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DEBRIS SLIDES AND FLOWS
ON ANAKEESTA RIDGE
WITHIN THE
GREAT SMOKY MOUNTAINS NATIONAL PARK,
TENNESSEE, U.S.A.

A Thesis
Presented for the
Master of Science
Degree
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Patrick T. Ryan, Jr.

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For Kathy, Tess, Maeve and Eve.
Thank you for your love, patience and support.

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ABSTRACT

Debris slides and flows along Anakeesta Ridge in the GSMNP have been investigated utilizing dendrochronology, aerial photography, erosion stations, precipitation data and rock-slope engineering techniques.

Based on the slides for which the specific dates of occurrence are known, the corresponding precipitation that was at least in part responsible for triggering the mass-wasting events varies from 1.5 inches of rain per day to 4 inches of rain for a 6 hour period. Unfortunately, intensity records are not available to provide at what rate the precipitation was delivered. Based on TVA precipitation records it was determined that 1273 storms (1 or more inches of rain per 24 hour period) occurred during the period of 1951-1987 in the general area of the Great Smoky Mountain National Park. An abundance of moisture is available to effect erosional processes.

Transportational processes operating on Anakeesta Ridge include creep, overland flow, and debris sliding. Additional slope modifiers include: needle ice, slaking, and bank slumping. Slope retreat is primarily accomplished through sheet wash, mass movement and tree throw. Appreciable amounts of fine sediment are moved downslope by slopewash. Tree-throw continues to operate proximally to the Anakeesta Ridge slide scars. Additionally, tree-throw is present at every breach of a ridge crest in the study area and is common along the unfailed slopes of Anakeesta Ridge. In terms of biogenic transport, tree-throw importance is clearly expressed by the microtopography created by decaying tree-throw mounds. Recent slide scar retreat has been nonexistent in some areas, primarily side-slopes, whereas retreat has been as high as 37.2 cm over a 7 month period in scar head areas.

Log jams act as an effective debris dam in slide-track constrictions until that time when the logs are structurally weakened by decay and can be overcome by a new debris torrent. The logs then become part of the ensuing debris slide and contribute to vegetal debris in the fan.

Periodic aerial reconnaissance of areas of interest may be sufficient to detect incipient slide development. The Anakeesta Ridge slides have developed through headward erosion; this was easily tracked through sequential aerial photographs. Incipient slides are followed by additional, ongoing sliding.

The compound slide scars on Anakeesta Ridge have increased in area and volume: 4,300 m² and 1790 m³ in 1953 to 128,000 m² and 84,100 m³ in 1987. The scar head and upper slide track areas are the primary debris volume contributors. Anakeesta Ridge is in a stage of accumulation in the mid-slope regions of the slide scars. The scar heads continue to erode headwardly, supplying material to the lower regions of the scar. Aerial imagery for Anakeesta Ridge indicates that slope failure is initiated in the mid-slope region. The upper slope segments average 43.4°, the mid-slope segments average 33.6° and the lower slope segments average 25.6°. Anakeesta phyllite has an abundance of release surfaces in the form of cleavage planes, joints, faults and bedding planes. The chute morphology is characterized by wedge failure planes as formed by the intersection of cleavage/bedding and joints.

Direct shear testing of Anakeesta phyllite yielded an internal friction angle (ϕ) of 58.2° and a cohesion (c) of 6134 pounds/foot². Utilizing these values, and the slope and failure plane orientations, a factor of safety range from 1.19 to 2.53 was generated.

A high percentage of the draws about Anakeesta Ridge are associated with debris sliding activity. Debris fans along the ridge, vary in approximate minimum dates of origin from 1749 to 1971 as determined from dendrochronological data. Tree coring data yields dates which range from 6 to 87 years between events and average 13.8 years. This length of time may represent the average amount of time required for slope ripening (weathering, accumulation) to occur, the time between major precipitation events or a combination of the two.

Anakeesta Ridge slope instability is due to the coincidence of weathering, regolith accumulation, tree levering against shallow root networks, and high precipitation events. Slides will continue to cause problems along U.S. 441, therefore a need exists for locational and temporal

predictors and a precise accounting of slide localizing factors. Debris sliding cannot be prevented, however, hazardous areas can be delineated, and risks be assessed. In this way, landslides will not be studied simply for forensic purposes. A debris slide in a particular area is not a one-time event. Portions of Anakeesta Ridge have failed in the past, and under the present climatic regime, will continue to do so. Landslide potential is limited only by the availability of excess water, steep slopes and material.

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CHAPTER I

INTRODUCTION

Rock slope failure in the Great Smoky Mountain National Park (GSMNP) has been a pervasive problem in the past, and increasing development of the southern Appalachian region enhances the likelihood of greater (future) road damages both in and outside the Park as well (Torbett, et al., 1986b). Anakeesta Ridge overlooks U.S. 441 and, as portions of the ridge have failed in the past, it is critical that a more precise accounting of slide localizing factors is made (see Plate 1, note breaching of soil and vegetation at ridge on crestline).

In the United States, Schuster (1978) estimates that direct and indirect costs of slope failure exceeds one billion dollars per year, while Sangrey (1985) estimates the losses to be upwards of two billion dollars per year. Landslides represent a major hazard, second to earthquakes (Leighton, 1976). Losses due to landslides can be dramatically reduced because landslides are considered one of the most potentially predictable of geologic hazards (Leighton, 1976). White and Haas (1973) feel that if landslide risk zones can be adequately identified, land use regulations could assure that development is commensurate with the level of risk. A need exists for a more precise accounting of slide localizing factors. Slope stability analysis is an important area of work for engineering geologists; due to dangers potential failures pose to life and property, accurate, rapid analyses of the problem are required (Thomas and Wu, 1985). While much has been learned about the causal mechanisms of landslides, much more knowledge is required for planning and design purposes (White and Haas, 1973).

Purpose of Study

The purpose of this study is to: 1) Identify, date and analyze the landscape as modified by past debris slides and flows, and the processes that change them. Research is in part based on



PLATE 1. View of compound slide scars looking north from Morton Overlook.

historical records; these records include aerial photographs, precipitation records and tree core data. Seven slide tracks were identified and dated. One set of compound slide scars was monitored and analyzed in terms of erosion rates and ongoing processes. 2) Monitor slope backwastage in order to extract rates of erosion. 3) Evaluate the factor of safety (FS) of slopes based on the limiting equilibrium method with supporting direct shear tests and discontinuity data.

If the various parameters of debris slide localization and distribution are better understood, areas of slide potential can be determined with more accuracy. This study enhances the knowledge of slope processes as they operate on the Great Smoky Mountains. The high diversity of forms and the complexity of interrelationships, as well as the practical relevance of landslides, can be recognized only by systematic and thorough study (Zaruba and Mencl, 1982).

Previous Work

Previous work in the general area includes a dissertation and paper by Bogucki (1970, 1976) on debris slopes in the Mount LeConte-Sugarland Mountain area and a thesis by Koch (1974) on debris slides in the Webb Mountain area. Clark (1973, 1984b, 1987) summarized work done and stressed the need for further research, especially in areas of applied geomorphology. Dilworth (1978) completed a senior project specifically on the debris slides on Anakeesta Ridge.

Methodology

Reconnaissance and map preparation of the field area was supported by aerial imagery. This imagery enabled a slide chronology to be established for several years beginning in 1953 through the most recent 1984 coverage. A series of figures were generated using the photographic coverage. This allowed a planimetric evaluation of the progressive areal increase of the slide scar area. Backwastage measuring stations were established along selected slide scar edges; measurements were taken over a three year period, documenting slide scar expansion. Profiles of all the

major slide tracks were undertaken using a Brunton™ pocket transit, based on profiling methods outlined by Gardiner and Dackcombe (1983). Direct shear tests were performed in rock samples so that the rock shear strength and internal friction angle could be determined. Precipitation data from area recording stations were obtained from the Tennessee Valley Authority in Knoxville, Tennessee. Bedding and cleavage relationships to slope failure were determined and graphically represented with the Schmidt stereonet (Hoek and Bray, 1981).

Location

The study area (see Figure 1) lies within the Great Smoky Mountains in the southern section of the Blue Ridge province (Fenneman, 1938). Anakeesta Ridge is located on the Mount LeConte and Clingmans Dome 7.5 minute U.S.G.S. Quadrangles. Anakeesta Ridge is an east-west trending ridge, 4.5 km², (2 mi²) area, that has been partially sculpted by landslides. The summit elevation range of the ridge is 1200 to 1800 m (4000-5988 ft); average elevation is 1650 m (5400 ft). Anakeesta Ridge is 3 km (1.9 mi) south of Mount LeConte, and 11.5 km (7 mi) southwest of Gatlinburg, Tennessee. The ridge is bounded to the north by the second order, dendritically patterned Alum Cave Creek; to the south by the third order Walker Camp Prong, which is paralleled by highway 441 (see Figure 2). Newfound Gap is 1.2 km (0.7 mi) to the south. The coordinate location of the study area is 35 37' 45" latitude and 83 25' 10" west longitude.

Stratigraphy

Much of the Great Smoky Mountains which span the boundary between Tennessee and North Carolina is underlain by the Ocoee Supergroup of Precambrian age (see Figure 3). This is a body of clastic metasedimentary rocks which have minor intercalations of limestone and dolomite, but no identified volcanic components or known fossils (King, et al., 1958). The Ocoee is divisible into three broad units of regional extent and contrasting lithologic character: the Snowbird Group, the Great Smoky Group, and the Walden Creek Group. The Great Smoky

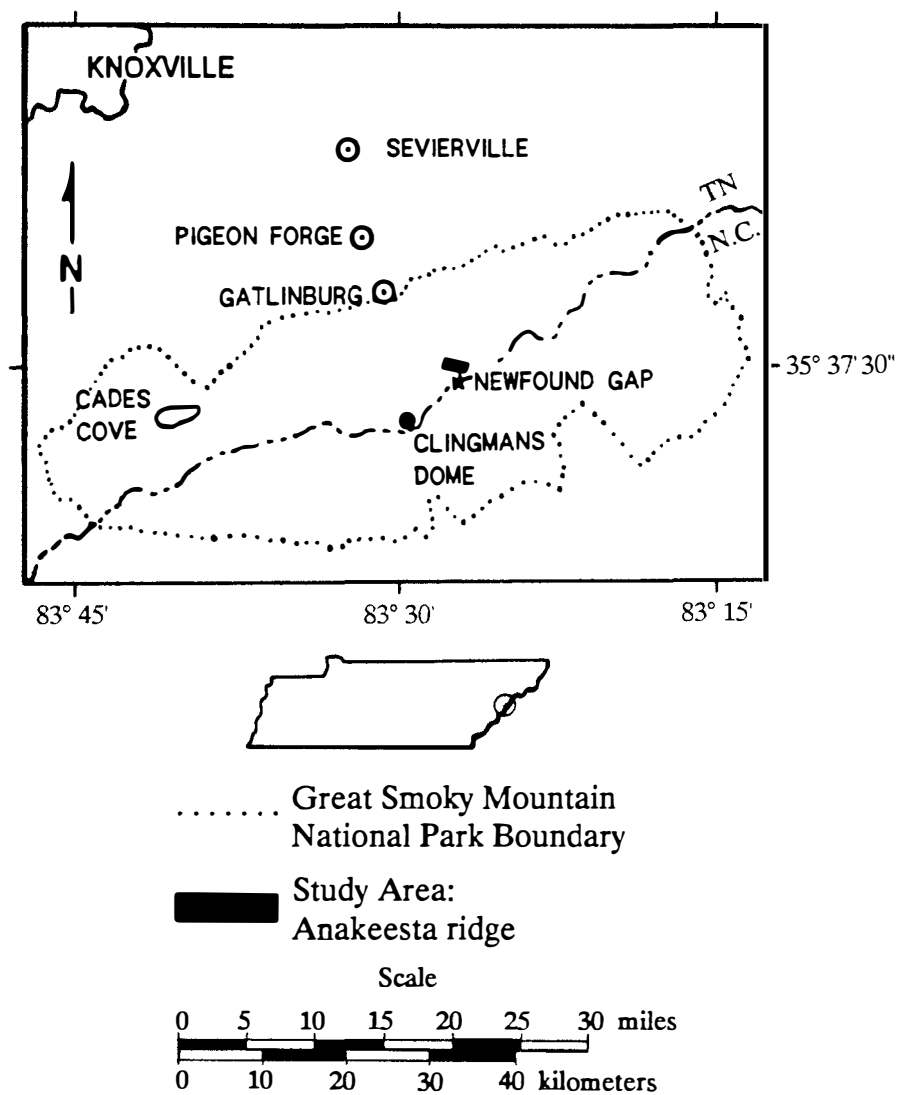
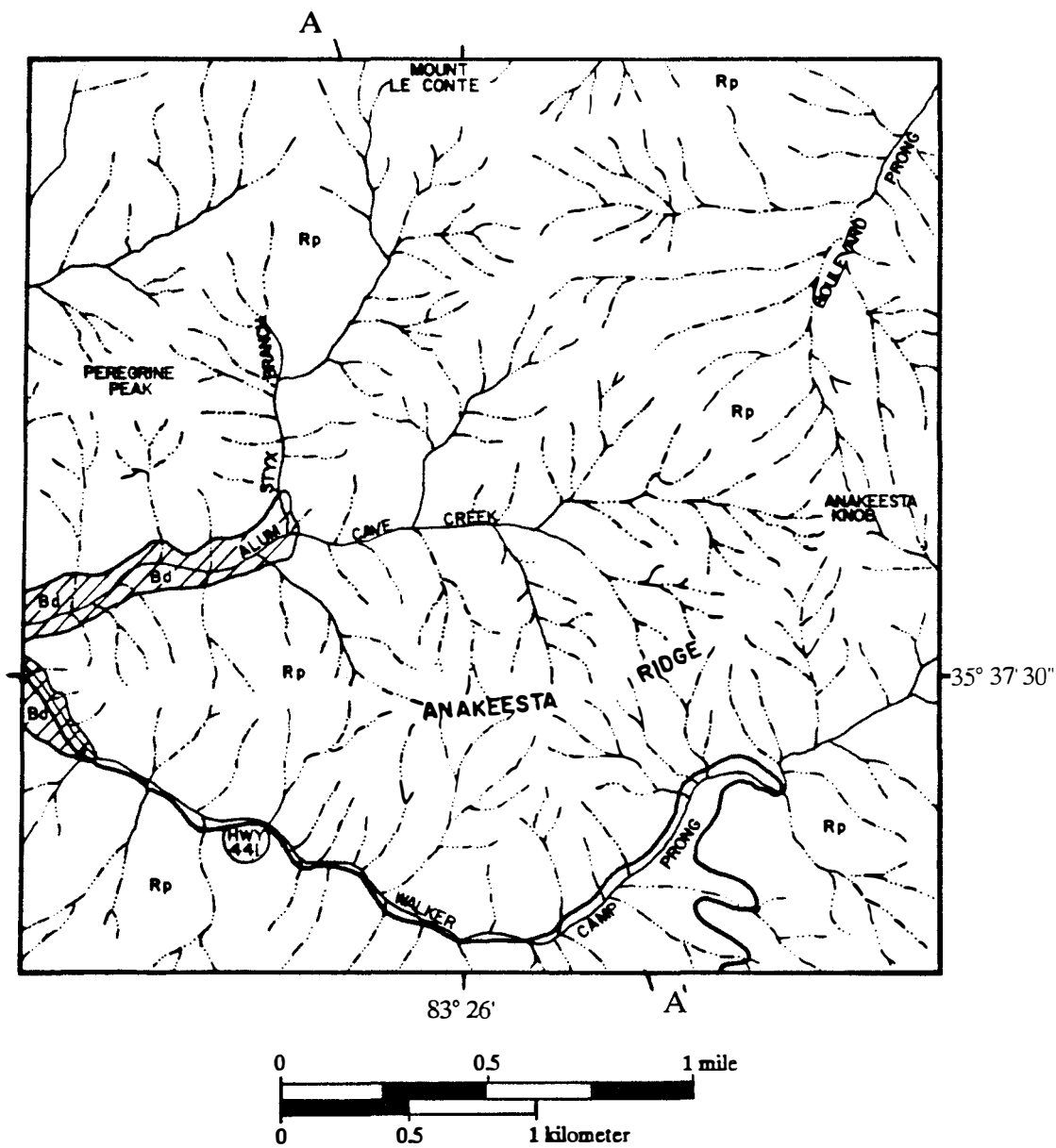


FIGURE 1. Location of study area.



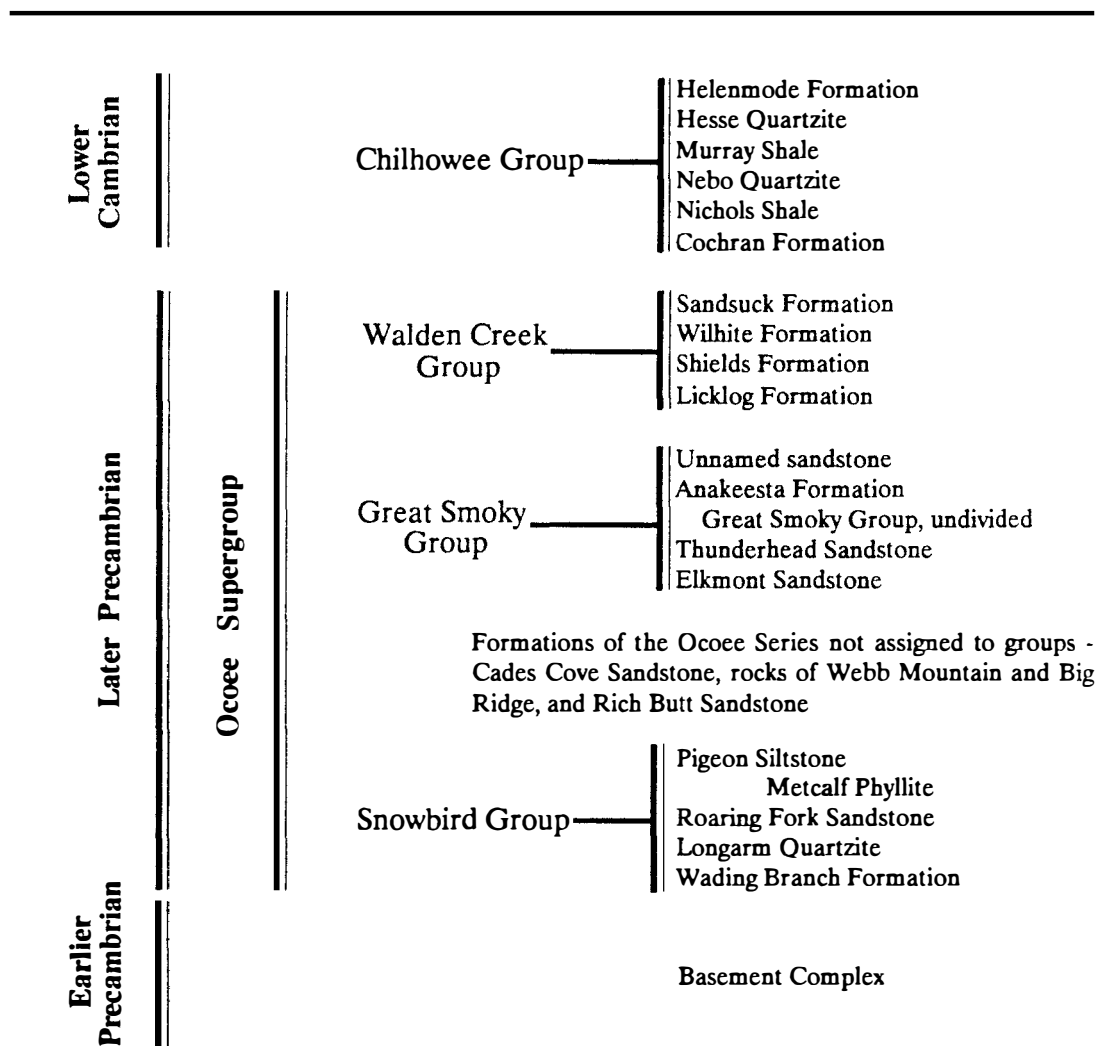
- | | |
|----|---|
| Rp | Rough mountainous land, Ramsey soil material. |
| Bd | Barbourville stony fine sandy loam, rolling phase soil. |

Source: Soil Survey Staff, Soil Survey of Sevier County, Tennessee. Series 1945, No. 1, issued August, 1956.

