strata. Thicker, coarser grained ones have hummocky-swaley cross strata and/or planar strata in their lower parts. Sandstone load casts occur immediately beneath some sandstone bodies.

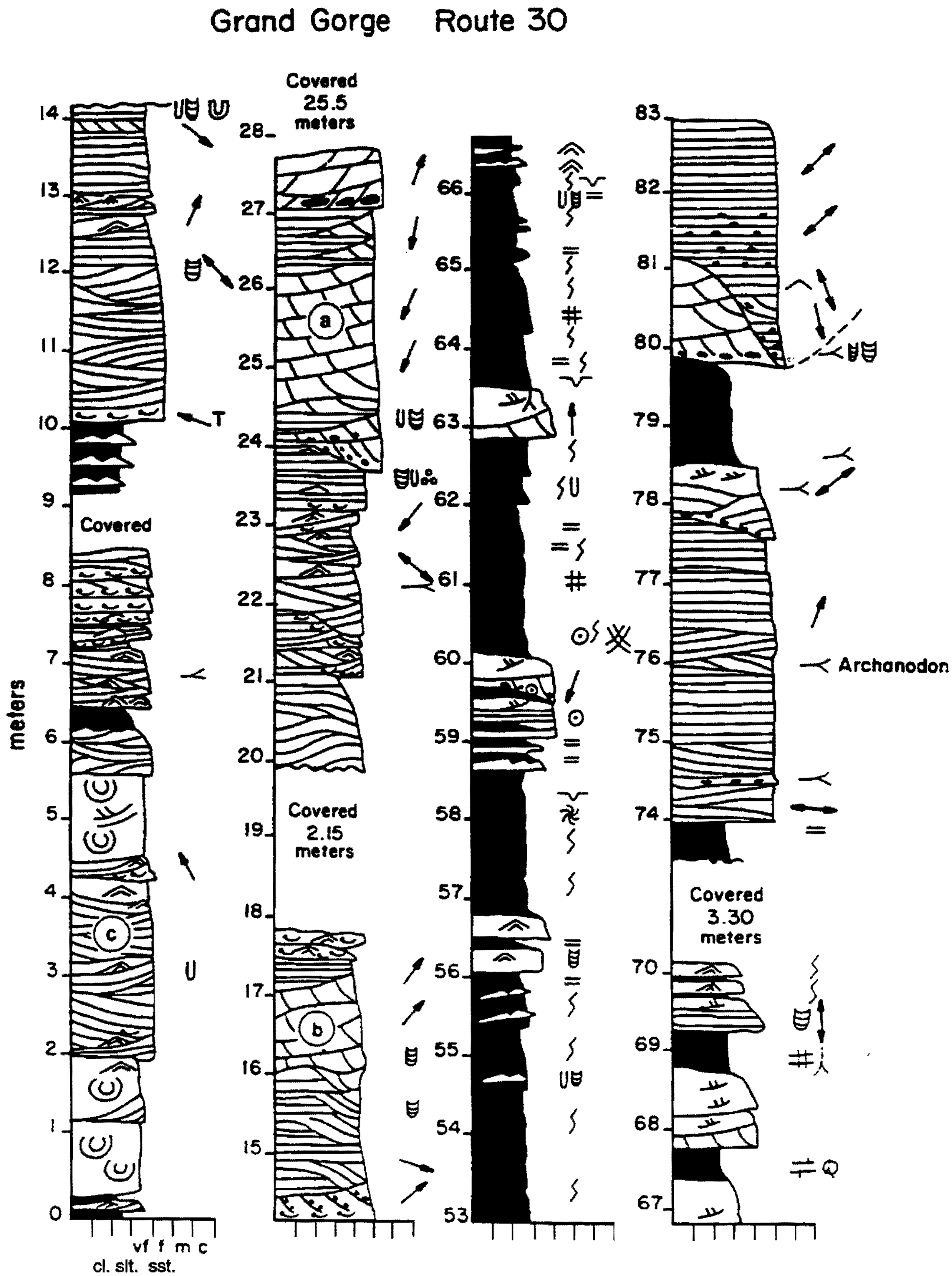

Figure 7. Sedimentological log of Rt. 30 roadcut, north-east end of Grand Gorge. Sandstone body (b) corresponds to that in Figure 5.

