Rock-Block Slide on Bare Mountain, Southern Onondaga County, New York

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Introduction

Special Type of Landslide?

A two-mile long series of contour-parallel notches mark slivers of Devonian bedrock that are separated with 30 to 100 ft of relative-vertical displacement from their *in situ* stratigraphic equivalents high on the east side of Bare Mountain, in southern Onondaga County, NY (Fig. 1); yet downslope exposures show no chaotic or disruptive features, nor do geomorphic forms hint of a landslide foot or toe (Fakundiny, 1997.) Either the displacements took place on a series of (1) closely spaced, shallow-dipping, listric faults that have their bottoms in deep bedrock or (2) slide planes that separate long, thin, mostly intact, sliver-like blocks with toes buried in valley fill. No glacial deposits or ice-scour features have been found in the back-head notch, on top of or on the valley-side slope of the blocks. Lack of glacial deposits high on the mountain side suggests that block movement occurred during the last stages of retreat of the Laurentian ice sheet or later. If hypothesis 1 is correct, listric faults in the Finger Lakes District of central New York may indicate latent seismic hazard, whereas, hypothesis 2 may indicate possible landslide hazards at the base of Bare Mountain. With only its top exposed the feature lying on the side of Bare Mountain is difficult to characterize. Structural interpretation of faults is usually made from limited exposures at their tops and geophysical or drilling data from their buried parts. Landslides, conversely, are usually totally exposed. The feature on Bare Mountain, however, is mostly buried by colluvial, alluvial and postglacial lacustrine deposits, thus, obscuring the geomorphology of its lower parts. We are not aware of any descriptions in the literature of a rock-block slide with a buried foot and toe, although Jane Gilotti and Joseph Hull (oral communication, 1997) have observed one in Greenland. Our studies to date favor rock-block sliding over Holocene faulting. Thus, the landslide hypothesis will bias the following discussion.

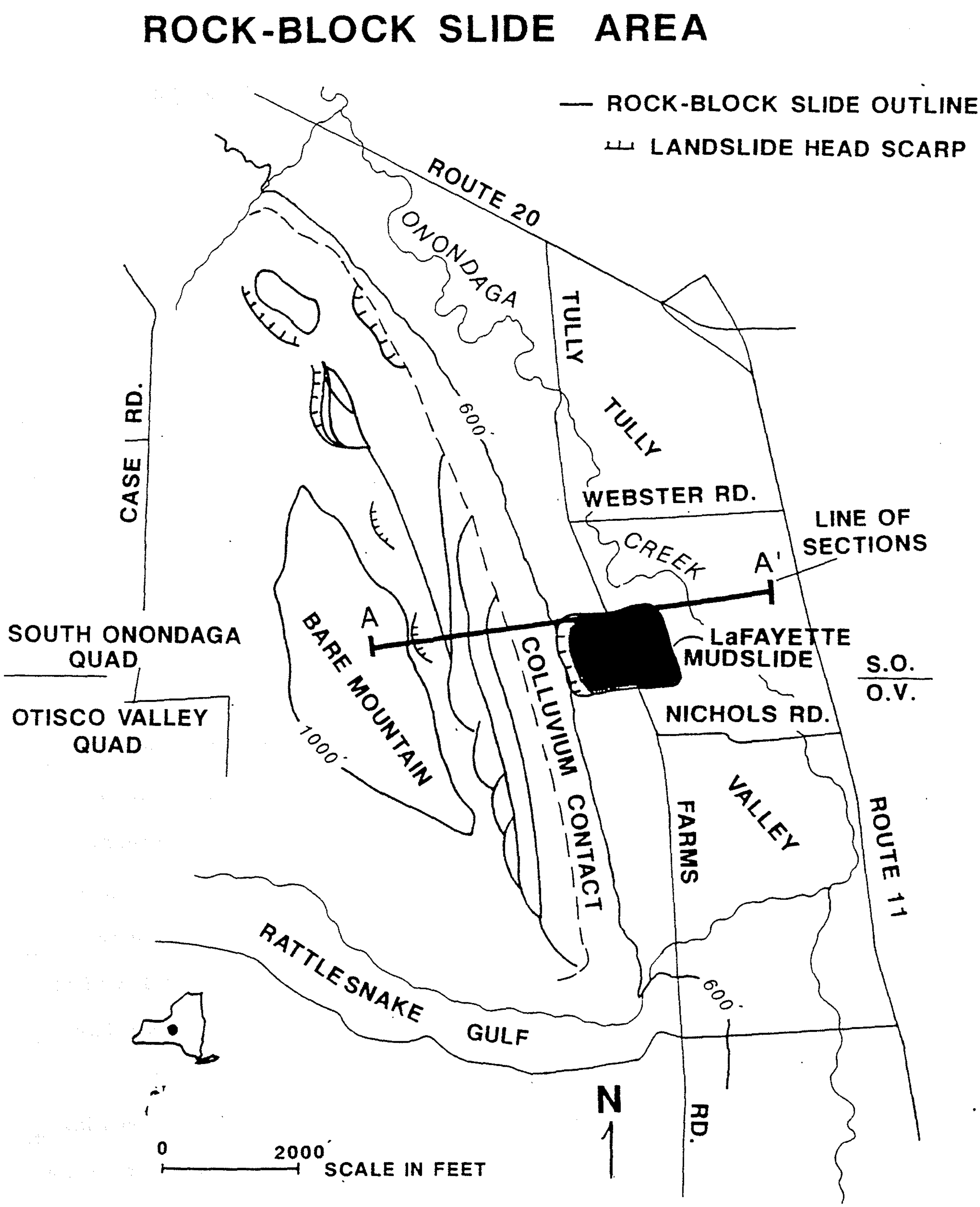

Figure 1. Location map of Bare Mountain rock-block slide in southern Onondaga County, NY. Section A-A[1] is location of Figures 4 and 5. S.O./O.V. is the South Onondaga-Otisco Valley quadrangle boundary.

Purpose and Scope

The purpose of the field trip and this essay is to show and describe what might be the first-recognized, large, rock-block landslide with a foot and toe that are presumably buried by proglacial lacustrine and postglacial alluvial and colluvial sediments. We say this on the bases of the limited literatuare search that we have been able to make. The relations between the geomorphology of the hillside where the landslide is parked and local glacial features places its time of movement from about 12000 to 15000 B.P.

The scope of this investigation was confined to only aerial photograph interpretation and cursory geologic mapping of marker beds within the Devonian bedrock sequence of central New York, as defined by Brett and others (1995) and Ver Straeten (1995.) Other types of studies could add information, such as: (1) drilling, which may soon be done in the search for potable groundwater; (2) refraction seismic profiling, which is not currently acceptable to the landowners; and (3) coring sediment-filled sags for pollin, which has yet to be undertaken. These types of studies await new resources for their undertaking.