A-A' Line of geologic cross-section for Figure 3.

FIGURE 2. Soil and drainage map of Anakeesta Ridge and surrounding area.

STRATIGRAPHIC UNITS OF GREAT SMOKY MOUNTAINS NATIONAL PARK AND VICINITY



Source: Philip B. King, Robert B. Neuman, and Jarvis B. Hadley, Geology of the Great Smoky Mountains National Park, Tennessee and North Carolina, Geological Survey Professional Paper 587 (Washington: Government Printing Office, 1968).

FIGURE 3. Stratigraphic units of the Great Smoky Mountains National Park and vicinity.

Group is located above the 4000 m (13,000 ft) Snowbird Group and below the 2400 m (8000 ft) Walden Creek group (Hadley and Goldsmith, 1963). The upper part of the Great Smoky Group consists of fine-grained, dark argillaceous rocks. The Anakeesta Formation is contained within the 7600 m (25,000 ft) Great Smoky Group.

Mineralogical analysis for the Great Smoky Group include the following results: pyrite, 0.3-2.4 percent; quartz, 5.5-48.4 percent; biotite, 1-15.8 percent; and chlorite, 0.9-10.9 percent (Hadley and Goldsmith, 1963). Garnet and Pyrite are readily observable in a recently exposed hand sample.

The Anakeesta Formation consists of fine-grained dark argillaceous rocks interbedded with metasiltstone. The Anakeesta Formation not only lies on the Thunderhead Sandstone with a conformable and gradational contact (see Figure 4), it also intertongues with the Thunderhead through a large stratigraphic interval (Hadley and Goldsmith, 1963). The Anakeesta Formation includes a variety of rocks: small-pebble arkosic conglomerate, graywacke, fine to coarse-grained feldspathic sandstone, gray chloritoidal slate, dark carbonaceous slate and phyllite (Hadley and Goldsmith, 1963). Intraformational chips and slabs of slate are abundant in many sandstone beds. Much of the rock unit consists of dark-gray indistinctly bedded pyritic and carbonaceous slate or phyllite (Hadley and Goldsmith, 1963). Both black slate and dark metasiltstone are commonly interbedded with fine to coarse-grained feldspathic sandstone and fine grained arkosic conglomerate. Sandstone beds, less than 0.3-3 m (1-10 ft) thick, are commonly interbedded. Most of the sandstone is lithologically similar to corresponding rocks in the Thunderhead Sandstone; it is poorly sorted and possesses graded bedding (Hadley and Goldsmith, 1963). Hadley and Goldsmith (1963), believe that much of the Thunderhead Sandstone on Clingmans Dome is stratigraphically equivalent to the lower part of the Anakeesta Formation in the vicinity of Newfound Gap.

The Anakeesta Formation is named for Anakeesta Ridge, a high spur between Mount

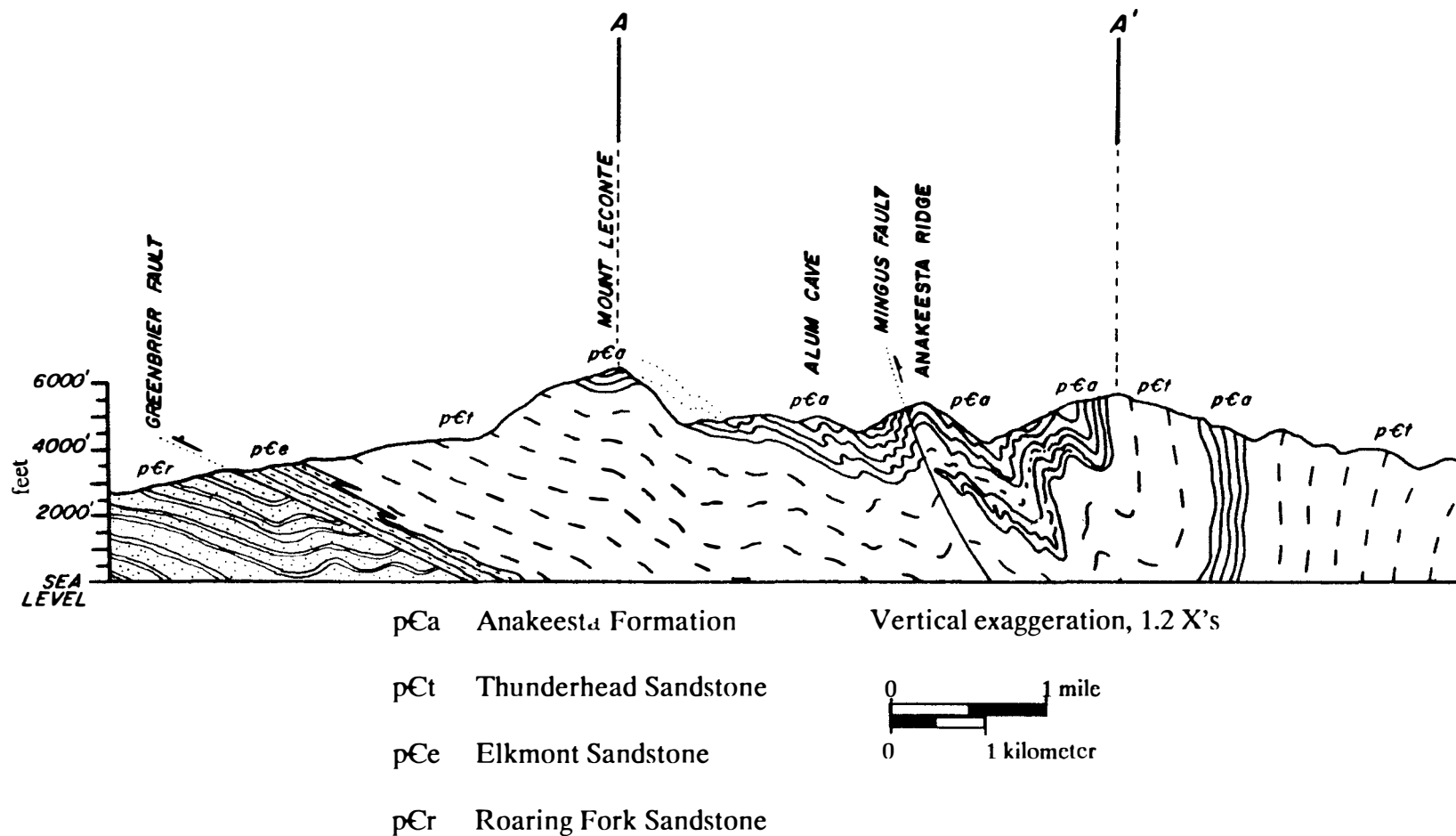


FIGURE 4. Geologic cross section (A-A' shown on figure 2).
Source: U.S.G.S. Professional Paper 587, Plate 1.

LeConte and Newfound Gap. Typical exposures occur along U.S. 441 from the base of the ridge up to Newfound Gap. Recent slope modification (winter of 1986) has exposed fresh Anakeesta phyllite in the rock slope immediately south of the Clingmans Dome road intersection on U.S. 441. Other good exposures occur at intervals along the Alum Cave Creek trail to the crest of Mount LeConte, at Cliff Top and at Myrtle Point. The areal extent of the Anakeesta Formation in the GSMNP along the Tennessee/North Carolina border is approximately 155 km². The topography on the formation is characterized by steep-sided ridges with serrate crests and craggy pinnacles of which the Chimneys near U.S. highway 441 west of Mount LeConte are good examples (Hadley and Goldsmith, 1963). Hadley and Goldsmith (1963) have also noted that large areas of the formation have been bared by fire or landslides, producing a rugged, spectacular scenery.

Structure

The west end of the Alum Cave syncline terminates abruptly against the transcurrent, right lateral Oconaluftee fault. This fault trends northwest-southeast and is located 2.5 km (1.5 mi) southwest of Anakeesta Ridge. This fault is followed by segments of streams and occupies prominent gaps in the intervening ridges (Hadley and Goldsmith, 1963). The lower tongue of the Anakeesta Formation southeast of Newfound Gap is displaced approximately 0.8 km (0.5 mi). The Mingus (see Figure 4) fault and adjacent outcrop belt of the Anakeesta formation is displaced by approximately 1.5 km (0.9 mi). The Mingus fault, a steep reverse fault (Hadley and Goldsmith, 1963), lies immediately north of the ridge and trends east-west. South of the Greenbrier fault, the rocks of the Thunderhead Sandstone and the Anakeesta Formation occur in a complex synclinorium. At Newfound Gap, interbedded argillaceous metasandstone and minor phyllite at the base of the main body of the Anakeesta Formation are overturned to the northwest at 50-55° (Hadley and Goldsmith, 1963). Structurally interesting aspects of the study area include quartz rich en echelon sigmoidal veins which stand in bas relief within highly silicious "rolls," and abundant boudinage within the Anakeesta Formation. A stereonet of the discontinuities in the

area reveals two areas of pole concentration (see Figure 5).

Slope instability factors include: bedding planes, mineralogy, stress history (cleavage), jointing, faulting, weathering history and present day forces of rainfall, tectonic activity and oversteepening through river undercutting, road construction and overall loss of toe support. Rock control (mineralogy and discontinuities) is dominant in shaping the Appalachian landscape (Mills, et al., 1987). Endogenic influences such as structure, lithology, and past and present tectonics, are contributors to geomorphic features (Mills, et al., 1987).

Delcourt and Delcourt (1985) note that the Anakeesta Formation is “particularly vulnerable to landslides along bedding planes on steep slopes.” In this study, the predominant planes of failure are joints, bedding and cleavage. Cleavage and bedding are very similar in orientation (See stereonet plots in Figure 6).

Soil

The mantle of surficial material in the Great Smoky Mountains consists of residuum, alluvium, and colluvium (King, 1964). A complex history of changing climate, weathering, soil formation, erosion and deposition is preserved in this blanket of surficial material (Torbett and Clark, 1984).

Soil development on hillslopes is typically poor, with weak horizon development (Clark, 1987). The lack of distinct horizons indicates that surficial mantle is relatively unstable (Bogucki, 1970). In West Virginia, Schneider (1973) has observed that in many slide scars, soils do not have pronounced horizons. These soils are considered inceptisols, as they lack eluvial and illuvial horizons; the Soil Survey Staff (1956) has classified the soils in the rough mountainous regions of the Great Smokies as Ramsey Soils (see Figure 2, page 6). Feldkamp (1984) more specifically classifies these soils as belonging to the Umbric Dystochrept subgroup. The soil, saprolite, and unstable rock are constantly being removed by slope processes, thus modifying the topography.

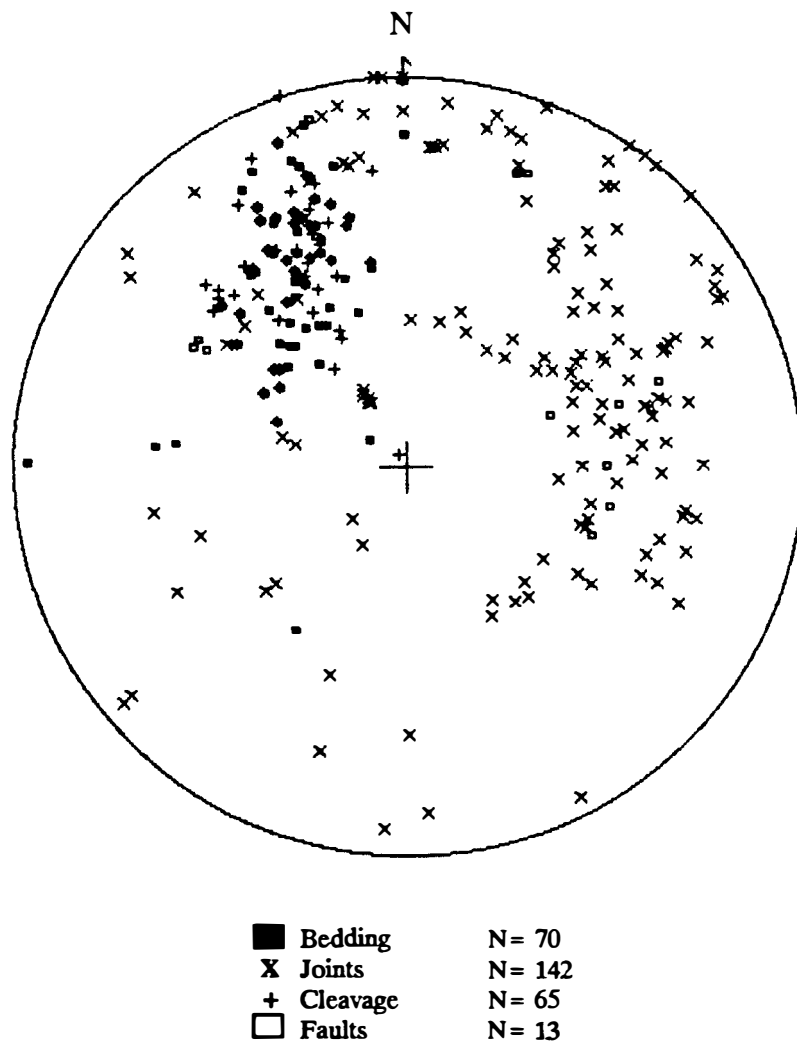


FIGURE 5. Schmidt pole plot of discontinuities in study area.

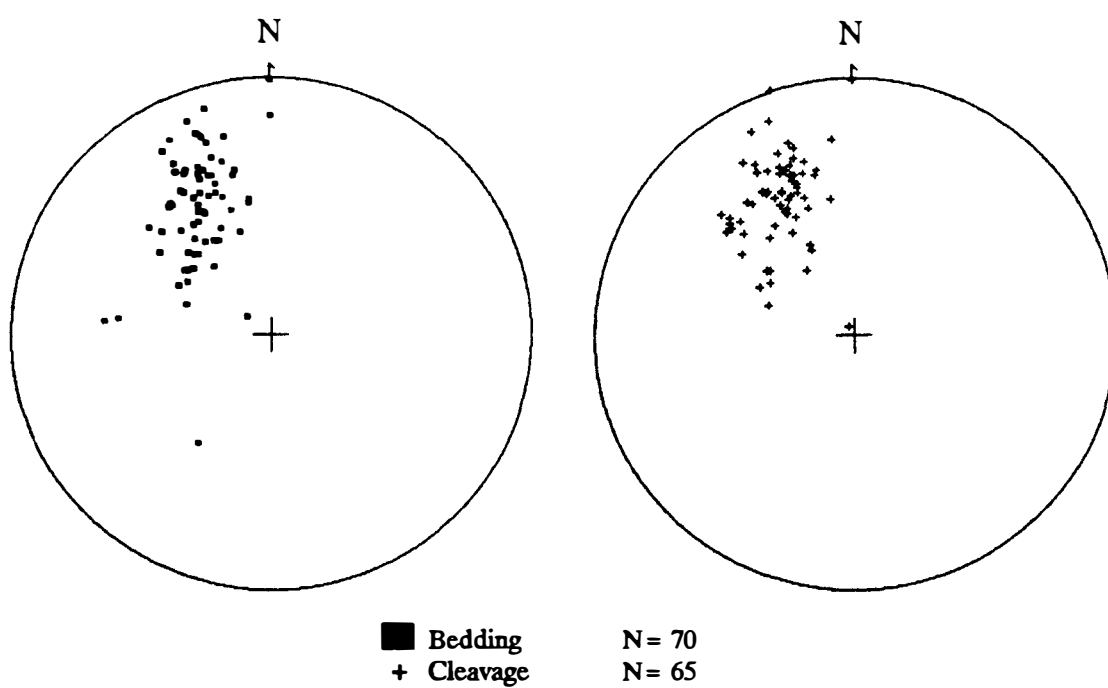


FIGURE 6. Schmidt pole plot of bedding and cleavage in study area.

Outcrop and boulders, mainly quartzite and phyllite, occupy 20-70 percent of the surface in most places. The quartzite lithotype dominates the boulder size category as the phyllite lithotype more readily breaks apart due to cleavage, joint and bedding discontinuities. Ramsey soil material occupies the space between outcrops and boulders. This material varies greatly in depth, color, and texture. Organic matter content is variable and depends mainly on the exposure, elevation, and vegetation. These soils are predominantly formed from quartzite and phyllite.

Clark (1987) finds that illite and kaolinite are dominant clay minerals with minor vermiculite and chlorite often present. Wolfe (1967) has found that soil over phyllite is commonly illite, intergrading to vermiculite, with minor kaolinite. Kaolinite is a nonswelling clay, a component of more stable soils. Bogucki (1970), in his work on the Mount LeConte slides, found illite, vermiculite and kaolinite as dominant clay minerals. The presence of a continuous clay layer along the soil and bedrock interface would decrease permeability and increase runoff. Due to the abundance of rock discontinuities and the dynamic nature of the surface, such a continuous layer does not exist.

Vegetation

Vegetation types and serial communities can be used as indicators of landslide age, degree of activity, type and component parts of a landslide (Crozier, 1984). The effects of time on debris slide recovery and revegetation were noted by Feldkamp (1984): "With increasing age, soil depth increased, bare rock cover decreased, and cover of vascular plants, bryophytes, and lichens increased on most parts of each debris slide." Feldkamp (1984) determined average soil depths available for vascular plant rooting-all are less than 14 cm (see Table 1). The unstable and disrupted drainage conditions associated with landslides favor colonization by light-tolerant, and fast-growing species; on exposed surfaces-of-rupture these are likely to be shallow-rooting, drought-tolerant species, while on the accumulation zone, they will be deeper-rooting moisture tolerant species (Crozier, 1984).

Table 1. Mean soil depths available for rooting of vascular taxa
(From Feldkamp, 1984).

