DEPARTMENT OF THE INTERIOR
U.S. Geological Survey

Roadlog and site description for the 1989 Southeast Friends of the Pleistocene field excursion: surficial geology of the New River Valley, southwest Virginia

Edited and compiled by Art Schultz
with contributions from

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Introduction

The 1989 SEFOP Field Excursion is centered in the scenic Valley and Ridge province of southwest Virginia (Figs. 1-5). The stops of this trip include a number of features associated with the most recent geomorphic evolution of this part of the folded Appalachians. The main emphasis of the trip is the morphology and genesis of a variety of surficial deposits (block streams, block fields, ancient debris flows) and giant ancient bedrock landslides. Other stops include a look at spectacular solution karst pinnicles, New River terraces and Mountain Lake. All these features will be placed into a framework of the evolution of the present landscape.

Bedrock in the Valley and Ridge province (Fig. 5) of the field trip area consists of a 4,000 m-thick stratigraphic sequence of carbonates and siliciclastic rocks ranging from Lower Cambrian through Lower Mississippian (Lowry, 1979; McDowell and Schultz, 1989). Folding, faulting and cleavage are prevalent (Lowry, 1979; Bartholomew and Lowry, 1979; Schultz, 1983; Schultz and others, 1986; McDowell and Schultz, 1989).

The New River crosses the Valley and Ridge approximately perpendicular to the regional strike of bedrock. The present New River is chiefly on Cambrian and Ordovician limestones and dolomites. This area is dominated by floodplain and terrace deposits and karst. Deep clay-rich residuum is prevalent in areas underlain by limestones and dolomites. Southwest and northeast of the central carbonate belt, topography is characterized by high, long, linear ridges supported by quartzite and sandstone with intervening valleys of shale and/or carbonate. Outcrops of erosion-resistant sandstone and quartzite occur along the steep upper slopes of the ridges. Extensive colluvium is found on the slopes below the ridge crests and is derived from outcrops of sandstone and quartzite. Floodplain alluvium, terrace deposits, and karst are found in the intervening valleys.

Day 1

The first day of the field trip will focus on large-scale slope failures on dip slopes. (These features are described in Schultz, 1986; Schultz and Southworth, 1989; Farrell and Schultz, 1989).

Mile 0.0
Leave the parking lot of the Holiday Inn Blacksburg, heading northwest on State Route 460 Bypass.

Mile 1.8
Price Mountain (see radio towers on crest) to the southwest is composed of Mississippian sandstone and shale that is structurally below Cambrian carbonates of the Pulaski thrust sheet, i.e., they are anticlinally folded and exposed in a fenster in the thrust sheet (Bartholomew and Lowry, 1979) (Fig. 3, cross section). Although the Mississippian rocks underlie rocks of the Pulaski thrust sheet, they are topographically higher because they are more erosion resistant than the overlying carbonates. These same rocks form the ridges to the north where they are the physiographic front between the Great Valley (really only a small equivalent of the Great Valley farther north) and the Valley and Ridge.

Mile 2.5
Virginia Polytechnic Institute and State University on right. The university is located on Cambrian dolomites of the Pulaski thrust sheet. In this part of the Valley and Ridge, the Pulaski thrust sheet is riding in a Mississippian decollement (Fig. 3) (Bartholomew and Lowry, 1979; Schultz, 1983). The belt of deformed carbonates forms a wide, relatively flat terrane more characteristic of the Great Valley in northern Virginia. Characteristic Valley and Ridge topography is further west where folded and faulted younger rocks occur. The campus is about 8 mi east of the New River. Mapping has shown occurrences of New River gravels in this area and ancient drainage reconstructions have been based on them (Bartholomew and Lowry, 1979; Houser, 1981; Bartholomew and Mills, in prep.).

Mile 6.9
Cross Pulaski fault expressed as small valley below Brush Mountain. Ascend Brush Mountain composed of southeast-dipping Mississippian and Devonian sandstone and shale. In the New River Valley, Brush Mountain is the only large mountain not held up by Silurian quartzite. West of Brush Mountain, all the high topography is controlled by the resistant Silurian quartzite. The easily eroded overlying Devonian shale and underlying Ordovician shale and limestone form valleys on either side.

Mile 7.8
Cross bedded sandstone of the Mississippian Price Formation on left and right roadcuts. These sandstones contain Mississippian coals. The Merrimac coal seam was reportedly mined along the slopes of Brush Mountain during the Civil War for fuel for the Confederate ironclad battleship Merrimac.

Mile 8.5
Crest of Brush Mountain. View to west of Gap-Sinking Creek Mountain.

Mile 10.6 Stop 1
Crest of Gap-Sinking Creek Mountain. This is a brief overview stop to review the important Silurian bedrock stratigraphy (Fig. 5). Silurian

rocks not only cap the highest topography throughout much of the Appalachian Valley and Ridge of southwest Virginia, but weathering of these rocks accounts for the majority of surficial deposits. This includes both colluvium and alluvium and giant bedrock landslides in this part of the New River drainage.

In this area, the Silurian rocks are composed of a three part lithostratigraphic sequence (Figs. 5-7); a lower Tuscarora quartzite, a middle shale and sandstone of the Rose Hill Formation, and an upper quartzite and sandstone of the Keefer Formation. Typically, rocks of the 200- to 250-ft-thick Keefer Formation form the entire dip slope of the mountain. The Rose Hill sandstones and shales are about 100 ft thick and usually crop out along the ridge crest. The quartzites of the Tuscarora range from 100 to 150 ft thick and form prominant scarps on the outcrop slope just below the crest. This topographic and stratigraphic setting is unique to this portion of the southern Appalachians, which may have an effect on local relief and on the distribution of giant bedrock landslides (Schultz and Southworth, 1989).

The majority of mountain slope colluvium on the antidip slopes is derived from weathering of the Silurian Tuscarora quartzite, and, to a lesser degree, from rocks of the Ordovician Juniata and Silurian Rose Hill Formations. The large bedrock landslides are composed for the most part of quartzite and sandstone of the Keefer Formation and some of the shale and sandstone of the Rose Hill Formation (Schultz and Southworth, 1989). The slip surface for the giant bedrock landslides is within the Rose Hill Formation. An unusual pyrite-cemented sandstone is present near where we believe the slip zone is located. Solution pits and caves(??) can be found in this sandstone, and this may be an important factor in localizing slope failure.

During detailed field work in this area, Jerry Bartholomew, presently with the Montana Survey, discovered a small deposit of alluvial gravels and cobbles that resemble material usually associated with the New River. This deposit has been important in reconstructing ancient drainage patterns through this gap on Gap-Sinking Creek Mountain (Houser, 1981; Bartholomew and Mills, in prep.).

Mile 11.6

Retrace route back east on Rt. 460 to valley bottom and turn east on County Route 621. Note sign showing continental divide. The intersection of State 460 and County 621 crosses an eastern continental divide separating New River drainage on the west from Roanoke and James Rivers drainage on the east. The New River flows northwest across the folded Appalachians to the Ohio and on to the Gulf of Mexico. The Roanoke and James Rivers flow off the eastern side of the folded Appalachians on to the Atlantic.

Mile 11.9-22.0

Depending on the weather, a number of overlook stops and short hikes will be made along the dip slopes of Sinking Creek Mountain to examine the geomorphology associated with giant bedrock landslides.

Giant Bedrock Landslide in the New River Valley
by
Art Schultz and Scott Southworth

In the course of routine field mapping, anamolous topography, bedrock structure and vegetation were identified as giant, ancient bedrock landslides (Schultz, 1986; Schultz and Southworth, 1989). Along the dip slope of Sinking Creek Mountain (Figs. 6-14) (view to west from the road), the zone of failures consists of about seven distinct slides over a distance of 30 miles. Within this zone of failures only one location on the dip slope remains intact. (OVERLOOK 1) Fortunately, this intact dip slope enables us to visualize the slope prior to slope failure and to compare it to the failed slopes both to the east and west (Figs. 7 and 13) Sandstone and quartzite of the Keefer Formation form the dip slope to near the crest of the mountain. Large v-shaped flatirons are present on this dip slope with the beds dipping to the southeast from 65 to about 25 degrees. Evergreen vegetation is limited to the very crest of the mountain and within the steep sided v-shaped hollows between the flatirons. Surficial material on the dip slope consists of sandstone diamictons on the lower slopes that are most likely the result of debris flows originating in the upslope hollows. Relatively small boulder streams are present in these hollows on the higher slopes. This material is presently being derived from weathering of escarpments on the edges of the flatirons on their incised edges. On the surface of the flatirons, very thin (at most, a few feet thick) sheets of quartzite are peeling along bedding planes and moving by gravity down the slope to form bouldery talus. This general setting is important to keep in mind when comparing the slope morphology with areas of known giant slides.

From the road at this location (OVERLOOK 2) (Figs. 8-10), the view of the dip slope shows several large slide complexes, one of which forms a striking topographic feature known as Huckleberry Knob (Fig. 10). This single block contains a minimum of 30 million cubic meters of rock and is one of the smallest individual slope failures on the dip slope in this area!! The evergreens on the frontal scarp of the block are growing above a 10 to 20 m quartzite escarpment of northwest dipping Silurian quartzites of the Keefer Formation lying above southeast dipping Devonian shales. This block has traveled about 1.0 km downslope, and has moved further than the slide masses on either side.

Detailed mapping and structural data indicate that the slides were emplaced as a series of intact sheets. Structural data (Fig. 11) show that emplacement was essentially directly down the dip of the slope with rotation over a "frontal ramp". The best preserved part of the failures is the piece that rests above the ramp. Examination of this escarpment (Figs. 8, 9, & 11) (OVERLOOK 3) shows that reorientation of joints has a profound effect on the distribution and type of surficial material on the dip slope. Emplacement of the large bedrock landslides involved both translational and rotational sliding (Fig. 12). Rocks of this landslide escarpment are dipping about 10 degrees to the northwest. This rotation during slide eplacement reoriented the joints and bedding so that the weathered blocks of the escarpment are more characteristic of the anti-dip slope, i.e., they are more square and blocky than sheet like. In this way, colluvium can be differentiated into material present prior to the large scale slope failures and material that post-dates the emplacement of the slides. Landslide derived colluvium is extensive on slopes below the failures and below the breakaway scarps. Colluvial material that predates slide emplacement consists of sandstone diamicton with weathered pebbles, cobbles, and boulders of sandstone from the Keefer and Rose Hill Formations. These deposits are probably the result of debris flows down hollows incised into the dip slope. These deposits predate the large bedrock landslides upslope, because colluvium derived from the landslides cuts acrosss and has overridden this material on the higher slopes.

The oversteepened dip slopes along the escarpments are typical in zones of rock block failures in this area. The oversteepened slopes are probably the result of both the landslide mass itself and the protection from erosion of the overridden shales. The capping resistant sandstones of the landslide blocks armor the slopes, slowing degredation with respect to the surrounding shale dip slope. Deep incision into the landslide block escarpments attest to their great age. Reconstructions of what these failures must have looked like prior to erosion, and using a 40mm/thousand year regional denudation rate places the slides well into the Miocene!!! Although we remain skeptical of such longevity, there is no doubt that these large slope failures have undergone a long period of weathering.

Although we have not been able to locate the slip surface on these slides in the field, there are exposures that are very close to this surface. One location (OVERLOOK 4) (Figs. 8 & 9) exposes sandstones and shales of the Rose Hill Formation dipping gently to the northwest. This exposure is about 10 m above the slide surface. Southeast-dipping sandstones of the Keefer Formation form the footwall below the slide sheet. Very little deformation is present in both the footwall and hangingwall, although elsewhere, sandstone breccias have been found.

Some evidence suggests that the large failures have been emplaced over a considerable period of time. The contrast in the topography between different slide masses on the slope (OVERLOOK 5) (Figs. 13 & 14) may be a function of their relative ages. The slide to the west is more dissected and weathered than the eastern one. However, since we know little about the emplacement mechanics of any of these old slope failures, it is possible that one of the slides may have disintegrated more during movement than the other. The slide complex on the slope here (OVERLOOK 5) consists of huge lumpy topography roughly consisting of two parallel series of topographic reversals and knob (Figs. 13 & 14). During the winter, evergreen vegetation on slide escarpments stands out from the surrounding maples and oaks that normally reside on the dip slopes. The evergreens prefer the sandy, dryer "flats" which are characteristic of the sandstone escarpments of the landslide blocks. A prominent cross strike escarpment is preserved on the dip slope (OVERLOOK 5) (Figs. 13 & 14), which is a lateral breakaway scarp.

Springs, topographic slope reversals, sink holes and anomolous vegetation are all common within the landslide masses. These areas are not typical of dip slopes and have enabled us to recognize many other similar features from air photos and topographic maps.

These topographic flats on slopes, which normally are much steeper and rugged, are unique environments which are taken advantage of by uncommon flora and fauna. Because they are usually much wetter throughout the year, we have observed a rich flora of evergreens and ferns, mosses, cattails, etc., and a fauna of snakes, lizards, frogs, turtles, crayfish, and other creatures not usually encountered at this elevation or on the dip slopes. At several localities we have found evidence of Indian camps as well as old fence lines and cabin foundations.

Stratigraphy of Sagponds associated with giant rock block slides:
Implications for Quaternary slope evolution:
Camp Tuckaway
by
K. M. Farrell

The ancient sag pond at the Camp Tuckaway locality occurs as a large circular depression behind the escarpment of the bedrock landslide at an elevation of approximately 2,580 ft (Figs. 13-17). At the present time, this sag pond is intermittantly wet; however, the stratigraphy of the sag pond fill reveals that it was once part of a large lake complex that formed along slope reversals following landslide emplacement. Datable organic material is completely missing from the sag pond sequence; however, the presence of clastic laminates mixed with gravel supports a perched lake hypothesis. Interestingly enough, the lake deposits appear to be overlain by a colluvial diamicton of unknown age.

The sag pond deposits were impenetrable by vibracoring because subaerial conditions enhance sediment dewatering and increase compaction so that the coring method doesn't work. Instead, the deposits were hand augered to a depth of 16 ft with a 3.5 in diameter sampling bucket at the sites indicated (Figs. 13 & 14). Augering was relatively easy through these highly gravelly deposits because most of the rock fragments were deeply weathered into unconsolidated clasts. Because hand augering destroys many primary attributes of stratification, and because weathering has altered the apparent lithology, special attention was given during logging to identifying the principal lithology, identifying matrix versus rock fragments, and determining whether clast-supported or matrix-supported fabrics were present.

Stratigraphy

The graphic log and text below summarize the five units observed in this basin infill (Figs. 14, 15, and 16).

1. Gravelly Sand: The deepest sediment encountered is a loose dark reddish-brown (10 R 3/4) medium sand to silty fine sand, with sporadic gravel-sized clasts. Above this is clast-supported(?) conglomerate with a silty to clayey, very fine sand matrix or interlayered(?) conglomerate and very fine silty sand. The rock fragment population throughout is dominated by moderately indurated blackish-red (5 R 4/2) sandstones of the Rose Hill formation. Rock fragments are surrounded by a "matrix" which is typically lighter in color (grayish-red 5 R 4/2). Locally, even the sand fraction consists of sand-sized fragments of the Rose Hill formation. This red gravelly sand grades upward into the overlying gravelly clay as the matrix changes from sand to clay in a matrix-supported gravel.

2. Gravelly Clay: This red gravelly clay, ranging in color from dark reddish-brown (10 R 3/4) to grayish red (10 R 4/2), includes numerous floating (matrix-supported) rounded clasts of deeply weathered rock fragments. Rock types are highly variable and include pale blue (5 Pb 7/2) claystone, yellowish-gray (5 YR 7/2)

very fine sandstone, mudstone and shale, grayish-red (10 R 4/2) fine sandstone, greenish-gray (5 GY 6/1) sandstone, mudstone and shale, moderate yellow (5 Y 7/6) very fine sandstone, and pale olive (10 Y 6/2) shale. Locally, the red clay includes swirling parallel white streaks, which could represent evidence for primary laminations. Locally, the clay appears to be sand-rich because of large numbers of powdered sandstone clasts. Laterally, the clay coarsens to a plastic clay-silt also with floating clasts of variable lithology. The clay-silt includes layers of clast-supported gravel with a clay-silt matrix and a layer of well-sorted granular-sized quartz.

3. Gravelly Clay-Silt Laminates: Upsection, the red gravelly clay is replaced by moderate brown (5 YR 4/4) to dark yellow brown (10 YR 6/6) laminated and fissil clay-silt which locally serves as matrix for zones of clast- and matrix-supported conglomerate. The sediment is typically very soft and smooth because both the matrix and weathered rock fragments are composed of silt and clay-sized particles. In the absence of color differences, it is difficult to differentiate the matrix from the pseudomatrix (rounded clasts of soft weathered mudstone and shale). Matrix-supported conglomerate is present where Keefer (grayish-orange 10 YR 7/4) and Rose Hill (dusky red 10 R 2/2) sandstone fragments are present as floating powdered sand patches (3-4 cm in diameter) in laminated silt. The unit becomes more and more gravelly upsection. Clast-supported gravel occurs where the silt is locally chocked full of mostly deeply weathered rock fragments, ranging in size from granule to cobbles. Locally, a layer of clast-supported gravel with a clayey silt to sand-rich matrix is cemented by hematite(?) into a hardpan.

4. Well-Sorted Fissile Sand: This unit consists of moderate yellowish-brown (10 YR 5/4) fine, well-sorted sand with a definite fissility due to silt-rich partings. It basally includes clast-supported gravel and internally it fines upward into the overlying soil profile. Especially near its base, it contains abundant deeply weathered rock fragments (5E). Granules of quartz are sporadic throughout.

5. Gley Soil: Dark yellow-brown (10 YR 4/2) mottled clayey silt that resembles a gley-type of soil typical of low lying floodplains gradationally overlies the fissile sand. Stratification is not present, and the unit is extensively bioturbated, rooted, and locally contains plant remains. Granules of quartz or quartzite are present throughout.

Geologic history and implications for slope evolution

Any interpretation of the sequence must account for the dominance of Rose Hill rock fragments in the lower part of the basin infill, the coexistance of fine-grained laminates and gravel in the middle part of the basin infill, and the absence of interstratified soil profiles.

The basal red gravelly sand, consisting of a loose mixture of Rose Hill sandstone fragments and quartz grains derived from disaggregated sandstone, was deposited in the basin after the rock block slide exposed fresh faces of the Rose Hill sandstone on denuded mountain slopes. Rain water runoff rapidly washed disaggregated sand grains and rock fragment talus into the basin (Fig. 17A). The preponderance of gravelly sand and probable clast-supported, poorly-sorted gravel with a clay to silty sand matrix at the top of the unit may represent a sudden influx of slope talus, a debris flow, or slope wash of some type.

Water probably began to pond soon after the rock block slide. A lake formed over the initial slope wash deposits because of impermeable understrata and possible plugged stream outlets. The red clay deposit with floating rounded rock fragments of variable lithology was deposited in quiet standing water surrounded by talus slopes (Fig. 17B). This lake environment was stable long enough for at least 3.5 ft of red clay to accummulate. The sedimentation rate is unknown. Stream valleys became entrenched in the slopes above the lake so that rock fragments of variable lithology were exposed and supplied to the lake (Fig. 17C). Gravel-sized particles could be added sporadically to the middle of a lake by ice rafting or dislodging by tree roots.

The fining-upward sand with silt-rich partings is best explained as bar and sheet wash deposits that formed as major storm activity allowed streams to transport and deposit a significant amount of well-sorted sand in the basin (Fig. 17D). There are no interstratified soil profiles here, so I suspect that all of the sand was deposited during a single event or entirely subaqueously. It is possible that the lake became subaerial after being choked with sediment. Perhaps coincident with this storm, outlets formed, dramatically emptying the lake and allowing a fluvial drainage to exist temporarily (Fig. 17E).

Since this time, the region of former lake has been subaerial. The basin floor is flat and relatively undissected although crossed by a small stream that exists through a major entrenched outlet. Surficial deposits consist of a dark-colored gley-type soil typical of floodplain deposits. It is fine grained (clay-silt), bioturbated, mottled, and unstratified. Besides the formation of this soil, the basin does not seem to have received sediment for a long time. The entire sequence is overlain by a colluvial diamicton of unknown age (Fig. 17F). The diamicton is composed entirely of Rose Hill sandstone clasts derived from the upslope landslide scar.

Retrace County 621 west to State Route 460, continue northwest on 460.

Mile 25.0

Crest of Gap Mountain. Entire Silurian stratigraphic section is exposed in this wind gap. View to west of Giles County, Virginia. Quartzites exposed on the limbs of regional-scale folds form most of the major topographic highs in Giles County.

Mile 27.1

Rounded slopes developed in the shale of the Ordovician Martinsburg Formation on left and in cuts on right. Spruce Run Mountain on right held up by the Silurian Tuscarora Formation in a syncline.

Mile 27.7

Excellent exposures of southeast-dipping limestones of Middle Ordovician age. Sinking Creek on right.

Mile 30.1

Turn left on County Route 730. Note the numerous sink holes developed in Middle Ordovician limestones.

Mile 33.6

Turn right onto Route 813, a dead end road to the suburbs of Eggleston, Virginia. Proceed to bottom of hill, park vans, and walk to the karst towers on the New River.

Stop 7 Eggleston karst

This stop not only gives us a chance to study an excellent example of joint-controlled karst features, but also allows us to view one of the most scenic spots in Giles County (Figs. 18 & 19).

The following description is taken from Historic Springs of the Virginia, a Pictorial History, by Stan Cohen, published in 1981 by Pictorial Histories Publishing Company, Charleston, West Virginia.

Located three miles west on S.R. 730 off U.S. 460 on New River in Eggleston, Giles County.

Eggleston Springs has known many names through the years--Gunpowder Springs, New River Sulphur Springs, Chapman's Springs and Hygeian Springs.

It was situated on the east bank of the New River opposite some magnificent geologic formations--towering, grotesque limestone cliffs.

Adam Harmon, one of the early settlers of the region, in 1849 was the first to build on the site, but he had no vision of using the spring water for commercial purposes. It was not only the early 1830s that any thought was given to establishing a resort here, and its first name--Hygeian Springs--was quite appropriate. Because of its highly sulphurous odor and taste it had been known also as Gunpowder Springs.

A Dr. Chapman took over management in 1853, and it was alternately called Chapman's Springs and New River White Sulphur Springs. He constructed a fairly large hotel in 1855, and the resort became a popular visiting spa for people from southwest Virginia and even as far away as South Carolina. The Virginia and Tennessee Railroad brought guests to Newbern in Pulaski County, and then a stage made the last 26 miles. Fishing in the New River became a favorite pastime for guests.

Unlike so many other resorts during the Civil War, the resort came through in good shape--but with no customers left. In 1867, Captain William Eggleston assumed control, and the spa was known as Eggleston Springs from then on. It was still hard to get to, but some guests discovered a relatively easy solution. From the railroad stop they could board boats for a leisurely 26-mile trip downriver to the spa. When the Norfolk and Western Railroad built a branch line along the New River to the West Virginia border in 1883, guests could take the train all the way and cross the river by ferry to the spa.

The old hotel was replaced by a new one in 1902. But because the Virginian Railroad moved into the New River Valley and its tracks came very close to the new resort, the hotel was moved further up the hill in 1906. With the coming of the new railroad, built mainly to haul coal from West Virginia to the Atlantic Coast, the peacful and serene setting of the spa was spoiled and business dropped off through the years.

The resort ceased operations altogether in the late 1930s, and the hotel and remaining dance hall were torn down shortly afterward. Nothing remains today; new homes have been built on the old resort site.

The village of Eggleston, which grew up around the springs, is still somewhat isolated although a new bridge constructed in 1979 provides much better access to this scenic part of southwest Virginia.

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Bedrock exposed in the 100 m-high pinnacles is in the Upper Cambrian Copper Ridge Formation (Fig. 18). Bedding is striking about N80E and dipping about 15 degrees to the southeast. Two conspicuous joint sets (a several less well developed sets) are present. One set strikes N80-85E and is vertical- to steeply-dipping to the southeast. These strike joints are continuous and closely spaced and occur from the river level to the top to the pinnacles. The other set trends N20-30W and is vertical- to steeply-dipping to the southwest. The three surfaces, bedding and the joints produce a three dimensional box-work of intersections that have selectively undergone solution during formation of the towers (Fig. 19). Within the numerous small caves and cracks, Late Pleistocene animal remains have been recovered.

Another interesting feature here at Eggleston is the "Eggleston Suck". During flood stage, huge whirlpools form in the New river below the pinnacles. During a flood in 1979, Schultz observed two large whirlpools and numerous smaller ones. At that time, the larger whirlpool was 200-100 meters in diameter and had a vortex hole at least 1 meter below the surrounding water surface. Several meter-high waterfalls were present along the northwest side of the river flowing upstream!! Numerous tales are told by the local people, including one in which a freight car was swept off the railroad tracks and into the whirlpool, never to be seen again. Estimates range from 20 to 50 meters deep for the suck. Occasionally, the larger whirlpools are outlined by slowly rotating floating debris during calm water periods.