Skolithos burrows (mm to cm diameter) are common in the tops of sandstone stratasets. *Arenicolites* and *Chondrites* also occur. Large (up to 0.1 m diameter) vertical, sandstone-filled meniscate burrows are characteristic of thicker sandstone bodies. Shelly fossils include brachiopods (*Spinocyrtia, Mucrospirifer, Tropidoleptus carinatus, Cupularostrum, Mediospirifer, Orthospirifer mesastrialis*), bivalves (*Goniophora, Grammysia, Actinopteria, Cypricardella, Paleoneilo*), *Tentaculites*, and crinoid ossicles. Most shell fragments are disarticulated, lie concave down along bedding planes, and show little evidence of breakage.

Interpretation of lower part

The thickest, cross-stratified sandstone bodies are interpreted as deposits of laterally migrating tidal channels (Johnson and Freidman, 1969; Miller and Woodrow, 1991; Bridge and Willis, 1994). Ebb-directed paleocurrents are dominant in the lower parts of sandstone bodies, whereas there is flood dominance in the upper parts, as is typical of asymmetrical, coastal tidal currents. The general absence of tidal-bundle sequences and rarity of current ripples on reactivation surfaces in medium-scale cross strata is an indication of strongly asymmetrical tidal currents where either the flood or ebb current is dominant during deposition.

Some of the sandstone bodies indicate the influence of wave currents or combined currents lower down, but predominantly unidirectional currents higher up. Interpretation of the origin of these sandstone bodies depends on their lateral transition to the other types of sandstone bodies, as observed in nearby outcrops of coeval strata (e.g., Stevens Mountain Quarry, Figure 20). Accordingly, these bodies are interpreted as channel-mouth bars that were prograding into a storm-wave dominated sea. Such bars may have been on the seaward side of tidal inlets (i.e. ebb-tidal deltas) associated with estuaries or barred coastlines, or associated with tide-influenced deltaic distributaries (Johnson and Friedman, 1969; Miller and Woodrow, 1991; Bridge and Willis, 1994). Landward-directed medium-scale cross strata may be associated with flood-tidal currents or may represent wave-formed swash bars. Ripple marks on dune/bar slip faces indicate periods of tidal slack water or reduced tidal currents.

The thinnest sandstone bodies dominated by swaley-hummocky cross strata are interpreted as more distal parts of channel-mouth bars that were dominated by storm waves and associated currents. The lack of mud suggests deposition above fairweather wave base. The occurrence of load casts near the base of these sandstone bodies suggests rapid deposition of sand on mud, as is common offshore from deltaic distributaries.

The shelly and pebbly strata at the top of the sandstone bodies are evidence of relatively strong shoreward-directed currents, possibly due to flood tides, "storm tides", or asymmetrical shoaling-wave currents. They appear to be associated with abandonment of the channel-mouth bar, relative reduction in sand supply, and marine transgression.

The sparse fauna, low degree of bioturbation, and dominance of *Skolithos* ichnofacies in the sandstones is typical of sandy deposits in coastal areas. Large, vertical meniscate burrows are similar to those reported in fluvial and coastal deposits elsewhere (Thoms and Berg, 1985; Miller, 1979; Bridge et al., 1986) and some may be due to upward escape of bivalves. The brachiopod-bivalve fauna is typical of Devonian nearshore communities (McGhee and Sutton, 1985; Sutton and McGhee, 1985).

The shales with interbedded sandstones are interpreted as nearshore marine deposits formed below fairweather wave base (details in Bridge and Willis, 1994). Sandstones were formed during storms by either wave currents or combined wave-unidirectional currents, whereas mudstones represent fairweather deposits or those accumulated below storm wave base.