Species	Taxa	Soil Depth (cm)
Yellow Birch	<i>Betula lutea</i>	6.6
Red Spruce	<i>Picea Rubens</i>	13.0
Fire/Pin Cherry	<i>Prunus pensylvanica</i>	11.3
Catawba Rhododendron	<i>catawbiense</i>	6.0
Blackberry	<i>Rubus canadensis</i>	6.9

Revegetation occurs in the mid to lower slide track on the colluvial debris. The debris consists of soil, vegetation, and Anakeesta Phyllite and sandstone. Bogucki (1970) notes that revegetation is generally most advanced in debris fans, with the following species dominating the revegetation process: Fire or pin cherry (*Prunus Pensylvanica*), Sweet birch (*Betula lenta*), yellow birch (*Betula alleghaniensis*), red spruce (*Picea rubens*), and blackberry (*Rubus canadensis*). Crandell (1958) indicated that mosses and liverworts are the pioneering plants on bare rock surfaces. Sand-myrtle (*Leiophyllum buxifolium*) follows, with minnie-bush (*Menziesia pilosa*), catawba rhododendron (*Rhododendron catawbiense*), or rosebay rhododendron (*Rhododendron maximum*); under the correct elevation, soil and moisture conditions red spruce (*Picea rubens*) or fraser fir (*Abies grandis*) may grow. Hupp (1983b) notes that pioneer species such as blackberry do not become a big revegetation factor until the second season after the mass wasting event. Pioneer species in Flaccus' (1959) study area included Yellow birch (*Betula alleghaniensis*). Blackberry and Green-briar are also very common. Daubenmire and Slipp (1943) found that the aridity of south-facing slopes slowed revegetation and limited it to marginal encroachment by vascular plants. In an elevated catchment zone on Anakeesta Ridge, conditions are moist enough to support grasses which require a constant moisture supply.

Hack and Goodlett (1960) determined tree species abundance on nose, side slope and

hollow areas-Yellow birch (*Betula alleghaniensis*) was found to be dominant in the hollows. The fact that Yellow birch (*Betula alleghaniensis*) dominates the slide track, in terms of tree species, is not too surprising because many authors such as Hupp (1983b) have found that the prominent canopy species in the area are the first and most common species establishing on the post-disturbance surface. Yellow birch (*Betula alleghaniensis*) is found on undisturbed mountainsides at elevations of 1067-1372 m, 3500-4500 feet (Stupka, 1964). Stupka (1964) reports that a large proportion of the mature trees along highway 441 between the tunnel overpass and Walker Prong are Yellow birches (*Betula alleghaniensis*). Many specimens exceed 10 feet in circumference.

Scott (1972) visited several debris slides that occurred in 1940; although originally eroded to bedrock, these scars were almost completely healed. Scott (1972) feels that typical debris avalanche scars within the Blue Ridge Mountains are nearly completely healed within 30 years of formation, especially at lower elevations (less than 1280 M; 4200 Ft). Flaccus (1959) determined that slides of approximately 130 years of age, are close enough to the climax forest so that the slides could not be detected /located exactly by means of vegetation. Revegetation of the September 1, 1951 Mount LeConte slide scars has been slow; after 38 years they are still obvious. This can be attributed to high elevation, steep slopes and the fact that the majority of these scars have bared bedrock. In agreement with Schneider (1973), revegetation on the slope head is more difficult, due to headward erosion, than along the chutes in which small trees and brush recover and thrive. The primary types of vegetation that grew on the compound slide scar are Yellow Birch (*Betula alleghaniensis*) and Blackberry (*Rubus canadensis*).

CHAPTER II

DEBRIS SLIDES DEFINED

Terminology

Classification is useful as a means of discovering degrees of order within data. Mass movement classification is not easy due to the combination of materials, slopes, climate and agents responsible for movement. While various classification schemes are available, terminology has been adapted from Clark (1987), Brundsen (1973), Crozier (1973), Varnes (1958) and Sharpe (1968, 1938). Varnes (1958) defines landslides as the downward and outward movement of slope-forming materials composed of rocks, soils, or artificial fills. Sharpe's (1938) classic mass-movement classification includes a variety of factors: kind of material, size, cause, rate of movement, water content, characteristics of internal friction, organization of material within moving mass, relation of moving mass to surface material and substratum. Sharpe (1938) makes a distinction between slides and flows based on the presence or absence of a slope plane separating the moving mass from the stable ground: In flowage, no slip plane is present and movement takes place by continuous deformation-flows consist of incoherent rock debris within a viscous (muddy) fluid which exhibits internal turbulence with discrete boundaries or marginal zones of shear; true slides move on a slope surface in which deformation is not continuous. Sharpe (1968) considers the moving mass a true landslide when water constitutes a small portion of the mass. As the water content decreases and the mass becomes more viscous, the slope required for movement steepens.

Definitions

Definitions ascribed to are utilized from work put forth by Clark (1987) and others as cited (see Figure 7):

Angle of internal friction- a value which indicates the extent that friction, induced by normal stress, contributes to shear strength.

Chute- A general term for hillslope areas where vegetation and soil mantle have been partially or completely removed by the processes of debris slides/debris flows. Chutes often extend from short distances below ridge crests downslope to debris fans or to channel-ways if fans are absent.

Cloudbursts- A torrential rainfall in the mountains. The Tennessee Valley Authority defines intense rain as having one or more inches of rain in one hour or three or more inches of rain in 24 hours (Bogucki, 1972).

Cohesion (C)- shear strength in the absence of normal stress. Cohesion is inherent to the strength of material.

Debris Fan- Fan or cone-shaped accumulation of predominantly water laid debris, mainly stones and gravel, some fines, and vegetational remains. Debris fans are the depositional termini of chutes in locations where chutes do not directly enter channel-ways; some debris fans, however, may be transitional to high-energy alluvial fans.

Debris Flows- Rapid mass movements involving rapid debris flowage, containing coarse-grained materials, and resulting almost invariably from intense precipitation.

Debris Slides- Rapid mass movements initiating along regular to irregular surfaces. The movement primarily involves soil and vegetation but may also involve layers of bedrock. Initial movement may be rotational or translational in nature, or may involve elements of both. Involves intense precipitation events on relatively steep slopes that possess a weathered soil mantle.

Effective Stress- The most important influence of water in a discontinuity, is the reduction of shear strength (t) due to a reduction of the effective normal stress (s) as a result of water pressure (Hoek and Bray, 1981): $t = c + (s - u)\tan\phi$.

Factor of Safety (FS)- FS is a ratio of the sum of resisting forces over the sum of the driving forces; when $FS > 1$, the slope is likely to be stable; when $FS < 1$, the slope is likely to be unstable.

Failure- The inability of the soil/rock element to withstand an applied stress rate. Failure is associated with large strains and/or a rapid decrease in the stress state which can be resisted by the soil/rock (ASTM, 1980).

Flow Track- That portion of the chute modified by debris flow activity accompanied by fluvial erosion, transportation and deposition.

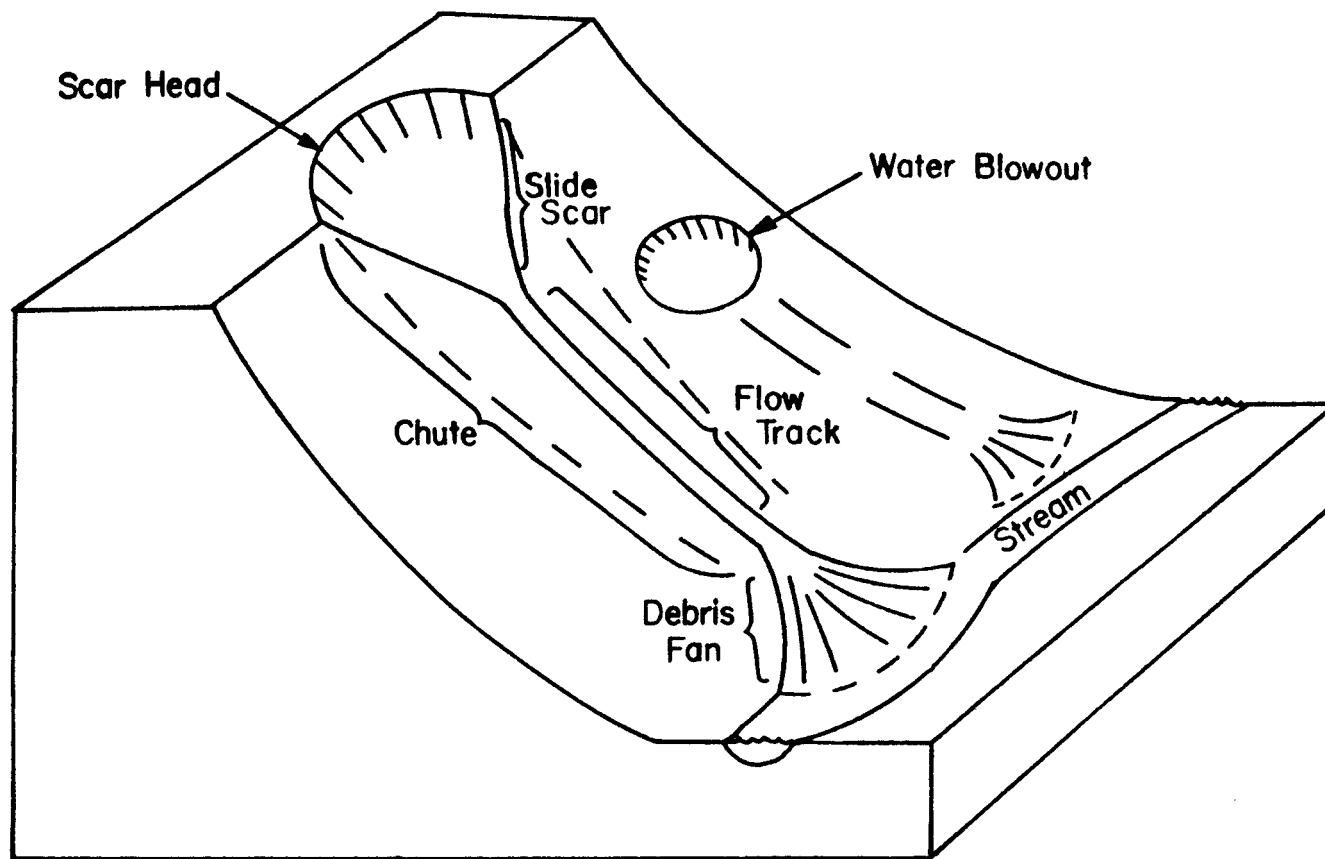


FIGURE 7. Schematic block diagram illustrating the terminology used for features produced by rapid mass movements. Source: Bogucki (1970) and Clark (1973).

Hollow- This is the central part of the valley which contains the stream head, this is a moist area in which the contours are concave outward; every point in this area converges toward the stream (Hack and Goodlett, 1960).

Internal Friction Angle (ϕ , $\emptyset$)- The stress dependent component of shear strength that is similar to sliding friction (Holtz and Kovacs, 1981).

Nose- The driest portion of a valley; includes ridge crests and slopes in which the contours are convex (outward). Running water within this area tends to diverge (Hack and Goodlett, 1960).

Sideslope- That portion of the slope which is adjacent to slide scar and flow track.

Scar Head- The heel print-shaped head of most slide scars (Schneider, 1973).

Shear Strength- The shear strength, t , of a material is a function of the applied normal stress, and the strength parameters of $\emptyset$ and C as related by Coulomb's equation: $t = \text{normal stress} (\tan \emptyset) + C$; (Holtz and Kovacs, 1981).

Slide Scar- The erosional depressions produced on hillslopes by debris slide activity. Slide scars are often transitional downslope to flow tracks.

Water Blowout- Erosional holes in hillslope debris mantle that show no evidence of ground cover breaks above or below the depression.

Pore Water Pressure

Ground water strongly influences the effective stress state in earth materials which in turn can precipitate hillslope instability (Terzaghi, 1950). The importance of pore pressure has been described by eye-witnesses to mass-wasting events. Eye-witness accounts indicate that the debris sliding occurs very quickly with the ground oozing then the entire hill sliding quickly; the whole hillside moves at the same instant with water spraying out from the moving earth (Williams and Guy, 1973). Iverson and Major (1987) have found that times of high ground-water head at the base of the landslide correspond well with times of rapid landslide motion. They also found that landslide motion is closely regulated by the direction and magnitude of near-surface hydraulic gradients and by waves of pore pressure caused by intermittent rainfall. It may take a period of time for the pore pressure to reach a critical mass-wasting point. The local resident's accounting of the slides suggest that the sliding was more frequent toward the end of the storm (Williams and Guy, 1973).

Iverson and Major (1987) have documented seasonal rainfall cycles producing long-period waves that modify basal pore pressures, but only after time lags that range from weeks to months; time lags can depend on antecedent moisture storage. "When seasonal pressure waves reach the landslide base, they establish a critical distribution of effective stress that delicately triggers landslide motion (Iverson and Major, 1987)." Through the use of piezometers, extensimeters, or inclinometers, or on-site precipitation gauges, Iverson and Major (1987) determined that single rainstorms typically produce short period waves that attenuate before reaching the landslide base. This study differs from the Iverson and Major (1987) study in that debris slides are not slowly moving masses and the Anakeesta slides are relatively shallow. Within imminent debris slides, soil pore water pressure is increased to the point where the shear strength of the soil is lost due to the reduction of effective intergranular stress. The moisture loading of the soil mass also contributes to the disturbing forces. The behavior of a soil mass is controlled not only by the external total stresses applied to a soil element, but also by the water and air pressures developed in the pores of the soil (ASTM, 1980). The controlling feature is the resultant interparticle forces which govern the shear strength. To understand more fully the shear strength behavior of soils it is necessary to understand the influences on shear strength of ground-water pressures and their variations with location and time. This is expressed for a saturated soil by the effective stress principle: the strength of a soil depends on the difference between the total stress and the pore water pressure-this difference is termed effective stress. The effective stress equation, $e' = e - u$, is possibly the most important equation in soil mechanics: (e' = effective stress, e = total stress, u = pore water pressure).

CHAPTER III

DEBRIS SLIDES DISCUSSED AND DESCRIBED

Debris Slide/Flow Evidence

Costa and Jarrett (1981) use as debris flow evidence, the presence of coarse, lobate, poorly sorted, unstratified, unconsolidated deposits with well-defined levees and terminal lobes. This evidence was utilized for identifying areas of debris slide and/or flow activity. Other forms of evidence included the presence of log jams, abraded flow tracks, reverse grading of debris fans, change in topography, and the presence of disturbance indicating species such as Yellow birch (*Betula lutea*) and Blackberry (*Rubus canadensis*).

Morphology

The Anakeesta Ridge slide scars are classic examples of compound slide scars (see Figure 8). The chute morphology along Anakeesta Ridge is characterized by wedge failure planes along the flow track and a combination of circular, planar, and wedge failure planes in the upper slide scar area. The Anakeesta Ridge slide scar heads tend to be elongate and transitional to the flow track.

The chute morphology between the Thunderhead Sandstone and Anakeesta Formation bedrock sites contrast in that slide chutes in the Thunderhead Sandstone are more box-like, with release surfaces governed by vertical joints; the Anakeesta Formation chutes are v-shaped, with release surfaces governed by cleavage, bedding, and joints.

According to Scott (1972) the shape of a debris slide scar can be attributed to the typical location of scars in a steep-walled valley/hollow: the linear chute is positioned along the valley

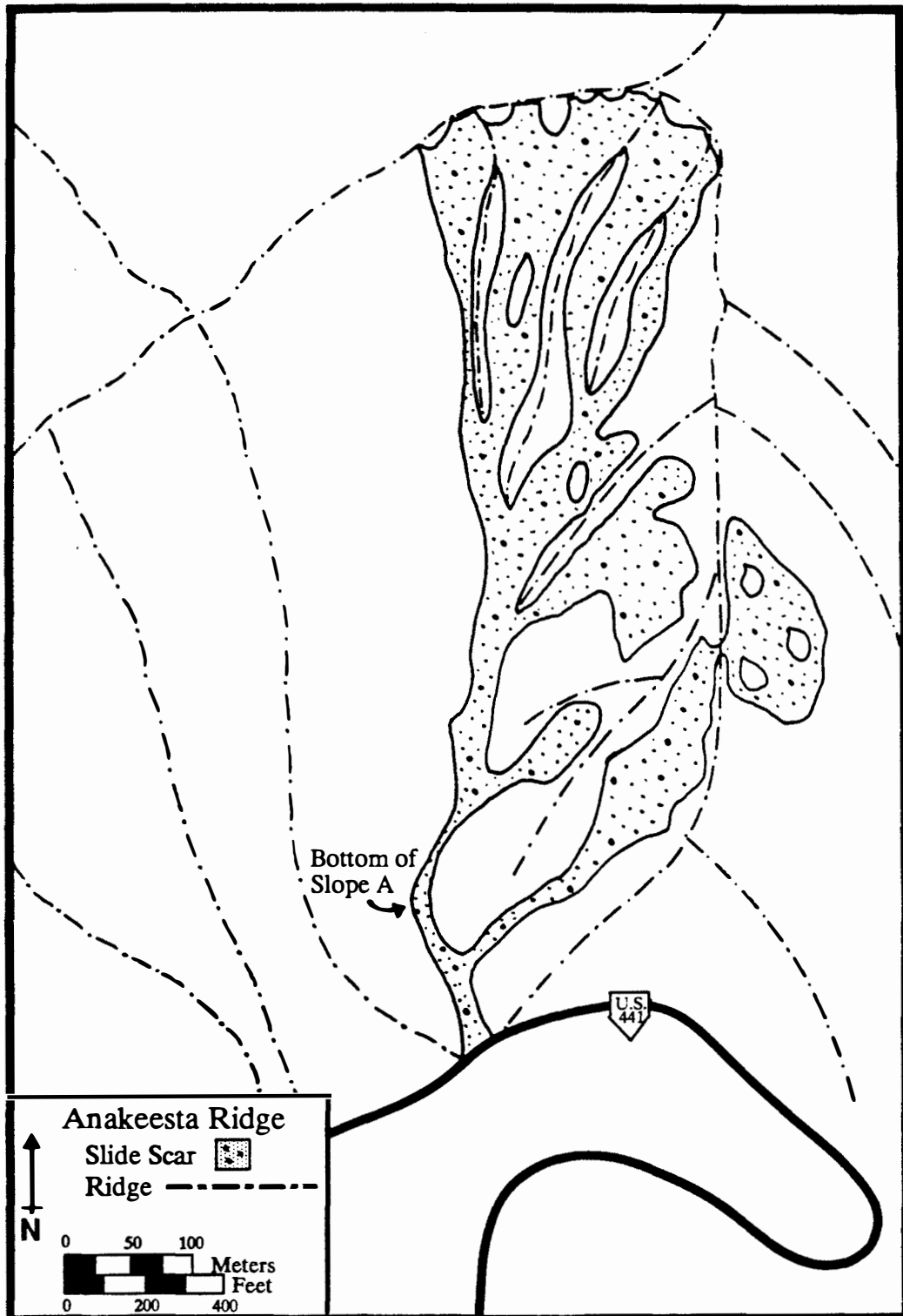


FIGURE 8. Compound slide scars of Anakeesta ridge.

bottom and the fan-shaped scar-head follows the hollow of the valley head. Shape therefore, is a function of topography.

Debris slides consist of three basic segments: the upper, erosional section, the (scoured) transportational gully section, and the depositional section (Scott, 1972). The scar head is shaped like a heel print (Schneider, 1973). The transportational gully section is usually the shortest segment of the debris slide in the Blue Ridge (Scott, 1972). If the slide track is of any length, it probably follows a path of least resistance that is not always perfectly straight; in these instances the flow mass will do some "bobsledding" around corners (Kuhaida, 1971). This is true of the Anakeesta Ridge slides (see figure 8, bottom of slope A).