Location

The east side of Bare Mountain faces the Tully Valley approximately 5.5 miles south of Syracuse, NY, in southern Onondaga County, central New York (Fig. 1) and lies across the border between the South Onondaga 7.5-minute topographic quadrangle to the north and the Otisco Valley 7.5-minute topographic quadrangle to the south. Bare Mountain is approximately 3 miles in length along a N25°W axis and 1.25 miles wide at its middle. The mountain rises approximately 1000 ft above the Tully Valley floor on the east and 280' above the saddle between it and Manus Hill on the west. Bare Mountain is bounded on the north by the Valley of the West Branch of Onondaga Creek and on the south by Rattlesnake Gulf, which carries an east-flowing tributary to the East Branch of Onondaga Creek in the Tully Valley. The political boundary between the towns of Otisco, to the west, and LaFayette, to the east, transects the mountain from north to south. The Village of Cardiff, which lies at the intersection of US Route 20 and State Route 11A, lies at the north end of Tully Valley across from the north end of Bare Mountain. All land is private, and access is restricted. Trespassing requires permission of the owners.

Problem

The features on the eastside of Bare Mountain have all of the characteristics of a rock-block slide (Cruden and Varnes, 1996), except that bedrock bedding does not dip into the valley. The problem with a landslide interpretation is that the block slivers are long (up to 2 miles), thin (~250 feet thick), intact, relatively unrotated, and have no exposed foot or toe--characteristics that we have not found described in the literature. Is this a new type of rock-block slide?

Bare Mountain has a plateau top with a cliff that stands 900 ft above the Tully Valley floor. The valley floor itself is the top of a deep fill of Pleistocene and Holocene till and lacustrine units interfingered with sidewall alluvial fans and colluvium (Kappel and others, 1996; Pair, 1995, and in prep.) The intact nature of these long, thin slivers along their length is unusual for rock-block slides. Few rock-block slides have been described in the literature, as shown by the lack of references to them in Turner and Schuster (1996.) Lack of disintegration suggests that movement was steady and slow and that blocks were gently eased into their current positions without severe disruption (Fig. 2.) Advancing glacial ice may have eroded and oversteepened the preglacially sculptured bedrock wall at the mountain's base, thus destabilizing the slope above. Unlike rock-block slides that have been described from other parts of the northern Appalachian Basin (Schultz and Southworth, 1989) that have slipped along valley-dipping weak beds, the Bare Mountain slide most likely moved along a valley-dipping joint system. Slip may have been gradual, possibly the blocks accommodated by being let down against wasting ice in the valley. Concomitant infilling of lacustrine deposits around the ice and over any remaining parts of the foot of the slide could have buttressed these blocks and secured them to the side of the mountain in a metastable or dormant state. Mudslides at the foot of the valley wall, such as the LaFayette mudslide (Fickies, 1993; Fickies and Brabb, 1989; Negussey and others, this volume) (Fig. 1), could relieve that restraining load and reinitiate downslope movement of the rock-block slide. If these features are, indeed, a rock-block slide, they form (volumetrically) one of the largest landslides recognized in New York State (>$2x10^8yd^3$), possibly in the entire northern Appalachian Basin (see Schultz and Southworth, 1989.) The morphology of the slip surfaces and their role in collecting and distributing upland surfacewater and groundwater may help interpret the artesian hydrologic systems of Tully Valley, in which mudboils abound (Kappel and others, 1996; Kappel, this volume).

Fault?

Other workers in the area have suggested the possibility that the block movement may have been along a tectonic fault or large gravity-induced fault system (William Kappel, oral communication, 1996.) Although faults have been mapped locally (References reported in Kappel and others, 1996), factors weigh against this interpretation, including: (1) no extensions have been mapped along strike in either direction from the exposed breaks; (2) interpretation of geomorphic features (discussed below) strongly suggests that the timing of fault movement would have to be latest Wisconsinan;