Return by Route 813 to County route 730, proceed west on 730.

Mile 34.2
This is a short stop to examine a high level New River terrace that is much altered by karst solution. The view to the northeast shows the karst pinnacles just visited on the previous stop (Fig. 18 & 19). New River gravels have been found in the fields above the karst towers. These are about 100 meters above the terrace level we are presently on. Within the road cut is a typical exposure of highly weathered New River alluvium. Rare clasts are of sandstone and quartzite with an occasional weathered metamorphic or igneous rocks derived from the Blue Ridge or Piedmont to the east. Mills and Wagner (1985) have studied these deposits along much of the length of the New River. Our last stop on Sunday will examine New River terraces in more detail.

Although not a good idea, you may wish to walk out on the Eggleston bridge for a spectacular view up and down the New River. Please watch out for cars. The speed limit is 55 mph along here. Note the steep river escarpments and the location of the terrace in the roadcut.

Mile 34.3
Cross New River. View through water gaps in Silurian rocks to the southeast. View of Eggleston to northwest.

Mile 36.4
The low hills through this stretch of road are held up by chert in the Cambrian and Ordovician dolomites. Throughout the years the farmers have picked up the chert that forms thin mantles on the slopes and piled it up along fence rows and in sinkholes.

Mile 39.2
The road is running along strike in a limestone valley of Middle Ordovician rocks.

Mile 41.4
Turn left onto State Route 100 and cross Big Walker Creek.

Mile 44.5
Turn right onto State Route 42. For the next several miles we will be running along a valley of the Saltville fault, a major southern Appalachian thrust fault (Figs. 2 & 3). The fault is usually just off to the northwest of the road (Figs. 20 & 21), which is in the southeast-dipping Honaker dolomite of the immediate hanging wall. Rocks in the footwall are overturned to the northwest and are of Middle Ordovician limestones.

Mile 48.1
Road cut through a 70 year old debris flow. This stop is a brief look at the texture of an old debris flow near its toe adjacent to a floodplain. The following description is taken directly from the Roanoke Times, August 26, 1920. The article was supplied by Mrs. Byrd Bryan, a local resident whose interest in geology stems from her geologist husband, who is employed by a major oil company.

DETAILED STORY OF CLOUDBURST

August 26, 1920

James Wright Displayed Great Courage in Rescue of Two Men and Two Women

The following detailed story of the cloudburst and resultant damage, as viewed by a Roanoke woman visiting near White Gate, was delayed in transit by the high water incident to the cloudburst:

White Gate, Va., Aug 29.--Today hundreds of visitors to the farm of W. B. Wright viewed the destruction wrought Thursday night by the most terrible cloudburst ever known in this section. Starting near the crest of the spot known as the Doubles of the mountain where Flat Top and Brushy Mountains intersect, the torrent swept down the densely forested mountainside tearing out by the roots all the great trees in its path, bringing down thousands of tons of earth and huge boulders, carving out a gorge from thirty to fifty feet deep, and at its base spreading out into a log and brush-strewn area a quarter of a mile wide, before it emptied into Walker's Creek.

What was the public highway became for a distance of half a mile an impassible barrier, stumps and huge rocks, entirely shutting off White Gate from the outside world. Friday there was no mail at all, and since that time the letters have been transferred by horseback.

Story of Heroism

A pretty story of heroism was brought out by the catastrophe. Hearing the thunderous roaring, the Wright family became alarmed for the safety of their neighbors, Paris Saunders and his wife, so James Wright hastened to their home. He found them all right, but hearing a cry of distress beyond, he discovered an automobile containing Miss Marjorie Miller and James Miller, of White Gate, and Miss Josephine Johnson and J. C. Dobson, of Knoxville, Tenn., who were returning from Pulaski, caught in the swiftly rising stream.

While young Mr. Wright returned for his horses to pull the car out, the occupants, suddenly deluged, waded through water up to their armpits seeking shelter in the old deserted Saunders cabin. There they watched a pile of huge trees and rocks as high as a house descend upon them with what looked like certain death. They attempted to escape but were hurled back and pinned against the cabin by the smaller debris.

With wonderful quickness of mind, Miss Miller made them all return to the stairway inside the cabin, from whence they saw that, as by a miracle, the old barn a few feet above the cabin had caught and held the great log mass.

When Mr. Wright returned with the horses they refused to enter the torrent, so after two hours of heroic struggling through the two hundred yards of madly rushing rock and timber, wading up to his waist or stumbling into deeper holes and gullies with nothing to guide him through the roaring darkness but a whistle from all but exhausted victims, he finally reached the cabin. Then began the perilous return journey, and as the others were practically helpless from exhaustion, Mr. Wright was obliged to carry them one after another back the way he had come, across a slippery log spanning the main gorge fifty feet above the torrent, and on to his home a half mile away where his mother and sister received them with a great fire, and steaming coffee. Soon the others were rolled in blankets and in bed, but young Mr. Wright, unable to telephone, hastened a mile further to assure Miss Miller's family of their safety.

Llewllyn Miller returned with him to look for the car, but up to the present time no trace of it has been found. The automobile party suffered severely from bruises and cuts, the cold, and shattered nerves, but all are recovering.

Another Car Accident

Half a mile below the scene of this accident a second car driven by Claiborne Bane, son of Charles Bane of Crandon, was upset, and demolished when a fill in the road bed gave way before another mountain avalanche. Mr. Bane barely escaped with his life.

Great damage is reported from the Sugar Run District by the same storm including the loss of cars, houses, and two sawmills.

Great Boulder Moved

The greatest single instance of the power and force of the water at White Gate was the moving of a boulder weighing about one hundred tons, a distance of a hundred yards down the gorge. Those of our young people who had been in "No Man's Land" in France say the scene of desolation bears a strong resemblance to that region of shell-torn earth and splintered trees, not one left standing. One soldier who was in the battle of the Muese-Argonne remarked that Mr. Wright's voluntary risking of his life in the horror-filled darkness marked as high an order of courage as was shown on the battlefields of France.

Mile 50.6

Late Pleistocene sag ponds associated with a large dip slope failure. This stop affords us the chance to examine several features associated with a very old landslide complex. Hike begins behind Mrs. Bryan's house and permission must be granted from Mrs. Bryan and her neighbor Mr. Vest before hiking to the sites. We are most gratefull to Mrs. Bryan and Mr. Vest for their help in all phases of the project.

HIKE TO LATE PLEISTOCENE SAG PONDS

Hike Stop 1

Climb steep slope to top of small knob that overlooks the floodplain of Big Walker Creek (Figs. 20 & 21). This knob is on overturned Middle Ordovician limestone in the immediate footwall of the Saltville fault. The view to the east is of Big Walker Mountain (Gap-Sinking Creek Mountain along strike) held up by Silurian quartzites. The linear ridges in the foreground to the east are southeast-dipping sandstone of the Cambrian Copper Ridge Formation. The meader bend in the eastern foreground was the site of an extensive Indian village. The view to the north is of the dip slope of Brushy Mountain, held up by Silurian sandstones (Figs. 20 & 21). However, the entire section is here overturned dipping about 35 degrees to the southeast. The steep slopes to the northwest are composed of shales and siltstones of the Ordovician Martinsburg Formation. This steep rounded topography is typical of the Martinsburg on this slope and is in great contrast to the slope to the northeast where a series (?) of massive slope failures in the quartzite and sandstone of the Keefer Formation have armored the slope and subdued the topographic incision. The massive slope failures here contrast with those seen on Sinking Creek Mountain, earlier on the trip. No recognizable slide sheets are present. This means either that erosion has completely destroyed the original shape of the slide or that this slope failure was somehow different. On this dip slope to the west are high anomolous double ridges, ridge crest depressions and bulges. Recently, Schultz and Southworth (1989) have described gravitational sags or sackung on this and other ridges in the New River Valley. Whether this feature is the remnants of one has yet to be determined.

Continue upslope following trail to large boulder of quartzite in stream gully.

Hike Stop 2

Here is an excellent example in the contrast of slope morpholgy between a colluvium-free shale slope on the west and a colluvium-covered slope on the east. Note the contrast in slope form and steepness. Continue across the bouldery rubble, cross the fence and into area of lumpy landslide topography. The downed trees on this slope are the result of hurricane Hugo in 1989.

Hike Stop 3

This area displays excellent, anomalous lumpy topography associated with the failure of this entire dip slope. Although we really don't know how these holes and swales form, they are classic landslide topography on a grand scale. The holes may have formed with loss of fine material within the landslide debris from long term erosion following landslide emplacement. Continue on trail to swamp.

Hike Stop 4

Late Pleistocene to present sag pond. This pond was located in the course of landslide investigation and has become a significant site in our studies of ancient bedrock landslides and on slope evolution. Vibracoring of the site was completed in the summer of 1987 and the cores have been studied for pollen, radiocarbon dating, and sedimentology.

Stratigraphy of sag ponds associated with giant rock block slides: Implications for Quaternary slope evolution: White Gate locality

by

K. M. Farrell

At White Gate, two sag ponds were identified on anomalously flat benches along the southeastward-dipping slope of Brushy Mountain at elevations of 2,575 and 2,675 ft above MSL, respectively (Figs. 20-22). At each site a scarped slope rims the northwest margin of the bench and the rock block slide lies downslope to the southeast. A cross-section through the mountain shows the scarp, sag ponds and landslide block relative to the underlying southeastward-dipping overturned Paleozoic section (Fig. 22).

In June 1987, both sang ponds were vibracored at the sites indicated (Fig. 22). The lower sag pond is significant to our studies of ancient bedrock landslides and slope evolution because the vibracores retrieved up to 5 m (uncompacted) of beautifully preserved peat from which we acquired three radiocarbon dates and the complete Holocene pollen record. The upper sag pond is interesting geomorphically but has not been dated. From the stratigraphy preserved in the cores, several inferences can be made regarding Quaternary slope evolution. Core descriptions and inferences are presented below.

Late Pleistocene to Holocene sag pond

Stratigraphy.--The lower sag pond includes two units that overlie a weathered rock profile:

1. Gravelly silty sand: Conglomeratic silty sand is present as scattered lobes in the peat and as the basal unit flooring the basin. It is extremely poorly-sorted and contains matrix-supported angular rock fragments up to 6 cm in diameter, which are concentrated at the top of the unit. The matrix for the most part consists of tan poorly-sorted very fine to fine silty sand and sandy silt mixed with numerous coarser sand grains. The matrix coarsens upwards within a sand bed. It is thixotropic; it liquifies, expels its water, and hardens to a brick-like sediment within minutes of being exposed to the air.

2. Peat: The peat is dark brown in color and is highly variable internally with respect to both composition and grain size. Lithologically it includes organic material and terrigeneous sediment. The organic material is present as (a) fine disseminates in terrigenous clay and silt, (b) stringy mats of roots, twigs, and reeds, (c) dense aggregates of granular appearing organics (reeds and wood chips?), and (d) sporadic tree trunks, branches, and twigs, wood chips, and seeds. Terrigenous sediment is present as (a) disseminated clay and silt in peat, (b) sporadic coarse grains (including sand-sized grains, granules and pea-sized pebbles, and (c) thin, bioturbated sand layers (<4 cm thick). Rock fragments include mostly white, fine-grained sandstone or quartzite. Lenses of the silty conglomeratic sand described above interrupt the

peat. Locally, the underlying conglomeratic silty sand grades upward into peat because of vertical biologic mixing.

Core Examination.--Several cores are available for you to exmaine. These are the same cores used to construct the cross-sections in Figure 23. The core labelled MT-4 was collected from the center of the sag pond and presumably penetrated close to the base of the sequence. Its total uncompacted penetration depth is 5.8m m. MT-4 includes 5 m (uncompacted) of peat overlying a basal sand 0.85 m thick (Figs. 23 & 24). Since this core had the greatest penetration depth and best recovery, it was sampled for pollen analysis and radiocarbon dating. A second core (MT-7, Fig. 23) includes a sand lobe (75 cm thick) that sharply overlies peaty silt and coarsens upward internally from sandy clayey silt with a basal pebble lag into well-sorted fine sand with matrix-supported rock fragments. In MT-8 (Fig. 23), the basal sand thins and has an indefinite boundary with the underlying weathered rock profile.

Radiocarbon dates.--Three radiocarbon dates were obtained from peat samples taken from core MT-4 (Fig. 24). The dating was done by Meyer Rubin of the U.S. Geological Survey in Reston, Virginia. The basal sample, collected at a depth of 440 cm (500 cm, uncompacted depth), yielded a radiocarbon date of 9860 ± 250 B.P. This date demonstrates that the entire Holocene section is preserved here and coupled with a total peat thickness of 500 cm (corrected for compaction due to coring) yields an average sedimentation rate of about 5 cm/100 years for this locality. The other two radiocarbon dates (Fig. 24) indicate that periodically the sedimentation rate increased during the Holocene infilling of the basin to a rate as high as 16 cm/100 years.

Pollen results.--Two peat samples collected from the surface and base of the peat deposit were anlayzed for pollen by Robert S. Thompson of the USGS (core MT-4). The following comments are paraphrased from Thompson's report: He found that both the surface and the basal samples were similar in overall compositon but had minor differences in minor taxa and in relative proportions. Since the basal sample included a very low abundance of pine pollen and completely lacked spruce and fir pollen, an age later than earliest Holocene is suggested because fir, spruce, and pine pollen are common in regional pollen spectra of late Wisconsin and earliest Holocene age (according to Craig, 1969; Maxwell and Davis, 1972; Watts, 1979; and Delcourt and Delcourt, 1986). At the nearest previously studied site (Potts Mountain Pond, Va.), spruce and fir disappear by approximately 9,000 yrs B.P. while spruce persists at higher than modern levels until 6,000 to 5,000 yrs B.P. Hemlock (Tsuga) dominates this profile between 10,000 and ca. 4,700 yr B.P., while oak (Quercus) is dominant after the latter date. Assuming that local site differences do not mask regional vegetation changes during the Holocene, comparisons with the Saltville profile indicate a middle to late Holocene age for the basal sample.

Geologic history and implications for slope evolution

The radiocarbon date at the base of the peat (earliest Holocene: 9,860 ± 250 B.P.) indicates that the rock block slide and subsequent formation of a small sedimentary basin behind it occurred prior to Holocene time. This date does not correspond well with the pollen age obtained by comparison of the pollen profile at this site to the pollen profiles at Delcourt and Delcourt's (1986) dated Saltville site (middle to late Holocene), but does correlate well with the Potts Mountain Pond site (Watts, 1979). The assumption is thus made that local site differences may indeed mask regional vegetation patterns.

The gravelly silty sand was deposited in the basin as debris flow lobes after the rock block slide exposed denuded slopes and prior to earliest Holocene time (Figs. 25A, B). This interpretation of the gravelly silty sand as debris flows is based on its extreme poor sorting, the angularity of the rock fragments, the concentration of large clasts at the top of the unit, the matrix-supported gravel and the thixotropic nature of the deposit. This, a late Pleistocene age is suggested for the deposition of the basal debris flow deposits. After the peat bog was firmly established about 10,000 yrs B.P., it has since received only minor incursions of debris flow (Fig. 25C).

The environment of deposition for the peat was highly variable locally throughout the existance of the swamp. The swamp consisted of vegetation clumps separated by zones of standing water. Sheet wash transported terrigenous sediment and detrital organics down-slope into pond-like areas. Clay, silt, and fine-grained detrital organics accumulated in standing water and became trapped within vegetation clumps, which provided a local standing crop of organic debris. Whole trees, branches, twigs, and seeds fell into the muddy organic-rich swamp deposits from the surrounding forest. Locally, vegetation clumps were very thick and well established for long time intervals so that dense aggregates of granular-appearing peat accummulated in situ with little terrigenous material added. Stringy mats of organic material probably represent rhizomes or some type of root system. Periodically, debris flows entered the swamp depositing dense lobes of gravelly silty sand.

After examining this site, proceed up trail to higher sag pond.

Hike Stop 5

Upper level sag pond. The upper level sag pond, which sits on the highest topographic bench on the slope, is located just below the breakaway scarp. It differs from the lower sag pond because thick, well-developed peat deposits are absent and replaced by terrigenous clays. The sequence was penetrated to a maximum depth of 3 m and its total thickness down to in situ rock is unknown. We have no radiocarbon or pollen dates on these cores.

Stratigraphy.--Terrigenous sediment predominated in the two short vibracores taken in the upper sag pond (Fig. 26). Three units were observed: (a) a conglomeratic cilty sand, (b) clayey peat, and (c) clay (Fig. 5B). Cores MT-9 and MT-10 are available for your examination. Detailed logs are shown in Figure 27.

1. Gravelly Silty Sand: The basal unit encountered consisted of the same tan-colored gravelly silty sand observed in the lower sag pond sequence. The sand was extremely poorly sorted, with matrix-supported angular gravel class floating near the top of a coarsening upward bed. Its total thickness is unknown.

2. Clayey Peat: A thin brown clayey peat sharply overlies the gravelly silty sand. It appears as a homogeneous mixture of clay and very fine particulate plant fragments with sporadic branches.

3. Clay: Several lithologies are present in the clay. Unstratified gray clay with scattered wood chips and plant debris forms the thickest part of the clay sequence. Locally, the clay is interlayered with laminations and flasers of detrital organics. A bed of clast supported clay balls is present.

Geologic history and implications for Slope evolution.--Because pollen and radiocarbon dates are not available, the relative age of these deposits can only be postulated by comparison with the lower sag pond's dated sediments. Based on the sedimentology the gravelly silty sand, presumably the basal unit, originated as debris flow deposits. After the period of debris flow activity, standing water conditions prevailed and terrigenous clay accummulated from suspension. Laminations attest to standing water conditions and alternating periods of deposition of detrital organics (Fall) and clay. Flasers indicate minor current movement and redistribution of detrital organics on the lake floor, probably after sheet runoff in the Spring. The clay ball conglomerate indicates that stream runoff ripped up the clay bottom of the pond and resedimented a mudball conglomerate. The reasons for the absence for a thick peat deposit here are unknown to us and may be related to ground water and surface runoff flow through the underlying formations at the site.

RETRACE ROUTE TO VANS

Mile 50.6
Leave White Gate. Retrace Route 42 east to State Route 100.

Mile 56.7
Turn right on State Route 42.

Mile 58.1
Cross through water Gap of Big Walker Mountain which is ridge as Gap-Sinking Creek Mountain to the east.

Mile 59.1
Excellent exposures of entire Devonian secton in roadcuts along here.

Mile 60.2
Cross over Brushy Mountain, held up by the Mississippian Cloyd Conglomerate. View to east across entire valley and Blue Ridge beyond.

Mile 61.3

Cross Pulaski fault again, exposed along Back Creek, at the base of Brushy Mountain. Highway crosses strike of Cambrian dolomites of the Pulaski thrust sheet. Low linear ridges are held up by thin sandstone lenses in the Cambrian Conococheague Formation.

Mile 66.7

Exit State Route 100 onto State Route 11, Dublin.
Proceed east on Route 100. Gently dipping limestones and dolomites of the Cambrian Conococheague Formation in outcrops both sides of road.

Mile 73.5

Turn left onto County Route 114.

Mile 75.7

Cross New River. Recent excavations have exposed several levels of terraces at this location as well as highly deformed Cambrian rocks of the Pulaski thrust sheet.

Mile 83.7

Turn left onto State Route 460.

Mile 86.7

Turn right in parking lot of Holiday Inn
END OF DAY ONE

DAY 2

Day 2 will focus on mountain slope deposits.

Mile 0.0

Leave parking lot of Holiday Inn and proceed west on State Route 460. Retrace route on 460 to top of Gap-Sinking Creek Mountain (see description on Day 1).

Mountain slope deposits near Mountain Lake, Virginia: genesis and geomorphology

by

Hugh Mills

Mile 21.1

Stop C. "Scenic overlook" pull off to south of Highway 460, (Figs. 4 & 31).

This high cut (Fig. 28) provides the best-exposed cross section of a boulder stream in the area, although it is smaller and steeper than the average boulder stream. The cut is less than 200 m from the crest of the mountain, which is held up by the Tuscorora Formation. Beds dip into the slope at about 45°. No cliff or ledge of Tuscorora occurs uphill. The boulder stream appears to head about 20-30 m above the Juniata-Martinsburg contact, so that the greater part rests on the Martinsburg Formation. Its slope angle is about 30°. Note the gentle concavity of the hollow cross section here. In such U-shaped hollows, intermediate diameters of the largest boulders commonly exceed 1.0 m. On most of the mountain flanks underlain by shale or other weak clastic rocks, V-shaped hollows form where caprocks supply few large boulders, whereas U-shaped hollows form where caprocks supply many large boulders. The large boulders appear to protect the flanks from fluvial incision that would form V-shaped hollows (Mills, 1989).

The boulder stream is underlain by a trough about 10 m wide and 4 m deep, cut into the underlying Martinsburg residuum and filled with colluvium (Fig. 28). A 3.2 m-high section of the trough fill reveals the following (Fig. 28). Boulders are underlain by a stony sandstone-rich diamicton which contains an increasing number of yellowish clasts of weathered Reedsville Shale downward. The sand content also decreases downward. There is a gradational contact between the diamicton and the Martinsburg residuum. Weathered _in situ_ shale occurs about 2 m below the base of this section and extends downward many meters. This vertical sequence is common beneath many boulder streams in the area.

Although colluvium in hollows may show substantial soil development, a high degree of weathering that would indicate great age is lacking. Hues typically are 7.5 YR and chromas 4-5 (Munsell), and sandstone clasts remain hard. A radiocarbon date on a bulk-sediment sample taken 1.8 m below the surface of this cut gave an age of 10,680 ± 320 YR BP (Beta-22688). As bulk-sediment dates generally are considered to be minimum ages, this result is consistent with the interpretation of the boulder stream as a late-glacial periglacial feature.

Long-axis clast orientations were measured at two medial and two lateral locations within the trough cross section, and yielded an interesting result (Fig. 28). The two medial fabrics showed orientations parallel to the long axis of the boulder stream. The lateral fabrics, however, showed orientations intermediate between this direction and a transverse one. Such oblique preferred orientations suggest that material is transported laterally into the trough from the side slopes as well as directly down the axis of the streams.

Seismic refraction was used at a number of sites in the study area to determine seismic velocities and thicknesses of surficial materials (Mills, in review). A seismic profile run 40 m east of this boulder stream showed 3.4 m of material with a seismic velocity of 365 m/s (i.e., "soil"), underlain by 11.0 m with a velocity of 1370 m/s (residuum and highly weathered bedrock), in turn underlain by material with a velocity of 2195 m/s (intact or slightly weathered bedrock). A profile along the axis of the boulder stream showed 4.1 m of 535 m/s (colluvium) and 15.9 m of 1220 m/s. This subject is discussed further in the following section.

Mile 21.4

Stop D. Scenic overlook platform on north side of Highway 460, across road from Stop C.

From the platform to the northwest, you may be able to see (depending on adjacent foliage) the southeast slope of Johns Creek Mountain, corrugated by U-shaped hollows.

The main feature to be seen here is the cut through the topographic nose across the road (Fig. 29). Note the trough-shaped reddish deposit of colluvium, about 6 m thick, capping the nose. The colluvium obviously is much more weathered than that at the previous stop. In such obviously ancient deposits, hues typically are 5 YR and chromas 6-7 (Munsell). The majority of pebble-sized sandstone clasts can be broken apart by hand. Age is uncertain, but probably is on the order of hundreds of thousands of years. The concave-up lower contact of the deposit suggests that it might be a former hollow fill, analogous to the hollow trough we just looked at. Although lower contacts of colluvial deposits are not exposed this well elsewhere, at many locations it can be seen that such contacts do not parallel the modern topographic surface, suggesting that the former surface of deposition was quite different from the convex-up shape of the present nose. These observations suggest that mountain-flank topography has changed considerably over time, rather than remaining in dynamic equilibrium, and it is at least conceivable that topographic inversion frequently occurs, with hollow floors coming to cap noses (Mills, 1981). The manner in which this inversion may take place will be discussed later.

Returning to the subject of seismic stratigraphy, I was able to delineate four broad categories of material by seismic velocity (Mills, in review):

1. <400 m/s - "soil", corresponding roughly to the solum, in which material is kept loose by biological and near-surface physical processes.

2. 400-800 m/s - unconsolidated, relatively unweathered colluvium, such as that seen at the previous stop. In boulder streams, such material may occur at the surface, the low-velocity zone being absent.

3. 800-2,000 m/s - old, weathered, partly consolidated colluvium, or else bedrock residuum or highly weathered bedrock.

4. >2,000 m/s - unweathered or slightly weathered bedrock.