The mapping of debris slide scars is subject to error due to photograph distortion, scale variation and measuring errors. These errors are minimized by field-checking the data and by utilizing familiar landmarks such as ridge crests and roads. Slide scars on Anakeesta Ridge have had large scale photogrammetric coverage since at least 1953. This coverage (see Table 2) has enabled nine stages of slide scars to be documented (see Figures 9 through 17).



TABLE 2. Photographic coverage for Anakeesta Ridge

DATE	SOURCE	CODE	SCALE
04-26-84	USDA	539-178 358205 HAP-81F*	B/W 1:62,000
04-26-84	USDA	539-179 358205 HAP-81F*	B/W 1:62,000
04-26-84	USDA	539-178 358205 HAP-81F	B/W 1:12,400
11-17-79	USDI	24 NPS GRSM 377-150*	Color 1:31,680
11-17-79	USDI	24 NPS GRSM 377-151*	Color 1:31,680
11-17-79	USDI	24 NPS GRSM 377-151	Color 1: 6,336
05-10-77	USPS	UAGII 3024 153.23-1266	Color 1:12,400
05-10-77	USPS	UAGII 3024 153.23-1265	Color 1:12,400
05-10-77	USPS	UAGII 3024 153.23-1264	Color 1:12,400
05-19-76	TVA	15:51 98126 34-44 H39000 GSVDUT	B/W 1:75,500
03-31-74	TVA	1444 EDT 157-4D-1X	B/W 1:24,000
10-15-73	USDA	A20 47155 273-15*	B/W 1:24,000
10-15-73	USDA	A20 47155 273-14*	B/W 1:24,000
03-30-63	USPS	1-108 GS-VARU*	B/W 1:31,680
03-30-63	USPS	1-109 GS-VARU*	B/W 1:31,680
04-29-61	TVA	165 4AR 13X	B/W 1:48,000
04-17-61	TVA	165 4AR 13*	B/W 1:48,000
04-17-61	TVA	165 4AR 12*	B/W 1:48,000
08-27-53	USDA	AOB-10L-106*	B/W 1:24,000
08-27-53	USDA	AOB-10L-105*	B/W 1:24,000