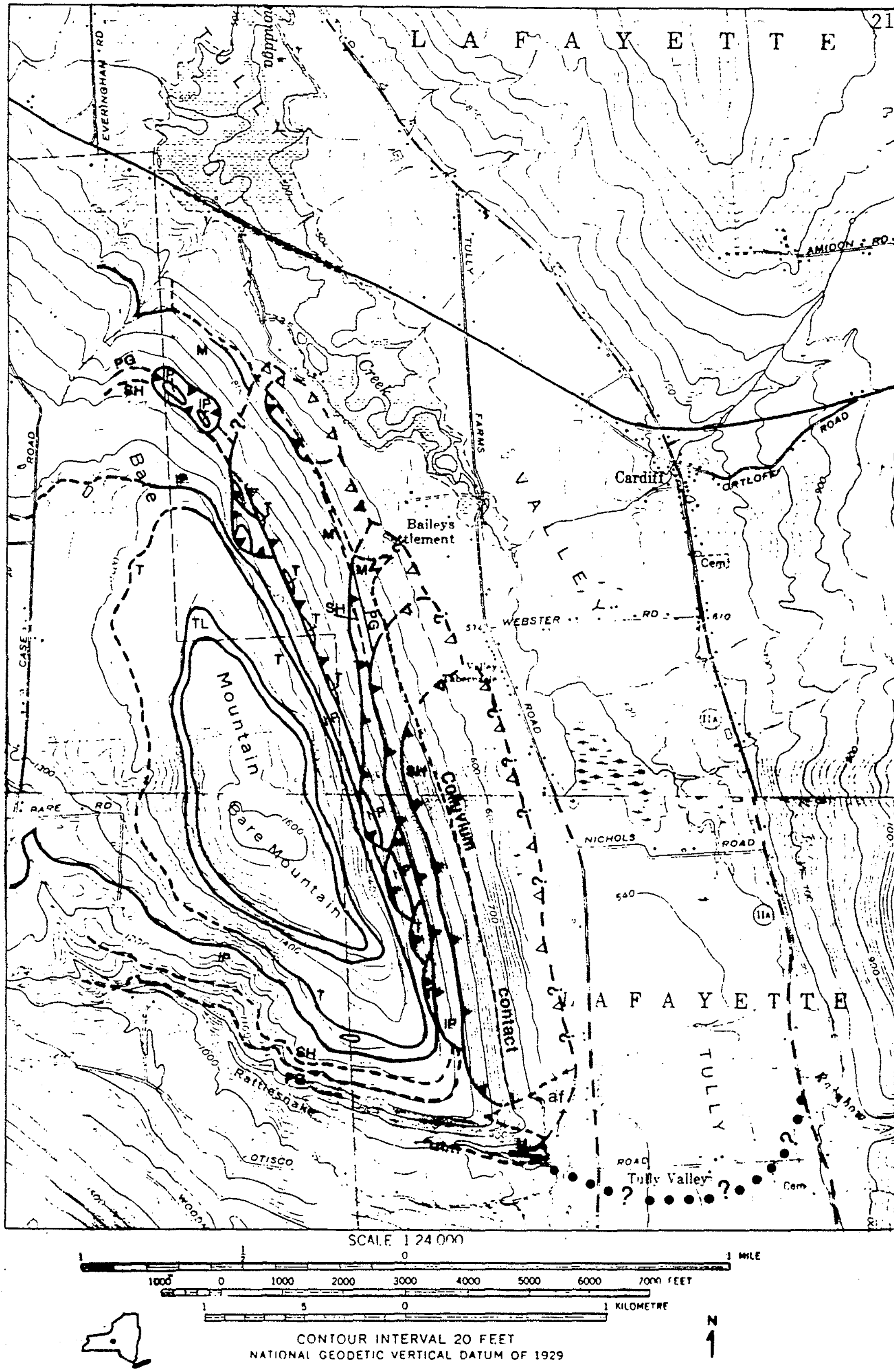

Figure 2. Geologic map of Bare Mountain showing landslide blocks outlines, colluvium contact, and several bedrock marker beds. TL=Tully Limestone; T=Tichenor Beds; IP=Top of Ivy Point Member; SH=Staghorn Coral Bed; PG=Peppermill Gulf Member; M=Mottville Member; at=diseccted alluvial fan. Solid lines with filled barbs are sliver ridge caps; dashed lines with open barbs are outlines of buried sliver toes. Dotted line with question marks is possible location of Onondaga Fm. subcrop below valley fill.

consequently, the movement took place in the modern stress field, which is compressive, with the maximum stress oriented ENE perpendicular to the strike of displacement planes and, thus, unable to promote either normal or strike-slip sheer movement; and (3) the map pattern (Fig. 2) is not typical of tectonically induced, regional fault systems in the Appalachian Basin (see Discussion below.) We, therefore, interpret the feature to be local and most likely the result of late Wisconsinan erosion and deposition upon failing bedrock strata, rather than tectonically induced faulting.

Bare Mountain structure

A block-rock slide interpretation for the Bare Mountain structure presents problems of interpretation also including: (1) its large size and intact shape, (2) the small amount of slip, (3) lack of an exposed foot and toe, and (4) timing. Note that this discussion will use the English system of measurements, consistent with the 7.5-minute topographic quadrangle scale and contour interval. The size and shape, >10,000 ft long, ≥250 ft thick, and ≥2500 ft along the 25° slope from head to toe makes the Bare Mountain feature larger than the rock-block slides in the Appalachian Valley and Ridge province of eastern North America (Schultz and Southworth, 1989). For terms used in this paper please see Appendix.

The vertical-slip component of movement, as measured by displaced, bedrock marker beds, ranges from 35 to 100 ft. The horizontal strike-slip component is not known, but is probably negligible. The unusual feature of the slide is the intact condition of the block components along strike (Fig. 2.) We would expect that $2x10^9 yd^3$ of rock crashing down a 25° slope for 100 ft would disaggregate and slump into a jumble of disoriented, rotated, jumbled blocks with no continuous mappable stratigraphy. At least 4 mappable slivers of the slide have internally coherent, traceable marker beds, that have little to no rotation of the bedding, except at the head ridges where small (<200 yd^3) rotational rock slumps locally occur. These commonly dip back into the head scar or gently to the south or southwest.

The foot and toe of the slide presumably is buried by proglacial, lacustrine and postglacial fluvial and alluvial units that were deposited in a lake in front of the receding ice lobe or on the postglacial valley floor (Kappel and others, 1996.) Burial and stabilization of the slide foot and toe took place before or during dewatering of the proglacial lake, presumably as the receding ice tongue uncovered lateral, melt-water channels across valley walls to the north (Hand, 1978; Gomes and Pair, 1997.) Timing of the movement can be roughly determined by geomorphology and relation to proglacial lacustrine and postglacial alluvial and colluvial sediments. Several colleague glacial geologists have visited the head of the slide and have expressed their opinions that none of the geomorphic features were produced by overriding ice or ice-contact, fluvial erosion that may have been associated with the Laurention ice retreat (Donald

Cadwell, G. Gordon Connally, Barbara Hill, George Kelley, Donald Pair, oral communication, 1996.) The interpretation that the morphology of these slivers at their heads is formed by post-glacial or pariglacial surface-water erosion, frost heave, and precipitation is consistent with the interpretation of Pair (in prep.) Thus, the most logical time for movement would be during ice retreat from the walls of the valley to allow for slumping, and at the time that high standing proglacial meltwater lakes drained along cross-channel fluvial systems (Hand, 1978, Jager and Wieczorek, 1994.) The latest stage of the Wisconsian would provide a unique time when: (1) some ice may have remained and foundered to buttress the slide and ease the slivers down the valley wall; (2) a proglacial lake still existed to form the sink for lacustrine deposits to cover the foot of the slide; and (3) pore-water pressure in the bedrock was still high from the pre-existing high-standing proglacial lake--all circumstance that could have aided the initiation of slide movement and eased the slivers into their current position. The age of movement would, therefore, be at the time of recession of the ice sheet from the Valley Heads moraine and when a slight readvance placed the ice tongue adjacent to the side of Bare Mountain (Gomes and Pair, 1997.) A dissected, hanging, alluvial fan that sits high on the southernmost part of the east face of Bare Mountain covers the southern end of the rock-block slivers and may help to determine the age of last movement (Fig. 2.)

Bedrock Geology

Mapping

The bedrock geologic map for the southern part of the South Onondaga and northern part of the Otisco Valley 7.5-minute topographic quadrangles was recently made by Brett and others (1996) and Ver Straeten and others (in prep) as part of the New York State Geological Survey-U.S. Geological Survey cooperative STATEMAP Program. Brett and colleagues are now mapping the Marcellus Quadrangle to the west and the Jamesville Quadrangle to the east.