The overlap between velocities of old colluvium and residuum was disappointing, since it meant that the thicknesses of old colluvial deposits could not be determined where they overlie residuum, which they usually do. For example, a seismic profile up the center of this nose, along the colluvial cap, yielded essentially the same results as profiles to the left and right of the cap: 2-3 m of soil overlying 12-17 m of material with a velocity of about 900 m/s overlying slightly weathered bedrock.

Mile 23.9

Continue west on 460 to the intersection of State Route 700. While passing the east end of Spruce Mountain, note the presence of several V-shaped hollows, in contrast to the U-shaped valleys that occur along the flanks of the mountain. This difference reflects a lower supply of large caprock-derived boulders here relative to the supply along the mountain flanks. Continue on 700 until about 0.1 mi short of the intersection with State Route 602.

Stop E. Northeast side of road (Fig. 32). Exposure of weathered colluvium in lower nose of Johns Creek Mountain.

The basic divisions of mountain-flank topography are hollow, sideslope, and nose. In a study of colluvium in these settings, I subdivided the hollow and nose environments (Mills, 1987, 1988). The hollow subdivision is discussed later. Noses were subdivided into "upper" noses closer to the mountain crest and "lower" noses farther from the crest. Noses with longitudinal slopes 14° or greater were classified as upper, those with slopes of less than 14° were classified as lower. Although colluvium in both environments is more weathered than that in hollows, lower-nose colluvium commonly is much more weathered than upper-nose colluvium. This stop shows an example of a lower nose. Note the red hues, relatively high clay content, and friability of sandstone clasts (with the exception of the hematite-cemented clasts derived from the Rose Hill Formation).

Mile 25.7

Stop F. Hollow on south flank of Johns Creek Mountain. Parking space is sparse.

This stop (Fig. 32) provides an interesting example of a relict deposit that testifies to the migration of the mountain crest. East of the hollow, the road cut reveals a deposit of Tuscarora boulders in a sandy matrix. The friable nature of many of the boulders indicates the antiquity of this deposit. Walking a short distance up the present-day hollow, however, reveals much different modern deposits. Virtually no Tuscarora boulders are seen; the mountain flank no longer supplies Tuscarora boulders to the hollow. A hike to the top of the mountain shows that there is no outcrop of Tuscarora at the crest. Since the time at which the relict deposit was laid down, the Tuscarora Formation has retreated beyond the crest.

Mile 28.6

Stop G. Overlook at southwest end of Salt Pond Mountain.
The valley before you is that of Doe Creek, which flows approximately along the axis of the Mountain Lake anticline. Figures 33 and 34 show the distribution of surficial deposits in this area; the 7.5-minute quadrangle is Eggleston, Virginia.

We are standing on the northwest flank of Salt Pond Mountain. At azimuth 010° is Bean Field Mountain (unnamed on topographic map), and to its west is Doe Mountain. Compare the U-shaped hollow form on the former mountain to the V-shaped vorm on the southeast flank of the latter. The stratigraphy and structure on both mountains is similar. Their flanks are underlain predominantly by the shale and siltstone of the Martinsburg Formation, which is in turn overlain by the Juniata Formation, with the Tuscarora and Rose Hill Formations capping the mountain. Strata dip into the slope at about 15-20°. The mountains differ greatly, however, in the supply of large Tuscarora boulders to their antidip flanks. On the south flank of Bean Field Mountain this supply is voluminous. On the southeast flank of Doe Mountain, however, the supply of large boulders is low. Although Doe Mountain is capped by Tuscarora, on the east end of the mountain the Tuscarora has eroded back from the head of the slope, which is held up by the Juniata Formation instead. Blocks of the latter are smaller and much less durable than those of the former, so that large boulders are few on the antidip flank. I suggest that this difference accounts for the difference in hollow form.

In contrast to the southeast flank, the southwest flank of Doe Mountain has numerous large boulders, and hollows are correspondingly U-shaped. One difference from the south flank of Bean Field Mountain, however, is that here many of the boulder deposits appear to be relict in nature, no longer being supplied by the caprock.

On antidip mountain flanks in the study area, there are three main surficial units: (1) younger, less-weathered sandstone-rich colluvium; (2) older, more-weathered sandstone-rich colluvium; and (3) shale residuum/bedrock. The distribution of these units is shown on Figures 33 and 34, and the distribution of boulder streams is shown diagrammatically. The south flank of bean Field Mountain is covered mainly by younger sandstone-rich colluvium; the east and southeast flanks of Doe Mountain chiefly expose shale residuum/bedrock at the surface. The southwest flank of Doe Mountain is underlain mainly by older sandstone-rich colluvium.

Older colluvium typically occurs on lower noses; some of this colluvium can be shown to be quite ancient, not only from its weathering characteristics but from its topographic position. For example, note the low hill about 600 m distant along an azimuth of 275°, with Tuscarora boulders visible on its surface. The hilltop is about 15 m higher than the immediately upflank topography, indicating that at least this much erosion has taken place since deposition. Such ancient deposits tend to be thin and mixed with shale residuum.

Even on mountain flanks largely covered by sandstone-rich colluvium, exposures of shale residuum/bedrock occur locally, chiefly on steep sideslopes and on high noses (only the latter are usually large enough in area to be shown on 1:24,000-scale maps). Presumably these high noses were once covered by sandstone-rich colluvium that has subsequently been removed by erosion, as indicated by the presence of scattered sandstone clasts on many of them. We are standing upon such a nose.

Three seismic traverses were carried out in the valley of Doe Creek to show the variation of regolith depth along the mountain flanks. One started about 75 m below us and ran to the northeast along the northwest flank of Salt Pond Mountain, one was along the south flank of Bean Field Mountain, and one was along the southwest flank of Doe Mountain (cleared area). Locations of these traverses are shown in Figure 6 of Mills (in review).

On average, seismic profiles showed regolith thicknesses in excess of 10 m, the greater part being residuum or weathered bedrock. This finding contrasts with the results reported by Carter and Ciolkioz (1986) for a site near the glacial border in Pennsylvania, which showed that young colluvium directly overlies relatively unweathered bedrock. This difference to an extent may reflect less-intense Pleistocene periglacial erosion in southwestern Virginia than in Pennsylvania. Topography generally was not a good predictor of regolith thickness; hollows showed greater thicknesses of young colluvium than did noses, but hollows and noses showed little difference in total regolith thickness. The largest systematic difference was found between hollow floors (or parts thereof) that seemed to be undergoing long-term downcutting and those that appeared to be relict features no longer associated with active drainageways. The former were underlain by an average of 5.5 m of weathered regolith, whereas the latter were underlain by a mean of 14.0 m, indicative of a greater depth of weathering and therefore probably a somewhat greater age. This is discussed further below.

Stop H Road cut through nose.

The road cut (Fig. 32) reveals that this nose has a contact between residuum and overlying sandstone-rich colluvium that is not parallel to the modern ground surface. On the left (north) side of the nose, colluvium is thick. To the right (southwest), colluvium thins until shale residuum occurs at the surface. The disparity between the form of this contact and that of the ground surface shows that hillslope topography has changed.

Mile 29.8

Stop I. Nose beneath high-tension lines (Figs. 30, 32, & 34).

View west along lines is of V-shaped hollows on southeast flank of Doe Mountain. View east shows shale nose, which farther upslope acquires a cover of sandstone-rich colluvium. Walk about 100 m back down road to hollow. Looking up the hollow (gradient 22°), note that the left (north) sideslope is steep and consists of shale or shale residuum. In contrast, the right sideslope is so gentle as to be nearly indiscernable, and is covered with large boulders and sandstone-rich colluvium (Mills, 1981).

I suggest that here the hollow is migrating to the left (north), cutting into the nose. This is accomplished by rare, intense rainstorms that provide runoff that moves down this lateral channel, undermining the shale sideslope, alternating with much rarer events capable of moving large boulders. (Such events probably are debris flows, although other mechanisms such as periglacial processes are possible). When the large boulders are moved they occupy that new strip of hollow floor formed by lateral fluvial erosion since the last boulder-moving event, thereby armoring it and restricting subsequent fluvial erosion to the margin of the bouldery hollow fill. Such hollow migration, downwards and sideways, eventually can result in topographic inversion of hollows and noses, producing colluvial caps on noses that appear unrelated to the present topography.

The boulder deposits in this area are of at least three different origins and probably vary somewhat in age (Mills, 1988). One origin is by the direct action of gravity on steep slopes, to include falling, rolling, sliding, and creep. We won't see any of these today, but there are good examples on the south flank of Butt Mountain along the deep valley of Little Stony Creek. These occur on slopes of 35° or greater, and their instability underfoot suggests that they are probably no older than late Holocene. The second and third origins I have already referred to: debris flows and periglacial depostis. The crux, I think, of the boulder-stream problem is discriminating between these two origins. Although this can by no means be done with confidence, I will suggest tentative classifications. Such classifications should be viewed only as hypotheses for future research.

Age, of course, is important. A late-Pleistocene age doesn't prove a periglacial origin, but it makes it somewhat more likely. Absolute-age dates are difficult to come by, but relative-age dates can help by showing that some deposits are older than others. The best such measure for boulder deposits that I have found is percent of cracked boulders; pitting and corner rounding were tried but were unsuccessful. Besides age, another possibility is sedimentary features diagnostic of certain depositional processes. Two important periglacial processes are frost heave and gelifluction (solifluction). Does either of these have sedimentary characteristics that distinguish it from debris flows? The answer is maybe. The sedimentary feature that I have looked at is clast orientation or fabric, both of the boulders and of the smaller clasts in the underlying colluvium (where exposures permit). Frost heave often orients tabular boulders into an on-edge position, an orientation much different from the flat-lying orientations produced by debris flows. In fact, I know of no process that can consistently produce on-edge orientations of large boulders other than frost heave. Therefore, I think that we can tentatively classify boulder streams, or parts thereof, with large numbers of on-edge boulders, as relicts of periglacial frost heave.

Gelifluction deposits are more problematic. One difference I have gleaned from the literature concerns fabrics of clasts in a matrix. Although the general fabric <u>patterns</u> of gelifluction and debris-flow colluvial deposits are similar, the strengths or consistencies of gelifluction fabrics are usually much greater than those of debris flows. The strong fabrics of the colluvium beneath the Gap Mountain boulder stream that we saw earlier, for example, I believe support a gelifluction origin. Within the study area, I have found large differences in strength of colluvial fabrics beneath boulder streams. The strong fabrics tend to occur in moist north-facing hollows closer to the mountain crest, whereas the weak ones tend to occur in drier hollows farther from the crest. On this basis, I classified hollows into types "M" (moist) and " D" (dry). Note that type M hollows, moister and drier than type D hollows, would have been more favorable settings for Pleistocene gelifluction. I have, therefore, tentatively classified colluvium in M hollows as gelifluction deposits.

The present hollow should be a type M, but exposures of colluvium are lacking. However, based on the presence of an unusual number of on-edge boulders, as well as a relatively large number of cracked boulders, I suggest that the part of this hollow floor covered with large boulders was probably emplaced during the late Pleistocene. This interpretation does not apply to the low part of the hollow cross-section adjacent to the shale sideslope, however. This location has less than one third the percentage of cracked boulders as does the rest of the hollow floor, and, in addition, the tabular boulders are generally flat lying. I suggest that this deposit has been deposited more recently, during the Holocene, probably as a flood or debris-flow deposit.

Concerning possible modern movement of boulders, over a five-year period, I was unable to detect any such movement, measured to within several millimeters (Mills, 1984, 1986a). It is true, however, that there are many trees tilting downstream on boulder streams. The downhill tilt of 30 trees at this location, for example, averaged about 14°. By measuring tree tilt at seven different locations, I tried to see if there was indeed a tendency for trees on boulder streams to tilt more than in other locations. The answer is yes, but there is such a large confounding variable that it is impossible to draw any firm conclusions. Namely, the tree types on hollow floors are almost completely different from those on noses. Here, for example, two-thirds of the trees are yellow birch (*Betula alleghaniensis*). This species almost never occurs on noses. The difference between hollow floors and noses, therefore, may be due simply to differences in tree types, i.e., yellow birch tend to tilt more.

Mile 30.5.

Stop J. Parking lot of Mountain Lake Hotel.
View Mountain Lake. This lake is perhaps the largest natural lake in the unglaciated Appalachians, and its origin has therefore stimulated substantial interest (Figs. 32-37). One theory is that Mountain Lake is a solution feature, an idea still supported by some. However, the lake rests on the upper Martinsburg Formation and Juniate and Tuscarora (see Fig. 30) which show no solution features elsewhere. Another theory is that the lake was formerly a stream valley, subsequently dammed by huge blocks of Tuscarora sandstone derived from ledges to the east of the lake (Fig. 35). This seems the most reasonable explanation, as the dam, or at

least the upper part of it, is indeed composed of Tuscarora blocks (some of which are visible from the parking lot). In addition, bathymetry indicates that the floor of the lake slopes northwestward as should the floor of a stream valley, with its deepest point (about 40 m) occurring just south of the dam. The question remains, of course, of why this is the only such talus-dammed lake in the Valley and Ridge province. Elsewhere (Mills, 1988) I have suggested that the combination of several characteristics of the setting, none in itself very unusual, may be responsible for this uniqueness.

Retrace west on 700 -- turn quick right on 613 down mountain.

Mile 30.7

Stop K. High cut in shale nose on south flank of Bean Field Mountain along State Highway 613 (Figs. 30-34).
Figure 8 of Mills (1981) shows a sketched topographic profile along the south flank of Bean Field Mountain; this is site 80. Although most of the south flank of Bean Field Mountain is covered with coarse sandstone-rich colluvium, this and one other high nose have shale and shale residuum exposed on their surfaces, along with a scattering of sandstone clasts. I suggest that such locations represent very old noses from which deposits of sandstone-rich colluvium have been almost completely removed by erosion.

The longitudinal slope of the nose is 25° and sideslopes are up to 35°. Because of thin soil, trees are poorly rooted. A late April snowstorm in 1978 dumped 635 mm of snow (measured at the Mountain Lake Biology Station several kilometers from here), causing many trees to topple, particularly on this nose. I located 30 tree-fall pits produced by the storm on this nose and its sideslopes and measured their volume. Assuming that half the volume of uprooted material moves downhill (the other half sliding back into the pits), and assuming a recurrence interval of 100 yr, a denudation rate of 13 mm/1,000 yr is represented. No uprooted trees on boulder streams in the area were found, indicating that boulder deposits provide a much more secure substrate for trees than do steep shale slopes.

This nose is also one of the five locations at which I measured downhill movement of surface clasts by taping from stakes over a five-year period (Mills, 1986a). The clasts on this south-facing slope moved downhill a median distance of 15 cm during this period. In contrast, on a north-facing nose with an identical slope angle on the north flank of Big Mountain, where clasts rest on sandstone-rich colluvium, the median distance was only 0.9 cm. Whether this large difference is due to the difference in substrate or aspect I am unsure.

Stop L. Low nose on south flank of Bean Field Mountain (figs. 30, 32, & 34). This nose (Mills, 1981) is a hollow floor that has become a nose due to the incision of hollows on either side of it. The initial stage of such topographic inversion is the formation of a "W-shaped" hollow, in which the center is higher and older than the lower areas on each side (Mills, 1981). Both W-shaped and asymmetric hollows (such as we saw at Stop I) are common on mountain flanks in this area. In both types of hollows, boulders on the lowest part of the hollow cross section tend to be younger than those on higher parts. In keeping with this finding is the observation that deposits with large numbers of on-edge clasts, presumably indicative of Pleistocene frost heave, rarely occur in the lowest part of the hollow cross section.

Stop M. Deep hollow on south flank of Bean Field Mountain (figs. 30, 32, & 34). This south-facing hollow (Mills, 1981) is somewhat drier than the one seen at Stop I; rather than yellow birch, basswood, sugar maple, and buckeye predominate. This would be classified as a type D hollow, and subsurface fabrics obtained from a drainage ditch near the road were quite weak. This fact, plus the flat-lying orientation of tabular boulders and the paucity of cracked boulders, suggests that the hollow-floor deposit here is probably a Holocene flood or debris-flow deposit.

The previous stop refers to hollows in which migration of sideslopes on one or both sides appears to be occurring. In addition to these, however, there are "ordinary" hollows with concave-up cross sections that appear to be undergoing long-term incision rather than lateral migration. It seems intuitively likely that the floors of such hollows would overlie a somewhat thinner section of weathered regolith than would the hollows engaged in lateral migration. Siesmic results supporting this idea are discussed under Stop G.

The present hollow does seem to be undergoing some lateral migration to the southeast, as suggested by a steep shale sideslope to the southeast and a gentler sideslope with numerous surface boulders to the northwest. However, as indicated by the depth of the hollow, the main long-term geomorphic tendency here seems to be downcutting. This interpretation is supported by the seismic evidence, which shows 5.2 m of young colluvium over unweathered bedrock. Presumably, the hollow has cut through whatever residuum and weathered bedrock once existed here, and there has been insufficient time to develop a new weathered zone.

Around the next bend in the road, park on right and walk west on State Highway 714. Near the intersection, note the cut through an upper nose, revealing a sandstone-rich colluvial cap which gradually becomes admixed with the shale residuum downward. This colluvium is more weathered than that in the hollows, but not greatly so. Continue up road to next stop.

Mile 31.5.

Stop N. Hollow on southwest slope of Bean Field Mountain.
Here (Figs. 30, 32, & 34) is a dramatic example of a convex-up hollow. It is a W-shaped hollow that is deeper on the right (southeast) than on the left (northwest) side. The central high is so strikingly lobate in nature that I don't rule out an origin other than topographic inversion. A periglacial origin has been proposed for such features; tabular boulders are generally flat lying, but one might attribute emplacement to gelifluction. Another possibility is a debris-flow lobe that ran out of shear stress. In any case, however the feature was placed, what is happening now is clear. A new drainage line has formed

on the southeast side of the hollow, thereby increasing the relief of the central high. Note that the clasts that seem to be in transit in this youngest part of the hollow floor are much smaller than those on the abandoned central part of the hollow floor. A seismic profile down the middle of the central high showed 2.6 m of young colluvium over 14.5 m of weathered regolith.

Continue up 714 around the next bend.

Stop O. Boulder stream with many on-edge boulders.
The numerous on-edge tabular boulders here (Figs. 30, 32, & 34) suggest relict frost heave. Note that the boulder stream is not along a present-day drainage line. A seismic profile showed that this location also overlies a thick section of weathered regolith. Why this stream should be so different from the nearby one at the previous stop, I can't explain, except to note that both may be periglacial relicts, and within a periglacial landscape gelifluction may dominate in some places and frost heave in others.

Return to 613 and start down the mountain.

Mile 35.5.
Turn left on State Route 460. Return to Holiday Inn.

Mile 54.1
Parking lot, Holiday Inn

OPTIONAL STOP
If time allows, a visit to a "classic" New River terrace locality will be included. On the way back to Holiday Inn, turn west on Prices Fork Road.

Mile 49.6
Turn west on Prices Fork Road, County Route 685.

Mile 52.8
Bear right onto County Route 652.

Mile 55.8
Turn left onto County Route 623.

Mile 57.0
Entrance to Whitethorn terrace locality. This stop gives us an excellent chance to view several levels of New River terraces opposite a present day steep cut bank (Figs. 36-38). Bedrock here is chiefly on the Cambrian Elbrook dolomite. Several studies have been done on the terraces at the locality (Mills and Wagner, 1985; Milles, 1986b) (Fig. 38). The highest New River gravels here are about 100 meters above the present New River. Examination of clasts show predominance of sandstone and quartzite with minor metamorphic and volcanic clasts. Other characteristics are summarized in Figure 39.

Retrace route to State 460. Return to Holiday Inn.
END OF DAY 2

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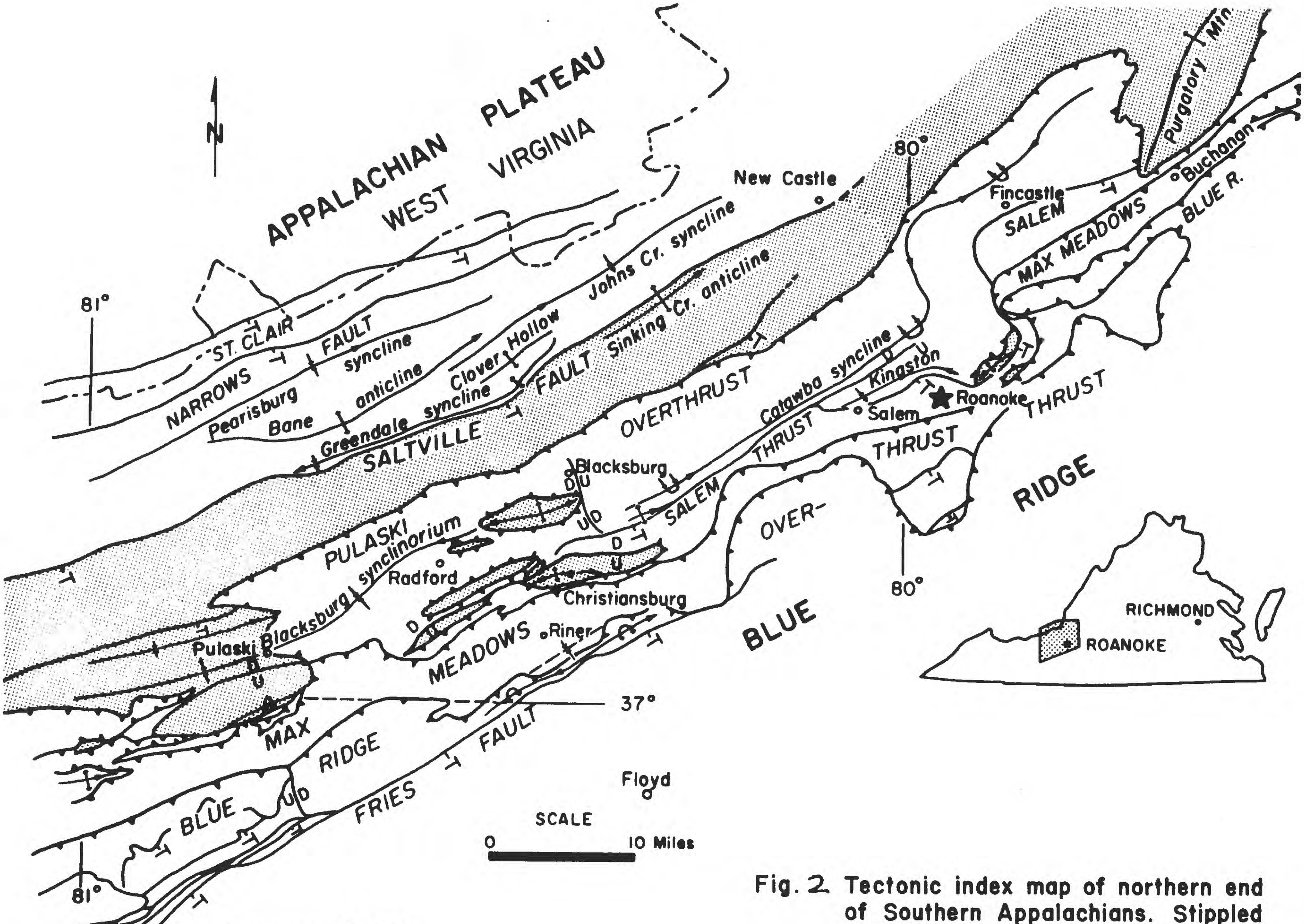
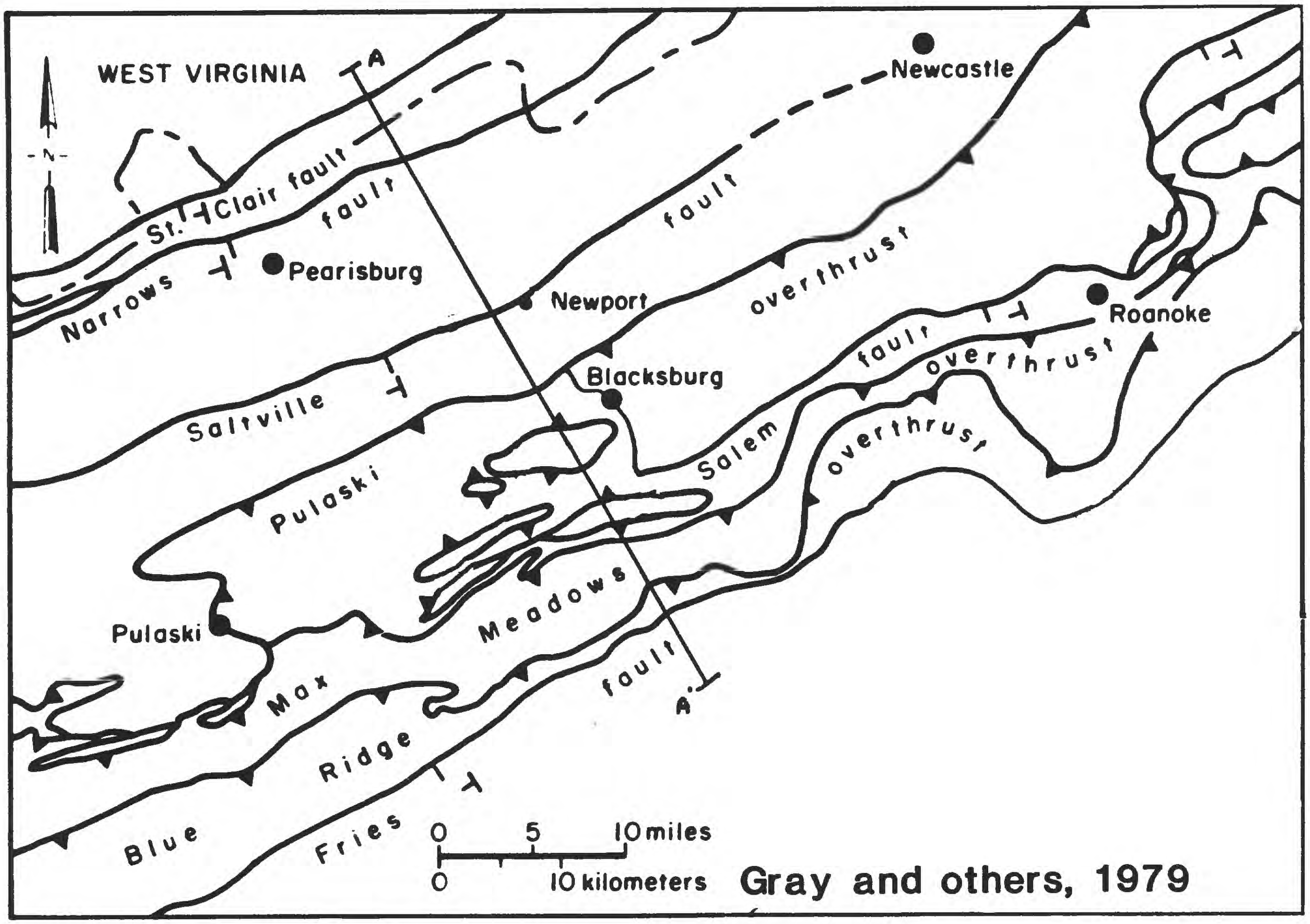
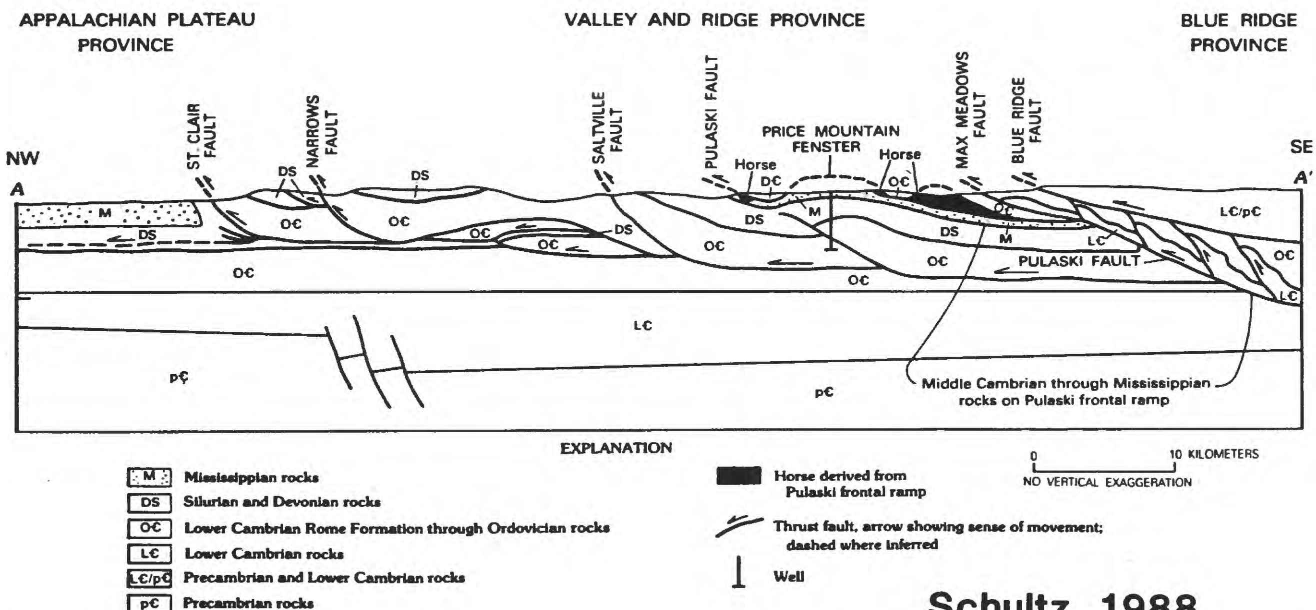
Figure 1. Satellite photography showing physiography of field trip area.