USDA: United States Department of Agriculture

USDI: United States Department of the Interior

USPS: United States Park Service

TVA: Tennessee Valley Authority

B/W: black and white

*: stereopair

X: due to poor quality or partial coverage photo not used

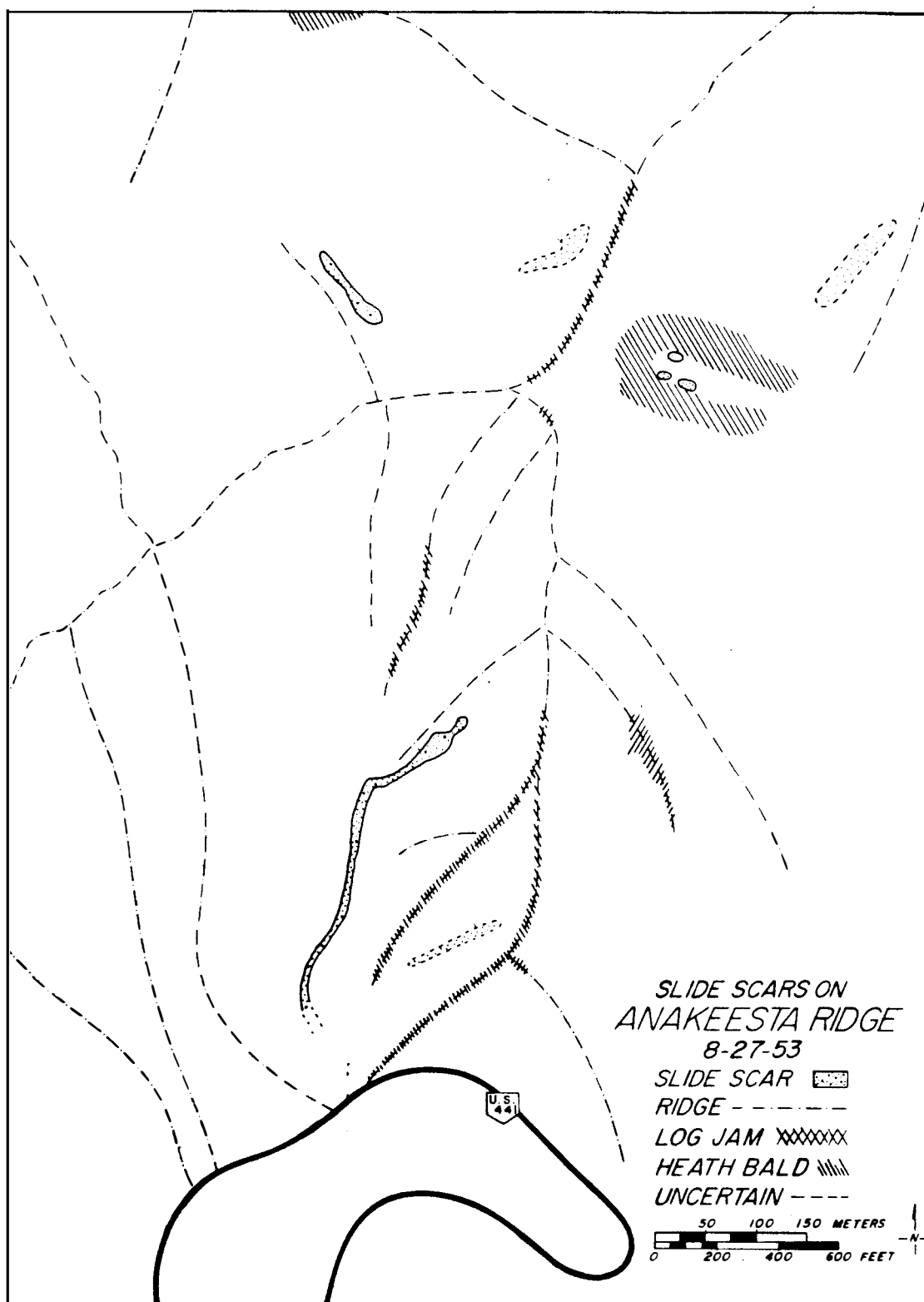


FIGURE 9. Slide scars on Anakeesta Ridge, 8-27-53

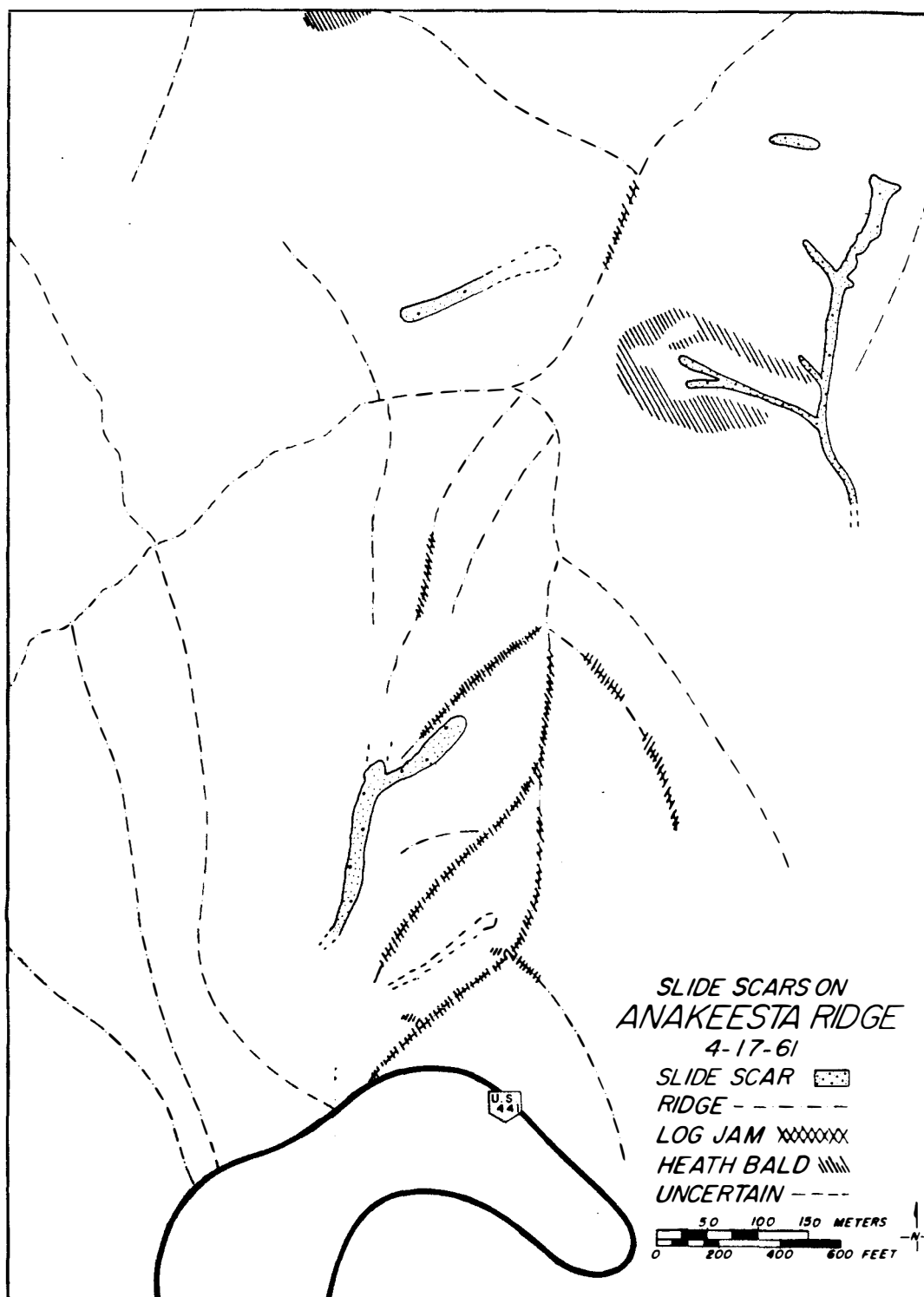


FIGURE 10. Slide scars on Anakeesta Ridge, 4-17-61

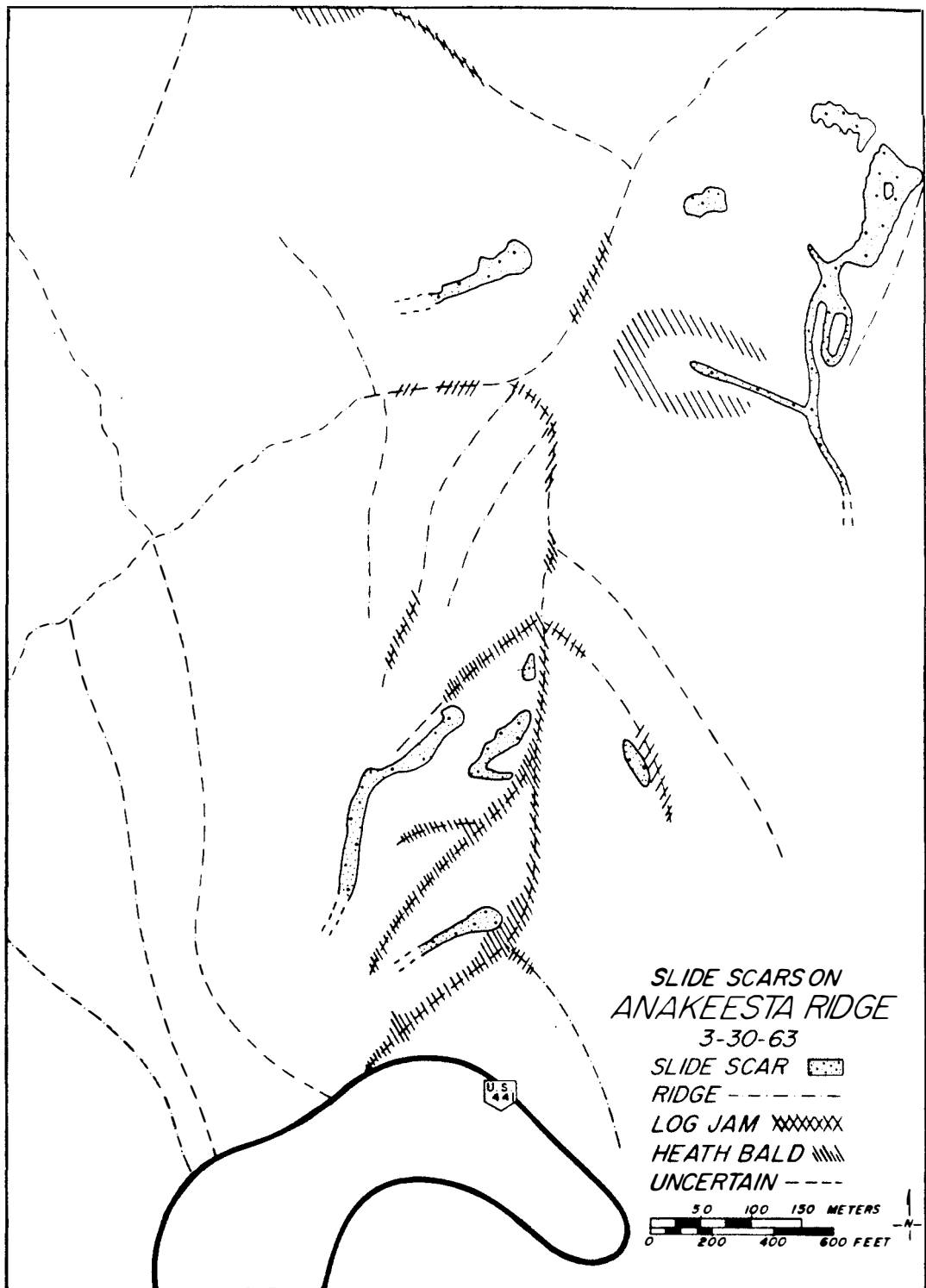


FIGURE 11. Slide scars on Anakeesta Ridge, 3-30-63

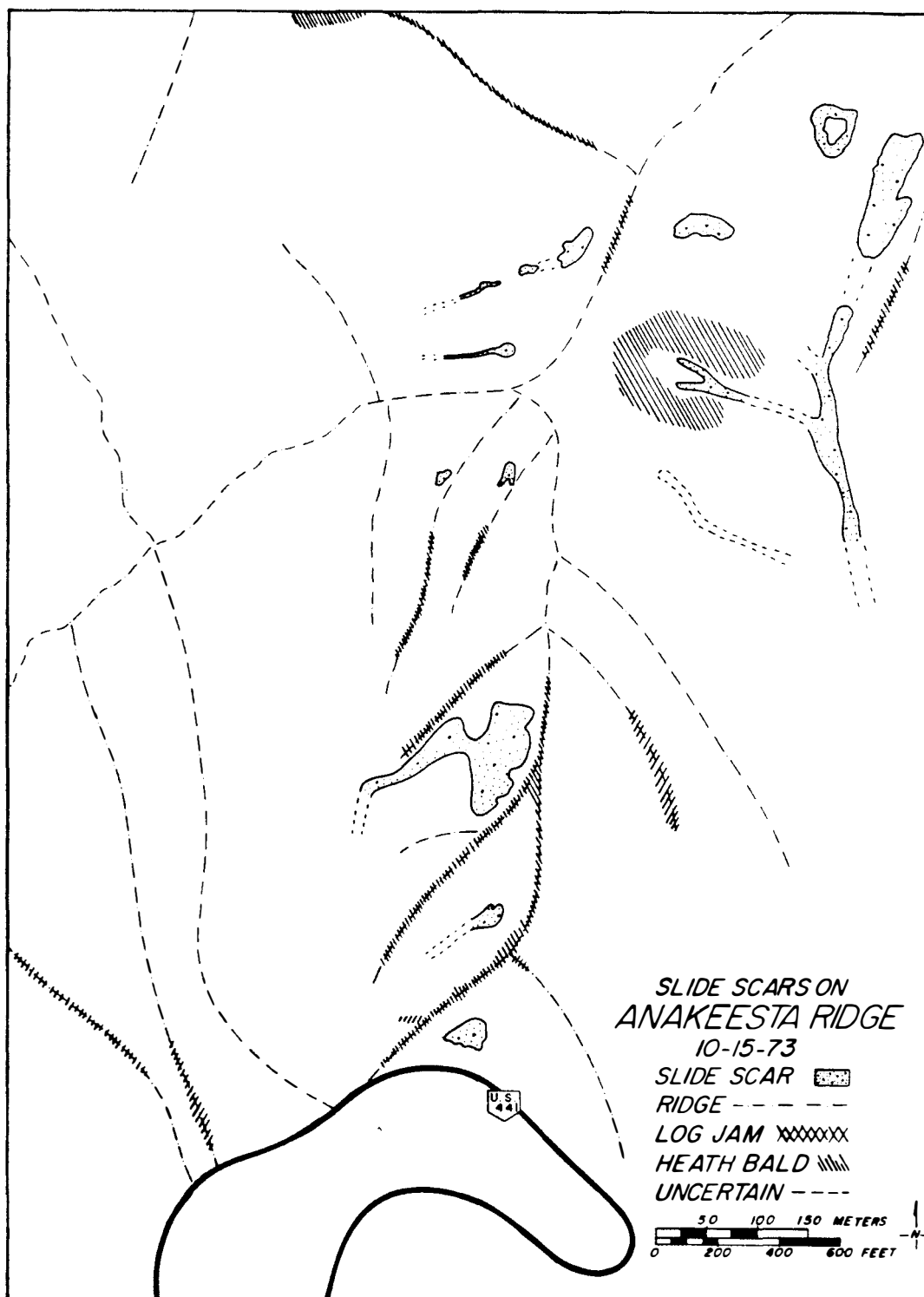


FIGURE 12. Slide scars on Anakeesta Ridge, 10-15-73

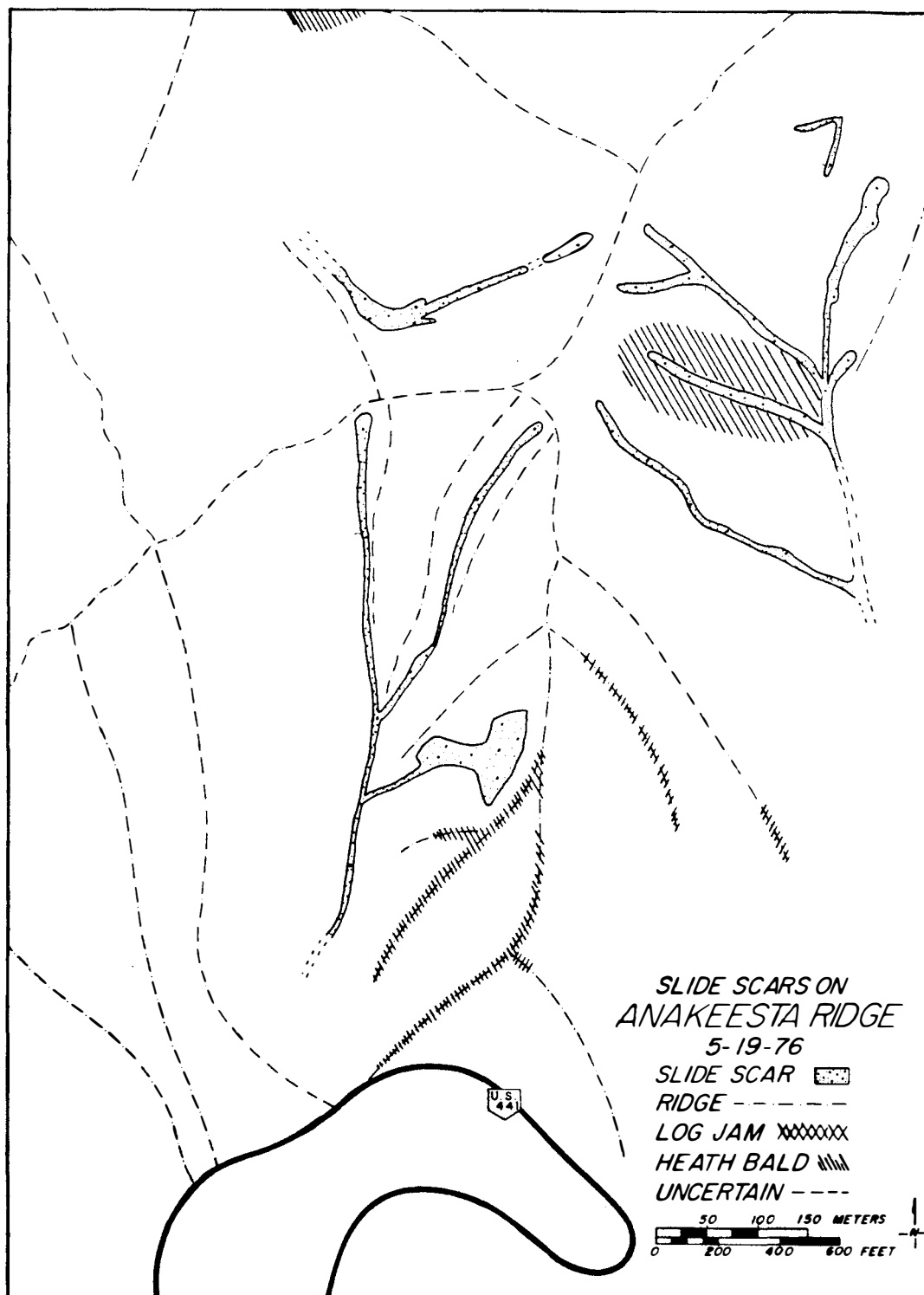


FIGURE 13. Slide scars on Anakeesta Ridge, 5-19-76

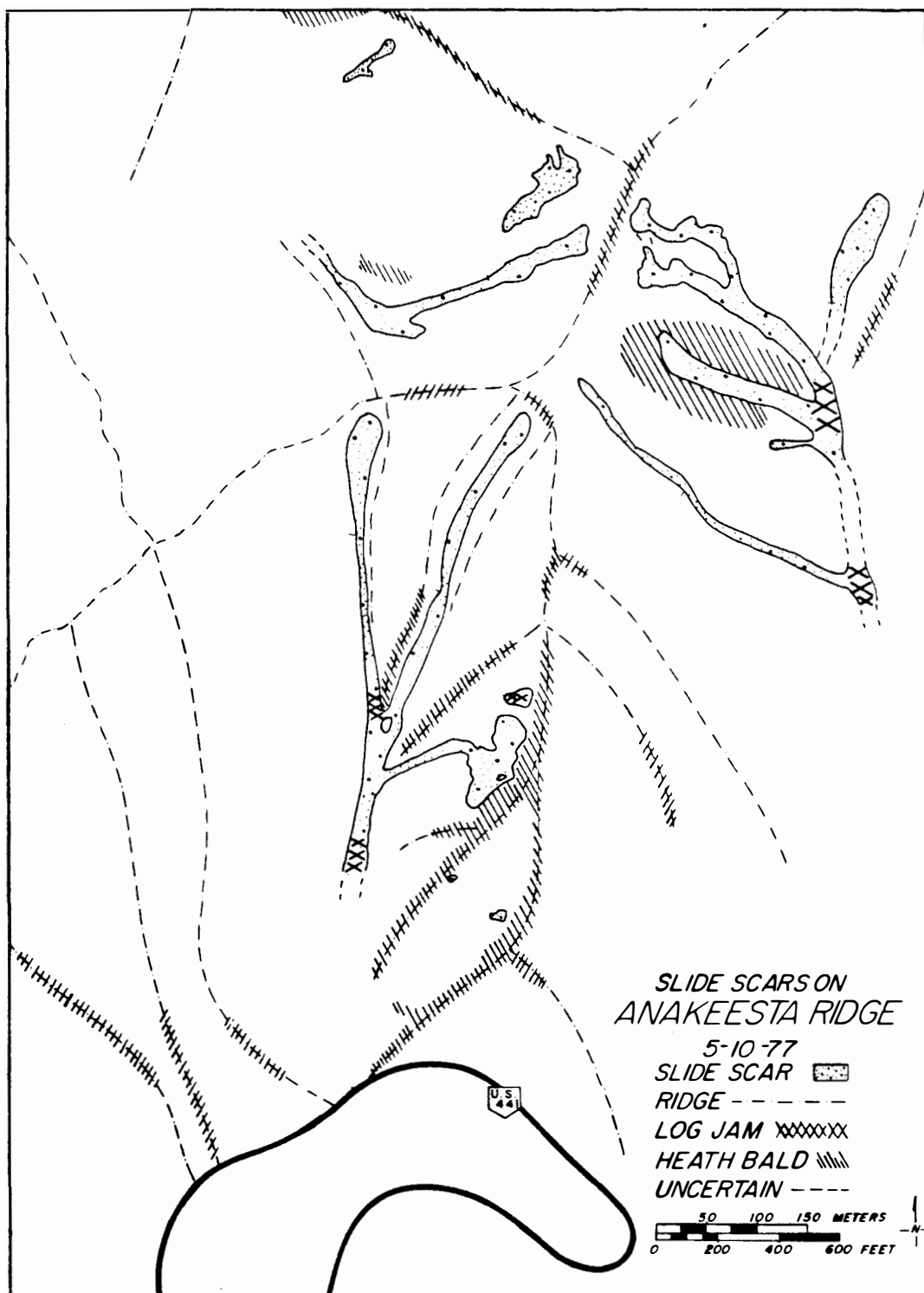


FIGURE 14. Slide scars on Anakeesta Ridge, 5-10-77

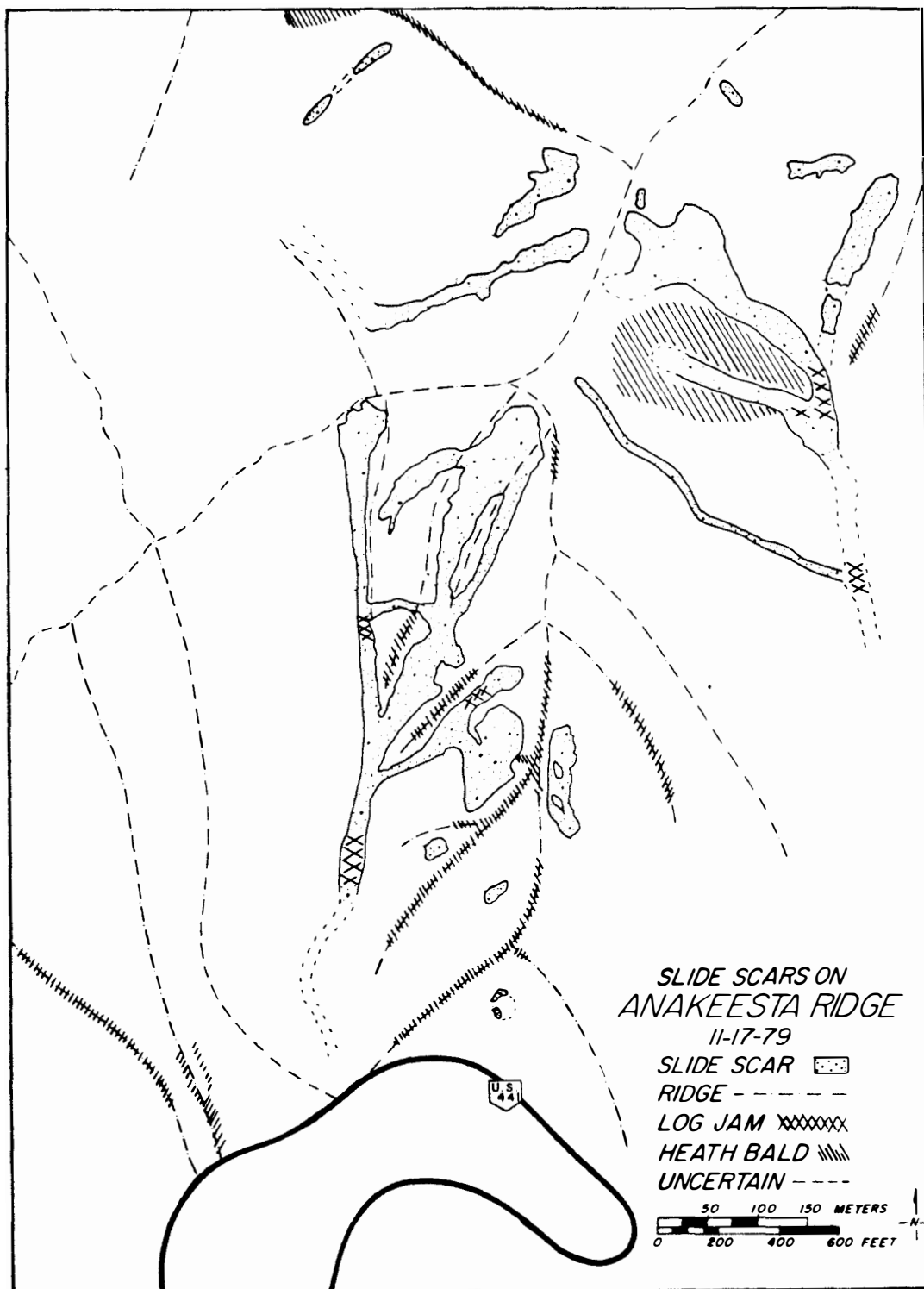


FIGURE 15. Slide scars on Anakeesta Ridge, 11-17-79

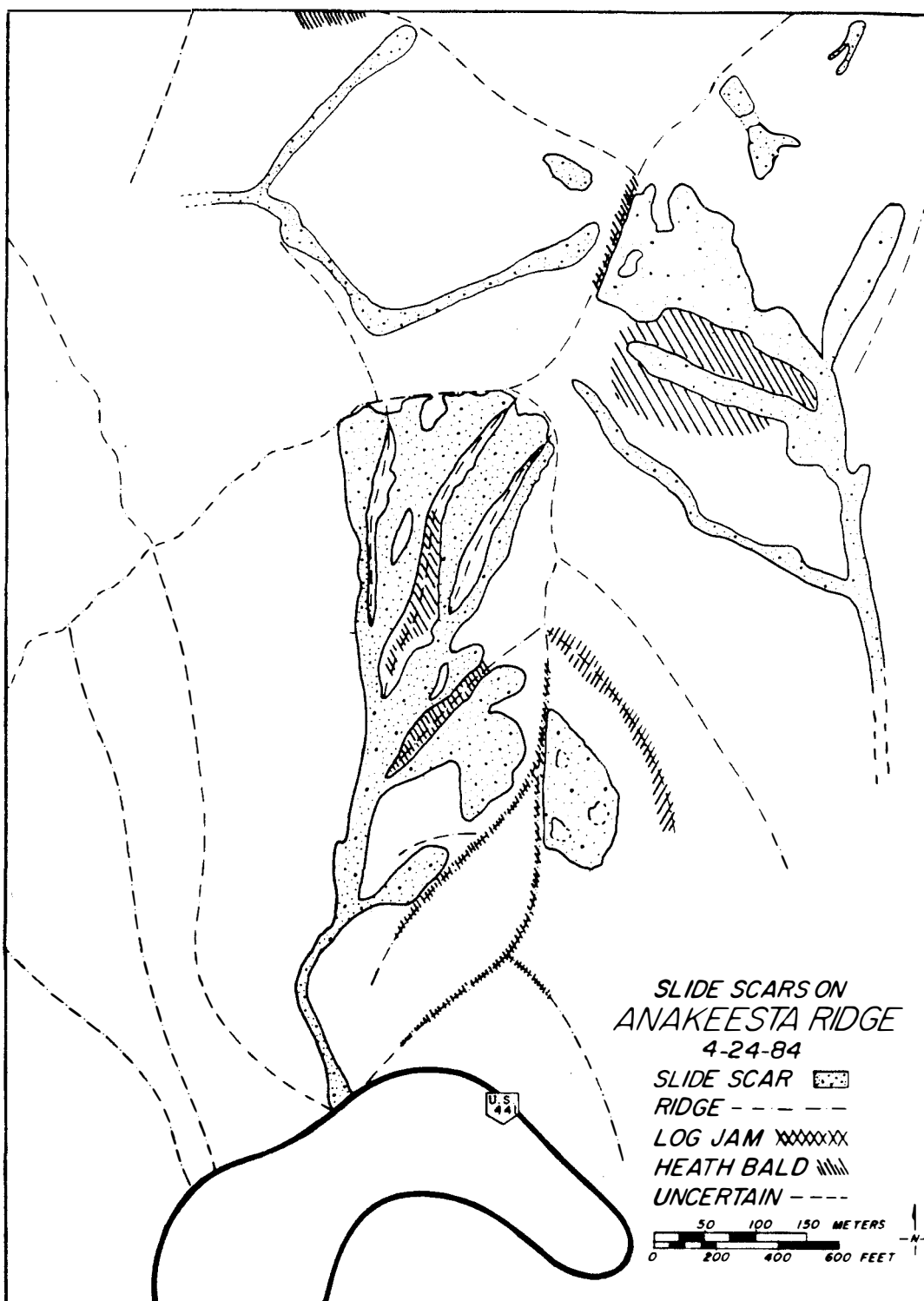


FIGURE 16. Slide scars on Anakeesta Ridge, 4-24-84

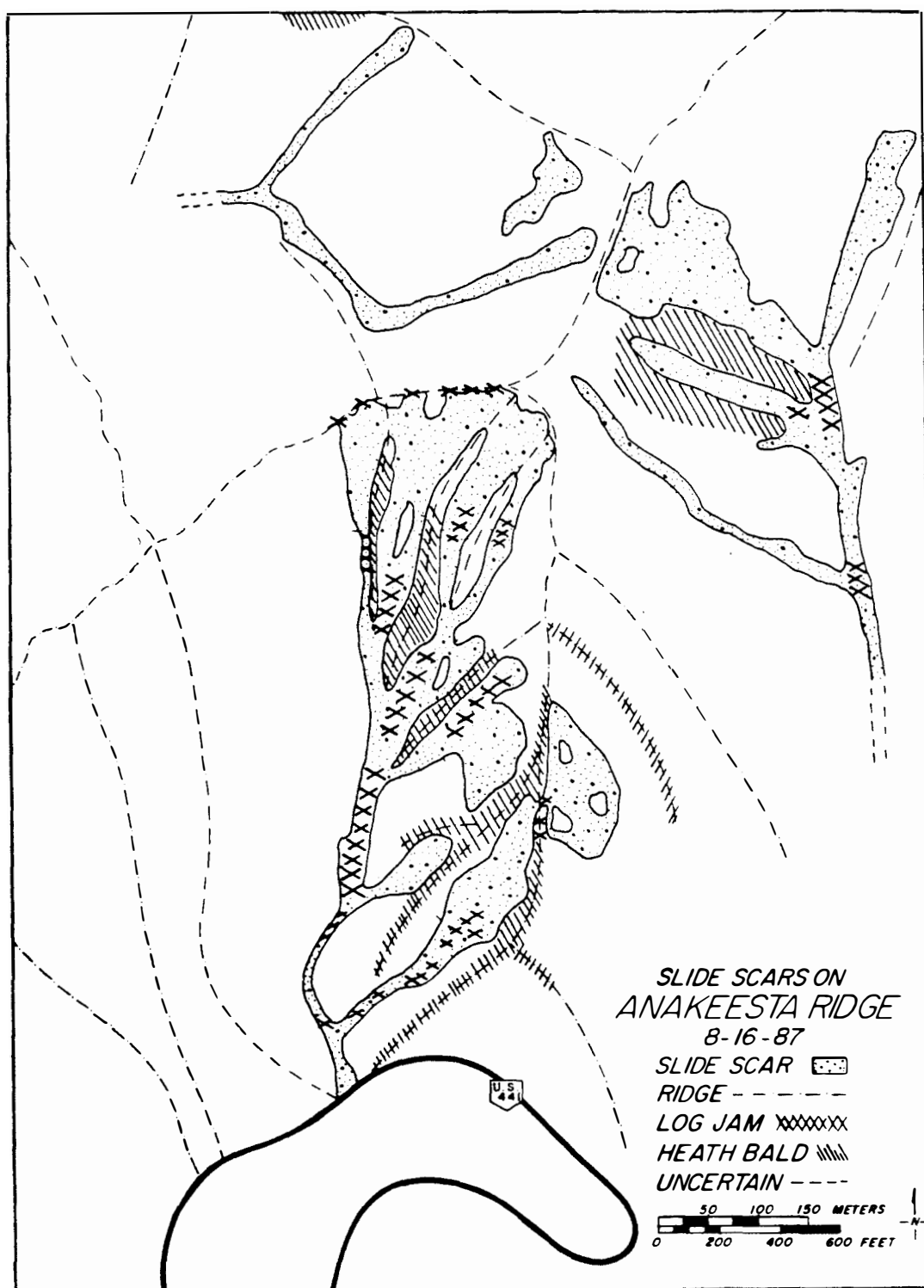


FIGURE 17. Slide scars on Anakeesta Ridge, 8-16-87

chronology (Hupp, 1983a). In areas where a debris flow removed vegetation and created sites for vegetation establishment, an approximate age for the debris flow may be obtained by coring the base of trees growing on the "new" surface (Hupp, 1984). Because the time between the mass movement event and plant growth is variable, only a minimum age may be estimated. Plate 2 is a picture of a tree impacted by a log jam. Botanical studies, when combined with geomorphic evidence, can provide a means for better estimation of magnitude and frequency of debris flows over a multi-year period (Hupp, 1984). The relatively short aerial photographic record can be extended by using dendrochronology to date earlier slides (Dietrich, et al., 1982). In conditions of stress such as drought, annual growth-rings may be eliminated altogether, however, within the historic period, dendrochronology remains a useful tool (Bowen, 1978). Hupp (1984) has found that dendrochronologic dating methods for debris flows proved consistent with available documented records of debris flows.

TABLE 3. Dendrochronology data (see figure 18 for location of tree coring sites)

SITE	MINIMUM SLIDE DATE
F2A	1919-1922
F2B	1929-1930
F2C	1792
F2D	1791
F3A	1749-1751
F3C	1837-1840
F4A	1952-1953
F4B	1951
F4C	1764-1769
F5A	1938-1941
F5BA	1942-1941
F5BS	1951-1952
F6A	1971
F7A	1932-1934
F8AA	1945-1951
F8AB	1932-1937



PLATE 2. Yellow Birch impacted by log jam at site F7.