Description of upper part

The lower Oneonta Formation (Figure 6, meters 53-83) comprises red, green and gray mudstones with subordinate red to gray sandstones. The mudstones range from fissile and relatively unbioturbated to intensely disrupted with blocky texture, slickensided clay-lined surfaces, horizons of calcareous nodules, and rare pseudoanticlines. Desiccation cracks, root casts and burrows are pervasive. The distinctive trace fossil *Spirophyton* is present but hard to find. Sandstone stratasets are sharp based, cm to dm thick sheets and lenses. The thinner ones have small-scale cross strata with wave- or current-rippled tops. Thicker ones tend to have medium-scale cross strata or planar strata at their bases. Centimeter-diameter, vertical meniscate burrows (*Beaconites*) are common, and root casts and calcareous concretions may occur at the tops of the sandstone stratasets. Groups of several sandstone stratasets with intervening mudstone may be arranged into meter-thick upward-fining or upward-coarsening sequences (e.g., Figure 6, 58-60 m). Two 4 to 6 meter thick sandstone bodies occur at the top of the sequence (Figure 6, 74-83 m). These have erosional bases (lined with intraformational shale clasts, plant fragments, and rarely the non-marine bivalve *Archanodon*) and large-scale inclined strata, that are channel filling in places (meters 80-83). Internal structures are medium-scale cross strata (mainly low angle), planar strata, and cm-diameter, meniscate vertical burrows.

Interpretation of upper part

Disrupted mudstones represent floodbasin deposits. Blocky textures, desiccation cracks, calcareous concretions and pseudoanticlines indicate repeated wetting and drying and formation of ped structure typical of calcareous vertisols. Fissile siltstones indicate relatively less bioturbation but repeated exposure, and may represent relatively high deposition rates and/or hypersalinity in ephemeral lakes. *Spirophyton* in fissile siltstones suggests more permanent standing water, possibly brackish (Miller, 1979; Miller and Johnson, 1981). Wave-rippled lenses indicate periodic wave activity. The thinner sandstones with unidirectional sedimentary structures indicate periodic sand deposition during floods, either as individual sheet-flood deposits within the floodbasin, or associated with levees or crevasse splays. Coarsening-upward sequences indicate progradation of levees or crevasse splays. Root casts and calcareous concretions in and above the tops of these units indicate reduction in deposition rate and soil formation. The thick sandstone bodies are interpreted as the deposits of river-channel bars and fills, with some tidal influence possible.

Discussion of meter-scale sequences in coastal deposits of the Catskill clastic wedge

Meter-scale sequences were interpreted by Bridge and Willis (1994) to record changes in relative sea level related to processes such as lateral migration, abandonment and filling of tidal channels; progradation of channel mouth bars and tidal flats; and filling of coastal bays. These sequences are parasequences using the terminology of Van Wagoner et al. (1988, 1990) and Mitchum and Van Wagoner (1991). As average Givetian deposition rates were very approximately 0.1 mm/yr, meter-scale sequences represent 10^4 to 10^5 years on average.

Sharp-based coastal sandstone bodies similar to those in the Schoharie Valley have been ascribed to deposition on prograding strandplains cut by estuaries or tidal inlets, and to wave-formed offshore bars or shoals (references in Bridge and Willis, 1994). Sandstone bodies with swaley-hummocky cross strata overlain by angle-of-repose cross strata have been interpreted as beach and shoreface deposits, their sharp bases being related to wave-current erosion during relative sea-level fall and so-called forced regression. The coarse grained strata on the tops of these sandstone bodies were taken as so-called transgressive lags associated with relative rise in sea level. In contrast, Bridge and Willis (1994) explained the sharp bases of some of the sandstone bodies as due to rapid progradation of storm-wave modified channel-mouth bars that formed rapidly as a result of channel diversion. A beach face origin of the sandstone bodies was considered unlikely due to absence of: (1) characteristic seaward dipping planar laminae; (2) alongshore directed cross strata interbedded with landward directed cross strata and planar strata typical of ridge and runnel systems; (3) eolian cross strata. Although beach deposits can be eroded during marine transgressions, and by tidal inlets, universal removal seems unlikely. The apparent absence of beach deposits in both transgressive and regressive sequences along a shoreline that clearly experienced strong storm waves is probably related to high deposition rates of sand and mud near the mouths of a complex of channels. Thus channel switching is a viable mechanism for producing the meter-scale alternations of channel-related sandstone bodies and mudstones. Upward-fining sequences representing filling of coastal bays and/or progradation of tidal flats could be formed in areas away from active channels, perhaps landward of abandoned channel-mouth bars (shoals) that were being reworked by wave currents.