Lowry, 1979

Fig. 2. Tectonic index map of northern end of Southern Appalachians. Stippled areas belong to Saltville fault block.





Schultz, 1988

Figure 3. Cross section *A–A′* showing major frontal tectonic ramp of Pulaski thrust sheet, the proposed source of horses exposed in fensters. Modified from Gresko (1985) and Woodward (1985).

Figure 3. Generalized tectonic map and cross section of overthrust belt near Blacksburg, Virginia.

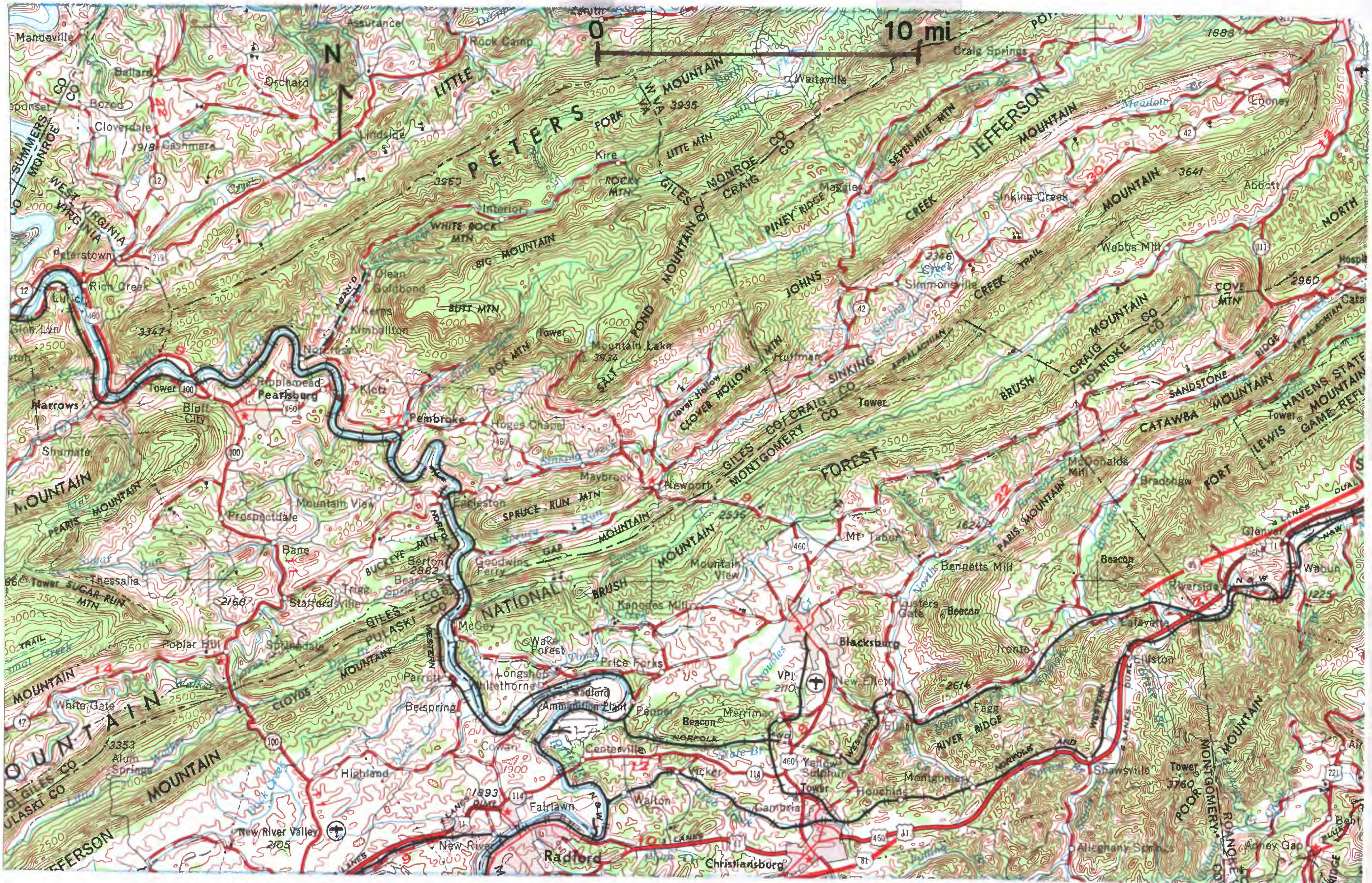
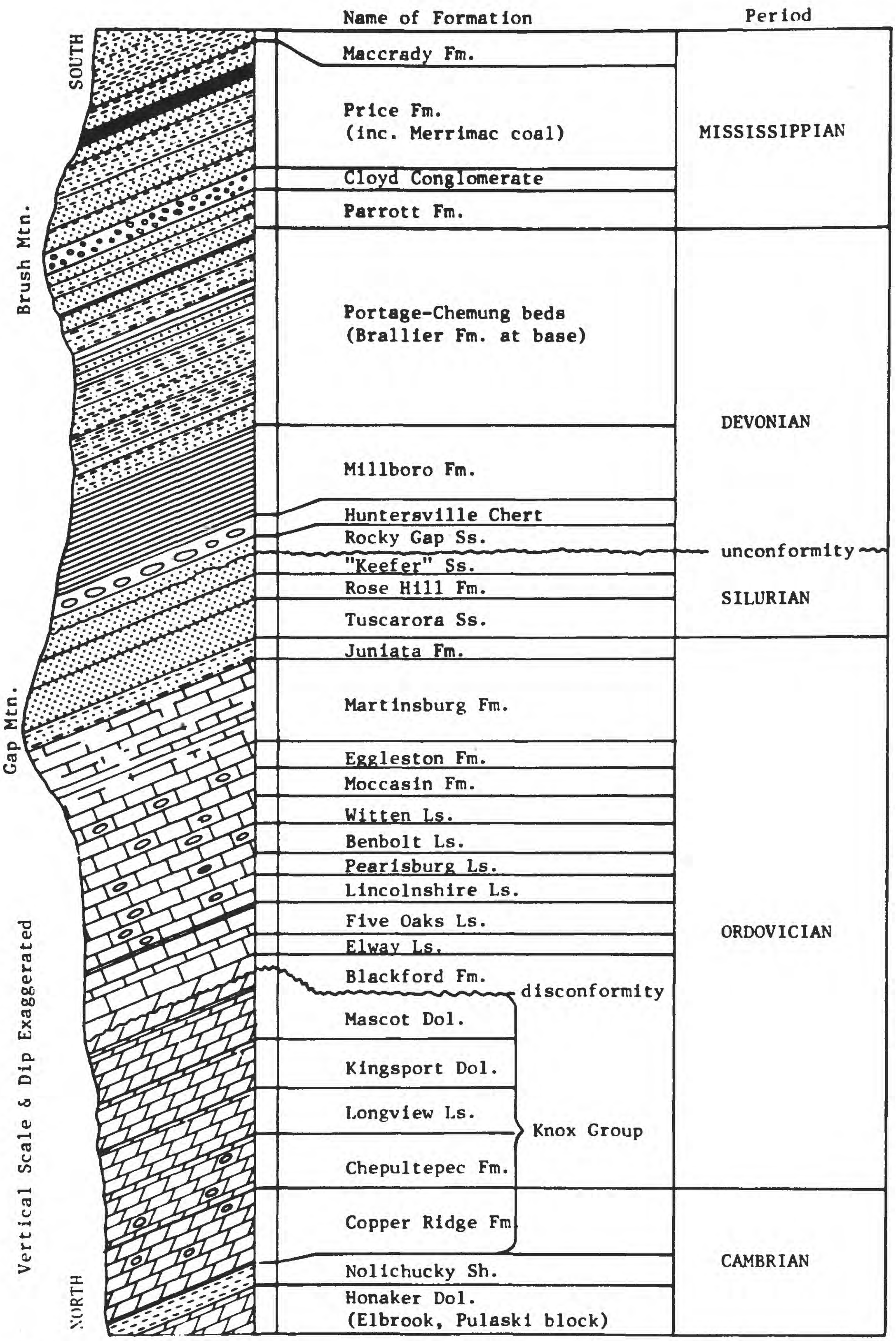


Figure 4. Road and topographic map of New River Valley, southwest Virginia.



Formations along New River SE. of Saltville fault (NW. limb Blacksburg synclinorium)

Gray and others, 1979

Figure 5. Bedrock stratigraphy and topographic profile across Gap Mountain, southwest Virginia.

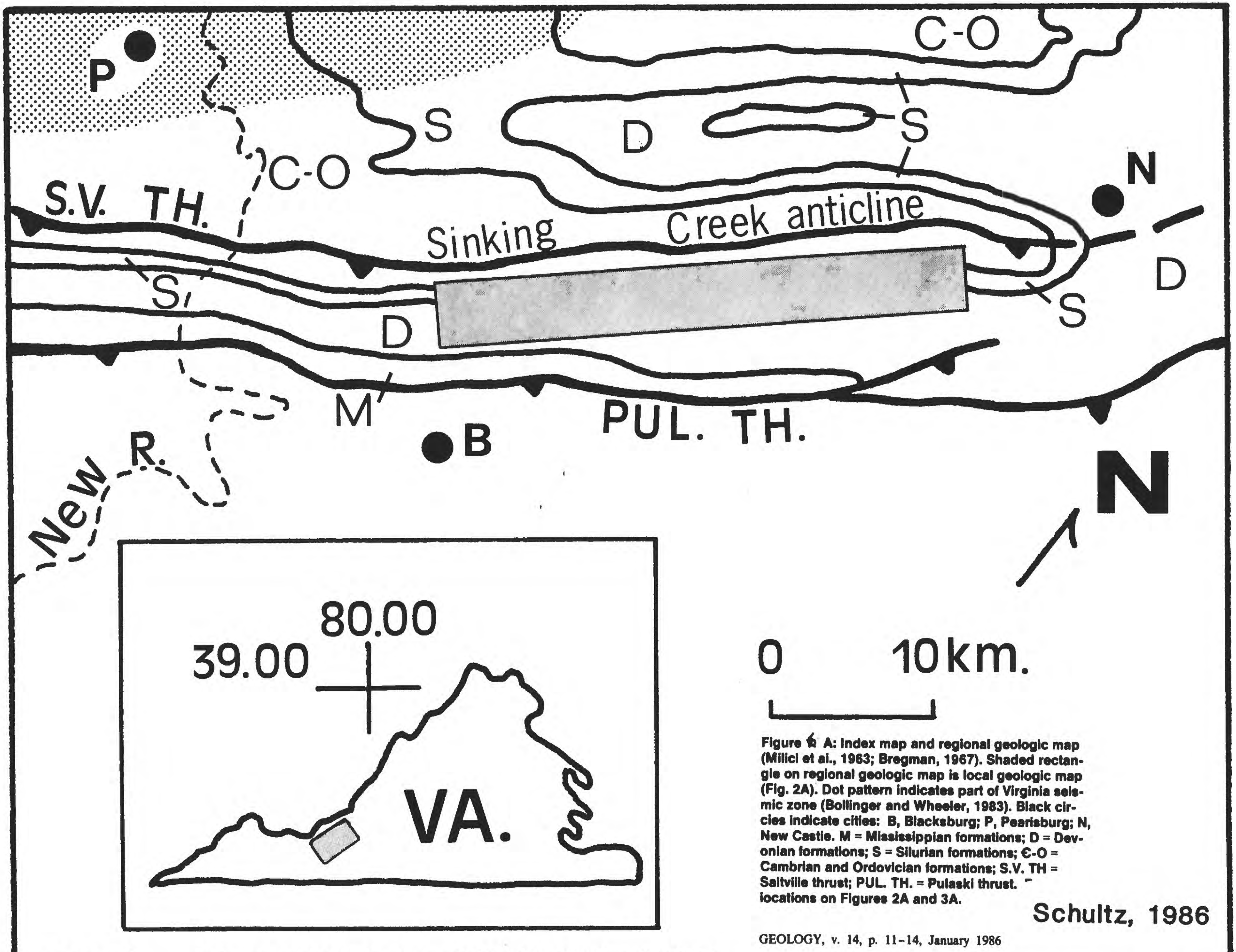


Figure 6. Bedrock geology in area of Sinking Creek anticline.

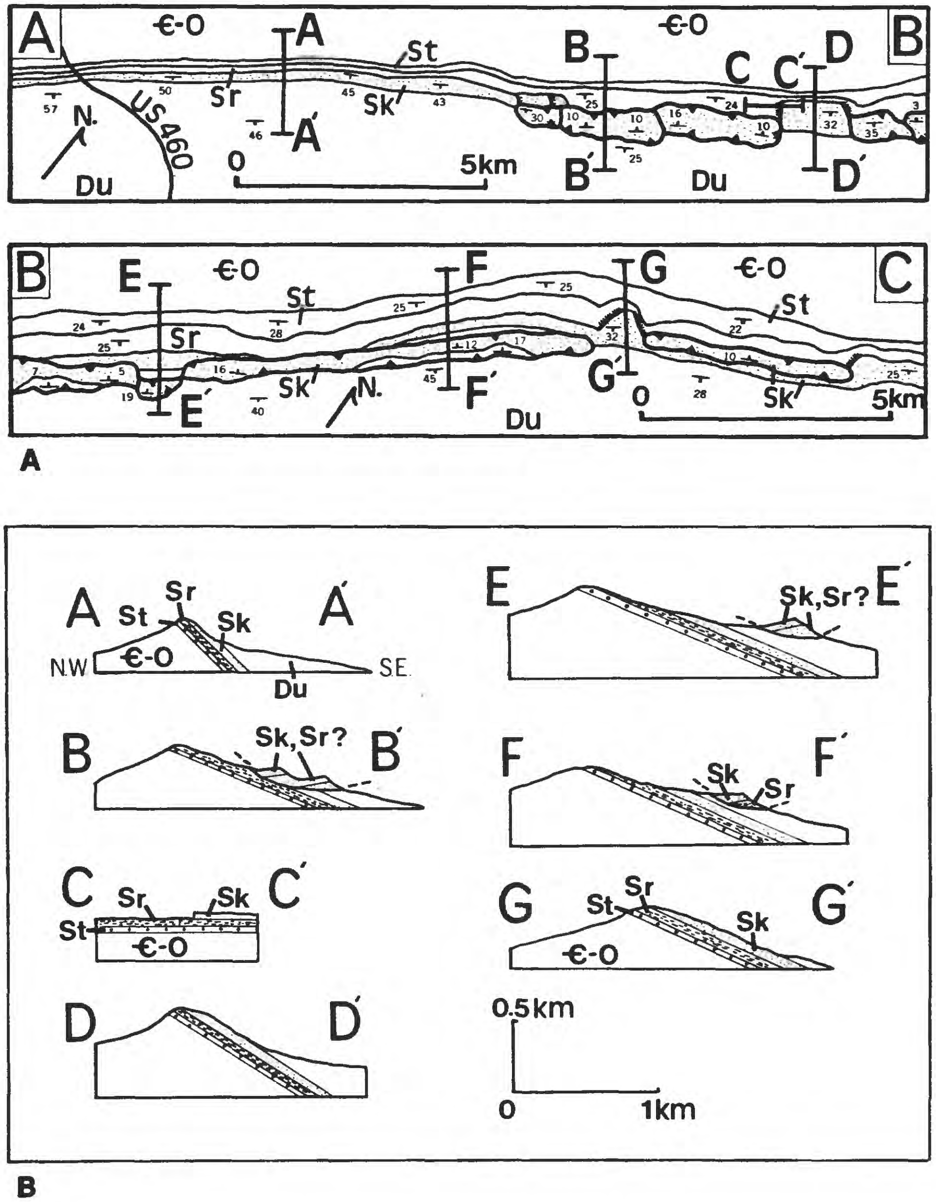


Figure 7. A. Geologic map and cross sections across Sinking Creek Mountain. Dev. = Devonian formations undivided; Sk = Middle Silurian Keefer Sandstone; Sr = Middle Silurian Rose Hill Formation; St = Lower Silurian Tuscarora Quartzite; €-O = Cambrian and Ordovician formations. Heavy lines with sawteeth show rockslide boundaries, teeth on upper plate; heavy lines with ticks = breakaway scarps, ticks on slide-scar side. B: Cross sections of Sinking Creek Mountain; symbols as in A.

Schultz, 1986

Figure 7. Sketch map of bedrock landslides and cross sections, Sinking Creek Mountain, southwest Virginia.

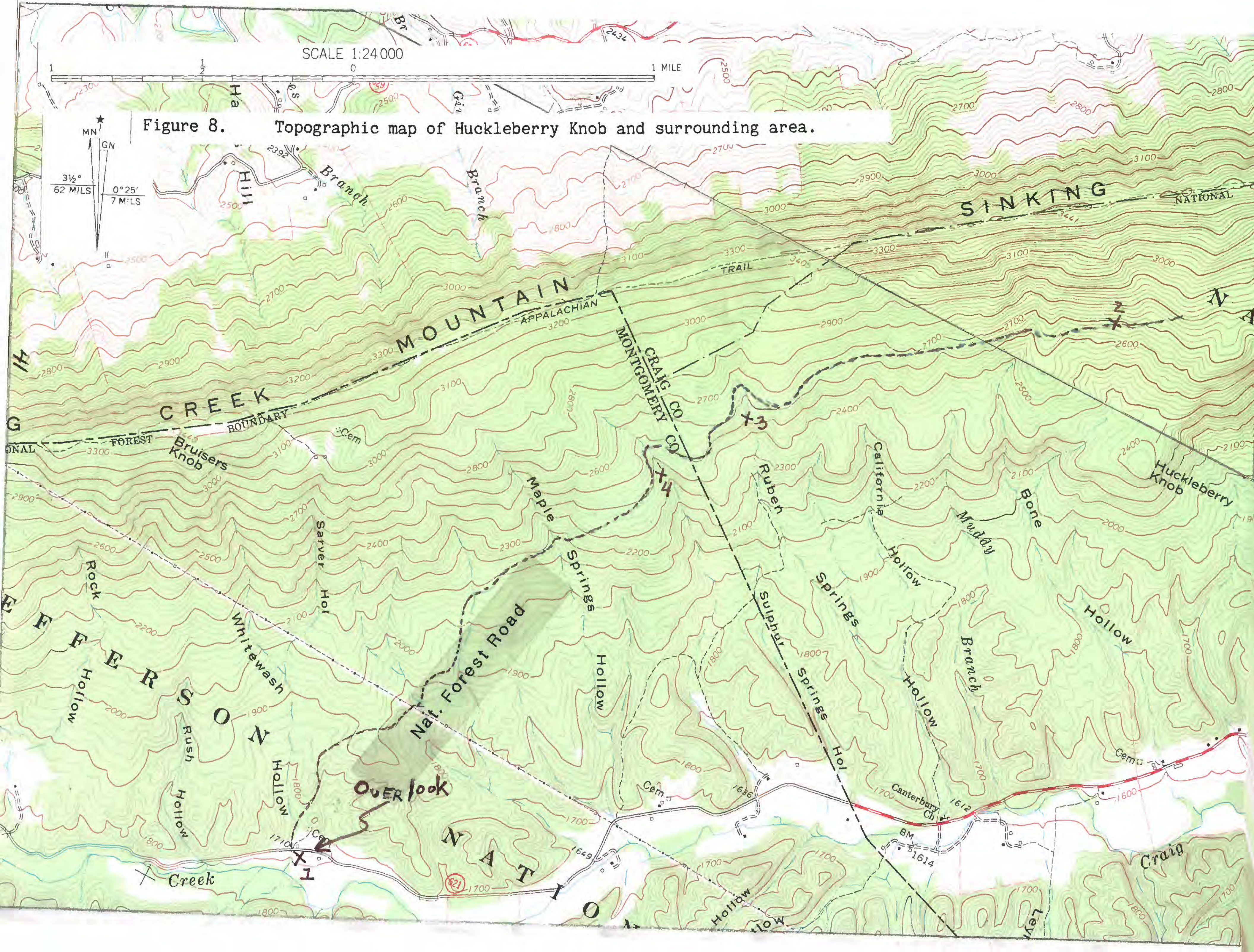


Figure 8. Topographic map of Huckleberry Knob and surrounding area.

Figure 9. Geologic sketch map of area shown in Figure 8.