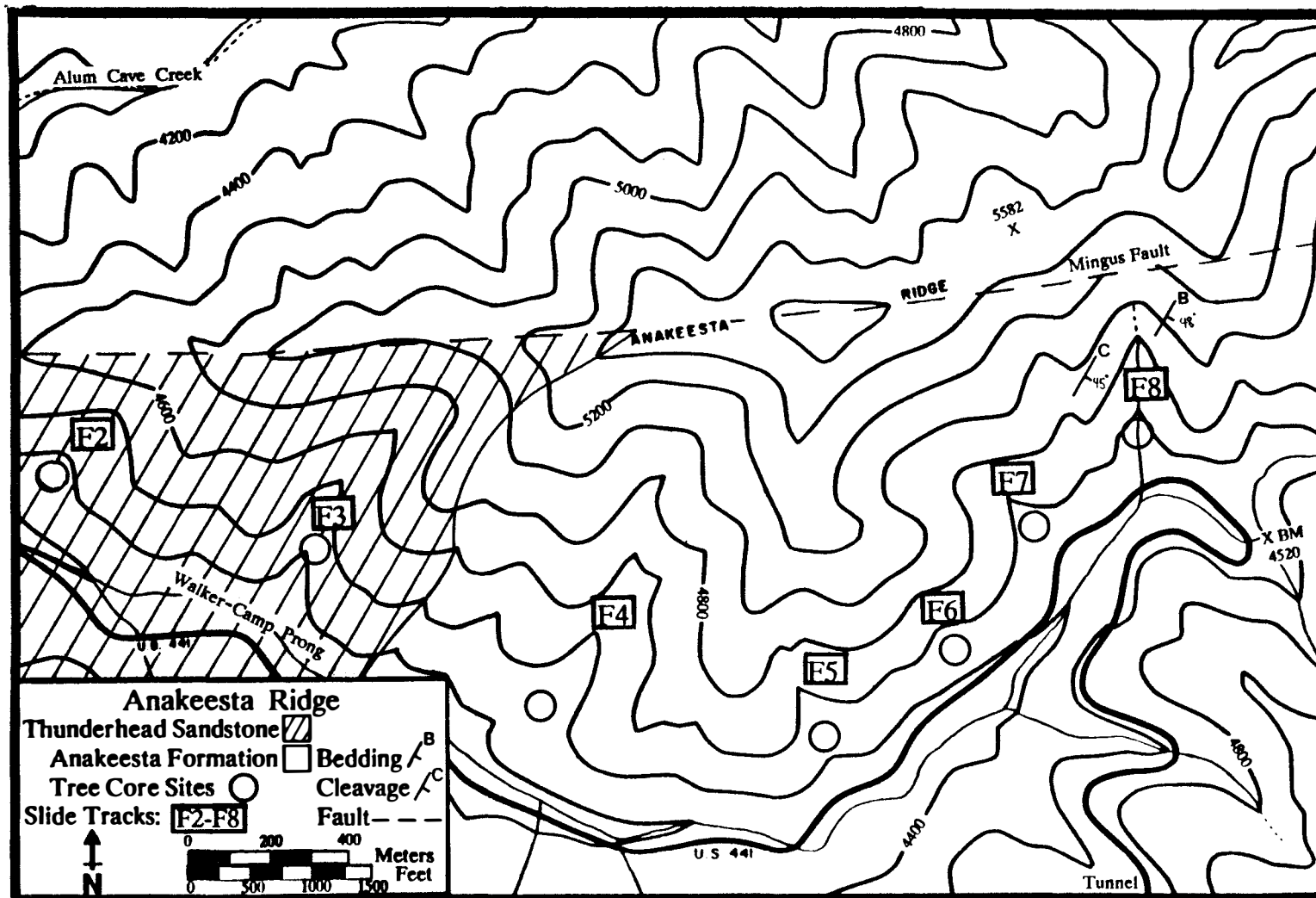


FIGURE 18. Location of tree core sites and slide tracks along Anakeesta Ridge.

Debris Volume

In determining debris volumes, adjacent hollows can serve as a model for preslide geometry. The majority of slide debris volumes were generated from the slide head. Observations of numerous Blue Ridge debris slide/flow chutes has led Scott (1972) to the conclusion that the average thickness of regolith removed is 1.22 m (4 feet); this is consistent with observations on Anakeesta Ridge. The slide scar area, and estimated volumes removed for 19 slide scars in the Alum Cave Creek watershed (Bogucki, 1970) ranged from 154 m² (1662 ft²) with 61 m³ (2160 ft³) of material removed to 10454 m² (112,532 ft²) and 3441 m³ (121,534 ft³) of material removed. Based on aerial photographs and field reconnaissance the compound slide scars on Anakeesta Ridge have increased in aerial extent and from 4,300 m² and 1790 m³ in 1953 to 128,000 m² and 84,100 m³ in 1987 (See Table 4).

TABLE 4: Slide scar development by area (m²) and volume (m³)

DATE	SLIDE SCAR AREA	SCAR HEAD AREA	DEBRIS VOLUME
08-27-53	4,300	1,470	1,790
04-17-61	12,000	3,850	4,690
03-30-63	19,600	10,800	13,200
10-15-73	21,400	12,600	15,400
05-19-76	25,700	12,500	15,200
05-10-77	44,400	21,900	26,700
11-17-79	68,900	39,000	47,500
04-24-84	116,000	50,000	61,000
08-16-87	128,000	69,000	84,100

Debris Fan

The debris fans involve a range of size and composition of materials, from clay to boulder size weathered and unweathered Anakeesta phyllite and Anakeesta sandstone, vegetation, and soil. Sedimentological characteristics common to fans formed by debris slides include extremely poor sorting, angular clasts and the absence of sedimentary structures (Wilson and Kochel, 1984). Debris fans can persist for thousands of years as topographic features and act as occasional

sources of sediment (Dietrich, et al., 1982). "The reason Blue Ridge debris flows are preserved and little affected by subsequent streamflows is that they are so coarse, normal stream flows are incapable of reworking the deposits (Costa, 1984)." This is true of the debris fan at the base of slide scar A, where elongate boulders up to 3 m (10 ft.) in length and approximately 3600 kg mass (7900 lb) have accumulated.

Velbel (1987) suggests that debris flows in the Southern Blue Ridge are derived from source materials that lack sufficient amounts of coarse clasts to form clast-supported debris flow deposits. Although the Anakeesta phyllite breaks apart quite easily, the Anakeesta sandstone member has proven to be strong enough to withstand a 430 M (1400 Ft) passage down the slide track A and maintain 2 plus meter (6.5 ft) lengths. This material does form sections of clast supported flow deposit. The underlying Thunderhead Sandstone also possesses enough integrity to form clast supported fan deposits.

Slope Angle

The short-term effect of a slide event is to make the ground surface irregular while in the long-term, landslides decrease the average slope angle until a "stable" slope angle is reached (Young and Young, 1977). Many researchers have measured slope angles in and about landslide areas seeking to determine this critical angle of stability (see Table 5). Scott (1972) feels that 30° is the approximate maximum angle of long-term stability of much of the surficial material covering the slopes of the Appalachians.

TABLE 5: Slide scar slope angles.

Slope Angle	Author(s)
35-44°, average = 40°	Bogucki (1970, 1976)
20-25°, (maximum range)	Beaty (1956)
30-32°, average = 30°	Bunting (1966)
> 15-20°	Costa (1984)
25-40°	Eschner and Patric (1982)
25-35°, average = 32°	Flaccus (1959)
38-43°	Hupp (1983b)
32-43°, average = 37°	Koch (1974)
8-19°	Lessing, et al. (1976)
> 35°	Neary and Swift (1984)
19-38°	Neary et al. (1986)
30-35°, average = 34°*	Scott (1972)
20-40°+	Sharpe (1938)
25-31°	Schneider (1973)
16-39° (26-31° most common)	Williams and Guy (1973)
43.4°, average upper slope angle	present study
33.6°, average mid-slope angle	present study
25.6°, average lower slope angle	present study

*: these data includes some road construction sites.

+: range of scar head angles.

Slope angles on drained, weathered debris are usually between 25° and 40° (Bloom, 1978) with slopes steeper than 40° usually barren of rock debris. Lessing, et al. (1976) determined that slopes which exceed 19° usually do not contain slide-prone soil. Slopes below 14° accumulate sediment (Neary, et al., 1986). Flaccus (1959) concludes that slides are not likely to occur on slopes that do not at any point exceed 25°.

Based on several slope angle measurements on the Anakeesta Ridge compound slide scars, the required steepness of slope for sliding ranges from 27° to 48°. Scar heads are at a high angle

(>40°), unsupported and unstable as a result of oversteepening from debris flow/slide activity. In this study all of the profiled slopes (A-G) are concave in profile (See Figure 19). The upper slope segments average 43.4° (range 47.5-36°), the mid-slope segments average 33.6° (range 38.5-27°) and the lower slope segments average 25.6° (range 35-15.5°).

This issue of what range of angles can be expected in the field is essentially one of determining what the angle of repose is for the study area debris. Because scar heads are generally in excess of 36° and barren of colluvium, and the transportational midsection averages 34°, and the toeslopes or debris fans average about 26° and are generally stable, the approximate angle of repose for Anakeetsa Ridge materials is in the range of 26° to 34°. For sake of comparison, an artificial, well-graded mixture of gravel with minor amounts of sand-sized material has an internal friction angle (loosely packed, so roughly the angle of repose) of 42° (Holtz and Kovacs, 1981).

Slide Azimuth

Writers such as Clark (1987), Hack and Goodlett (1960) and Beaty (1956) have found that north-facing slopes experience greater numbers of slide events. They attribute this to the microclimatic influence of slope aspect. Clark (1987) expects slopes with northwest, north, northeast and easterly exposures to retain soil moisture as they are shielded from direct solar insolation. Williams and Guy (1973) found that slopes facing north, northeast, and east suffered the greatest number of avalanches. They believe this may be because the lack of direct sunshine left these slopes with a greater prestorm moisture content and (or) the wind may have driven a greater amount of rain onto these slopes. The northeast facing slopes are wetter than slopes facing other compass directions because they are more protected from the drying effects of the sun and prevailing winds (Hack and Goodlett, 1960). The exposure to higher and more prolonged antecedent moisture conditions are experienced by shady slopes and those with a north-northeast aspect (Crozier, 1986). The slope aspect for landslides in a West Virginia study (Lessing et al.,

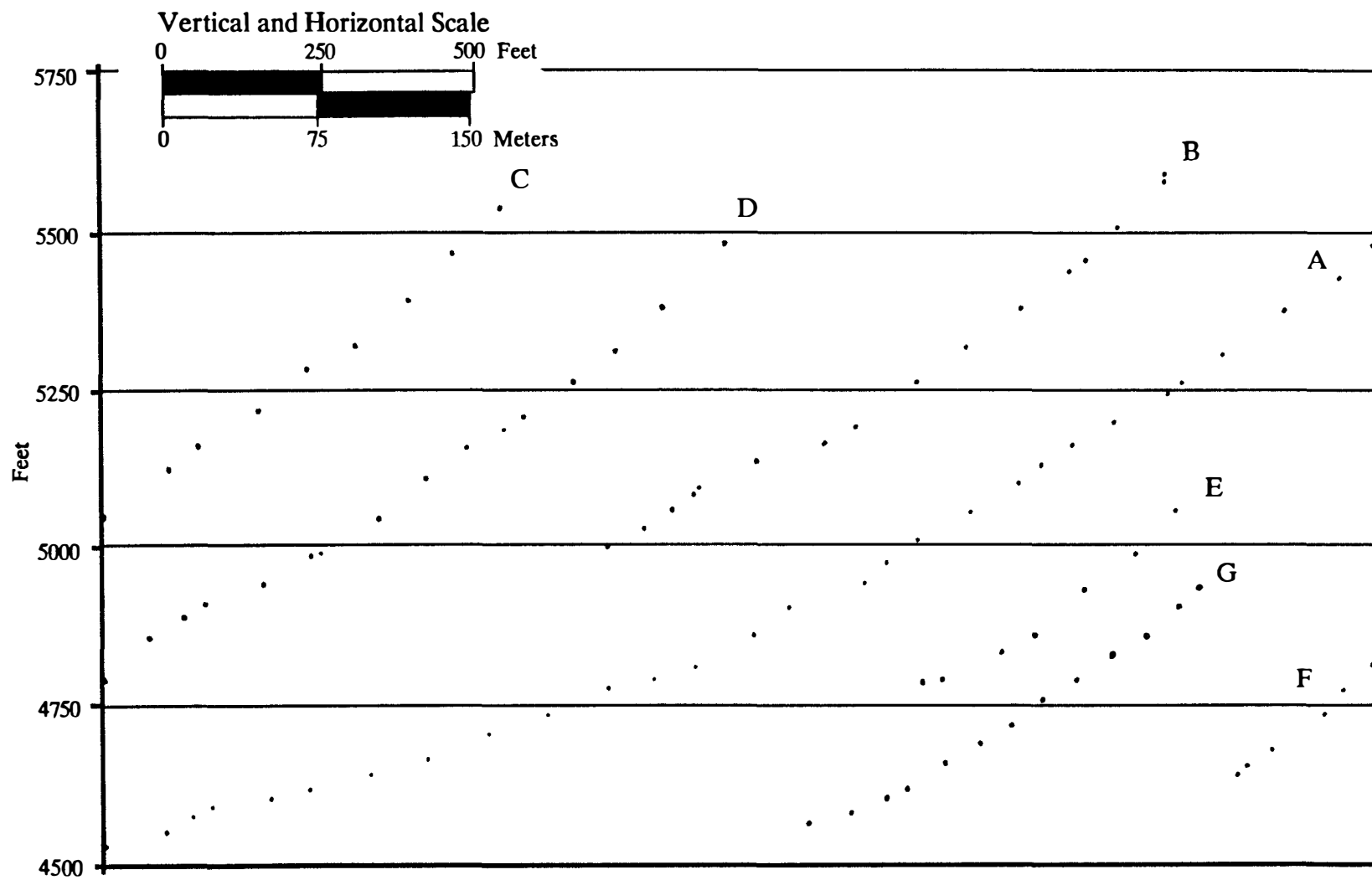


FIGURE 19. Anakeesta Ridge slide scar profiles

1976) covered all points of the compass although slightly more slides occurred on the northeast slopes. Beaty (1956) found that 70 percent of the landslides observed in his study occurred on slopes with northerly and easterly orientation. These slopes orientations receive less effective insolation and therefore would remain wetter following rain. Pomeroy (1982) observed a dominant northwesterly to northeasterly trend. Schneider (1973) and Koch (1974) found that south and southeasterly facing slopes experienced greater slide frequency. This is true of the Anakeesta Ridge slides. This can be attributed to the fact that cleavage, bedding and joint preferential release surfaces, are dipping to the southeast. Flaccus (1959) in his classic study of 135 debris slides in a granitic terrain concluded that compass exposure of slope has no significant effect on slide occurrence.

In this study, all of the investigated slopes possess a southern exposure. It is possible that slope orientation at this high elevation (over 1370 M; 4500 Ft) is less important than at lower elevations, because of the overall abundance of moisture in the Great Smoky Mountains.

Slide Location

With exception to those slides which occur immediately at roadcuts, the slides on Anakeesta Ridge have occurred in isolation from human influence in terms of removal of toe support. Human influence on vegetation through pollution is another question. Hack and Goodlett (1960) have found that it is common for almost all pronounced hollow sites to fail in intense events only. Slaymaker, et al. (1972) have found that 31/48 landslides originated in well-marked drainage depressions or seepage hollows; 14/17 were related to road construction. Bogucki (1970) has found that slides are most numerous in low order valleys. 81 percent of the Anakeesta Ridge debris slides/flows originate in drainage depressions. This is to be expected because water is concentrated in hollows.

Location of Slide Initiation

Salient factors concerning the location of future slides include hollow position and water concentration. Flaccus (1959) favors slide initiation on the steep upper slopes with movement initiated by sliding at the head. Williams and Guy (1973) and Schneider (1973) agree that debris slides tend to be initiated in the zone where the local gradient is the steepest. Scott (1972) feels that it is more logical to assume that incipient slope failure occurs in the central slope segment, followed by an uphill progression of debris dislodgement. This makes sense from a hydrological point of view that a slide originates in the mid-slope area: not only are the slope angles sufficiently high, in the 30° range, but there would exist a sufficient distance for a hydrologic head to generate pore pressures high enough to effect a slide; with failure of the mid-slope, the upper-slope would be unsupported such that failure for that area would either take place immediately or at least be in a position of instability for future failure. Sequential aerial imagery for Anakeesta Ridge indicate that slope failure is initiated in the mid-slope region, with the upper-slope placed in a position of instability for future failure.

Scott (1972) has noted that very few scar heads breach the ridge crests. He feels this is due to a diminishing uphill catchment area which is responsible for a sufficient water supply for soil saturation and subsequent movement. Along Anakeesta Ridge, however, the ridge crest is breached in no less than 7 places. The breaching of ridge crests is generally found in association with tree-throw. This indicates that pore pressure is not the primary factor in continued scar wastage. I contend that porewater pressure does not play a significant role after the initial event because the slope has made the undrained to drained transition. The slope will regain its undrained condition once revegetated and weathered material accumulates.

Failure Plane

The position of the failure plane is not always mentioned when debris slides are discussed. In fact, many times it is assumed that the failure plane occurs at the rock-soil interface. In this

study, this was not the case: the failure plane occurred below the rock-soil interface within the weathered Anakeesta phyllite along cleavage and bedding planes and joints. The relationship between soil mineralogy and slope failure is unimportant due to the fact that the majority of slope failures occur below the rock/soil horizon. Bedding, joints, and cleavage planes serve as release surfaces in the capacity of preestablished planes of failure. The failure planes are an orientational coincidence between slope and structure.

Rock Type and Slide Frequency

Feldkamp (1984) determined that 8 of the 9 debris slides in the Mt. LeConte area occurred on the Anakeesta Formation; the other slide is underlain by the Thunderhead Sandstone. Bogucki (1970) found that 63 percent of the September of 1951 slides occurred on Anakeesta phyllite as opposed to 37 percent on Thunderhead Sandstone. The majority of wedge failures along Tennessee highways occur in the Blue Ridge province and are associated with shale or slate lithologies (Moore, 1986). In a study of central Appalachian Plateau province landslides (Outerbridge, 1987) slope failures are closely related to the underlying lithology; shales are principally involved in earth flows and debris slides. Bloomer and Werner (1955) found that the great majority of debris avalanches in the Blue Ridge region in Central Virginia occurred on slopes underlain by gneiss and granite. Kujansuu (1971), however, found that only in exceptional conditions, such as prominent relief, prevalent jointing and weathered rock, have slides of crystalline schists or plutonic rocks occurred.

Any sulfidic rock, especially those sedimentary rocks formed in anoxic conditions that contain iron-disulfides are subject to the generation of acid by weathering through excavation (Byerly and Middleton, 1981). Rocks of the Great smoky Group and Ocoee Supergroup have the highest potential for creating acid problems (Byerly and Middleton, 1981). Weathering through acid generation weakens rock and the acids generated kill vegetation; any root strength supporting the slope is then lost. On a regional scale in the Blue Ridge Province, the following groups and

formations are mapped on a scale of 1:500,000 as “sulfidic” slate (North Carolina Geological survey, 1985): the Great Smoky Group (413 km², 157 mi²), the Boyd Gap Formation (67 km², 26 mi²) and the Wehuttu Formation (457 km², 174 mi²). These mapped units represent a large package of pyritiferous rock (937 km²; 357 mi²). Due to the acid potential and cleavage, a slope that consists of “sulfidic” slate and possesses the requisite planar or wedge failure orientation, may be a good candidate for slope failure.