Stratigraphy

Detailed stratigraphy for both quadrangles near Bare Mountain (Fig. 3) was determined by Carl Brett, Gordon Baird, and Charles Ver Straeten (Brett and others, 1996; Ver Straeten, 1995.) The bedrock section that may be incorporated in the slide possibly extends from below the Helderberg and Tristates Groups, up through the Onondaga Formation, and the Hamilton Group. The Tully Formation limestones and the section above may have been involved, but are now eroded from the top of the block sliver. Within Bare Mountain the bottom of the exposed section is in the Cardiff Member of the Oatka Creek Formation (Marcellus Formation of previous authors), and the top exposes Geneseo

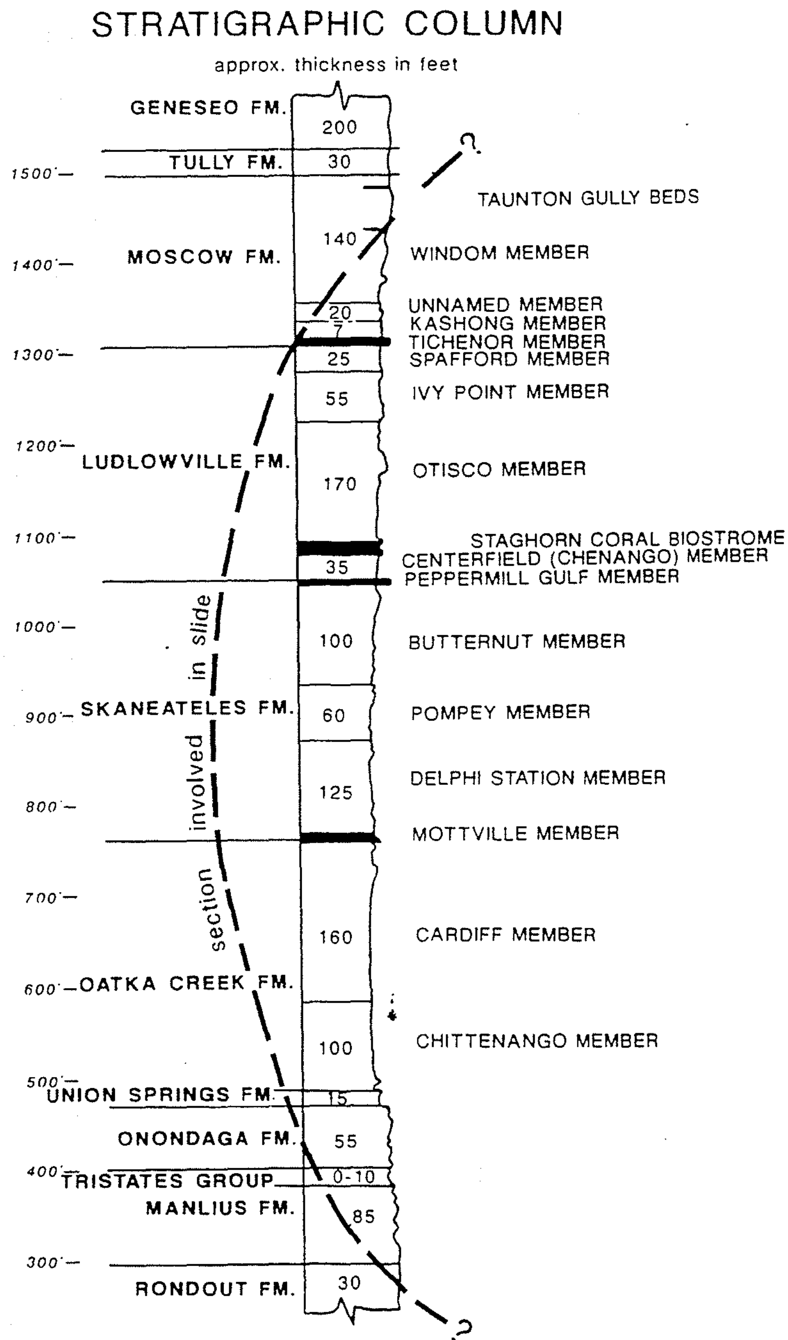

Figure 3. Stratigraphic column of Upper Silurian and Lower and Middle Devonian bedrock in the South Onondaga and northern Otisco Valley 7.5-minute quadrangles, showing formations, members, and marker beds. Also shown is section that was breached by the rock-block slide. Elevations are on Bare Mountain. Section from the top of the Manlius Formation downward to the top of the Rondout Fm. is the Helderberg Group. Section from the bottom of the Oatka Creek Formation to the top of the Moscow Formation is the Hamilton Group. Geneseo Formation is the bottom of the Genesee Group.

Formation shale. Of note are the marker beds that were used to map the slide, which include: Union Springs (in drill hole), Mottville and Peppermill Gulf bed of the Centerfield Member, the Staghorn coral biostrome, the tops of two siltstone tongues of the Ivy Point Member of the Ludlowville Formation, and the Tichenor Member of the Moscow Formation.

Regional Folds and Faults

Within the South Onondaga and Otisco Valley quadrangles the bedrock generally dips <5° to the southwest with generally a N70W (160°) strike. Local, open WNW-tending folds, NW-to N-trending monoclines, and a rare E-W-trending rupture of rock at Nedrow (ramp thrust from a deep decollement?) prevail. A north-trending monocline or fault lies west of Bare Mountain and lowers the bottom of the Tully Formation from an elevation of 1560 ft on Manus Hill to 1520 ft on Bare Mountain, against the grain of the regional dip. Another buried monocline (or fault) trending ≈N35W (325°) lies under Tully Valley and lowers the base of the Ludlowville from 1100' on the north end of Bare Mountain, with a projected elevation of about 1160' down to 1100' 1.5 miles up dip on the eastern wall of Tully Valley.

Joints

Joints control the form and attitude of the intact headwall-line scarp of eastern Bare Mountain above and behind the rock-block slide. In the scarp below the Tully Formation, joints can only be recorded from the Taunton Gully Beds (upper Windom Member) and the Tichenor Member of the Moscow Formation. In the slide blocks, joints can be measured in any exposed unit from the Ivy Point Member of the Ludlowville Formation downward. Not enough rock is exposed to allow for a statistical study of joints on the hillside, but studies elsewhere in central New York (Engelder, 1985; and Engelder and others, 1997, and references therein) show that most vertical joint sets that have been mapped in central New York State are part of a systematic regional manifestation, or genetically related to those regional systems. The vertical joints were presumably generated by regional, fluid-hydraulic pressure that resulted from tectonic loading, or from local overpressured hydraulic systems. Dipping joints are different from the vertical sets by being systematically related to local topography and are interpreted as late (post-Pleistocene?) unloading joints (Engelder, oral communication, 1997.) These "unloading" joints mostly dip toward the valley at angles less than or equal to the valley-wall slope. We propose that sets of the valley-ward dipping joints, like the ones exposed high on the east side of Bare Mountain, control the orientation and geometry of the slip surfaces beneath the slide blocks. At several headwall-line exposures a few conjugate pairs to valley-dipping joint sets dip into the hillside at comparable angles.