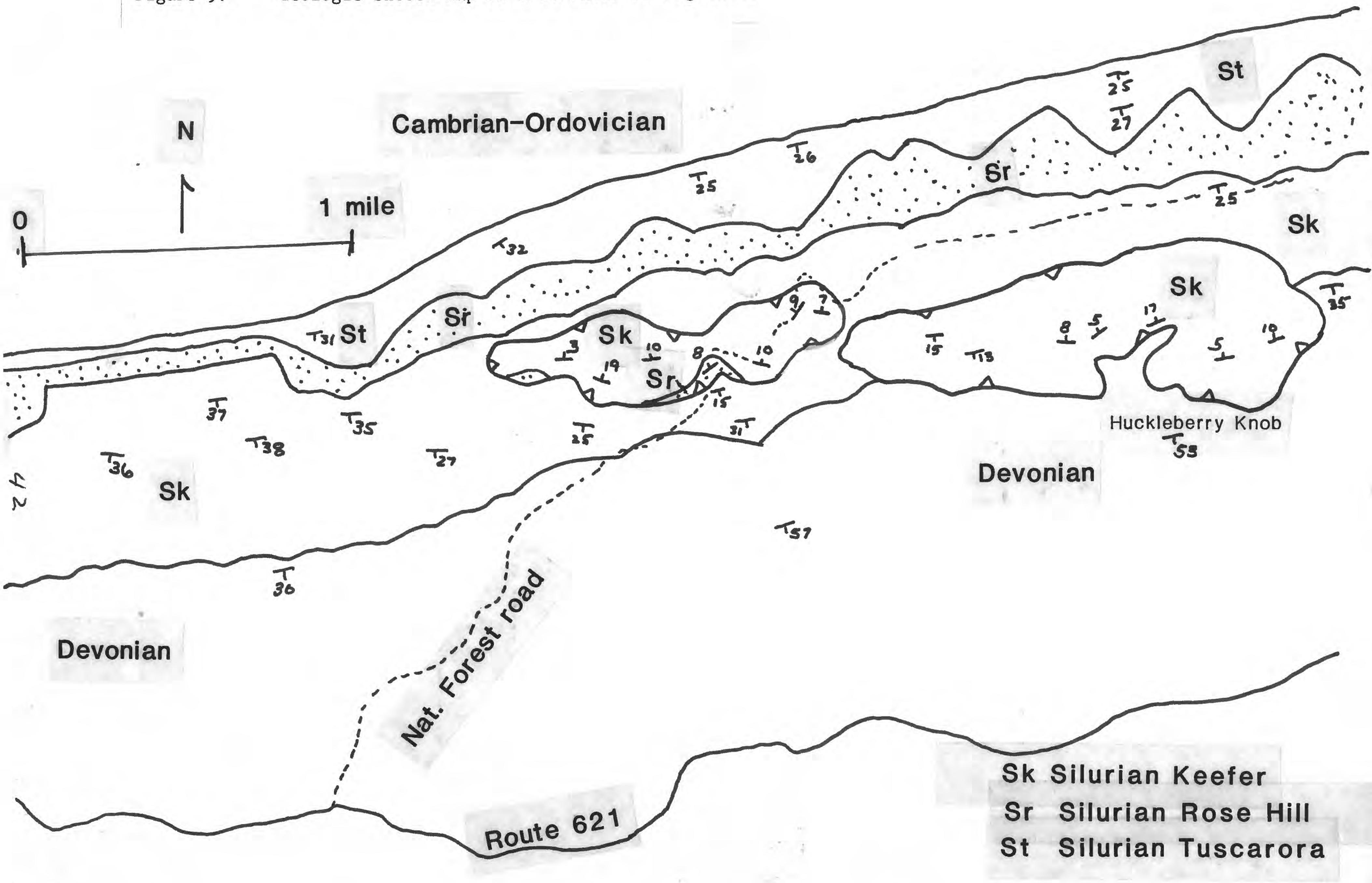
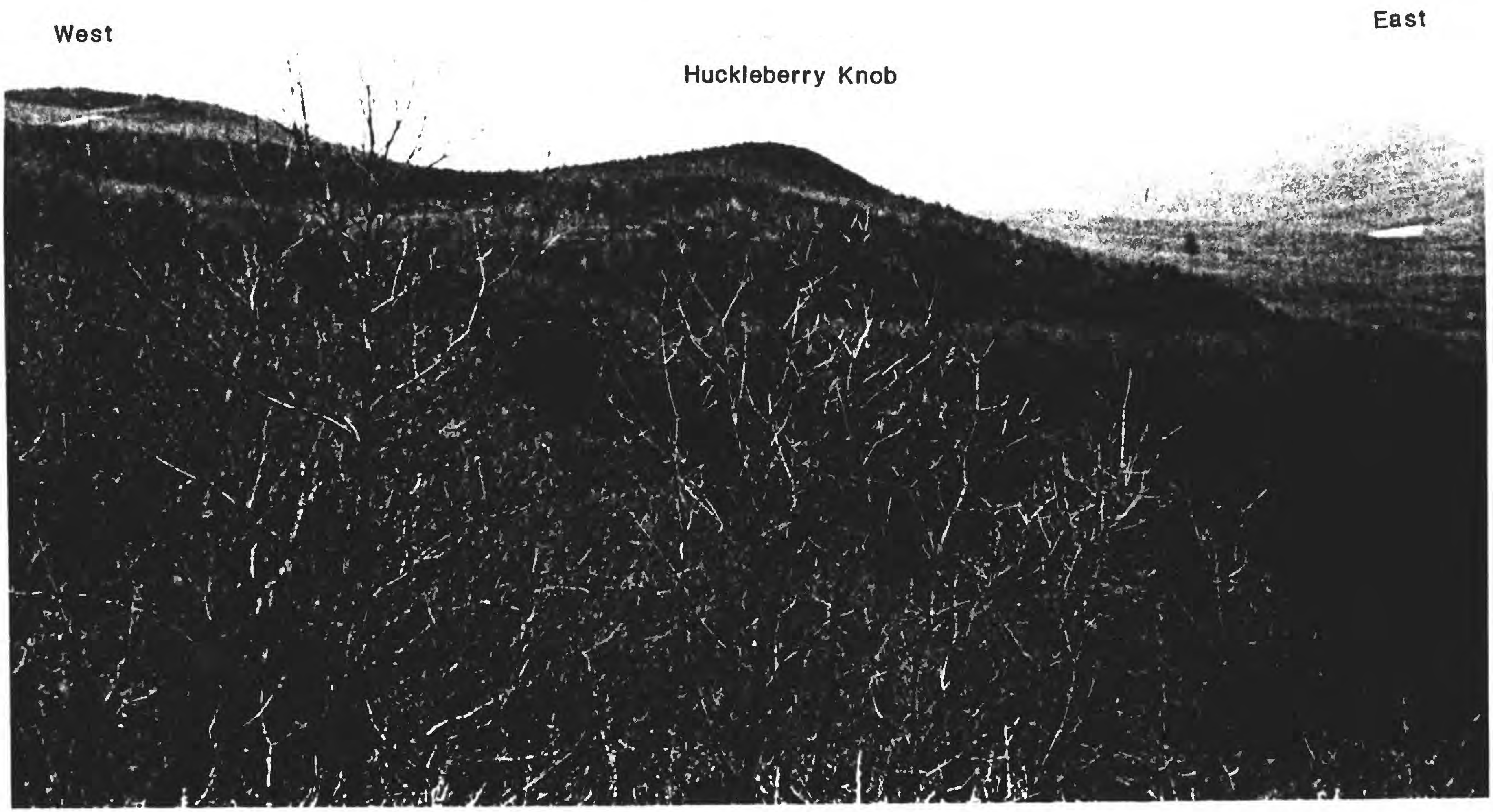


Figure 10. Photograph of Huckleberry Knob as seen from overlook on landslide escarpment.



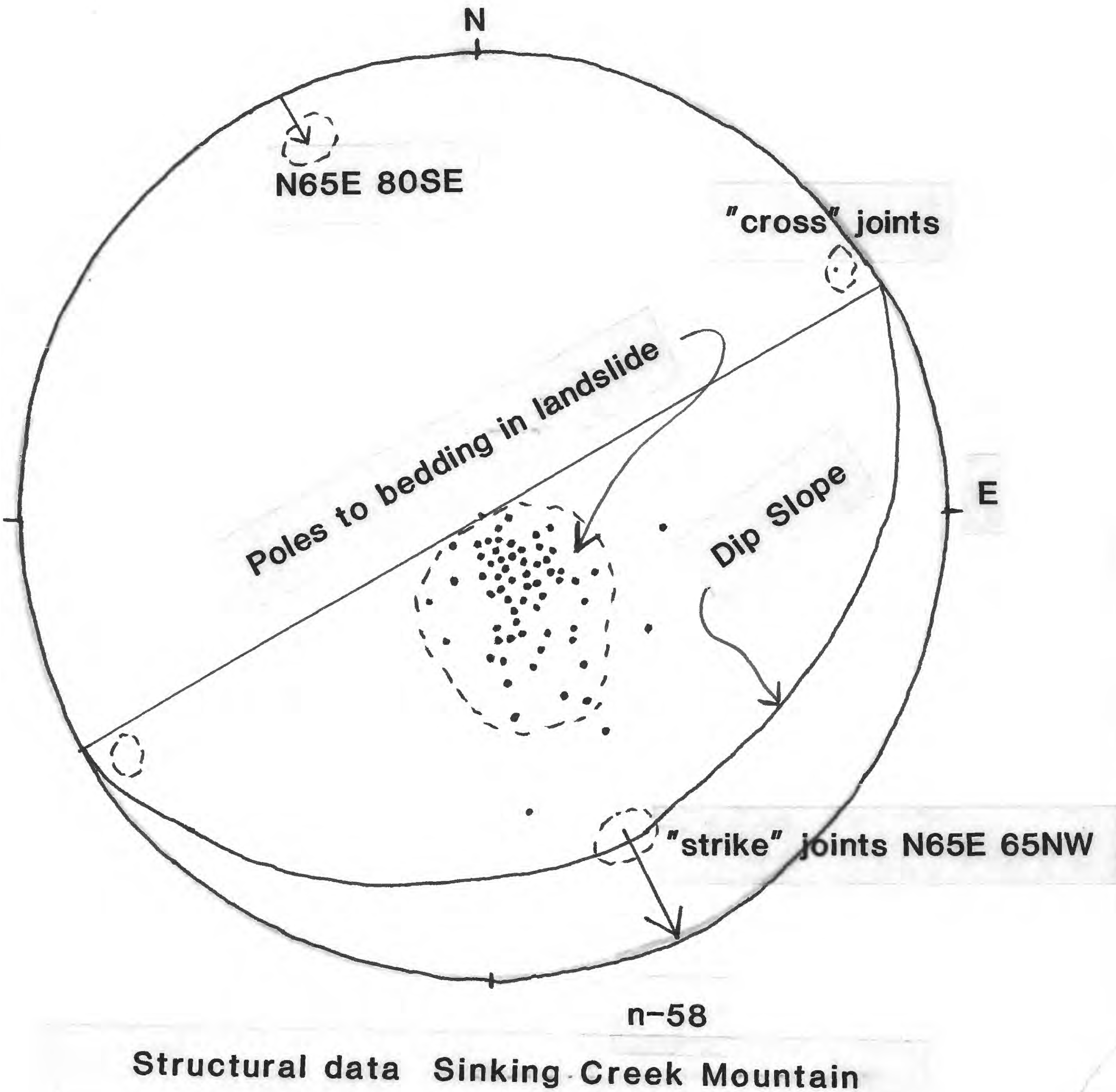


Figure 11. Poles to bedding shown on dip slope plane with joint orientations shown. Equal area net.

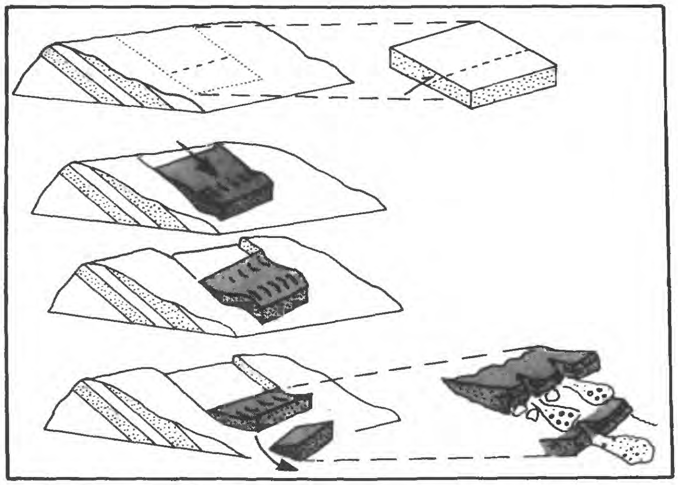


Figure 1Ꝣ. Model of rock block slide and rock slump emplacement. Failed block is about 2 km long across frontal scarp.

Schultz and Southworth, 1989

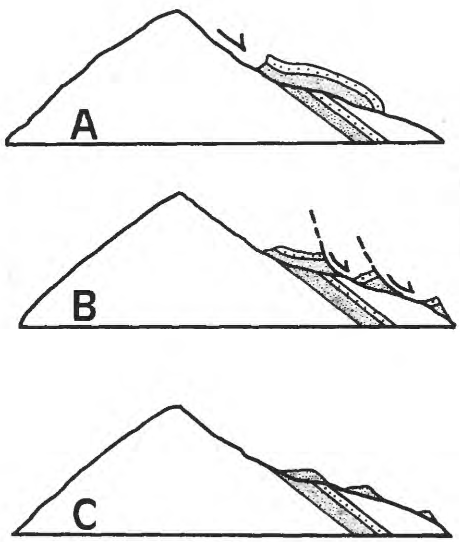


Figure 1Ꝣ. Generalized model (in part modified from Braddock, 1978) for Sinking Creek Mountain rock slides. A: Detachment and downslope transport. B: Slumping of front part of slide. C: Present-day topography after erosion.

Schultz, 1986

Figure 12. Model for giant bedrock landslide emplacement.

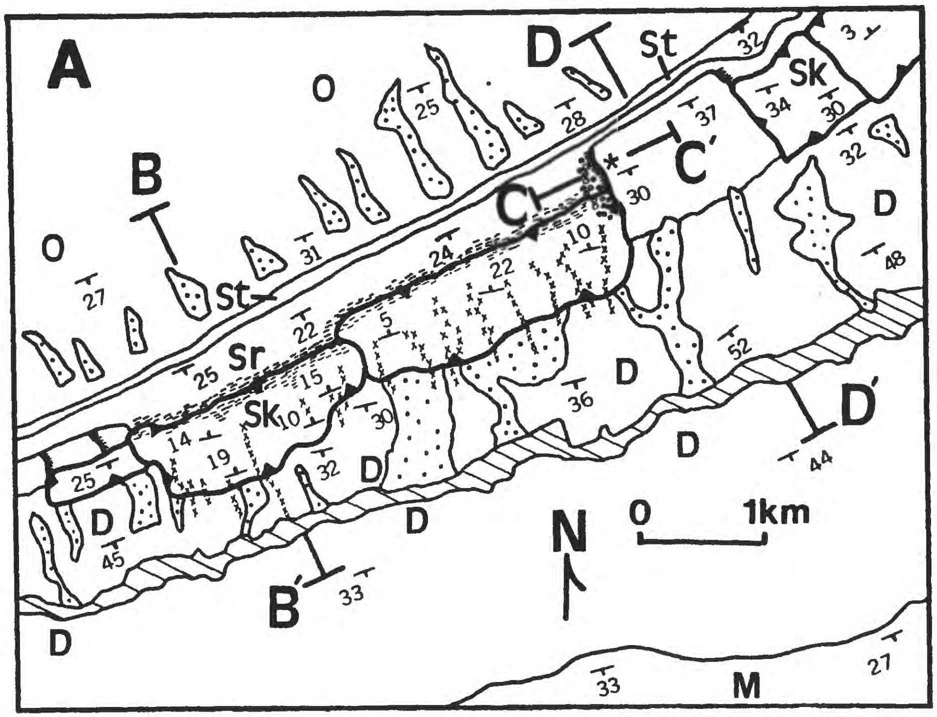


Figure13. A: Detailed geologic map of part of slide complex; cross sections as in Figure 2A. O = Ordovician formations; D = Devonian formations; other formation letters as in Figure 2. X = boulder stream and boulder field colluvium derived from rockslide escarpments; open circles = boulder stream and boulder field colluvium derived from lateral breakaway scarp; dashes = colluvium derived from slide scar (Rose Hill float); dots = colluvium present prior to major rockslide emplacement; diagonal rules = recent alluvium of valley bottom. B: High-altitude air photograph of part of slide complex shown in Figure 3A. Asterisk in photo and in geologic map are in same position, next to east-west power line and just east of breakaway scarp.

Tuckaway site

→o

Schultz, 1986

Figure 13. A. Geologic map of landslide complex near Camp Tuckaway.

B. Air photo of area shown in A.

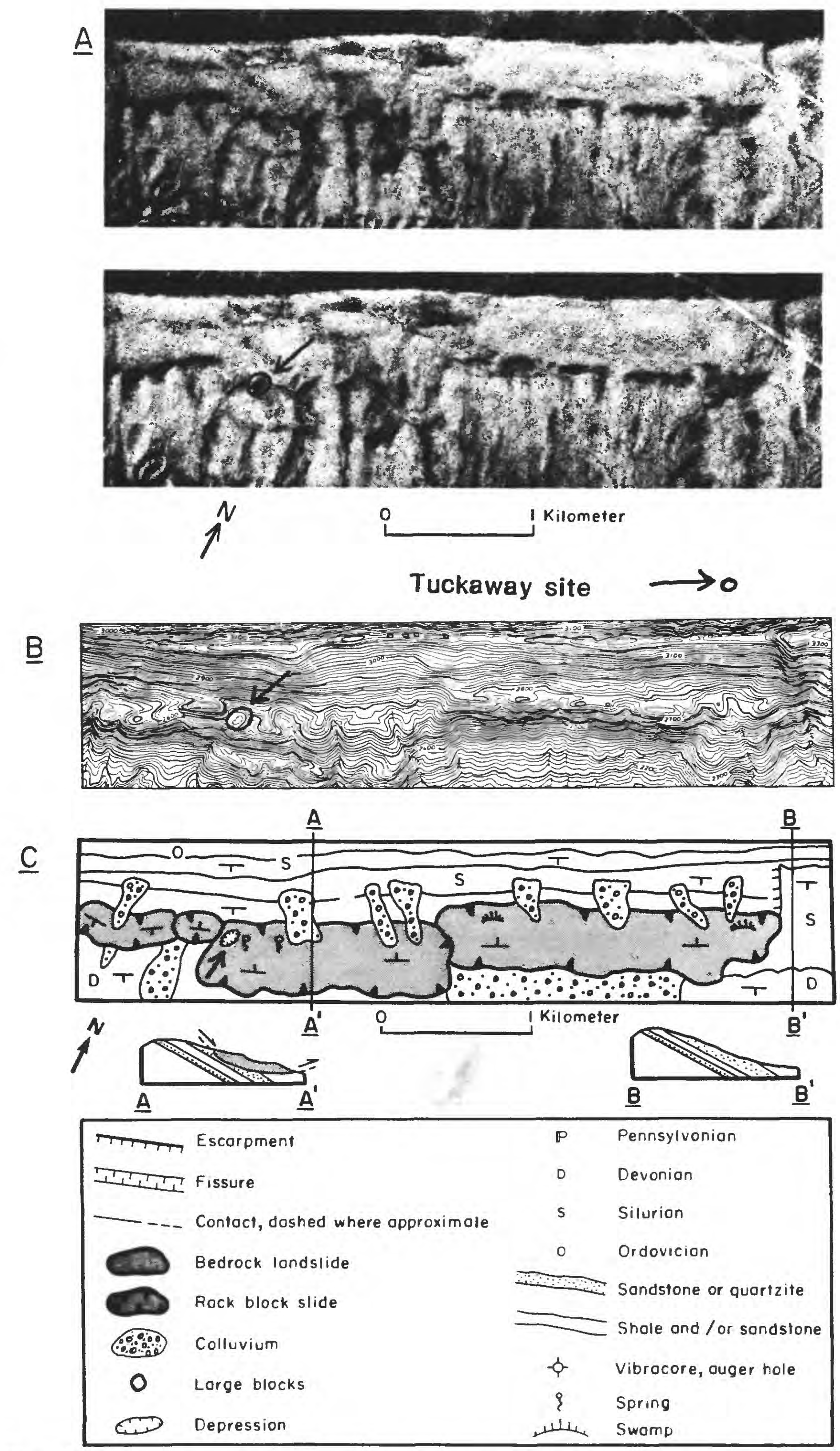


Figure14 Rock block slides and rock slumps on Sinking Creek Mountain, Virginia. A. Stereo aerial photographs. B. Topographic map. C. Geologic map and cross sections.

Schultz and Southworth, 1989

Figure 14. Stereo airphotos, topographic map and geologic map of rockslides near Camp Tuckaway.

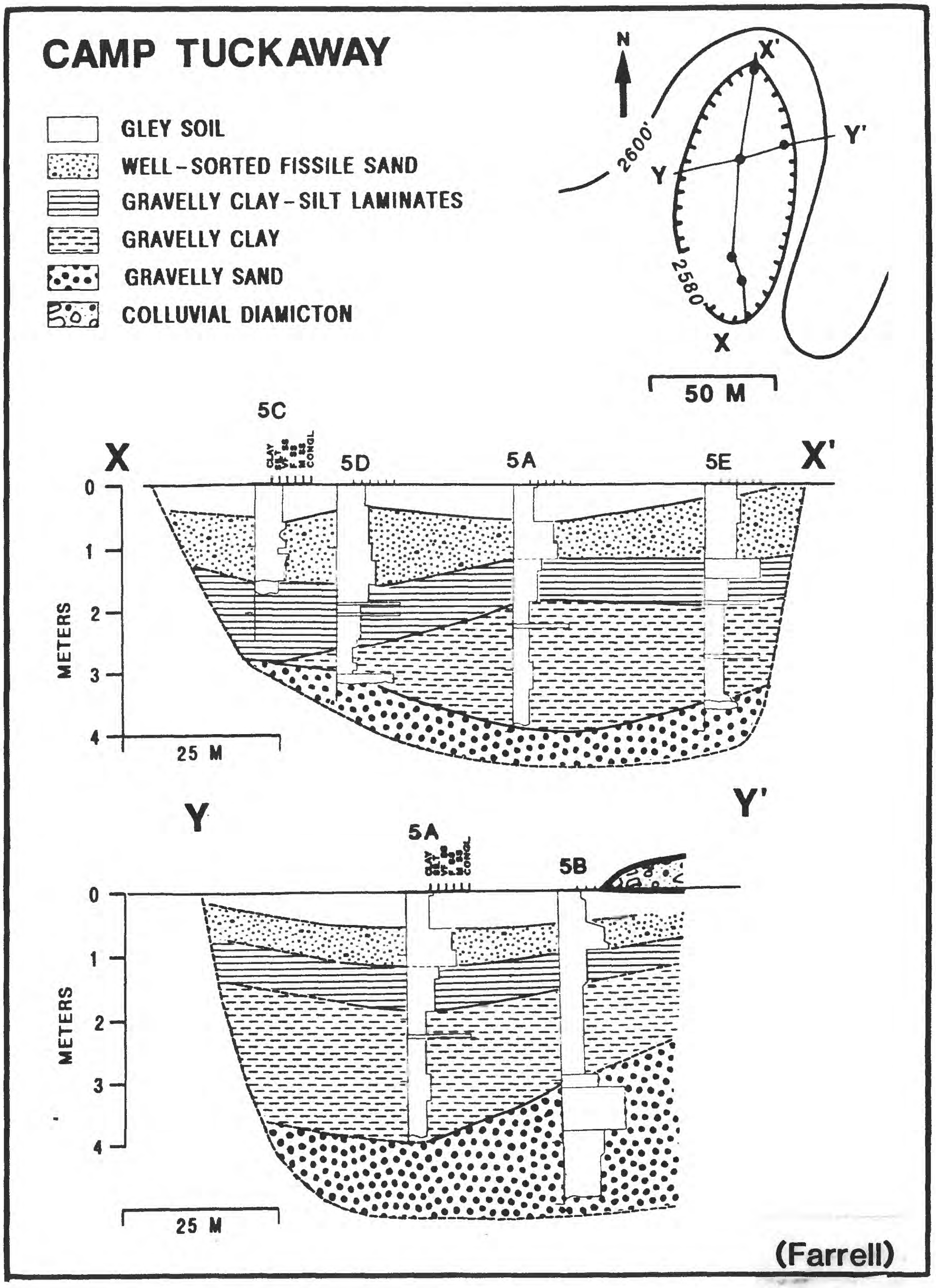


Figure 15. Hand augered bore hole sites and cross sections at Camp Tuckaway sag pond locality.