Ongoing Slope Processes

Several slope-modifying processes are continuing to operate on the slide scars of Anakeesta Ridge. Transportational processes operating on Anakeesta Ridge include creep, overland flow (fluvial transport), and debris sliding (Ryan and Clark, 1988); additional slope modifiers include: needle ice, faunalurbation, slaking, and bank slumping. Needle-ice is active on barren debris. Pitty (1971) determined that piprake (needle-ice) is capable of lifting blocks of material weighing up to 9.5 kg (20.9 lb). Bank slumping occurs along the slide chutes.

Water is seldom an important factor (Bell, 1983) in causing rockslides; freeze-thaw action, however, is an important cause. The barren slide scar is now exposed primarily to slabslides from this activity. Slope retreat is accomplished by sheet wash, mass movement and tree throw. Appreciable amounts of fine sediment are moved downslope by slopewash.

Weathering to depths of at least 1 m (3.3 ft) was observed at a recent (Winter of 1986) slope failure at Newfound Gap within the pyritiferous Anakeesta Formation. Weathering is part of the processes that Crozier (1986) refers to as “slope ripening.”

In the humid-temperate zone more material is moved downslope by creep than by surface wash (Young and Young, 1977). Erosion at the scar head is active, minimizing the development of soils. The debris is moving “by creep and other slow mass-wasting processes into hollows, where it accumulates until at rare intervals it is flushed from the hollows by debris slides (Mills

et al., 1987).” The rate of near surface soil creep in humid temperate climates is about 1-2mm/year (Young and Young, 1977). Creep includes the imperceptibly slow near-surface movement, of freeze-thaw displacement and raveling and spalling of surface debris. In the channelway, the action of creep is relatively unimportant (Hack and Goodlett, 1960).

Steep slopes in the Smokies are covered with a relatively thin residuum as the steep slopes do not permit a thick accumulation. Production of colluvium, however, exceeds the rate of removal by fluvial processes. Most slope failures (Lessing, et al., 1976) in the Allegheny Plateau of west Virginia “do not involve bedrock, but are confined to the soil, colluvium, or weathered rock veneer.” Much of the ground surface is occupied by loose or friable deposits including soil (Hunt, 1986). Large areas of mountainous terrain are typically covered by a thin mantle of loose debris (Hack and Goodlett, 1960). Tree-throw continues to operate proximally to the Anakeesta Ridge slide scars.

Tree-throw is present at every breach of a ridge crest in the study area and is common along the unfailed slopes of Anakeesta Ridge. In a transect of Anakeesta Ridge, 73 isolated tree-throw sites were observed. In terms of biogenic transport, tree-throw importance is clearly expressed by the microtopography created by decaying tree-throw mounds (Dietrich, et al., 1982). Tree-throw may perhaps be the most under-rated geomorphological agent on certain hill slopes (Hack and Goodlett, 1960).

Several slopes have been monitored with scar head and scar flank wastage stations to determine the rate of backwasting and the relative soil and vegetational mat stability (see Figure 20 and Table 6). Erosion that took place along scar heads was measured by using either indentations in the bedrock immediately downslope or measurement was taken from a large tree in the area (see Figure 21). The most recent slide scar, G, exhibits the greatest amount of erosion at 37.2 cm. The ridge of slope G was breached during the period of study (see Plate 3). Note in Plate 3 the truncation of an older debris slide/flow fan (lower left-hand corner) by more recent events responsible for main chute (from photograph center to lower right-hand corner). Tension

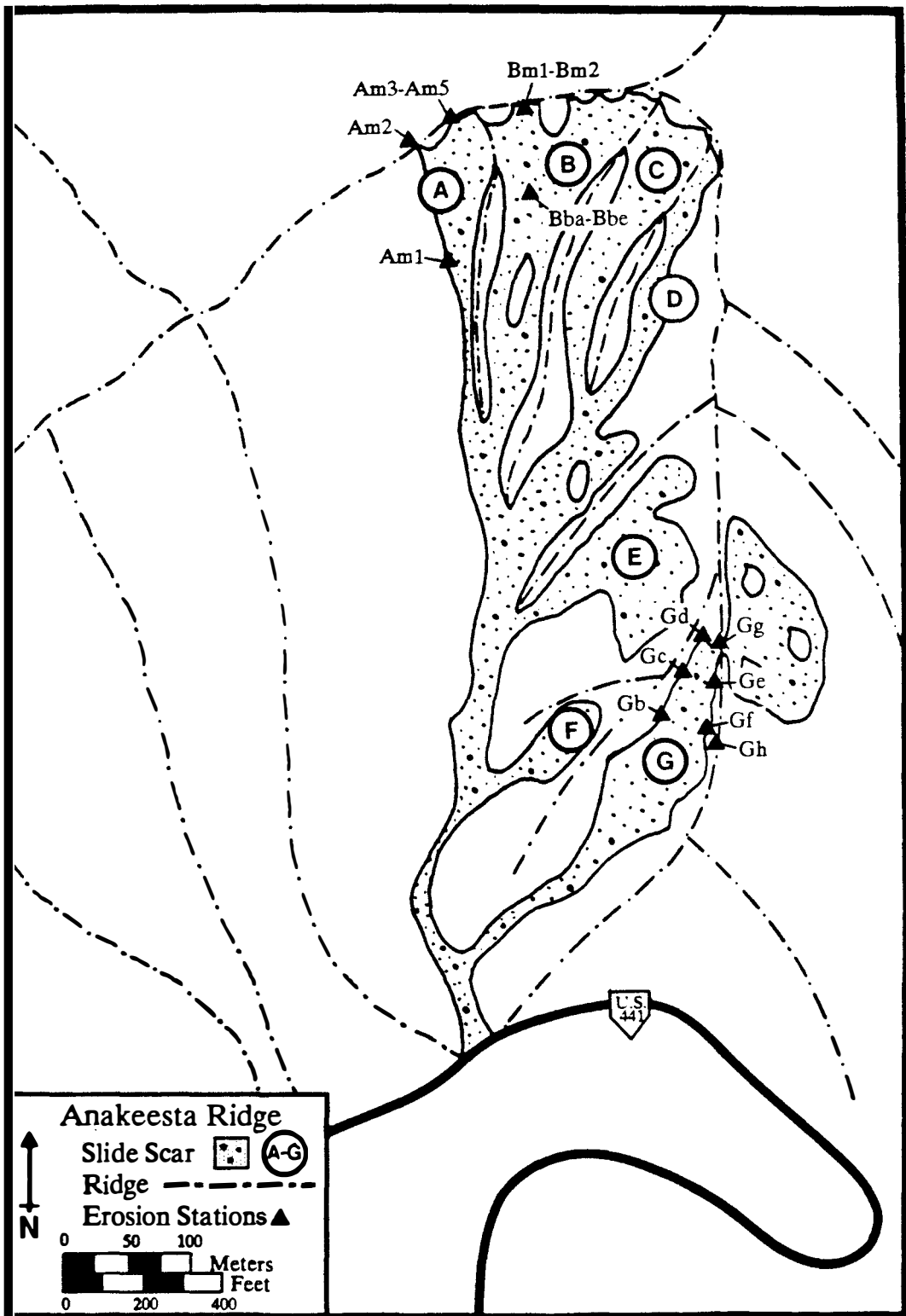


FIGURE 20. Anakeesta Ridge slide scar erosion stations.

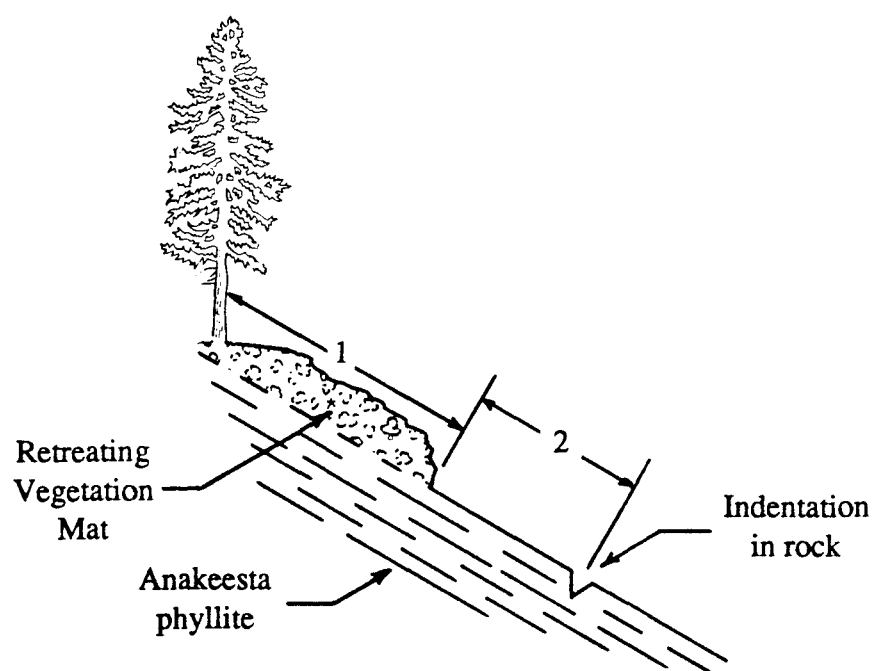


FIGURE 21. Two methods for measuring retreat of vegetation.

TABLE 6. Slide scar erosion rates.

Retreat of Slide Scar along Scar G (cm)						
Date	Stations					
	B	C	D	E	F	G
11-23-86*	-	-	-	-	-	-
11-29-86	0.0	0.3	3.0	1.5	12.5	1.5
12-04-86	0.0	0.9	0.0	0.0	15.8	0.0
02-04-87	0.0	0.0	0.0	0.9	10.4	0.0
05-01-87	#	#	24.4	3.6	14.0	0.0

Retreat of Slide Scar along Scars A and B (cm)							
Date	Stations						
	AM1	AM2	AM3	AM4	AM5	BM1	BM2
12-07-86*	-	-	-	-	-	-	-
12-14-86	12.5	0.0	0.0	1.2	0.9	0.0	0.0
01-08-87	1.5	0.0	0.0	9.1	0.0	18.3	0.0
02-04-87	0.3	+	+	+	0.0	0.0	0.0
06-30-87	16.2	+	+	37.2	0.0	21.3	0.0

* Erosion Station Established -

Station lost due to slaking of phyllite, no discernable movement of scar edge, however.

+ Data not collected



Plate 3. Breach of ridge crest at head of scar G.

cracks in vegetational mats are present along all scar-head fringes, and are ubiquitous along vegetated slopes. As opposed to continuing scar-head wastage, the chute flanks are more stable as indicated by undisturbed vegetation growth. Several hanging blocks located in the scar-head area of scar B were monitored from 12-17-86 to present and no discernable movement was detected as measured along major joints along the blocks.

Vegetational Effects

Vegetation is thought to have a key role in slope stability for a variety of reasons. Holmes (1917) wrote that soil will become more thoroughly saturated in a short period of time in a forest-covered area as opposed to a cultivated slope where water would run-off. Flaccus (1959) agreed with Holmes (1917) indicating that

it might be claimed with more justification that maturing forests would tend to increase slide susceptibility. There is first of all the weight of the forest itself added to potentially unstable slopes. Secondly, the change from little or no forest cover to mature forest should involve increase in infiltration capacity, reduced runoff, and consequently increased water retention during storms. This might increase susceptibility by adding more weight at the same time that the shear resistance of the soil is reduced by saturation.

Hack and Goodlett (1960) add that frictional resistance of vegetation and its ponding effect during heavy precipitation is important factor in debris slide origination. Flaccus (1959) provides support to this argument when he found in the White Mountains 93 of 105 debris slide sites occurred in unlogged areas. On the other hand, however, Scott (1972), and Schneider (1973) conclude that a healthy forest cover will reduce the incidence of debris slides. Bishop and Stevens (1964), Swanston (1969), and Crozier (1986) all document slope instability resulting from deforestation.

Crozier (1986) introduces to this discussion the concept of a cohesive versus a cohesionless slope. A cohesionless slope, as defined by Crozier (1986) is composed of a well drained, competent material such as a jointed sandstone. A highly jointed rock slope would represent a

drained slope. This type of slope has the ability to drain under loading conditions. Conversely, a cohesive slope is poorly drained, and consists of incompetent material such as weathered and slaked phyllite. For the rest of the discussion, the terms cohesive and noncohesive will be referred to as undrained and drained, respectively. Crozier (1986) indicates that roots and the corresponding vegetation mat can render a drained slope, undrained. The more complete the vegetal cover, the greater the amount of material displaced: this is due to a dense network of roots that will trap large amounts of unstable material (Crozier, 1986). Drained slopes that become covered by a dense root network may “convert” to a undrained slope type and root wedging may reduce internal friction to a value that is less than the slope angle (Crozier, 1986). Birot (1962) determined that the force set up by a plant with a root diameter of 10 cm, and 1 m long is capable of moving a 41 ton block. Eyles, et al. (1978) however feel that root networks at least temporarily offsets any reduction in strength resulting from wedging.

Due to the shallow-rooting nature, large tree height to root depth ratio of the trees along Anakeesta Ridge, tree throw is a problem, which in turn initializes sites for erosion. Tree throw is common on Anakeesta Ridge along slide tracks and within the forest along the ridge. Trees were not sheared, but uprooted, this reflects the high tree height to root length ratio. The shallow, spread out root system contributes to this phenomenon; these sites may concentrate water to the subsurface and possibly initiate sliding. Tree levering may augment root wedging whereby a tree may act as a simple lever that transmits stress to the root system (Crozier, 1986). With this in mind, the trees on steep slopes in the Smokies may actually have an optimum height at which tree levering is excessive.

Deep and extensive root systems from trees promote slope stability, by mechanically securing the slope but also by removing large quantities of moisture via transpiration (Lessing et al., 1976); weight is decreased and water pressure is reduced. Boring, et al. (1981) found in clear-cut areas of mixed hardwood conifers of the Appalachian forests that a living hardwood root system remains after clear-cutting; the frequency of shallow landslides increase 3-10 years after

logging-this is the time gap between root decay and plant regeneration. Scott (1972) in his study of 1700 mass wasting sites, found no evidence that clear-cutting was a major factor contributing to eastern debris avalanching. Plant root systems contribute to the stability of hillslopes in several ways (Boring, et al., 1981): vertical anchoring is effective in stabilizing relatively thin soils and larger structural roots can provide buttressing or soil arching action between trunks (Boring, et al., 1981).

Beneficial effects of forests (Varnes, 1984) include: retaining moisture in leaves, branches, and eliminating water via evapotranspiration, vegetal debris on forest floor; all reduce runoff and hence erosion.

Anaheesta slopes are both undrained and drained: Barren, well drained and jointed slopes do not permit a build-up in pore pressure, thus they have a lower failure probability; some slopes are covered with vegetation, have an accumulation of regolith and a low drainage capacity; this portion of the slope is subject to tree leveraging and an increase in pore pressure.

Log Dams

Log jams will develop over the course of a sliding event at slope breaks and at turns and constrictions in slide tracks. Consequent revegetation challenges further debris passage. Subsequent colluvial sediment fill results in an capable dam. Log jams act as an effective debris dam until that time the logs rot into loss of structural integrity.

In an Idaho study, Megahan (1982) found that logs lost 97 percent of their effectiveness within 6 years of emplacement; big trees have better longevity and decay rates are slow for logs in a wet environment. Megahan (1982) also indicates that storage behind obstructions is an important component of the overall sediment routing through forested drainage basins. In fact, organic-debris dams are a major factor in regulating the loss of particulate matter from forested ecosystems (Likens and Bilby, 1982). Megahan (1982) determined that logs were the most

important type of obstruction because they had the greatest longevity and stored the greatest amount of sediment—an average of 15 times more sediment was stored behind obstructions than was delivered to the mouths of the drainages as annual average sediment yield. Dietrich, et al. (1982) developed a sediment budget to analyze the relative importance of physical processes in the transport of organic and inorganic matter through a steep, forested drainage basin and compared this budget with one for a less steep drainage basin forested with a younger stand. They found that forest vegetation plays a prominent role in many phases of sediment transport and storage which includes the transport of soil by windthrow of trees. Log dams are important to the overall erosion-sediment budget for forested drainage basins.

Organic debris dams are rare in streams larger than third-order but are a common and important feature of headwater streams in forested areas (Likens and Bilby, 1982).

Log dams are an effective way of slowing the fury of a debris slide, the dam, however, becomes a liability once the logs have rotted, thus losing their yield strength and contributing to the ensuing debris torrent.

CHAPTER IV

TRIGGERING FACTORS

Factors which influence the stability of slopes include: rock strength and structure, bedding, slope angle, landform, slope shape, soil depth and strength, clay mineralogy (Sidle, 1985). Factors which cause mass movements include structure, lithology, hydrogeological conditions and stage of morphological development of area (Zaruba and Mencl, 1982). Landslides do not occur in isolation; they are a product of their environment and in turn they influence its condition (Crozier, 1986). The action of a triggering factor (Crozier, 1986) only partially explains the cause of a landslide; destabilizing factors may have brought the slope into a condition where a minor change of regular occurrence could initiate movement.

Explaining the cause of a landslide involves a number of factors, however, if there is enough sediment, the slope is relatively steep (over 25°), and enough water is available, then a slide will occur. Depending on the condition of the slope and its susceptibility to failure, a variety of earth magnitudes could trigger a slope failure. Two main types of triggering thresholds may be identified for any given level of mass movement: (A) Minimum probability threshold (PT_n), below which the defined level of mass movement does not occur and above which it may occur under certain conditions; (B) Maximum probability threshold (PT_x), above which the defined level of mass movement always occurs whenever the PT_x is equalled or exceeded (Crozier, 1986). Wolman and Miller (1960) state that "almost any specific mechanism requires that a certain threshold value of force be exceeded." Landslides can be viewed as a threshold phenomenon (DeGraff and Romesburg, 1980). Bull (1980) considers a threshold as a balance between opposing tendencies; the part of the system being looked at may be considered to be at a threshold or equilibrium condition when the ratio of stabilizing forces to destabilizing forces is equal to 1. Bull

(1980) feels that the chief advantage of the threshold approach is to focus attention on feedback mechanisms and complex responses that are likely to cause the mode of system operation to change. Coates and Vitek (1980) believe the idea and conceptual base that thresholds offer can serve as a common denominator to link and unite geomorphic processes and landforms.

Crozier (1986) comments that most slopes are stable, or marginally stable most of the time, therefore an actual landslide represents a transient condition infrequently attained by the slope. Crozier (1986) recognizes the main difficulty lies with determining the full range of stress changes that can be brought about by transient factors such as climate and earthquakes. These require a full record in order to make an accurate assessment of the situation. Sharpe (1938) recognized that two conditions exist, active and passive: passive conditions include steep slopes, thin permeable soil, structure, and a high precipitation climate; active conditions include cloudbursts, earthquakes, and anthropogenic impact. Terzaghi (1950) uses the same type of categorization but uses the terms internal and external causes: internal causes are mechanisms within the mass which brought about a reduction of shear strength/shear resistance to a point below the external forces imposed on the mass by its environment, thus inducing failure; external mechanisms are those outside the mass which are responsible for overcoming internal shear strength, those that effect an increase in shear stress, thereby causing failure, include an increase in slope weight, removal of toe support, and earthquakes.