Vertical joints recorded from the headwall are: N-S (000/90), N25E (025/90), N60E (060/90), N80E (080/90), N30W (120/90), and N10W (170/90). Dipping joints in the headwall-line scarp are: N30E35SE (030/35), N30W55SW (150/55), N55W40NE (205/40), N30W75NE (330/75), N30W45NE (330/45), and N20W45NE (340/45). Near-vertical joints in the slide blocks are: N30E (030/90), N65E (065/90), N60W86SW (120/86), N30W (150/90), N25W (155/90.) The only dipping joint recorded in a slide block was N70W65SW (110/65.) Just south of Bare Mountain in the Unadilla 15-minute quadrangle, Parker (1942) recorded the N-S, N25E, N60E (Set III), and N60W sets and interprets them to be regionally distributed.

Groundwater

The Middle Devonian shale sequence between the limestones of the Onondaga and Tully formations generally has low permeability and low groundwater yields. Water flows mostly along joints and disaggregated bedding. The Tully Formation has high yields (Fickies in Brett and others, 1995.) The Geneseo Formation, capping Bare Mountain, has geomorphic features that are reminiscent of land surfaces over internal drainage; that is, on aerial photographs the plateau at the mountain top shows only hummocky ground without any recognizable stream courses. The Onondaga Formation and carbonate units immediately below have the highest yields of groundwater, and confine the flow to joints and solution cavities. The groundwater flow and quality in the Salina Group shales and evaporates are influenced by the dissolution of salt units and the perforation of units by brine production wells. (See Kappel, this volume and Kappel and others, 1996.)

Engineering Geology

Bedrock engineering characteristics are given by Robert Fickies (in Brett and others, 1996.) Here Fickies noted the poor to fair cut-slope stability of all of the shale sequences in the strata on Bare Mountain. These are all abundantly jointed. Fickies also notes the fair to good cut-slope stability of the Tully and Onondaga formations and underlying limestones and dolostones of the Manlius and Rondout Formations. These characteristics allow for failure in the shales, especially where interbedded strong carbonate units are breached and undercut.

Surficial Geology

Mapping

The surficial geology of both the South Onondaga and Otisco Valley Quadrangles was mapped most recently by Pair (1995, in prep., and this volume) as part of the New York State Geological Survey-U.S. Geological Survey cooperative

STATEMAP Program. Pair mapped these and the Tully quadrangles in 1995-1997 and is now continuing in the Marcellus Quadrangle to the west of the South Onondaga Quadrangle and the Jamesville Quadrangle to the east.

Stratigraphy

Pair (in prep., and this volume) has mapped seven Pleistocene units for the two quadrangles: glacial till; ice-stratified drift; fluvial silt, sand, and gravel; glaciolacustrine silt and sand; alluvium overlying lacustrine silt and clay; ice-contact sand and gravel and glacio-lacustrine silt and clay; and 5 Holocene units: alluvial fans; alluvium; colluvium; organic deposits; and artificial fill. The Pleistocene glaciolacustrine silt and clay; their overlying alluvium, and Holocene colluvium and alluvial fans are of interest to the history of sliding. A cross section of valley fill and bedrock (Kappel, 1996), shows the interfingering nature of alluvial and deltaic glaciolacustrine deposits along a line extending east from Rattlesnake Gulf. One coincidence that should be noted is that, in the Tully Valley, colluvium is only mapped at the lower parts of the rock-block slide. Colluvium, alluvium, and possibly glaciolacustrine silts and clays, bury the foot and toe of the rock-block slide. Also note that the alluvial fan at the south end of Bare Mountain (Fig. 2), which is dissected at Rattlesnake Gulf, overlies the southernmost extent of the rock-block slide.

Groundwater

Groundwater flow has greatly influenced the production of geologic hazards within the surficial deposits of the Tully Valley (see Kappel, this volume, and Negussey, this volume.) The following is summarized from Robert Fickies (in Pair, 1995.) Under relatively common circumstances, coarse-grained, non-cohesive, stratified sand and gravel deposits have groundwater yields that very widely from fair to very good and have widely varying quality. Fine-grained, cohesive, stratified clays and silts have very poor to poor groundwater yields and may be high in dissolved salts. Till deposits have low groundwater yields, some of which fluctuate seasonally.

Engineering Geology

The engineering characteristics of surficial deposits are described by Fickies (in Pair, 1995) and are subdivided into upland deposits and valley-fill deposits. The relevant units to consider at the rock-block slide are upland deposits and the glaciolacustrine sediments of the valley-fill deposits catagory.

Till covers part of the plateau on top of Bare Mountain and, contrary to what Fickies presents as general characteristics, locally accepts precipitation rather than developing a prominent overland fluvial runoff system. Depth to bedrock varies, but is shallow. Colluvium covers the lower, eastern flank of Bare Mountain with an upper contact against rock and rock-block soil at the 800-900 ft contour interval. The lower part of the colluvial apron overlies alluvium and glaciolacustrine silts and clays down to about the 600 foot contour. Precipitation that falls on the bedrock and soil above the contact sinks directly into the slide blocks and joins the groundwater system. The resulting small drainage area developed on the colluvium does not provide enough overland flow to erode deep gullies, but rather the small amount of runoff has produced open, downslope swales.

Glaciolacustrine sediments lying above the projected, buried toe of the rock-block slide include silt, sands, and clays. Fickies (in Pair, 1995) reported that the silts and clays have low plasticity, may be varved, and may be poorly drained. The water table may be at or near the surface during and after rains (Dawit Negussey, oral communication, 1997.) These deposits may have engineering properties at or above the liquid limit below the water table and can behave as a viscous liquid. Slopes on these deposits may be marginally unstable and susceptible to landslides where greater than 10°. We believe that saline groundwater may have exacerbated the condition by reducing the attractive forces between clay particles.