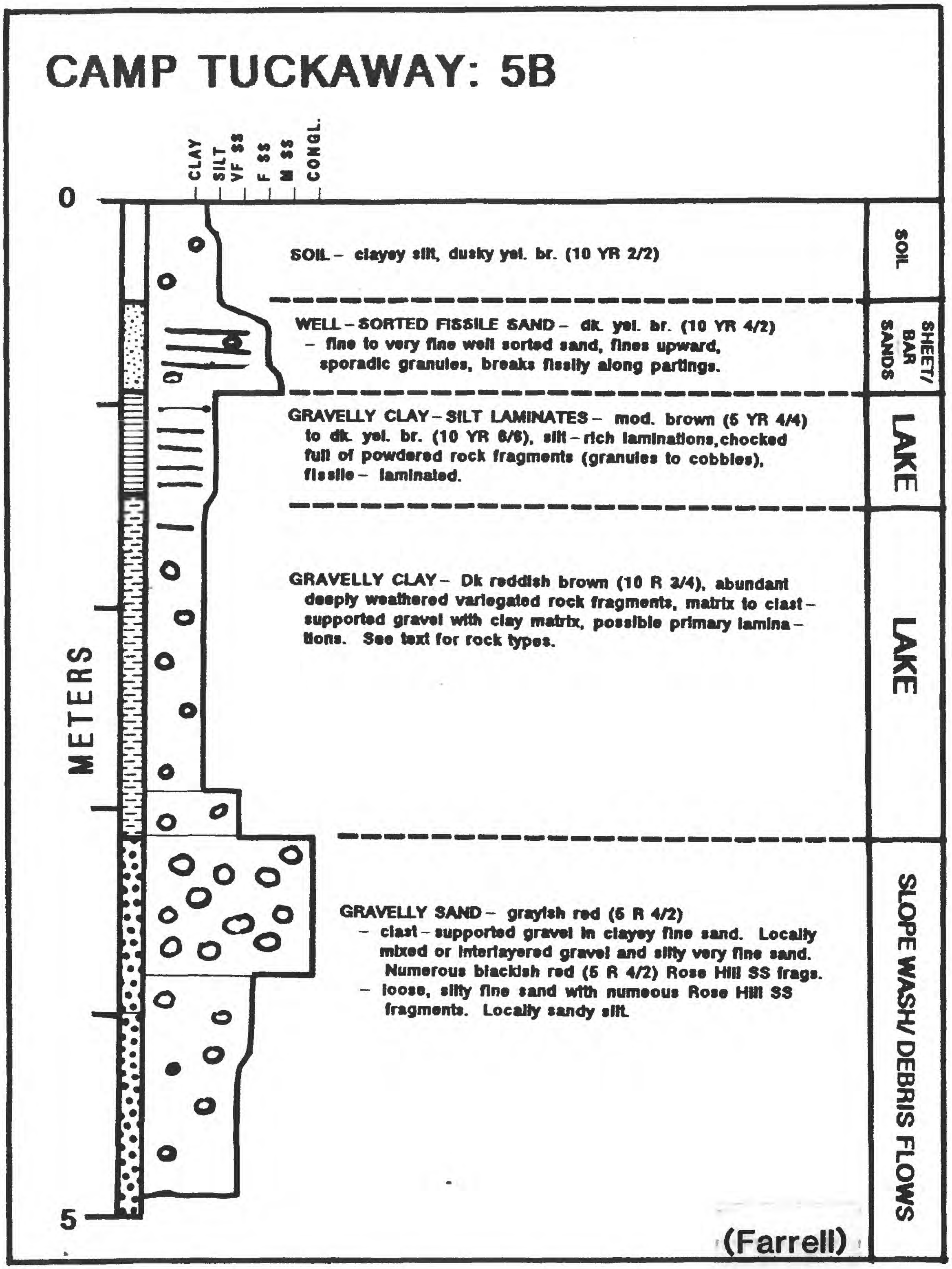


Figure 16: Graphic log for site 5B at Camp Tuckaway locality.
5B is a hand-augered bore hole on cross-section Y-Y' in Figure 3.

6. Graphic log for site 5B at Camp Tuckaway locality.

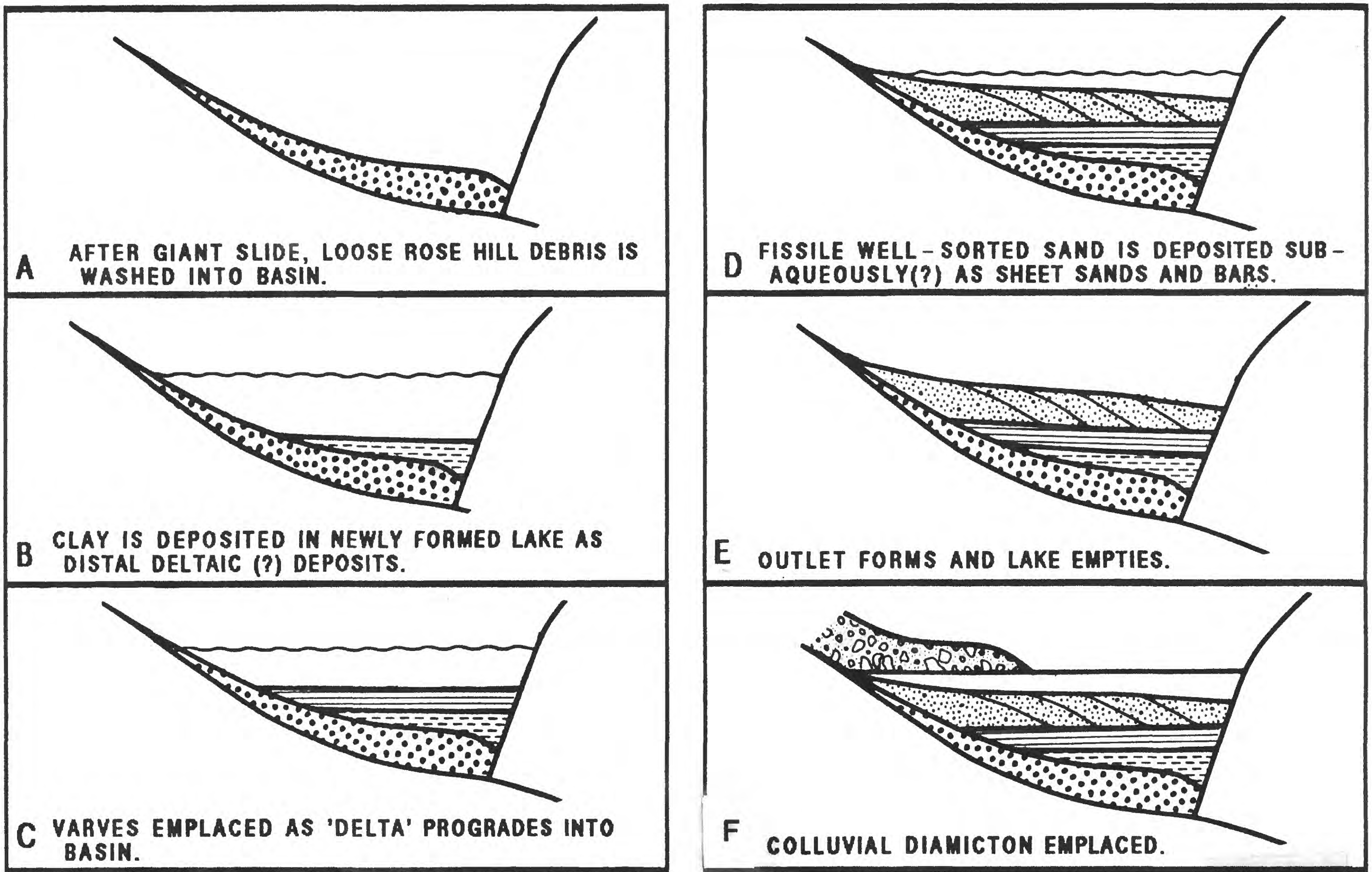


Figure 17. History of sag pond infill at Camp Tuckaway locality.

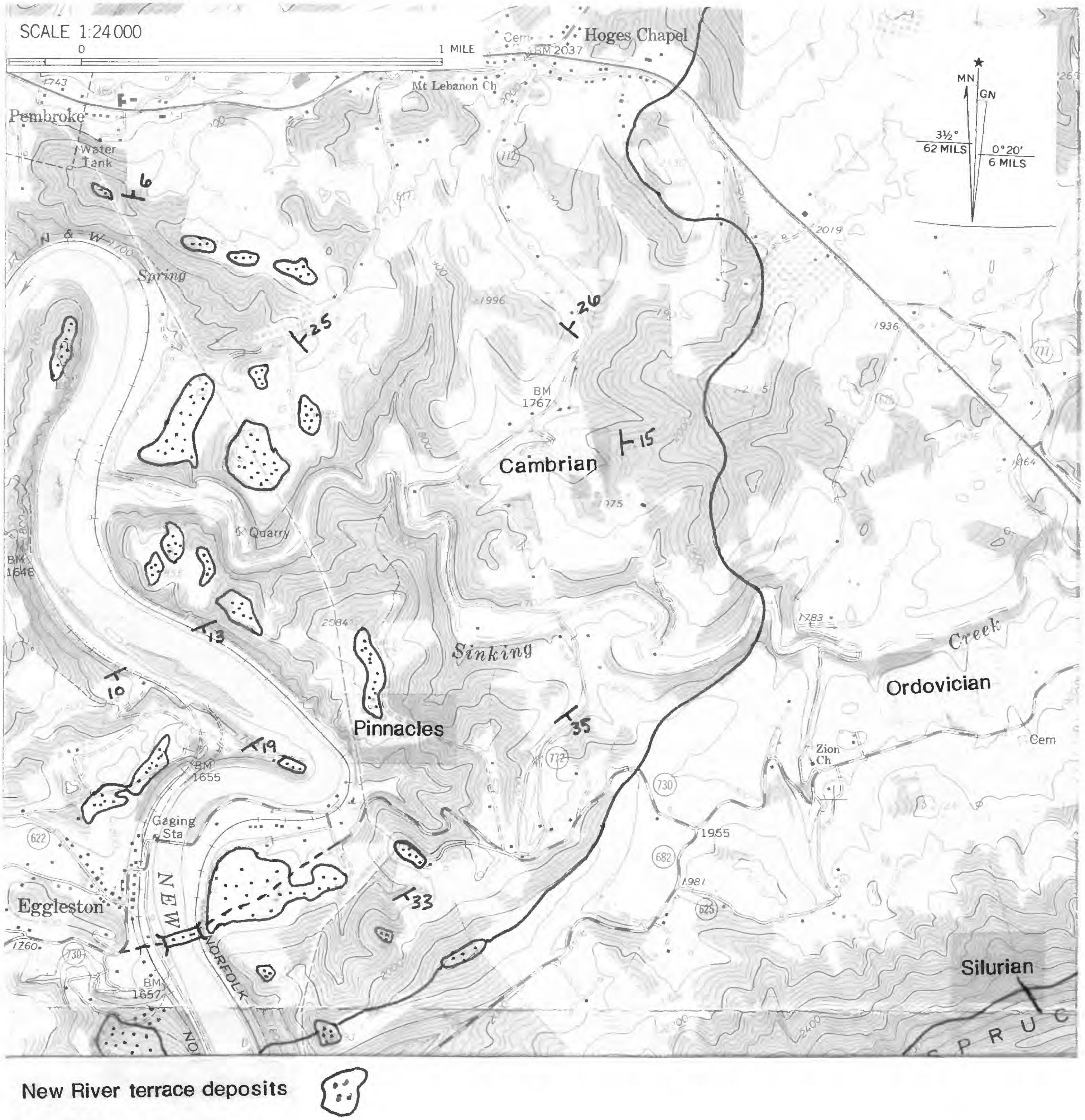


Figure 18. Simplified geologic map showing terrace deposits and topography at Eggleston, Virginia.

Figure 19. Photograph of pinnacles at Eggleston, Virginia.

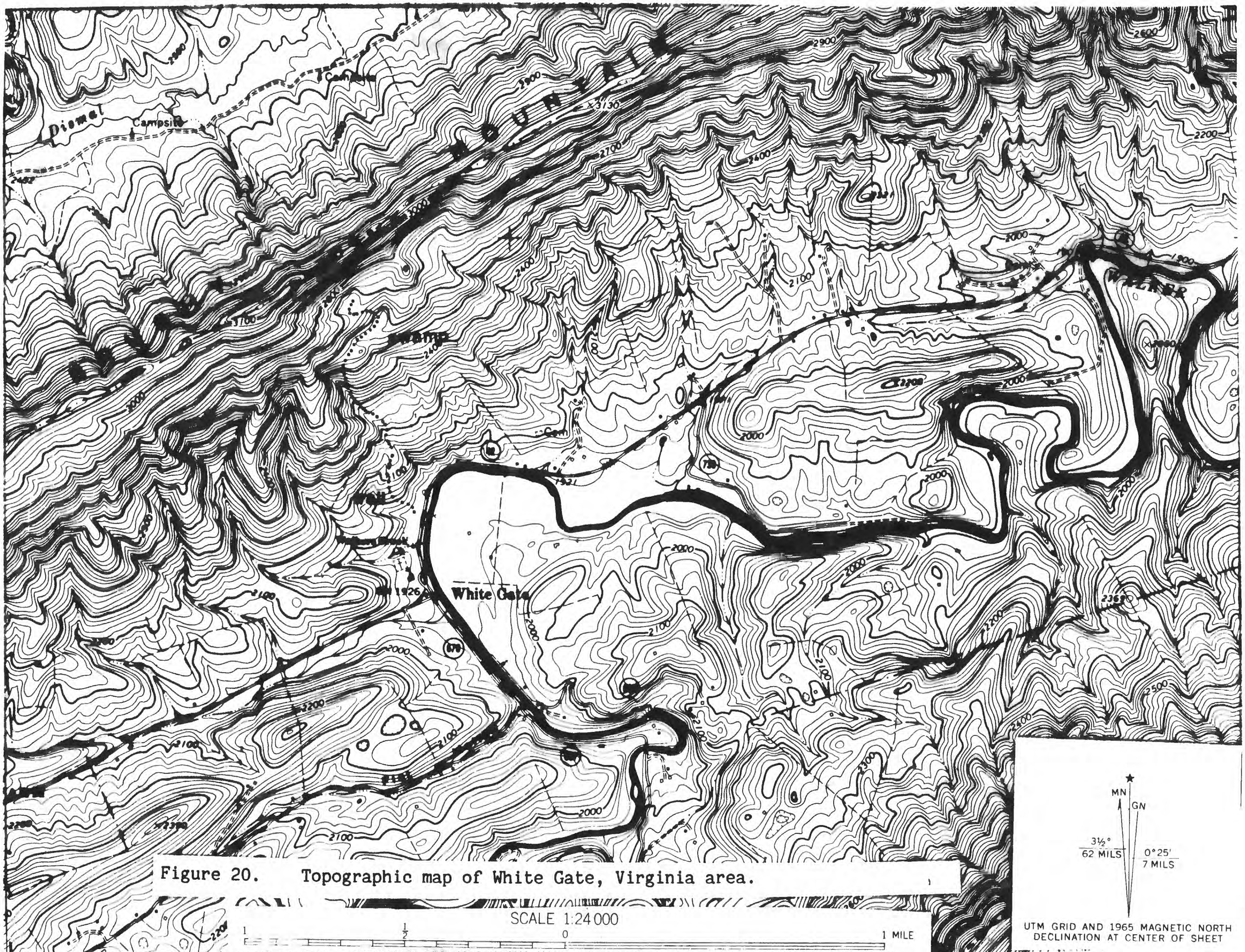


Figure 20. Topographic map of White Gate, Virginia area.

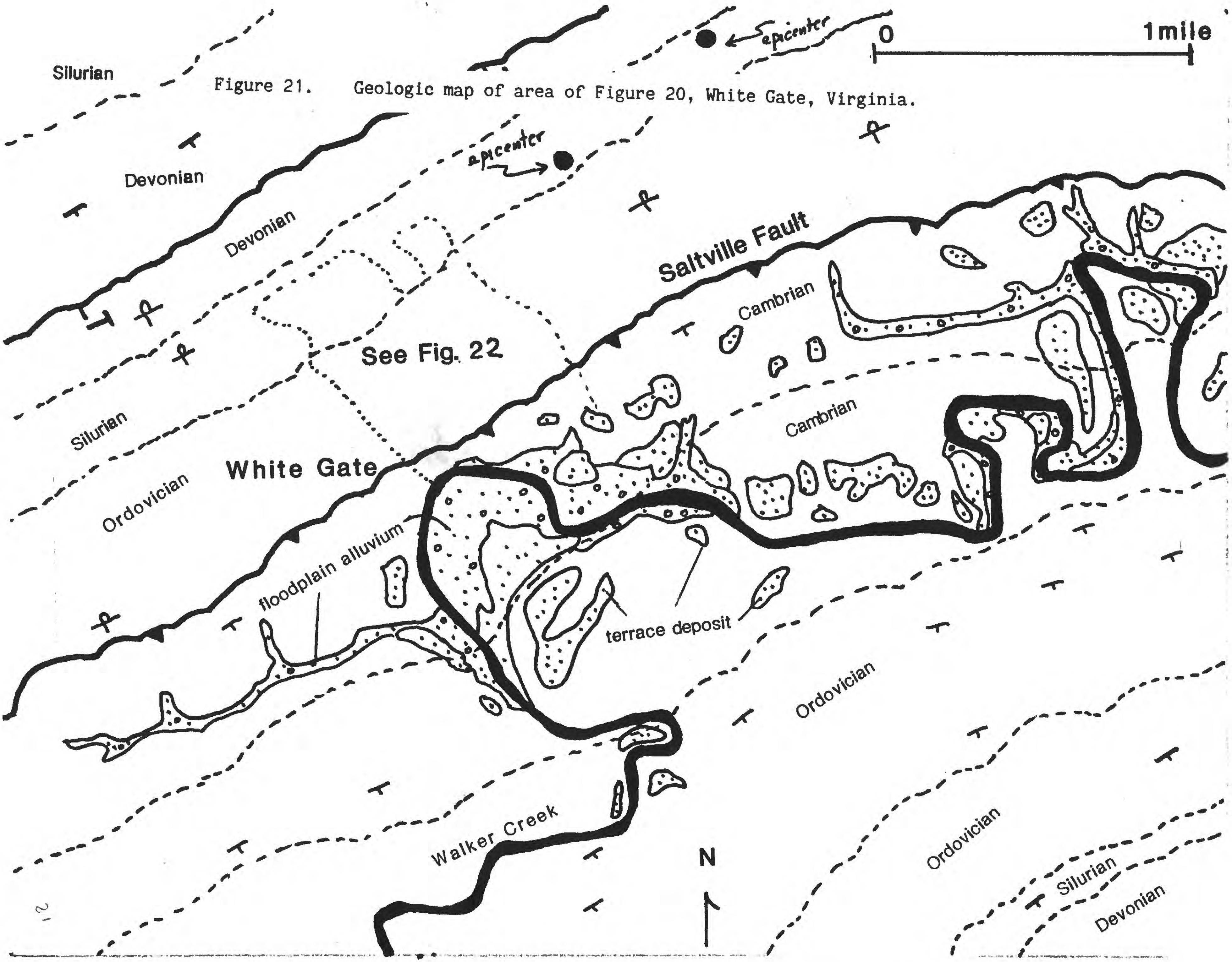


Figure 21. Geologic map of area of Figure 20, White Gate, Virginia.

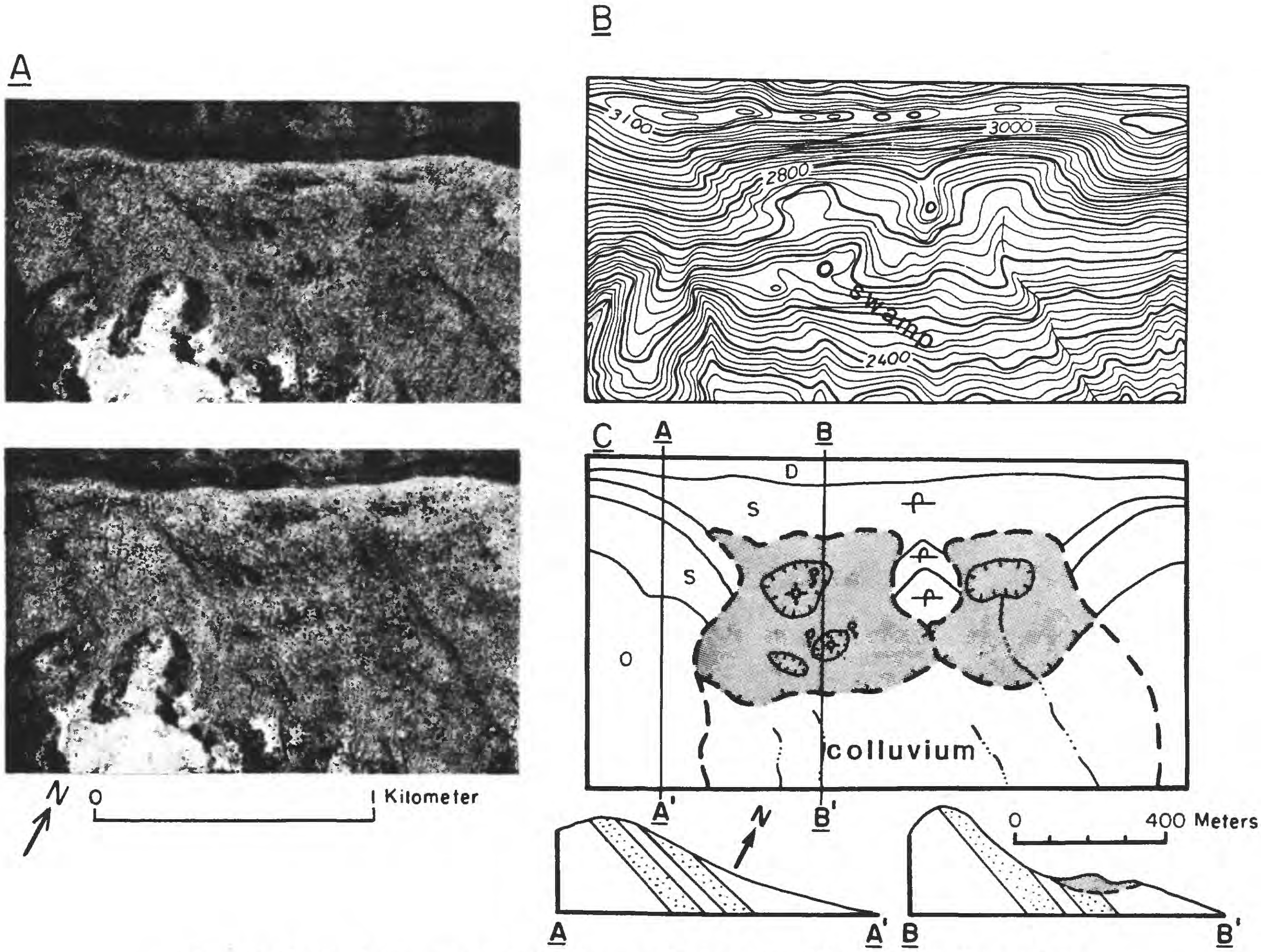


Figure 16. Erosional remnants of paired sackung on dip slope about 1 km northeast of the sackung shown in Figure 15. A. Stereo aerial photographs. B. Topographic map. C. Geologic map and cross sections. Symbols as in Figure 3.

Schultz and Southworth, 1989

Figure 22. Stereo aerial photographs, topographic map, geologic map, and cross sections of slope failure at White Gate, Virginia.