Estuaries and sand barriers with muddy bays on their landward sides are expected in coastal areas during relative sea level rises (references in Bridge and Willis, 1994). Although there is evidence for washovers from sandy coastal shoals, there is no clear evidence for barrier beaches. Furthermore, no evidence was found for ravinement during regressions and estuarine filling during transgressions. The lack of evidence for muddy intertidal channel deposits with large-scale inclined heterolithic bedding, and for tidal-bundles in cross-stratified sands, indicates that the North Sea tidal flats may not be good analogues for these deposits (contra McCave, 1968; Johnson and Friedman, 1969). The strongly asymmetrical, ebb-dominated currents possibly record the dominance of river flow during deposition.

Other Givetian and Frasnian coastal and marine deposits in New York and Pennsylvania are broadly similar to those exposed in the Schoharie Valley, and interpreted depositional environments have several features in common: (1) sandy, tide-influenced channels; (2) shallow bays and tidal flats where mud and sand were deposited; (3) rarity of beaches; (4) storm-wave

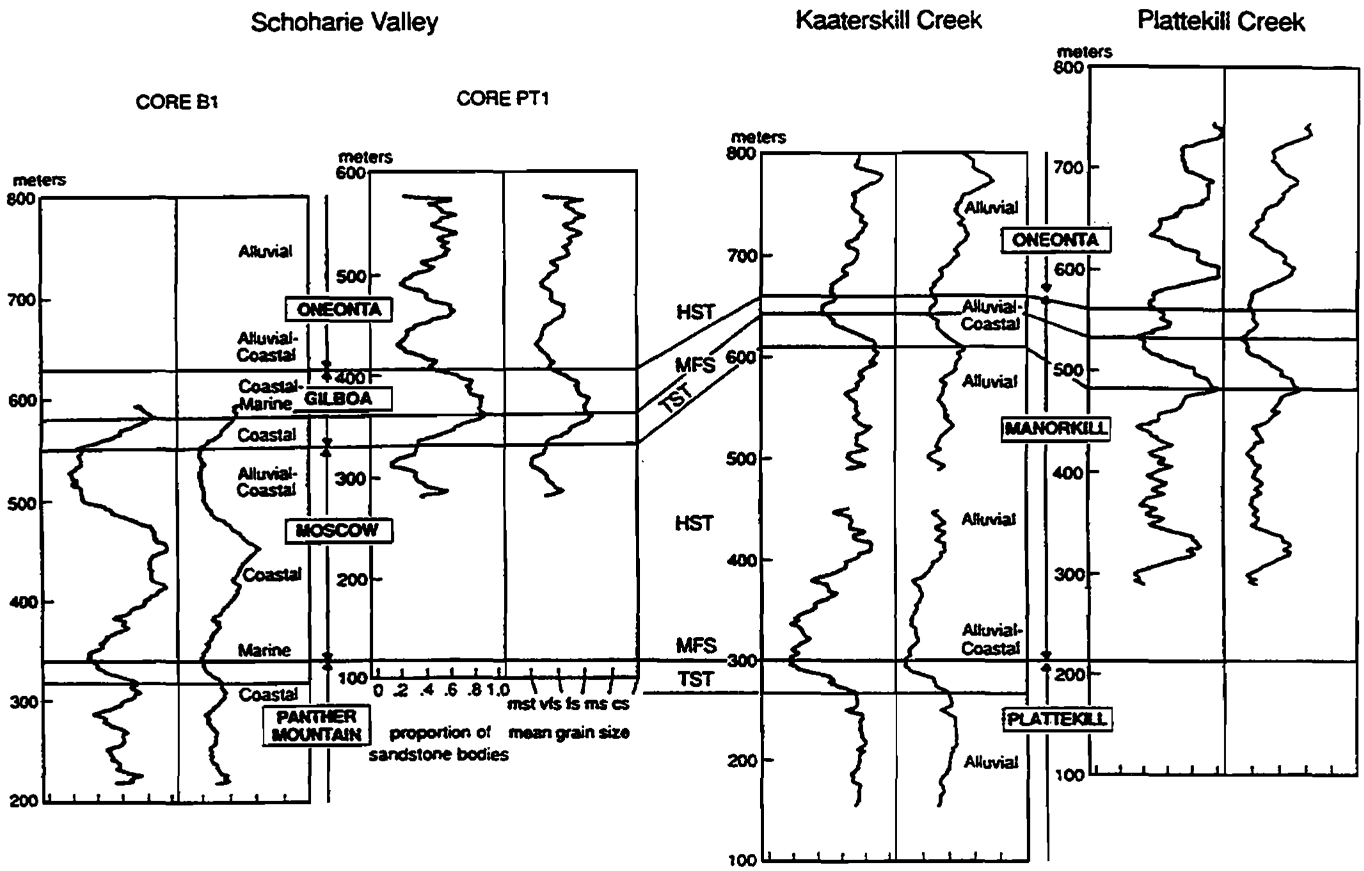

Figure 8. Correlation of Schoharie Valley Formations with those in Kaaterskill and Plattekill Creeks (from Bridge and Willis, 1994). TST = Transgressive Systems Tract. MFS = Maximum Flooding Surface. HST = Highstand Systems Tract.

domination of the marine shelf (references in Bridge and Willis, 1994). Much of the variability in the deposits from different areas could be explained within the context of a wave- and tide-influenced deltaic coastline. Meter-thick sequences similar to those in the Schoharie Valley but in the Frasnian part of the Catskill clastic wedge have been interpreted to reflect eustatic sea-level changes associated with Milankovich climatic cycles (Van Tassell, 1987). Such an interpretation is considered unlikely here. Such an allocyclic control could only be justified if sequences could be traced laterally for distances greater than would be expected for individual delta lobes, and indeed across the whole basin. However, correlation of sequences for more than a few hundred meters has proved difficult in these types of rocks (Miller and Woodrow, 1991; Duke and Prave, 1991).