Geomorphology

Topographic and Geologic Traverse along Cross Section

The topographic map (Fig. 2) does not have the precision to show the back-head depression (notches) and composite, head-ridge morphology. Although 20-foot contour intervals should reveal the notches and ridges, the aerial photographs that were used to make the topographic map must have been taken during summer when full foliage obscured the detailed morphology. Before the rock-block slide can be adequately described a precise and accurate topographic map, preferably at a 5' contour interval, must be constructed. In the meantime, the map presented in Figure 2 will have to suffice. A west-to-east traverse from the top of Bare Mountain to the break in slope at the valley bottom across the midpoint of Bare Mountain illustrates the complexity of the slide and the difficulty in determining the mechanistic cause and timing (see Fig. 4.) The traverse starts at the top of Bare Mountain (elevation >1600') in Geneseo Formation shales that are vegetated with grasing crops and sparse woods. A small cliff of exposed Tully Formation (limestones) rings the top of the Mountain at about 1580-1540'. Below the cliff is a bench or secondary plateau cut on the top of the Taunton Gully Beds of the Windom Member of the Moscow Formation. The Taunton Gully Beds are jointed and form a vertical cliff, in places more than 100' high, with talus below. This is not a head scarp, but a post-sliding, back-wasted cliff that

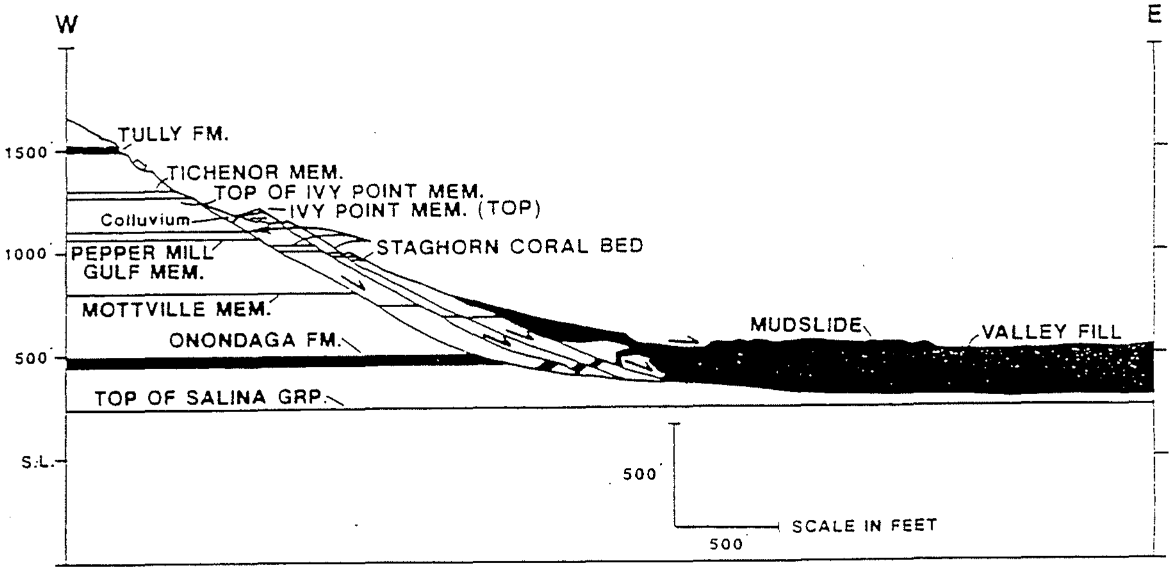

Figure 4. Cross section showing preferred interpretation of structure on the east side of Bare Mountain.

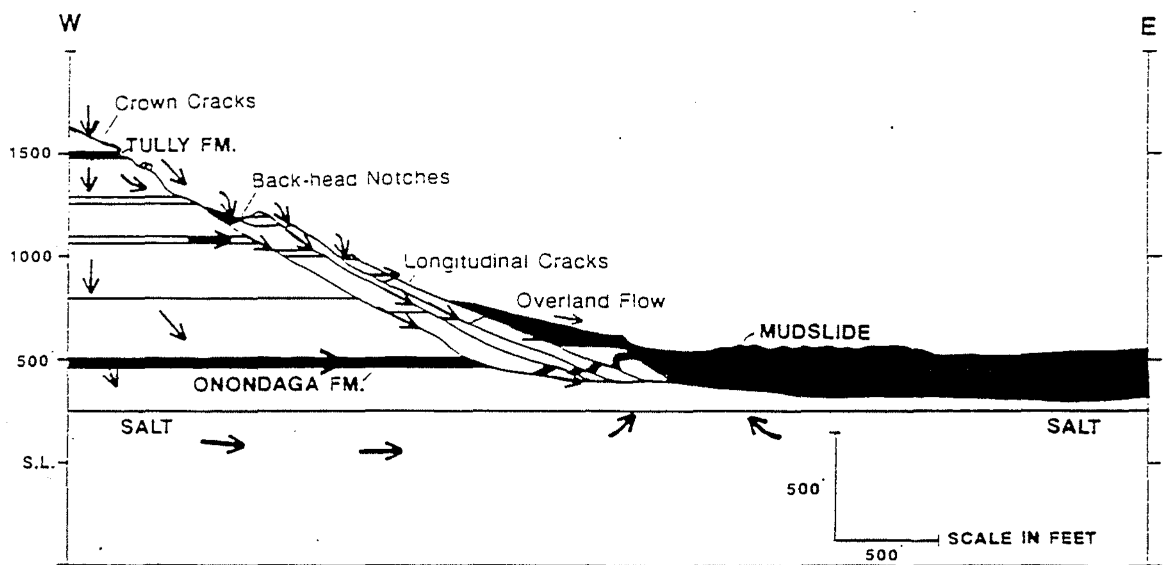

Figure 5. Cross section showing possible groundwater migration routes.