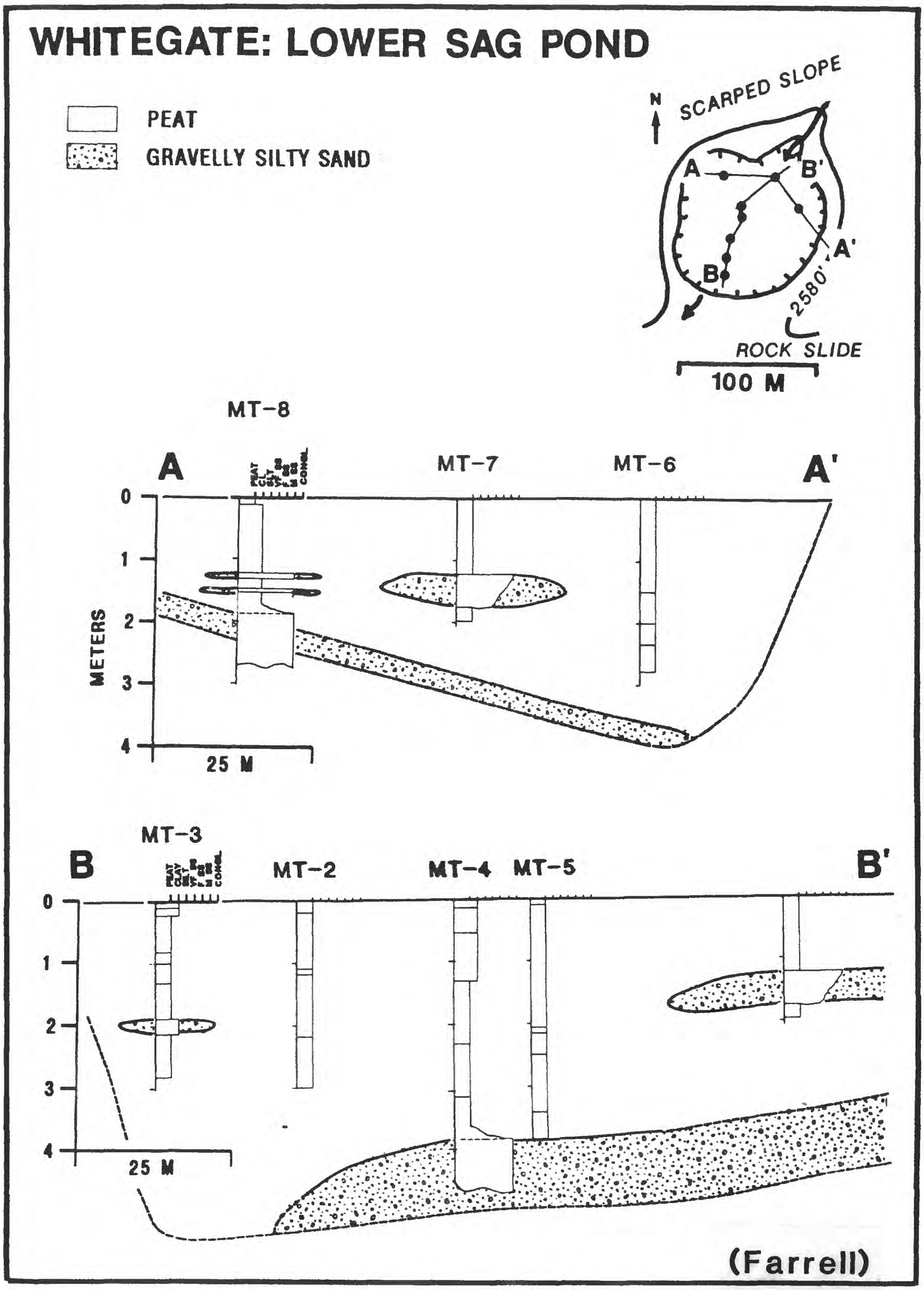


Figure 23. Vibracoring sites and cross sections for lower sag pond at White Gate locality.

23

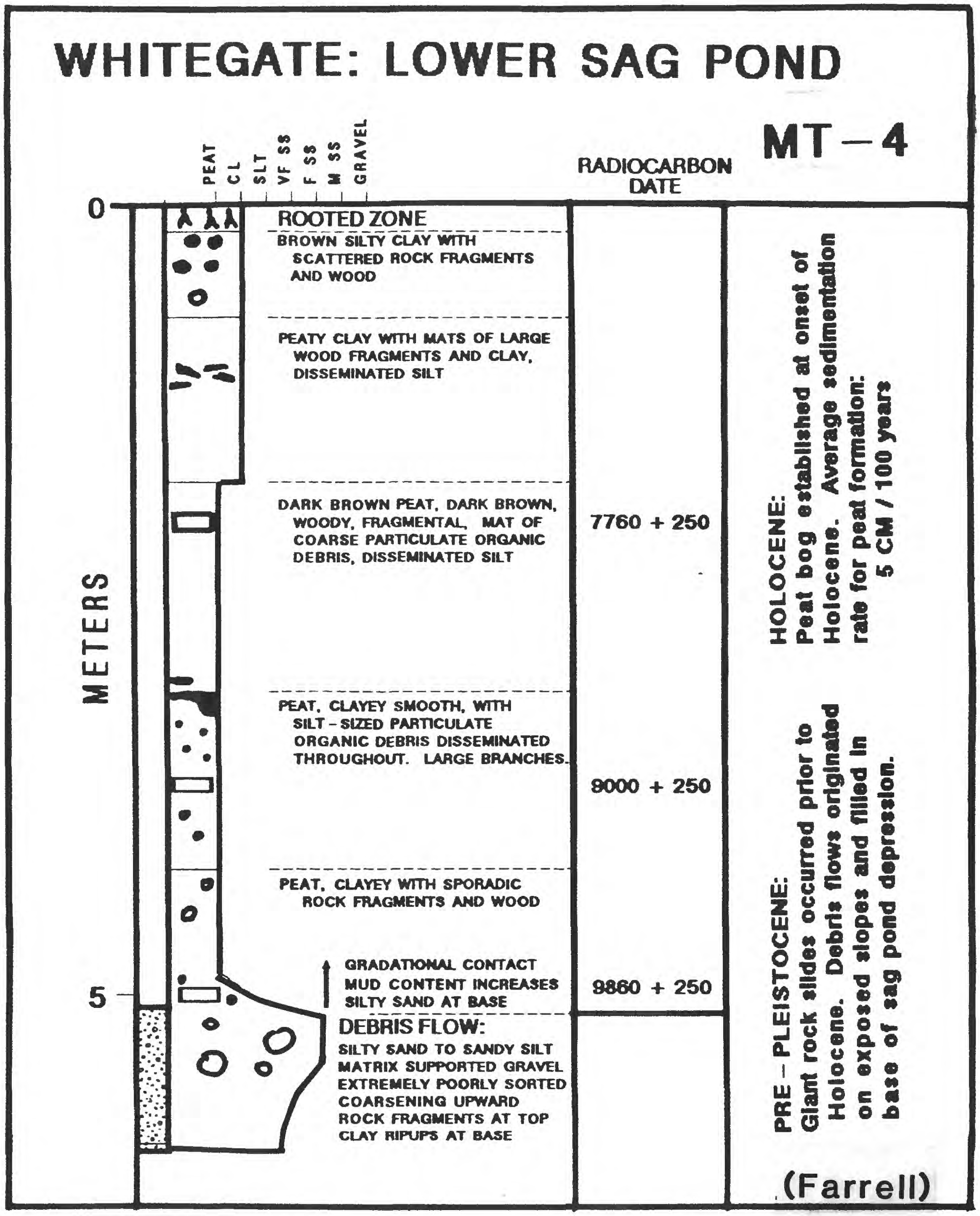
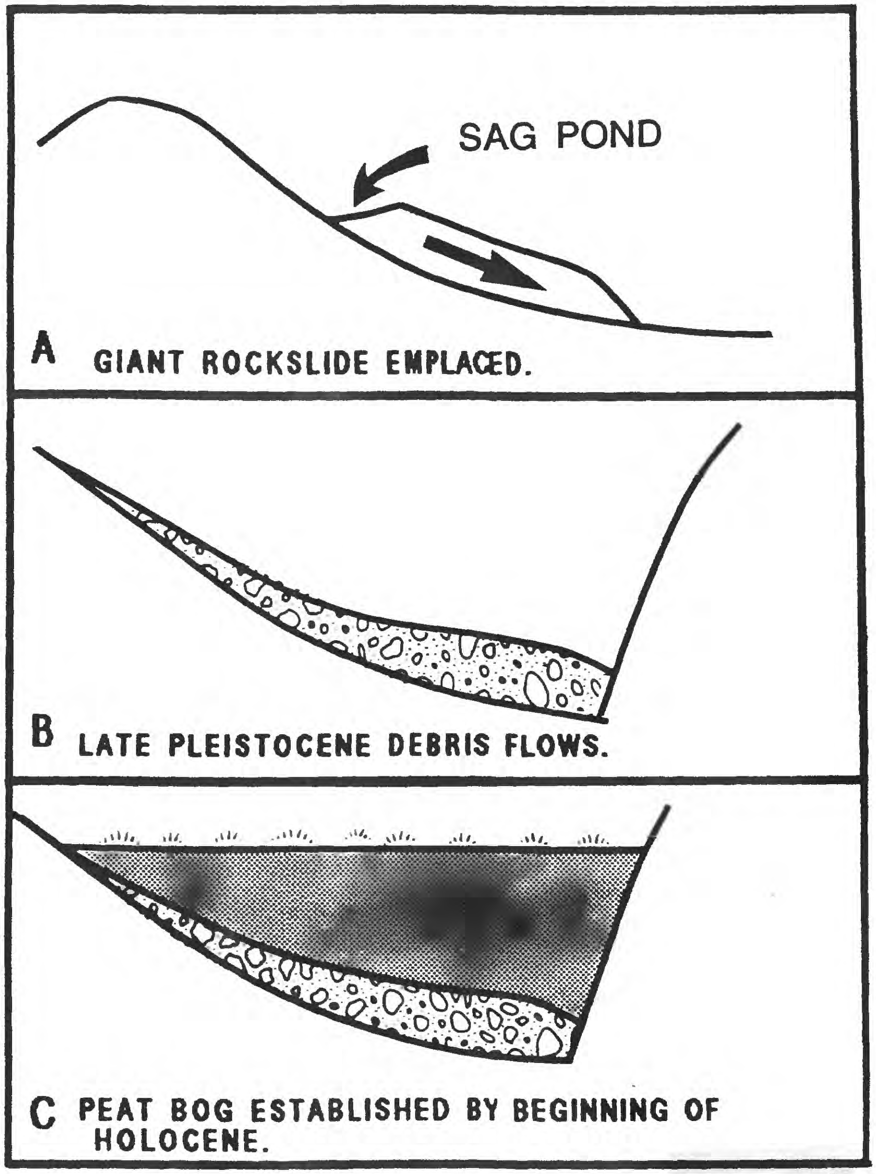


Figure 24: Graphic log for MT-4, vibracore from the lower sag pond at the Whitegate locality. MT-4 is part of cross-section B in Figure 2.

Figure 24. Graphic log for MT-4 vibracore from lower sag pond at White Gate locality.

24



(Farrell)

Figure 25. Geologic history of the lower sag pond infill at the White Gate locality.

25

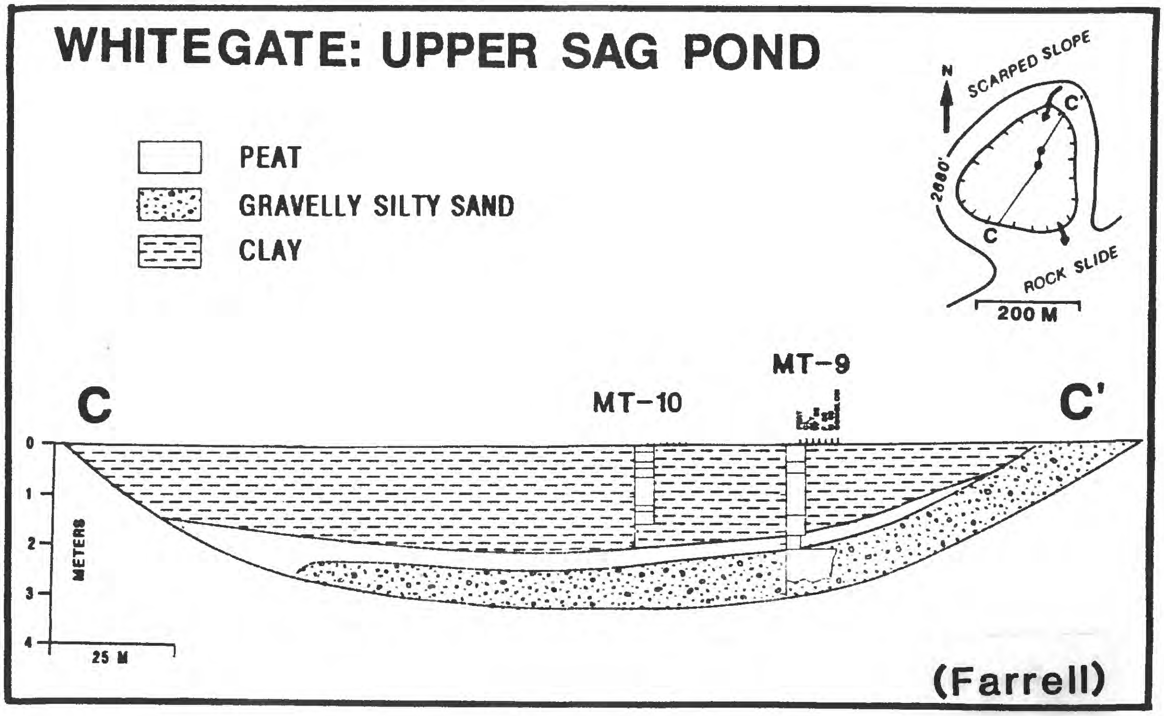


Figure 26. Vibracoring sites and cross section for the upper sag pond at the White Gate locality.

26

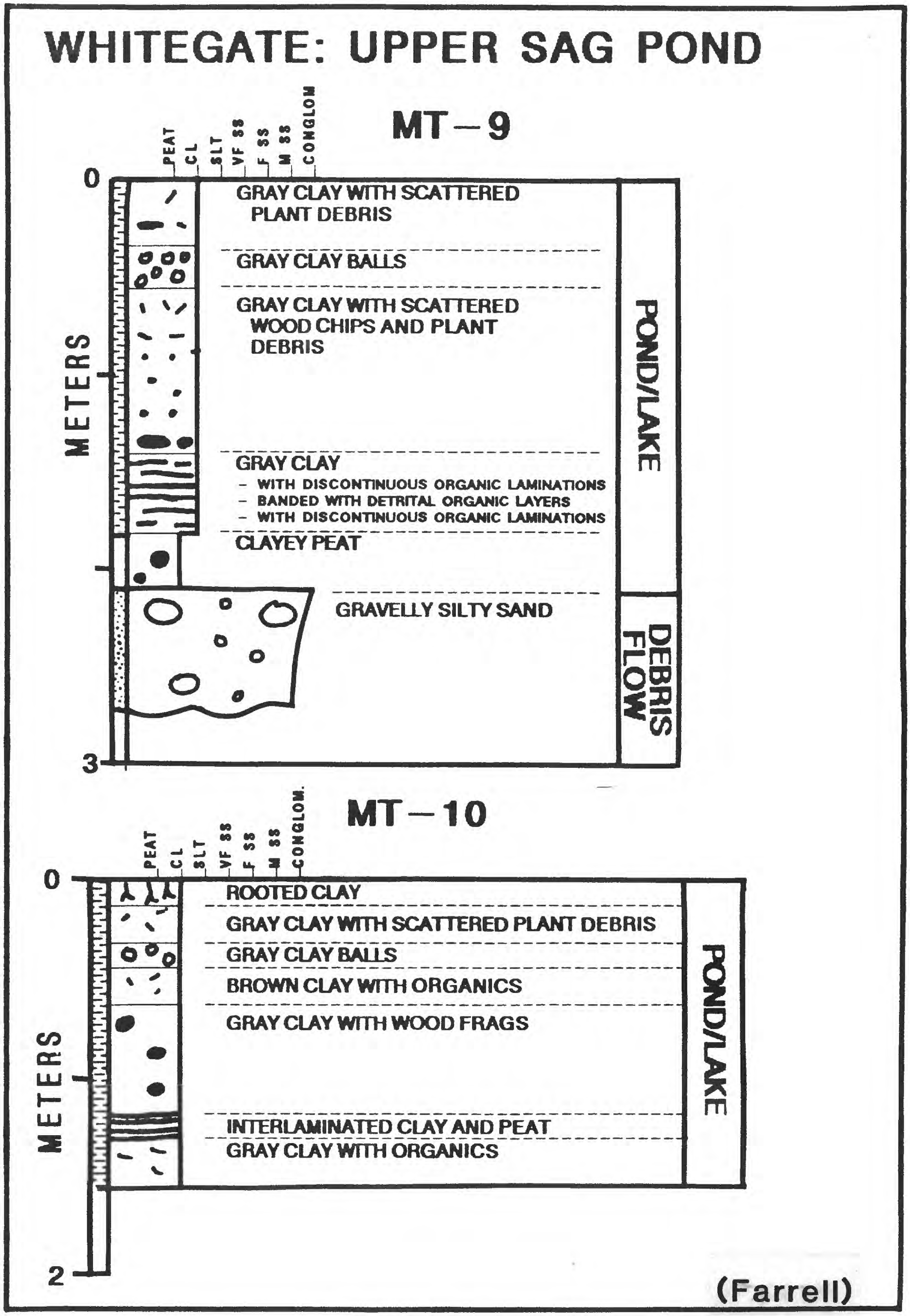


Figure 27. Graphic log for vibracores taken in the upper sag ponds at the White Gate locality.

27

FIGURE 28—Site 100 boulder stream (type B1). Exposure reveals a colluvium-filled trough eroded into Martinsburg Formation residuum (partly obscured by slumping) beneath the surface boulders. An outline of the trough cross section is shown in figure 38. The thickness of the surface boulder layer is greater than appears here, for most of the boulders at the top of the cut have fallen down. Site location is shown on figure 2.

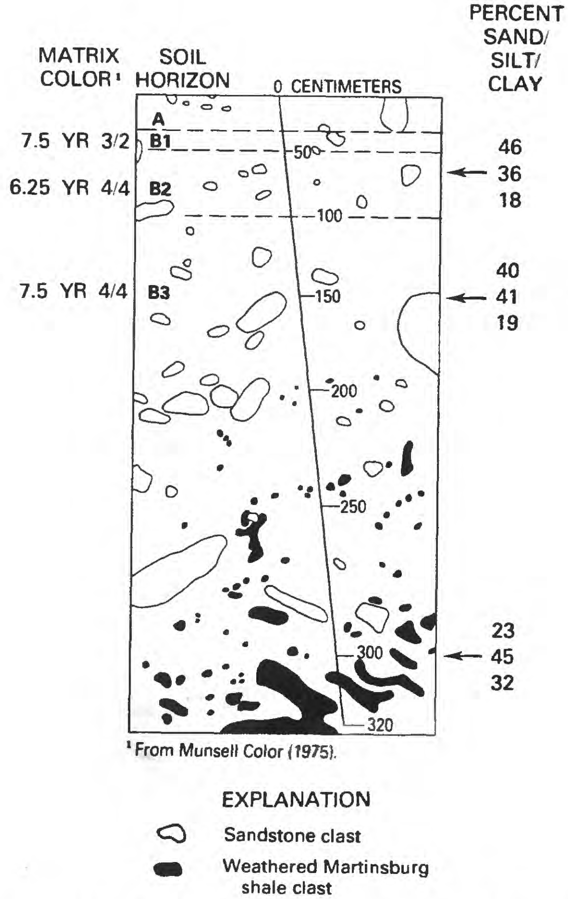


FIGURE 28—Close-up of vertical section on right side of trough cross section at site 100 (fig. 2). This sketch was made from a photograph taken at an angle, so that vertical scale is compressed toward the top. The oblique scale was drawn from a tape-measure included in the photograph. The B2 horizon, defined on the basis of color (reddest hue), occurs approximately between 50 cm and 100 cm on the tape. Above the 200-cm mark there are no shale clasts; such clasts increase downward until, below the 300-cm mark, they are the only ones present.

Mills, 1988

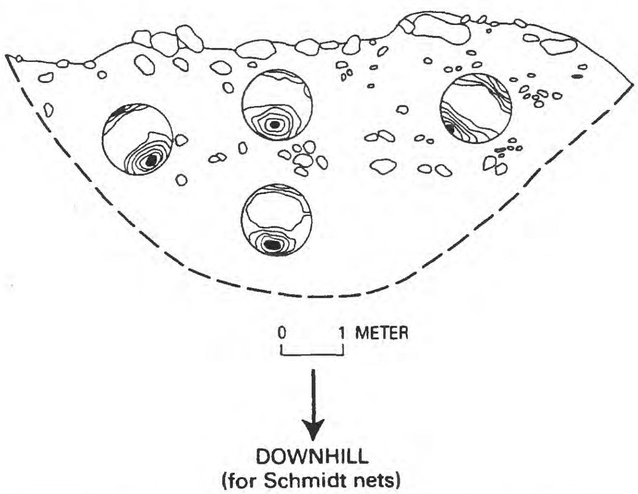


FIGURE 28.—Cross section of trough beneath site 100. This cross section can be discerned vaguely in figure 34. Each Schmidt net shows the fabric pattern based on long axes of clasts in colluvium at that location. All nets have been rotated so that the downhill direction (parallel to the long axis of the boulder stream) is toward the bottom.

Figure 28. Photograph, vertical section, and cross section of boulder stream on west side of Gap Mountain.

Photo 3. Cut through nose on Gap Mountain in Newport quadrangle, exposing concave-up contact (dashed line) between colluvium and underlying weathered shale. Maximum displayed thickness of colluvium is 5 m.

Mills, 1987

Figure 29. Photograph of roadcut through nose on Gap Mountain.

Figure 30. Simplified bedrock map of Mountain Lake area.

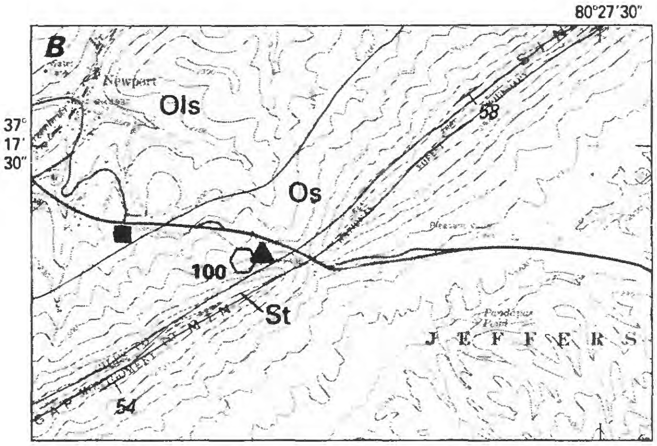


FIGURE 2.—Simplified bedrock map of study area, modified from Eckroade (1962). On the north flank of Big Mountain, contacts are so closely spaced that formations have been combined. Locations of study sites are shown; only those sites referred to individually in the text are numbered. Often, multiple samples were collected at one site. Sites in the Newport quadrangle, which is outside the study area, are shown on the inset map. A few sites in the Eggleston and Interior quadrangles were outside the map area shown here.

EXPLANATION

Dm	Millboro Shale (Devonian)
Dh	Huntersville Chert (Devonian)
Drg	Rocky Gap Sandstone of Swartz (1929) (Devonian)
D	Devonian units, undivided—Includes Millboro Shale, Huntersville Chert, and Rocky Gap Sandstone
Sk	"Keefer" Sandstone (Silurian)
Srh	Rose Hill Formation (Silurian)
St	Tuscarora Quartzite (Silurian)
Ss	Silurian sandstones other than Tuscarora Quartzite—Includes "Keefer" Sandstone and Rose Hill Formation
Oj	Juniata Formation (Ordovician)
Ome	Martinsburg, Eggleston, and Moccasin Formations (Ordovician)
Ols	Ordovician limestones, undivided (Middle Ordovician)
Os	Ordovician sandstones, siltstones, and shales—Includes Juniata, Martinsburg, Eggleston, and Moccasin Formations

EXPLANATION OF MAP SYMBOLS

Contact—Approximately located

U
D
High-angle fault—Concealed. U, upthrown side; D, downthrown side

Anticline—Showing direction of plunge where known

Overturned anticline—Showing direction of dip of limbs

Syncline—Showing direction of plunge where known

Strike and dip of bedding

Inclined

⊕ Horizontal

TYPES OF SAMPLE LOCATIONS

● Alluvium

○ Type D hollows

▲ Type M hollows

▽ Side slopes

□ Upper noses

■ Lower noses

⬡ Boulder streams

Mills, 1988

Figure 31. Map explanation and simplified geologic map of Gap Mountain along State Route 460.