Discussion of 100-meter scale sequences in coastal and fluvial deposits of the Catskill clastic wedge

The Gilboa and Oneonta Formations in the Schoharie Valley can be correlated biostratigraphically with fully marine strata to the west and fluvial strata to the east (Figures 3 and 8). The correlations are approximate because the resolution of biostratigraphic zones is on the order of a million years at best. McCave (1969) associated mudstone-dominated fluvial sequences with marine transgressions based on the assumption that rising sea level would lead to increased rate of floodplain aggradation and higher preservation potential of overbank mudstones. This concept has been used by others more recently (e.g., Shanley and McCabe, 1993; Wright and Marriott, 1993). However, floodplain aggradation rate is only one of several controls on proportion of channel-belt deposits in alluvial successions (Mackey and Bridge, 1995). Within the 100-meter scale sequences represented by the Manorkill and lower Oneonta Formations, the mean grain size and proportion of channel-belt sandstones, and the size of individual channels, increase upward (Willis and Bridge, 1988). These trends were interpreted by Willis and Bridge (1988) to record progradation of seaward-fining distributive fluvial systems. Dating of these strata is not accurate enough to determine if the upward increase in channel-sandstone proportion is related to decreasing accomodation space and sediment accumulation rate in the way suggested by Shanley and McCabe (1993). However, by analogy with the Miocene Siwalik strata in the Himalayan foredeep, high proportions of channel-belt sandstone bodies are associated with high subsidence and deposition rates, increasing channel sizes, and progradation of megafans (Willis, 1993; Khan et al., 1997; Zaleha, 1997).