has retreated from the head-scarp position and is controlled by prominent vertical and valley-dipping joints. We are informally calling this a headwall-line scarp in the fashion of a "fault-line scarp" that is used for back-eroded fault scarps. Small, modern, rotational rock slumps and rock slides in the cliff and talus reveal other marker beds within the Moscow Formation, such is the basal Tichenor Beds, in sparsely distributed localities at about 1300'. Talus from the shale of the Windom Member (Moscow Formation) fills much of the linear depression behind the first head ridge. To the south, large blocks of Tully Formation (limestone) have toppled or have been let down without much rotation onto talus from the Moscow Formation. The in-filling talus along the back-head notch has internal drainage and local sag holes, some of which are plugged with clay. One such sag south of the traverse has a small pond. This filled sag would be a favorable locality for pollin studies. Talus covers the Otisco, Ivy Point, and Spafford members of the upper Ludlowville Formation and extends to the foot of the back slope of the head ridge of the uppermost block slivers. The head ridges stand 15 to 40 feet above the back-head linear depression (notch) and are themselves covered by talus below their tops. The top of the head ridge is held up by either the Tichenor Member of the Moscow Formation or the top of the upper cycle of the Ivy Point Member of the Ludlowville Formation. Thus, the head ridge can have a cap that varies 30' in stratigraphic position, consistent with the variation in elevation of the cap. If one leaves the traverse at this point and ventures north along the top of the head ridge, one can travel .6 mi and not leave this 30 ft interval of the stratigraphic section nor vary in elevation more than about 50 ft. Returning to the traverse, locally the head ridge is segmented and small blocks from the cap of the ridge have sloughed off either outward or inward such that the rotated blocks have large dips of bedding that trend outward toward the valley or inward toward the main scarp. Many head-ridge segments will have slight dips of bedding to the north or south, possibly reflecting the small rotation of the slide slivers along their length. On the head ridge and the valley-side slope, the vegetational stress during spring and fall is evident from the apparent poor health of the trees, probably a symptom of lack of water. Farther down the valley-side slope the other ridge crests have similar stressed vegetation. Species of trees appear to vary also in correspondence with the location of back-head notches, where surface water sinks. At least two other notches and ridges are encountered while traversing down the slope. The most prominent notch is between 900' and 1000' and is capped by the Staghorn Coral Biostrome of the Otisco Member of the Ludlowville Formation. Below the Staghorn Coral, easily recognizable markers that are exposed on the slope where colluvium has been eroded through include the limestones of the Peppermill Gulf Bed and the Mottville Member at the top and bottom, respectively, of the Skaneateles Formation. Generally below about 800' elevation the slope is covered by colluvium, and bedrock is obscured, except in a gully near

Bailey's Settlement where the Skaneateles and Oatka Creek formations are exposed and provide a stratigraphically lower opportunity to determine total movement of the slivers. Evidence of transverse ridges or cracks, radial cracks, or toes is not apparent on the valley-side slope from the top of the lowest slide sliver to the valley floor.

The unnamed gully north of Case Road at the north end of Bare Mountain (Fig. 2) has a relatively complete, in-place, exposed section, from the Cardiff Member of the Oatka Creek Formation up to the Joshua coral bed within the Otisco Member of the Ludlowville Formation. This section and exposures in Rattlesnake Gulf provide the intact, undisturbed stratigraphic sequence used for reference to the displaced stratigraphy in the block slivers.

Discussion

Problems of Timing and Mechanism

Bedrock on the east side of Bare Mountain has slipped into the Tully Valley with apparent vertical displacement from 35 to >100' along 4 or more slivers of relatively intact rock. Dipping and vertical joint sets have influenced the location of the slip planes and the present morphology of the slope. Yet the amount of mass that has been transported should have left a morphological expression of the displaced foot and toe. The only place for the zone of accumulation material to reside is within or under valley fill and colluvium, a situation that makes this rock-block slide unusual and questions the mechanisms, history of movement, and the current state of stability, especially for those slices immediately above the LaFayette mudslide where some restraining valley-fill cover of the foot and toe has been removed.

Timing

If we assume that most of the major motion of the rock-block slide occurred in a single episode and that a single mechanism was involved, then we can closely pinpoint the time of slip. Lack of evidence of glacial scouring or ice modification high on the slope or at the ridge caps of the slivers and lack of ice-contact glaciofluvial erosion in the back-head depressions places the timing of the event after the uncovering of the slope by the Laurentian ice sheet. At the valley floor, the foot of the slide is covered by glaciolucustrine sediments with no morphological evidence that the toe shoved or warped the lacustrine bedding. This would place the event before or during the dewatering of the proglacial lake. Thus, the ice sheet would have backed away from the slope, but not far enough to have uncovered the lowest up-ice, cross-ridge dewatering cross channel. Pair (oral communication) places this time at 12000-15000 b.p.

Mechanisms

Mechanisms for movement must consider the processes for destabilizing the lower slopes by either undercutting or increasing the slope's angle. Loading the top of the slope (by ice?) is possible, but improbable. Kappel (oral communication, 1996) indicated that the cross-valley subcrop of the Onondaga Formation below the valley fill is located somewhere near Rattlesnake Gulf (See Fig. 2.) Consequently, the Onondaga Formation has been removed by glacial ice or preglacial fluvial erosion at the base of the east side of Bare Mountain. The mechanism that is finally chosen should incorporate the removal of the Onondaga Formation bench, and must also explain how the long, thin, slivers of rock could have moved vertically downward over 100' without any substantial disruption, breakage, or block rotation.

Several mechanisms can be speculated for triggering the slide, once the slope is destablized. Among them are earthquakes and dewatering sequences (see Cruden and Varnes, 1996.) Speculation about earthquake occurrences 1200 years ago is beyone the scope of this paper, except that the following scenarios may have happened in conjunction with seismicity, either regionally distributed tectonic earthquakes or hydraulically induced, high, local, proglacial lake-induced, hydrostatic pressures.

One possible mechanism for slow block movement considers contact with stagnent, wasting ice. In this speculative scenario, huge blocks of ice become detached and remain in front of the retreating ice sheet while bracing the block slivers against the intact part of the mountainside. Gradual wasting reduces the confining weight on the toe and gently lets the slivers descend. Meanwhile proglacial lacustrine deposits build up and replace the ice as the confining and bracing mass. This hypothesis requires a delicate balance between the mass of the grounded ice and lake levels. It might be expected that the ice block would float off of the toe before the lake sediments could be deposited to a necessary confining depth.

Gomes and Pair (1997, and oral communication, 1997) have postulated a readvance of the ice lobe in Tully Valley on the basis of geomorphic features of the valley sides and a buried moraine (Kappel and others, 1996,) presumably with a proglacial lake waxing and waning in front of it. The first advance might have undercut the slope at the Onondaga Formation bench and produced the sub-ice conditions that would prepare the slope for failure. The retreat would remove the confining pressure on the toe and allow for slip. Proglacial lacustrine sediments then begin to cover the toe. The readvance could compact the glaciolacustrine sediments and remove any morphological irregularities on the lower slopes and the top of the lake-bottom sediments. Glaciolacustrine sediments would continue to build until the slide is stabilized. The slide would then be placed in a metastable condition as the ice retreats and the lake drains.

A third scenario involves proglacial lake conditions without ice. Jager and Wieczorek (1994) and Grasso (1970) discuss the formation and drainage of proglacial lakes in the Tully Valley. Jager and Wieczorek (1994) also discuss the mechanisms for slope failure during the drainage of lakes and dewatering of the adjacent valley-wall groundwater system. The rise of Lake Otisco/Cardiff to an approximate elevation of 1253-1292 ft would pump up the pore pressure of the groundwater system in the bedrock of the lake walls. Sudden dewatering to lower stages such as Lake Heath Grove at 1023-1059 ft elevation could initiate the failure and produce incipient movement. The process involves removing the hydrostatic head of the lake while slow-draining materials, such as the Devonian shales of the hillside, still retain high pore-water pressure. Further dewatering to the first Lake Marietta stage (919-955 ft) could then allow the high pore-water pressure of the bedrock slope to destablize the slope and detach the slide. The drop from Lake Otisco/Cardiff to Lake Heath may have triggered the rotational block slumps that lie at the cap of parts of the highest ridge crests.