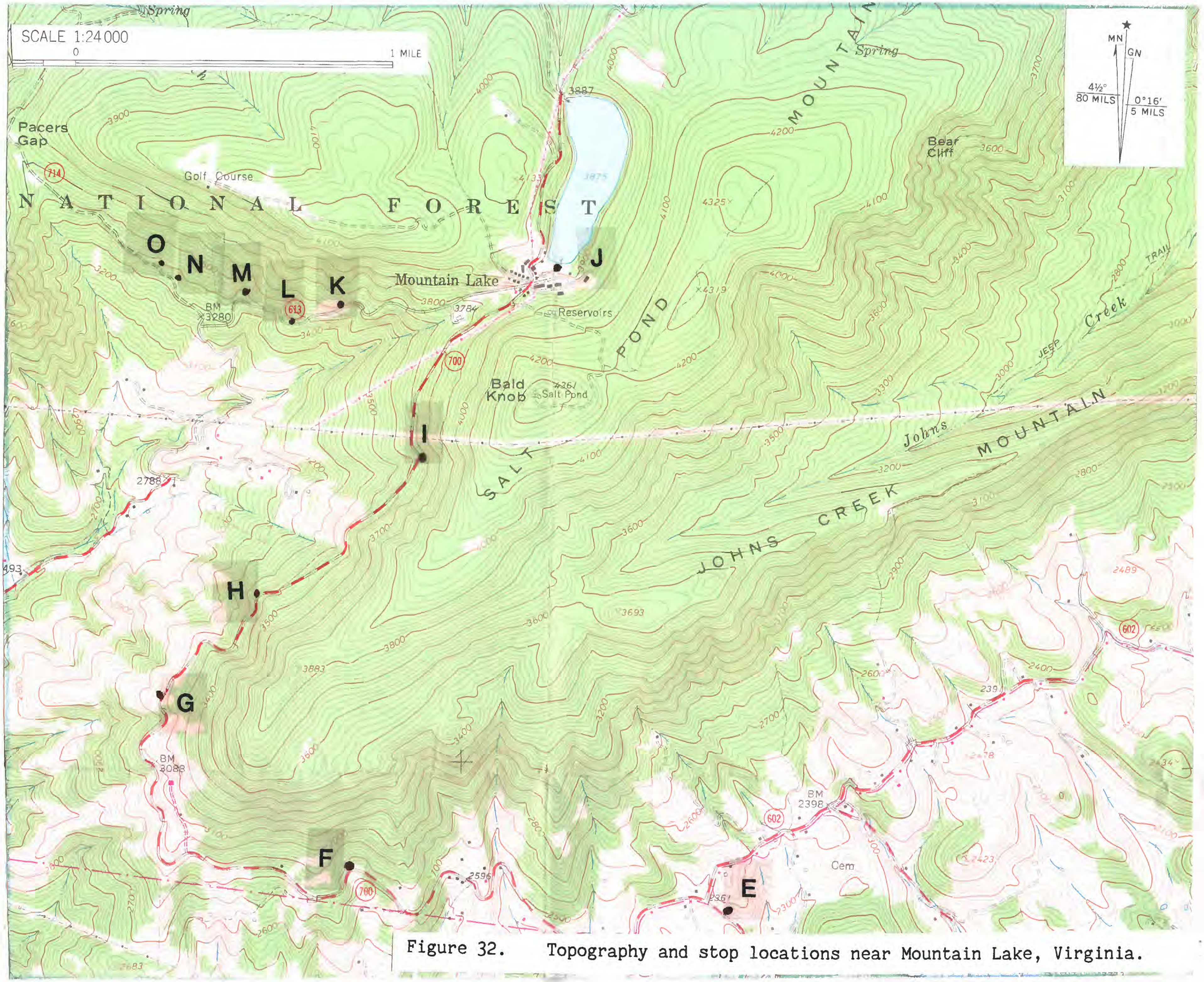


Figure 32. Topography and stop locations near Mountain Lake, Virginia.

Figure 33. Surficial geologic map, western half, in the Mountain Lake area.

EXPLANATION

Residual Regolith

- S — Shale residuum or bedrock (Unit 1)
- L — Limestone residuum or bedrock (Unit 2)
- R — Sandstone residuum and slightly transported colluvium (Unit 3)

Transported Regolith Colluvium

- Y — Younger, less weathered sandstone-rich colluvium (Unit 4)
- Older, more weathered sandstone-rich colluvium (Unit 5)

Transported Regolith Alluvium

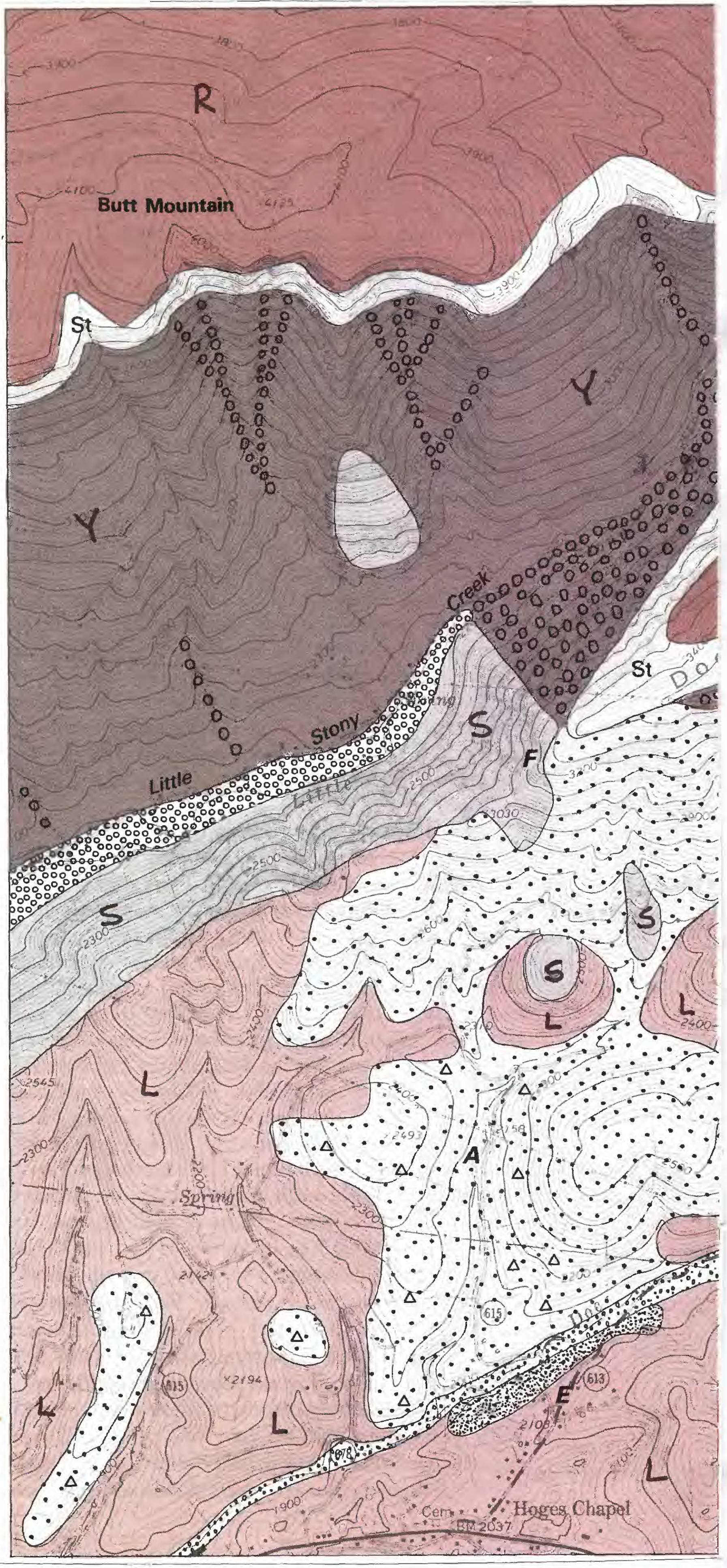
- Fine-grained younger alluvium (Unit 6)
- Medium-grained younger alluvium (Unit 7)
- Coarse-grained younger alluvium (Unit 8)
- Older, more weathered alluvium (Unit 9)

Other Symbols

- △ Areas having abundant chert fragments
- Boulder streams and other dense accumulations of large boulders
- St Tuscarora Quartzite (Silurian)

0 — 1 MILE

0 — 1 KILOMETER



Mills, 1988

FIGURE 33—Surficial geology of northern part of Eggleston quadrangle. Concentrations of boulders associated with older, more-weathered sandstone-rich colluvium are indicated by solid circles to stress rarity. Tuscarora Quartzite outcrop (St) is shown for outcrop slopes only. Map units, designated numerically, are discussed in greater detail in text. Letters *A–F* indicate specific locations discussed in text. Southeast corner of map, south of Johns Creek Mountain, is after Fiedler (1967).

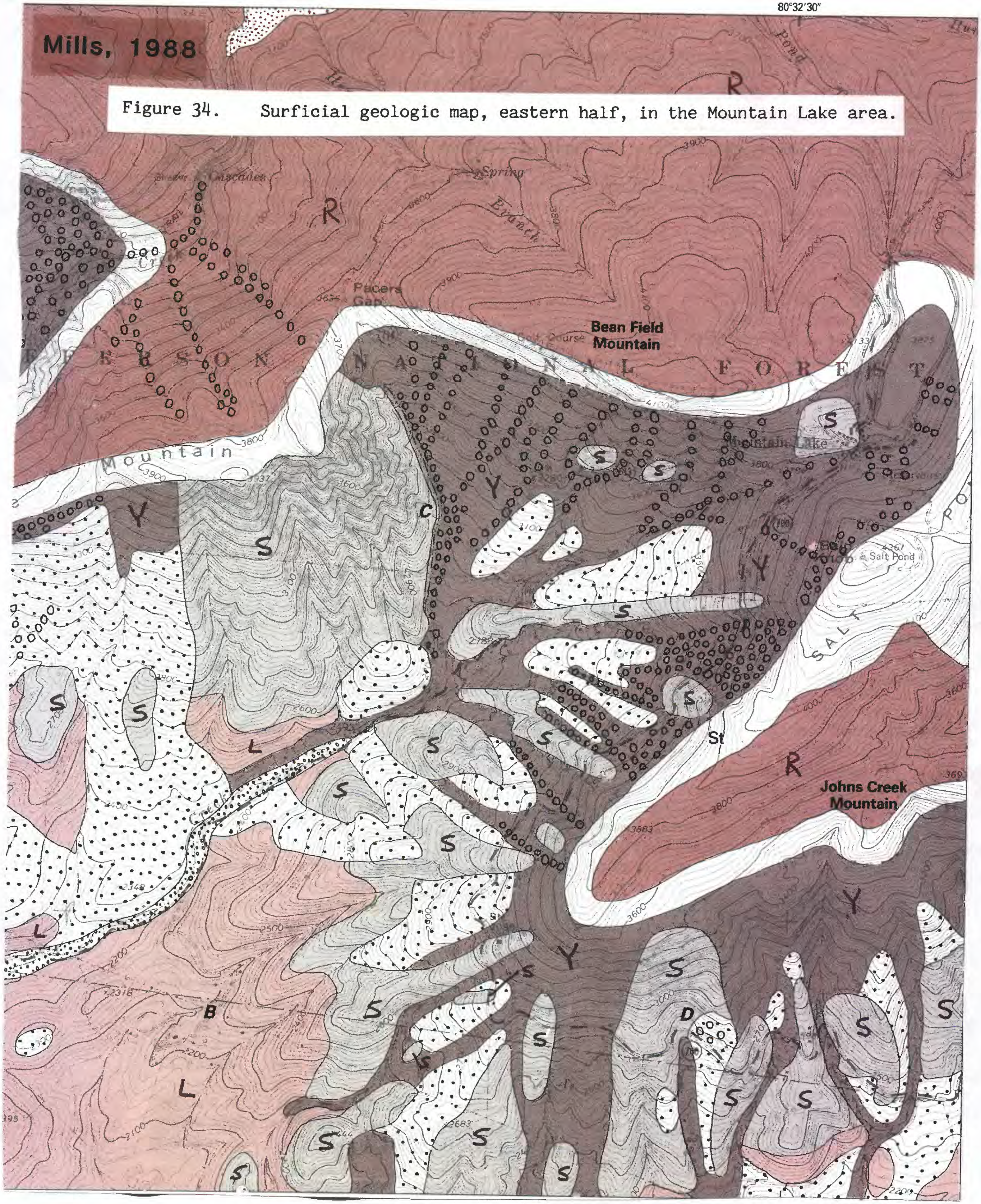


Figure 34. Surficial geologic map, eastern half, in the Mountain Lake area.

FIGURE 35—North end of Mountain Lake. Arrow indicates large boulders that form the natural dam. Boulders were derived from Tuscarora Quartzite ledge a short distance upslope. Outlet is behind boathouse.

Mills, 1988

Figure 35. Oblique aerial photograph of north end of Mountain Lake.

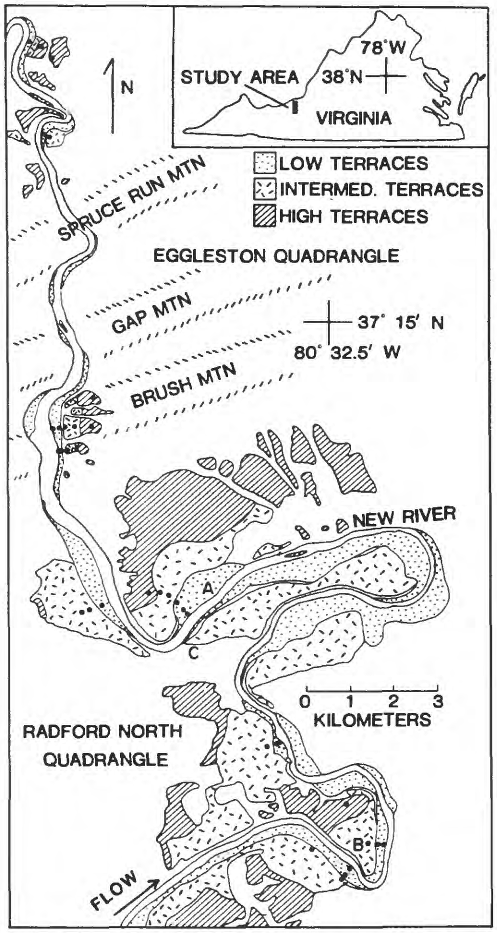


FIG.36—Overlay showing the distribution of New River alluvial deposits in the Radford North and Eggleston 7.5′ quadrangles. The boundary between the two quadrangles is at 37°15′N latitude. Low terraces (including flood plains) are those less than 25 m above the modern river level, intermediate terraces are those 25–50 m above, and high terraces are those greater than 50 m above. Solid circles show location of sediment sampling sites. Letters A, B, and C mark locations referred to in text.

Mills and Wagner, 1985

Figure 36. Map showing distribution of terraces along New River in Radford North and Eggleston quadrangles, southwest Virginia.

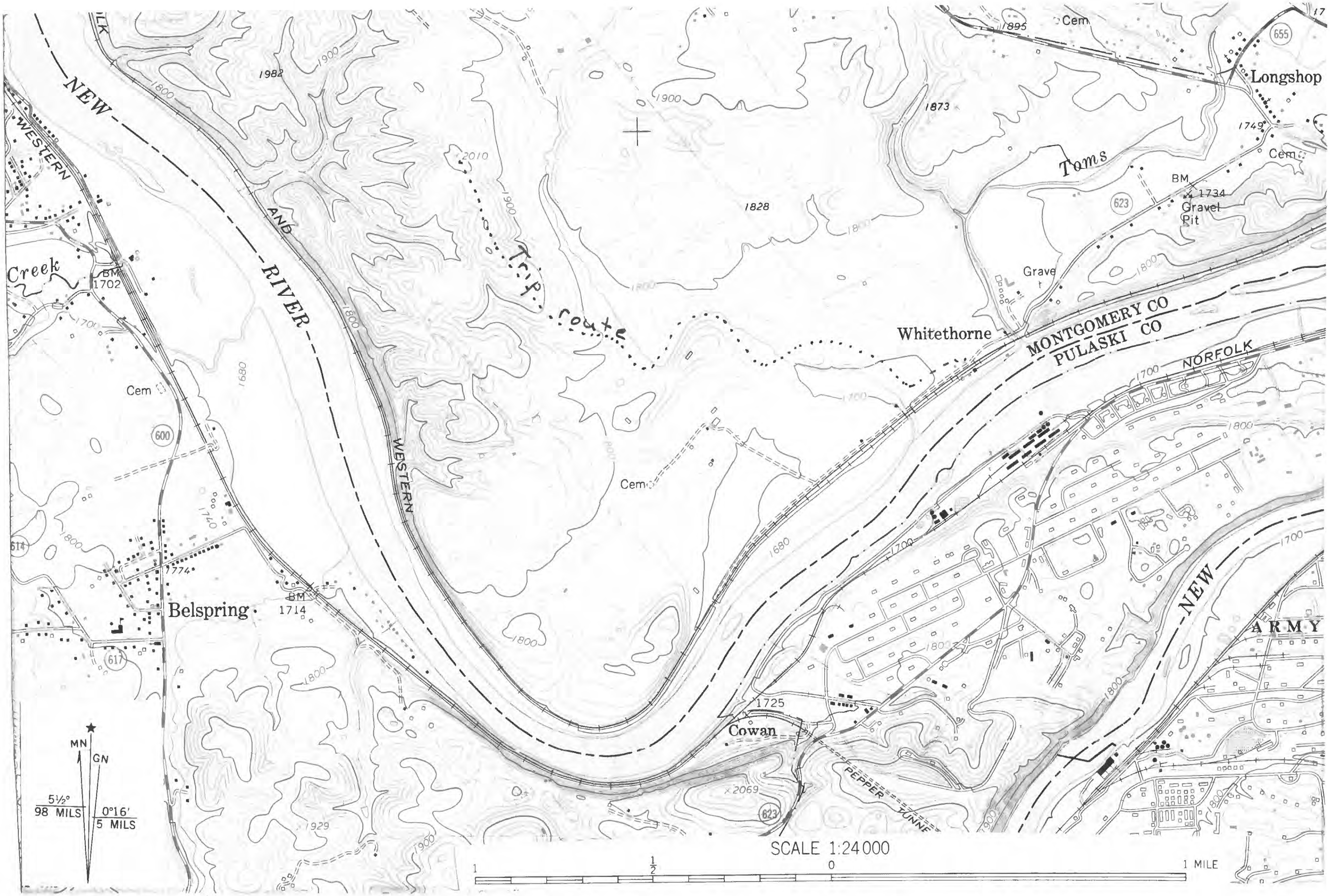


Figure 37. Topographic map of Whitethorne area on New River, southwest Virginia.

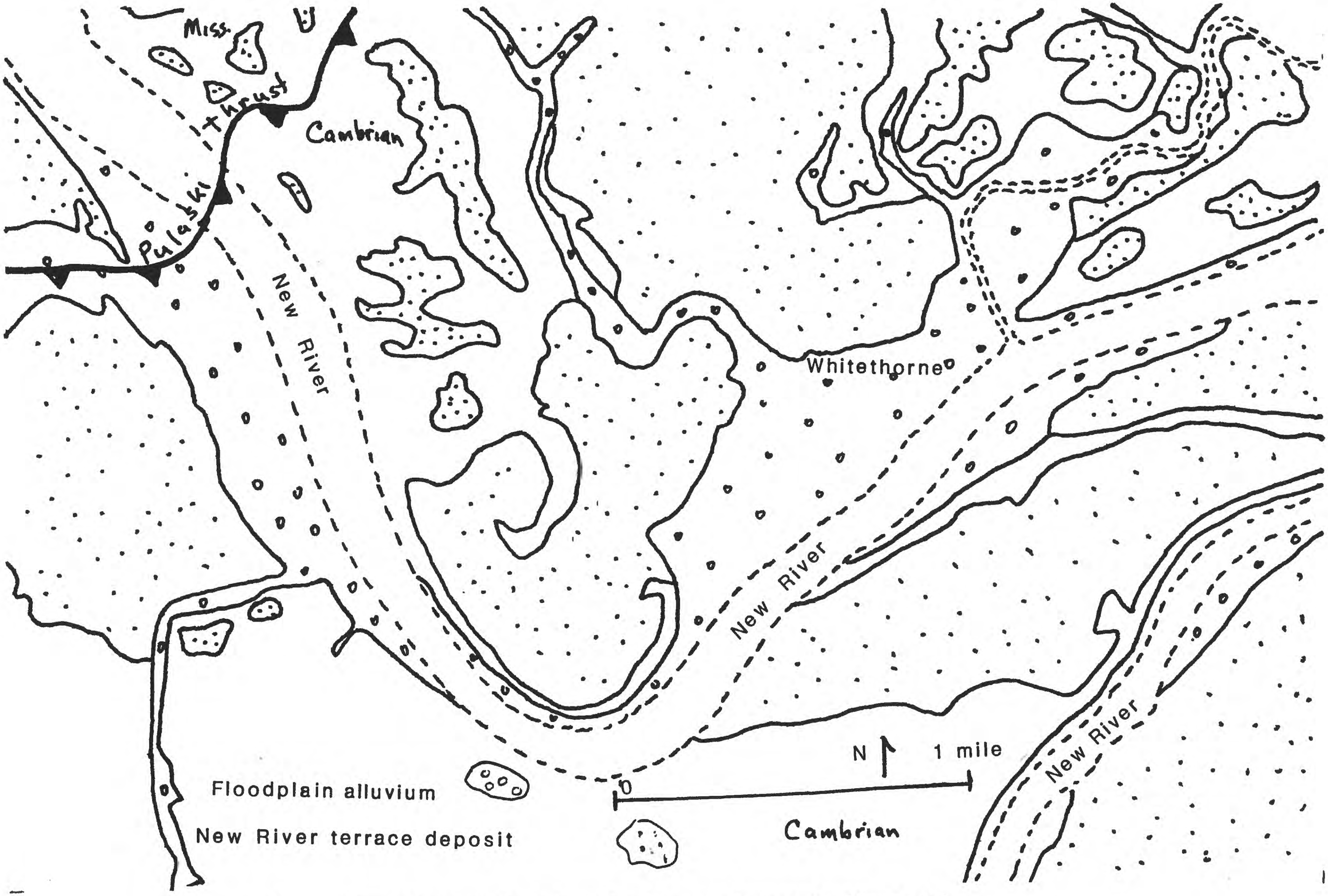


Figure 38. Simplified surficial geologic map of area in Figure 37, Whitethorne, Virginia.

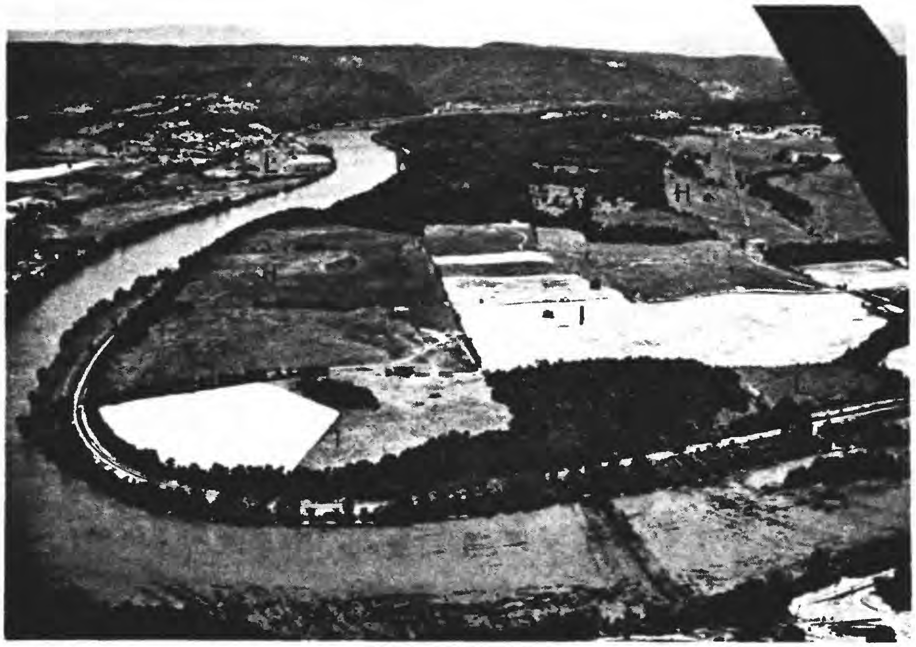


FIG. 2.—Oblique aerial photograph showing typical terraces of the New River. Taken near location C of figure 1, looking north-northwest toward water gaps. L indicates low terraces and flood plains, I indicates intermediate terraces, and H indicates high terraces.

TABLE 1

CHARACTERISTICS OF TERRACES AND TERRACE DEPOSITS

	Terrace Level (AMRL)		
Characteristic	< 25 m	25–50 m	> 50 m
State of original surface	Largely intact	Dissected, but remnants remain	None remains; rolling, irregular topography
Sinkholes	Rare	Numerous	Numerous
Stream valleys	Rare	Numerous small valleys	Numerous small and large valleys
Type of sedimentary structures	Primary only	Primary and secondary	Mainly primary
Jointing	Rare	Some	Numerous
Hue of B horizon	10YR–7.5YR	7.5YR–2.5YR	5YR–2.5YR
Texture[a] of B horizon	Sand to clay	Mainly clay	Mainly clay
Sandstone-clast weathering	0–20% friable	20–60% friable	60–80% friable
Crystalline-clast weathering	Most hard, though feldspar chalky at higher sites	Most decomposed	Rare; all decomposed
Occurrence of iron-manganese oxides	None below 7 m AMRL; films on peds and clasts	Thin layers common, as well as films on peds and clasts	Thin layers and joint fillings common

[a] Textural terms are from USDA Soil Classification.

Mills and Wagner, 1985

Figure 39. Oblique aerial photograph of Whitethorne locality and table of terrace characteristics.