McCave (1969, 1973) and Johnson and Friedman (1969) postulated that thin marine limestones (e.g.,Tully) were deposited during marine transgressions, because sea-level rise would cause trapping of sand and mud near the coast, starving deeper marine areas of sediment. In contrast, Brett and Baird (1985, 1986, 1990) interpreted such limestones to represent shallowing and sediment starvation-winnowing events following sea-level lowstands. According to Brett and Baird (1990), the upper Moscow Formation is represented in marine areas by a Type 2 sequence boundary that is overlain by the lower Tully Limestone (Figure 3). No corresponding erosion surface has been recognized in coeval coastal deposits. The upper Tully Limestone apparently rests on a transgressive erosion surface (Brett and Baird, 1990), and may therefore correlate with the transgressive lower Gilboa Formation in Schoharie Valley (Figure 3). However, relative sea-level changes are not necessary congruent in different parts of foreland basins, such that cross-basin correlation of regressive or transgressive strata does not necessarily imply coeval events.

Many authors favor eustatic sea-level changes as the main control on the large-scale sequences (references in Bridge and Willis, 1994). The transgressive deposits of the Gilboa Formation are associated with a widely accepted eustatic sea-level rise that can be recognized worldwide (House, 1983; Johnson et al., 1985). However, eustatic sea-level changes must be moderated by tectonically induced changes in subsidence and uplift rates and sediment supply. To establish whether eustasy or tectonism is the dominant control on sequence development in foreland basins, it is necessary to establish the ages of sequences, and the relative thickness of transgressive and regressive deposits, in different areas of the basin (Jordan and Flemings, 1991). It is very difficult to know if rising and falling sea level are coeval or otherwise in different parts of the marine basin, because of dating limitations. Also, it is very difficult to predict the timing and rate of deposition in different parts of the basin in relation to tectonic uplift of the adjacent mountains and development of the peripheral bulge. Jordan and Fleming's (1991) model suggests that the thicker regressive strata relative to transgressive strata in this region points to a tectonic control of some of the large-scale sequences.

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Appendix: Road log for Schoharie Valley field trip

Travel from Binghamton to Oneonta on I88. Take exit 15 in Oneonta and head east on NY23 towards Grand Gorge. Approximately 3 miles past Grand Gorge, turn left down a gravel road towards Hardenburgh Falls. Cross the bridge over Bear Kill and park. Walk east down track on north side of the creek to exposures adjacent to the bridge. **Stop 1: Hardenburgh Falls.**

Return to NY23 and turn left (east) towards Prattsville. Travel approximately 1.5 miles to the bridge that crosses Schoharie Creek. Turn sharply left (north) immediately after crossing the bridge and proceed along the road that goes along the eastern side of Schoharie Reservoir. Travel approximately 6 miles as far as Gilboa-West Conesville Central School. Turn right immediately past the school, travel to the back of the school and park near the playing fields. Walk southeast to the far corner of the playing fields and find the trail leading up the hillside to the quarry (about 10 to 15 minutes walk). **Stop 2: Stevens Mountain Quarry.**

Return to the main road by the school, turn left and travel south for approximately one mile to the bridge over Manorkill. Park on the south side of the bridge and find the trail leading down to the reservoir. **Stop 3: Manorkill Falls.**

Travel north approximately 1.5 miles (past the school again) and cross Schoharie Creek at Gilboa. The display of tree trunk casts is on the left (south) side of the road. **Stop 4: Gilboa.**

Continue on this road for about 1 mile until it meets NY30. Turn left (south) on NY30 and travel approximately 1 mile until the road starts to climb up the west side of Pine Mountain. Park at the base of the hill near the first outcrops. **Stop 5: Grand Gorge, Route 30.**