Another sliding mechanism, which will not be discussed here in detail, involves the gentle stopping of block slide movement by its plowing into uncompacted glaciolacustrine sediments with gradual enough deceleration to allow the slivers to remain intact. Later, lake-bottom water currents and gravity would have smoothed and flattened the lake bottom surface and removed evidence of disruption. This hypothesis could be tested by determining whether a significant section of early, varved clays is disrupted.

The final possible circumstance involves the downslope movement of frozen ground with a basal surface parallel to the mountain's slope. The frozen surface layer that would lie parallel to the mountain slope, with some given thickness, could dam groundwater behind and raise the pore pressure to failure levels (see Cruden and Varnes, 1996, for discussion).

Faulting

A fault interpretation would require a sloping, graben-like structure, which would have to die at both ends of the Mountain. Yet, the greatest displacements lie at the ends and are contrary to most accounts of faulting, where the greatest amount of displacement occurs near the fault's center. Whatever the fault geometry that is suggested, it must account for greatest displacements at the ends of the graben. A fault interpretation also implies Late Pleistocene-Early Holocene seismic activity in the central part of New York State great enough to generate >100' of displacement at the

Paleozoic bedrock surface. Such a large, post-glacial, seismically active, tectonic feature has not been described anywhere else in the eastern United States. This interpretation would demand an immediate, intensive, and comprehensive study. The available evidence is not compelling enough to initiate such a required study.

Groundwater Flow

The groundwater regime of the Tully Valley maintains artesian springs in the valley floor with saline water (Kappel, this volume). Springs also sit at the break in slope at the base of Bare Mountain. The logical source of the artesian head would be precipitation on the mountain. Contributors to the groundwater system include water: (1) from internal drainage that is located on the top of the mountain where crown cracks may receive local precipitation, (2) in the back-head notches, and (3) in any longitudinal cracks that may exist on the upper slopes of the slivers above the upper contact of the colluvium apron. Local capture depletes the overland flow, which thirsts the vegetation and fails to erode the slope at normal regional rates.

Groundwater flow is presumed to move along joints and bedding planes, especially in limestones and dolostones (Fig. 5.) Such a system could provide the current artesian head and saline concentrations in the valley floor.

Geologic Hazards

Bare Mountain and its presumed groundwater system provide several geologic hazards, which include: rock-block slides, rock slides, and other slope failure types on the bedrock surfaces; mudflows on the lower slopes; artesian water systems in the valley; and parched surface conditions on the upper slopes. One speculation combines these conditions and suggests that each contributes to the others: the initial drainage caused by the geometry of the rock-block slide feeds the saline artesian system that possibly triggered the LaFayette mudslide, which, in turn, has lessened the load over the buried foot and toe of the rock-block slide. The unloading may have placed the slide into a metastable condition that could allow the slivers to be reactivated, especially if another wet winter and spring occurs.

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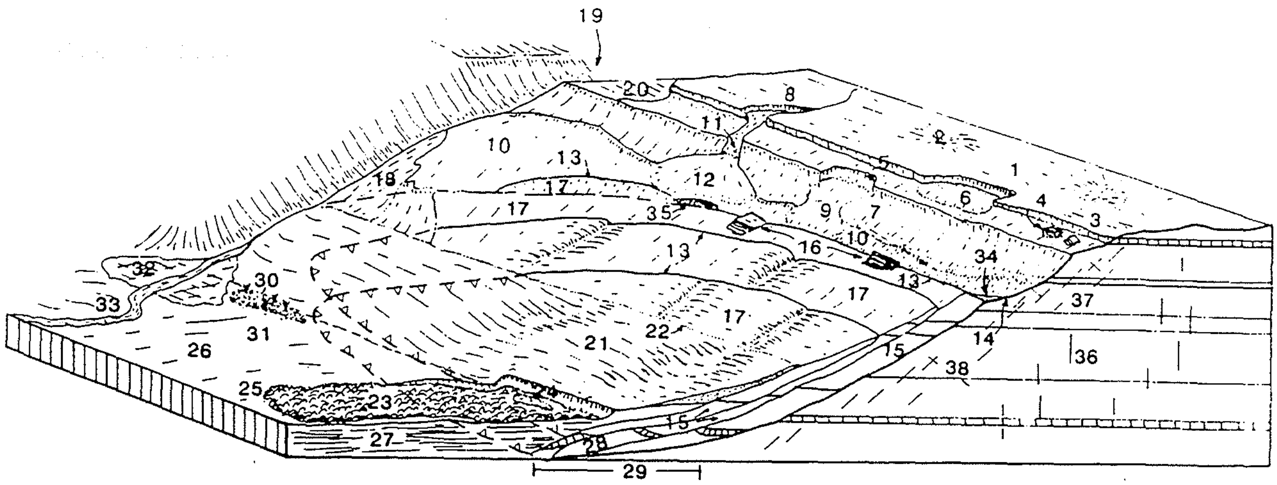

Appendix. Hypothetical block diagram showing nomenclature used in this paper. 1=plateau crown; 2=depression; 3=crown cracks; 4=topple block; 5=plateau cliff (Tully Limestone); 6=cliff slump; 7=headwall-line scarp; 8=stream-dissected crown; 9=headwall-line slump; 10=headwall colluvium; 11=headwall waterfall; 12=headwall alluvial fan; 13=block (sliver) ridge crest; 14=back-head notch; 15=rock-block sliver; 16=rotated ridge-crest slump block; 17=front or valley-side slope; 18=hanging alluvial fan (delta?); 19=gulf; 20=secondary plateau; 21=colluvial apron; 22=colluvium contact; 23=mudslide; 24=mudslide headwall; 25=mudslide toe; 26=valley floor (flood plain); 27=lacustrine deposits; 28=rock-block slide toe; 29=rock-block slide foot; 30=spring; 31=wetland; 32=alluvial fan; 33=stream; 34=back-ridge colluvium; 35=sag and sink hole; 36=vertical joints; 37=valley-dipping joints; 38=joints conjugate to valley-dipping joints. Dashed, barbed lines show presumed toe of slivers where buried. (Partly taken from Varnes, 1978, and Cruden and Varnes, 1996.)