Crozier (1986) invokes different terminology for the categories set up by Sharpe (1938) and Terzaghi (1950), he uses the stability state terms, preparatory and triggering factors: preparatory factors are those factors which dispose the slope to movement without actually initiating it, these factors tend to place a slope in a marginally stable state; triggering factors are factors which initiate movement thus shifting a slope from a marginally stable to an actively unstable state. There exists a temporal variability of factors (Crozier, 1986) that can be categorized as: passive-those slow to change such as weathering, change in slope angle and slope height; and transient-active, fast-changing factors such as fluctuation of water in the slope.

The incidence of debris slides may be a coincidence of cycles, which, when in phase result in a mass movement. Cycles to be considered would include precipitation patterns, weathering, and optimum tree height/levering in conjunction with the yield strength of the tree-root network.

Intense Precipitation Events

Next to gravity, water is the most important factor in slope instability (Varnes, 1984). Rainfall is generally accepted as one of the chief factors controlling the frequency of landslides; climate, topography, permeability and structure influences the resulting magnitude (Zaruba and Mencl, 1982).

In the study area, there is an estimated 100 year recurrence interval for storms that shed 3-4 inches of rainfall per hour (Yarnell, 1935). Storms of marginal intensity would be more likely to trigger major slope failures in cases where the soil is previously saturated whereas dry soils would stand less of a chance for mass movement events. Wedge failures occur most frequently in late winter and early spring in association with east Tennessee's wettest period and winter thaw (Moore, 1986). The late winter and early spring are also associated with one of the highest periods of freeze-thaw cycles.

Scott (1972) recognizes that rainfall and soil water content is capable of the greatest relative fluctuation of many of the slope stability factors. It has been documented (Bogucki, 1970) that intense summer rainstorms, cloudbursts, are the most common cause of debris slides in the Appalachian Highlands. At least 93 percent of the slides are known to have occurred during heavy rain (Flaccus, 1959). Clark (1987) notes that the concentration of slides in the southern section of the Blue Ridge can be explained by their proximity to precipitation intensity maxima. Of the storms in which intensities were recorded, 24 are cloudbursts or thunderstorms, 5 involve frontal storm systems, and 5 are hurricanes (Clark, 1987). Bogucki (1970) observed that of 11 occurrences of multiple debris slide sites south of the glacial border in the Appalachian Highlands,

8 are associated with extensive storm systems with the other 3 related to intense, localized storms. Scott (1972) goes as far to say that "every known instance of debris avalanching within the Blue Ridge Province has been accompanied by a rainfall of exceptional severity."

Schneider (1973) has found that debris slides occur most frequently toward the end of periods of heavy precipitation. Pierson (1977) found that variation of piezometric head was most closely associated with 24 hour rainfall and that debris slides occurred with rainfalls between 5-5.5 inches (130-140 mm) which included 6 hour bursts of at least 2-3 inches (50-75 mm). The cloudburst from the storm of September 1, 1951 over the LeConte-Sugarland Mountain area of the GSMNP resulted in the formation of more than 100 debris slides (Bogucki, 1970): Clingmans Dome precipitation station recorded a 24 hour total of 2.87 inches; the Cataloochee Ranch precipitation station recorded a 24 hour total of 4.80 inches, and for 3 consecutive days leading up to the September 1 event, rainfall was 1 inch or greater. A bucket survey indicated that the cloudburst was about 4 inches in less than 6 hours. Eschner and Patric (1982) contend that during "exceptionally heavy rains, debris avalanches may occur anywhere in forested mountains of the eastern United States," but they are most probable when 5 or more inches of rain fall in a 24-hour period on slopes 25° to 40° where soils are less than 0.91 m (36 inches) thick.

Much mass movement occurs and most regolith landslides are initiated during intense precipitation events or during lesser precipitation events associated with prolonged wet periods. The periods generally result from persistent antecedent rainfall (Crozier, 1986). Koch (1974) and Moneymaker (1939) indicate that the origin of the debris slides are mainly the result of the saturation of residual material. Crozier (1986) notes that changes in water content can quickly affect the stability of slope material by increasing pore water pressure, imparting lubrication, increasing weight, and by decreasing "effective" cohesion.

Beaty (1956) noted that "virtually all discussions of landslides stress two factors: the accumulation of a thick mantle of decomposed material and the presence of excess water in the

ground. Landslides will occur if these two conditions prevail, regardless of slope declivity, type of vegetational cover, parent material of regolith, or any other environmental characteristic." In fact, rainfall events (Caine, 1980) may be so intense as to override structural, slope azimuth and soil moisture conditions.

The question of how much precipitation and what kind of intensity is required to effect a debris slide is one which has been long asked by a plethora of workers. The minimum amount of precipitation necessary to produce debris slides is difficult to determine as the amount of precipitation is site variable and there is variation of pre-storm soil moisture conditions (Scott, 1972). There is also considerable difficulty in obtaining storm intensity records as very few precipitation recording stations are equipped to administer such data. Recording stations are typically few and far between in regions that are susceptible to debris sliding. In this study, only four weather stations were available above 4000 feet. Crozier (1986) indicates that day-of-event rainfall provides only an approximation of the critical soil water status during a landslide event. Bogucki (1970) suggests a relatively dry soil provides high initial infiltration capacity and reduces surface detention; a moist soil may not have yielded such severe slope damage.

Bogucki (1970) has outlined several hydrological factors which may be involved with mass movement processes: water as a lubricant, elimination of surface tension, weight of water, pore water pressure and overland flow. Hunt (1986) indicates that most gravity deposits develop by becoming lubricated with water or ice. Terzaghi (1950) however, feels that water actually acts as a nonlubricant when in contact with many common minerals. The surface tension of the moisture provides various degrees of cohesion to soil in which voids are partially moisture filled. The lubrication of water and ice may be important once a landslide has been initiated in determining the run of a debris mass. Terzaghi (1950) also feels that weight of water is not a significant destabilizing factor. Shanks (1954) indicates that a water surplus exists throughout the year at all

elevations above 3800 feet. Terzaghi (1950) also adds that there is always enough moisture in humid climates in the ground to act as a lubricant for movement, therefore the idea that slides develop principally due to the lubricating effect of water is not true, but rather due to the consequent increase in pore water pressure.

The importance of pore water pressure is recognized by all authors who seriously address underlying debris sliding mechanisms. Flaccus (1959) feels that at least some of the White Mountain slides are best explained using three of Terzaghi's effects: reduction of cohesion, increase in unit weight, and rise in pore water pressure. Hutchinson and Bhandari (1971) find that if groundwater is already within the zone subject to surcharge, this sudden loading will prevent drainage and excess pore water pressure will develop, decreasing shear strength. Debris accumulating at the head of a mudslide for example, is an important mechanism in promoting downslope movement. Loading reduces drainage, increases pore pressure, decreases shear strength and increases shear stress such that shearing will be induced or downslope movement will be accelerated. The development of pore water pressures requisite for debris sliding is of course tied to intense precipitation events.

Subsurface Hydrodynamics

In most landslides (Bell, 1983) groundwater constitutes the most important single contributory cause. Chamberlin (1972) identifies interflow as the process of water movement parallel to the soil surface; this is caused by boundary conditions sufficiently restrictive to prevent normal vertical infiltration and percolation to a water table. The two main sources of runoff to streams are in the form of direct and indirect runoff. Shortly after a storm a type of underground, storm period flow, also called subsurface stormflow, occurs most commonly in forested areas (Whipkey, 1965). Whipkey (1965) has found that undisturbed forest soil is generally covered by organic litter which protects the soil surface and keeps it permeable to water infiltration. Roots, old root sites,

burrows and structural discontinuities also provide zones of permeability. Infiltrated water travels virtually unimpeded through coarse textured soil/rock. When a finer texture is reached, the water travels laterally over this zone of relative impermeability; at high subsurface stormflow, this will be a zone of increased pore pressure (Whipkey, 1965). Interflow in the form of subsurface stormflow has been invoked for some time as the dominant process contributing to quick stream response following precipitation events in forested regions (Chamberlin, 1972). The barren slide scars would offer an ideal surface for this quick response. I feel that the effective operation of overland flow is limited temporally to those months of the year when the debris is snow-free, unfrozen, and when high precipitation summer events occur; spatially to those surfaces unprotected by vegetation and coarse debris; and lithologically to the finer fractions of the debris. Goodell (1972) has found that the use of land for wood production, through its inherent reduction of forest cover and evapotranspiration, results in the increase in quantity of water yield-the slide scars represent a tremendous reduction of forest cover.

Swanston (1969) in his study of Alaskan slides, found that the slides occurred along drainage concentrations. Sidle (1985) observed that convex slopes disperse subsurface water and tend to be more stable than the water concentrating concave slopes like v-notched gullies. Because the influx of water generally triggers slide action, debris slides usually occur where gullies or depressions channel water into a relatively narrow zone (Lessing, et al., 1976). Their data indicate that 69 per cent of the landslides studies occurred on concave slopes. This is easily explained as these natural configurations concentrate water. "More failures are found on laterally and longitudinally concave slope segments, indicating that topographically influenced flow concentration is an important factor in determining relative slope stability (Mills, et al., 1987). Field observations by Williams and Guy (1973) confirmed that many debris avalanches (85 percent) did take place where indentations or incipient channels already existed on the hillside. Debris slides and debris flows often originate in topographic hollows at stream heads (Dietrich, et al, 1982). It

is possible that the hollows in the study area are structurally initiated as faults or intense fracturing is present along each slide track. Based on the amount of present water flow, at least seasonal streams were present on the preslide Anakeesta hollows.

Destabilizing Factors

The action of a triggering factor (Crozier, 1986) only partially explains the cause of a landslide, more significantly, destabilizing factors bring the slope to a condition in which a minor change of regular occurrence could precipitate a movement. The search for destabilizing factors should be focused on those which possess the greatest rate of change. Destablizing factors include: weathering, frost-wedging, root activity, tectonic upliftand fluvial toe removal. Depend- ing on the severity of freezing, frost wedging can affect the outermost zone of a slope to a maximum depth of about 2 m (6.5 ft).

The effect of surcharge (weight) is dependant upon the stress/strain properties of the slope material, permeability and the presence or absence of cohesion (Crozier, 1986). Natural processes which produce a variation in surcharge include (Crozier, 1986): precipitation, mass movement erosion/deposition, volcanic extrusives, thrust faulting, change in vegetation and variation in atmospheric pressure.

Geological Localizing Factors

Geological localizing factors include lithology, degree of weathering, and structure. When bedrock has been exposed by a landsliding event, the exposed rock is subject to more chemical weathering, this in turn sets up the remaining slope for rock sliding or other mass wasting processes. If triggering events are separated by long intervals there may be sufficient time for slope-ripening to occur (Crozier, 1986). An estimation of slope-ripening can be derived from Newfound Gap roadcut information. This roadcut was opened in the early 1960's (Walker, 1989),

and it has failed translationally as recently as the fall of 1986. It is interesting to note that NewFound Gap, in the area of the recent failure, is comprised of Anakeesta phyllite. Crozier (1986) has observed in inherently unstable regions, major landslide episodes may occur on an average of once every 5-6 years. Weathering effects a reduction in strength of slope material such that the competence of the underlying material is reduced, thus losing its ability to withstand superincumbent weight of overlying material (Bell, 1983). Scott (1972) indicated the importance of rock type in slide-prone areas because parent material is integral to the type of soil developed; a mineralogically more basic rock type possesses a higher weathering index, thus it would produce, more quickly, a thickness of potentially unstable regolith. For the 30-35° slopes of the Blue Ridge Mountains, Scott (1972) feels that a soil depth of several feet is necessary for debris avalanching to occur. This author disagrees: while it is important to have material to transport for a debris slide, that material does not have to be in the form of soil. In fact in the mountainous environment a thick soil is not to be expected; suffice it to say that rock weakened by weathering processes and discontinuities provide an abundance of debris.

CHAPTER V

CLIMATE

Climate as opposed to tectonics is the major controlling factor of Appalachian fan deposits (Velbel, 1987). The effects of cloudbursts illustrate one way in which the landform and vegetation equilibrium is constantly being upset by change (Hack and Goodlett, 1960). Intense precipitation of short duration is apparently the result of convectional activity during the warmer months, with the higher elevation rain gauge group experiencing approximately 8 times as many one hour-short duration intense storms as the lower stations (Bogucki, 1972). Kochel and Johnson (1984) and many others have noted the importance of debris flows in the geomorphic evolution of the region and the role of climate as driving force of erosive/sedimentation episodes.

The orographic effect on precipitation is one source of increased climatic energy (Slaymaker, et al., 1972). For example the average wind velocities increase with altitude. In many parts of the world low pressure cells, particularly typical cyclones, are the major source of landslide-triggering rainstorms (Crozier, 1986). Frontal convergence, deep troughs of low pressure and the orographic enhancement of moist, ascending air have also been responsible for triggering rainstorms. The steep slopes exceed the angle of repose of unconsolidated surficial material. Debris avalanches (Scott, 1972) are restricted to humid, mountainous regions. The climate of the Appalachians is conducive to intense, heavy precipitation. Associated with this are high pore pressures and advanced weathering of surficial materials, which are then available for mass transit. Slides (Flaccus, 1959) are associated with all types of rainfall, but are more numerous in the widespread 2-3 day storms such as hurricanes; the requisite rainfall intensity is difficult to determine as this will vary from place to place, depending on slope, mantle, drainage, and soil moisture. When critical weather conditions are attained, landslides commonly occur in regional clusters (Crozier, 1986).

341 intense rainfalls (Koch, 1974) have been recorded in and about the GSMNP from 1937-68. Assuming that 1 inch of rain per day represents a minimum condition, all rainfalls exceeding this magnitude for local precipitation stations for the period from 1953 to 1987 were documented (see Figure 22). In this study it was determined that 1273 storms occurred during the period of 1951-1987 in and about the GSMNP (see Table 7).

TABLE 7. Number of storms in and around the GSMNP from 1951-1987 as determined from precipitation stations listed in Figure 22.

	Amount of Rainfall per 24 hour period (inches)				Total
	1-1.99	2-2.99	3-3.99	4+	
Storms	989	209	59	16	1273
Ave/yr	27	5.6	1.6	0.4	34

Yarnell (1935) found that the number of intense rainstorms are greatest during the months of June, July, August and September. May through November are the months when high intensity, convective and cyclonic storms are most abundant in the east (Eschner and Patric, 1982). In the GSMNP intense rainfall is more prevalent in the months April through September (Bogucki, 1972). It is Koch's (1974) opinion that a slide-producing storm may occur at any time and most probably in the summer months. Clark (1987) suggests that the majority of catastrophic rainfalls occur in June, July, and August because a typical winter air column could not hold requisite moisture volumes. Intense summer rains are divided nearly equally between the one-hour short-duration type and the 24 hour long-duration type (Tennessee Valley Authority, 1937).

In the higher elevations of the Smoky Mountains, annual precipitation may be as high as 128 inches per year such as in Coweeta, N.C. (TVA Precipitation Annual, 1979). Clingmans Dome,

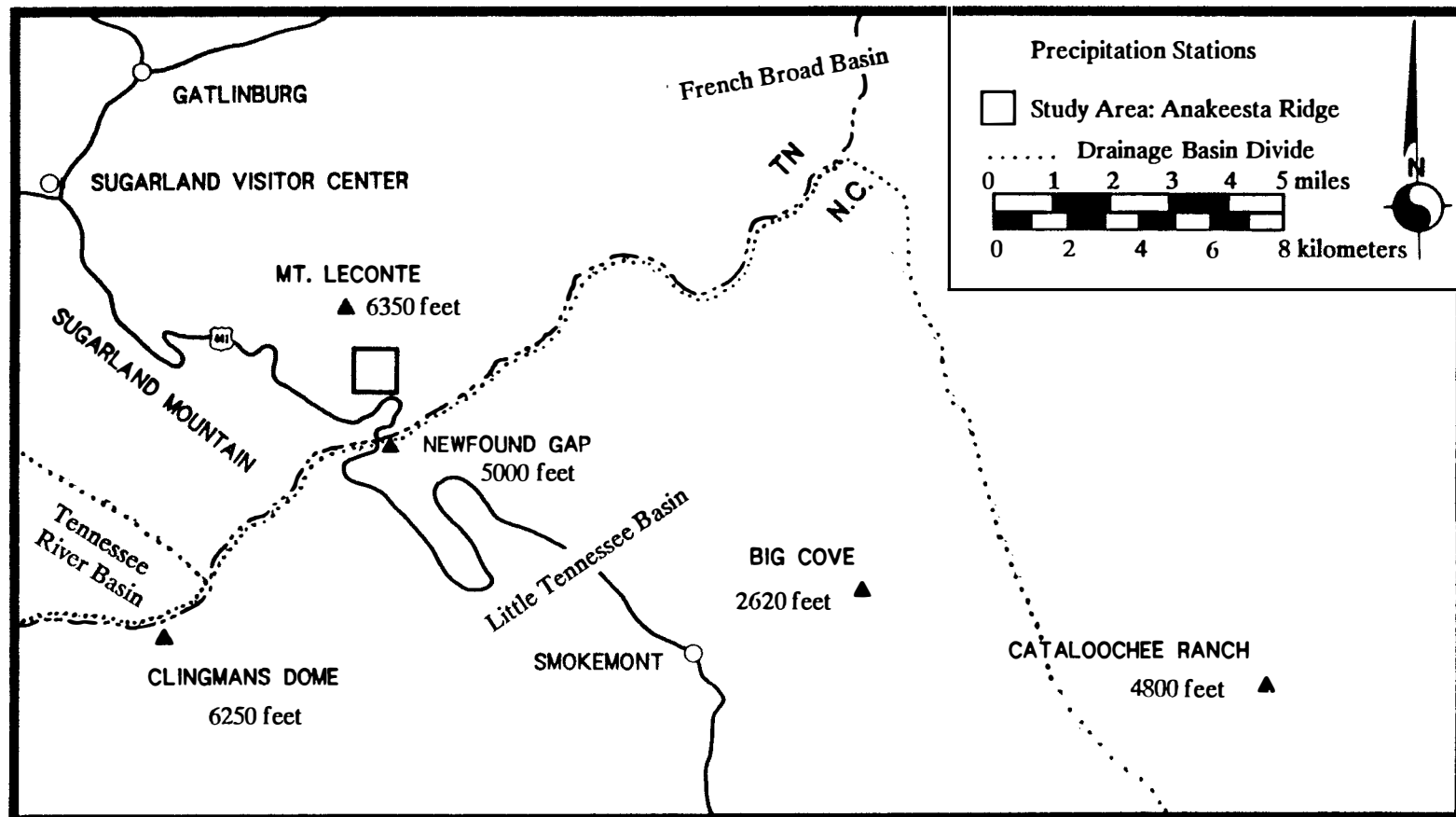


FIGURE 22. Precipitation gauge stations and drainage basins.

NewFound Gap and Mt. LeConte annually top 80 inches per year (TVA Precipitation Annual 1979). Precipitation changes markedly in the 445 to 1524 m (1460 to 5000 ft) of relief on the Great Smoky Mountains, in fact, precipitation increases by 50 percent where the 1372 to 1524 m (4500 to 5000 ft) elevations are reached (Shanks, 1954). The mountain climates above an altitude of 610-762 m (2000-2500 ft) are extremely humid and fall into the rain forest or perhumid class (Shanks, 1954). At the lower elevations such as Big Cove, precipitation averages less than 60 inches of rain per year (see Table 8).

TABLE 8. Average precipitation for selected stations in the GSMNP

Station	Elevation (ft)	Average Precipitation (inches)
Clingmans Dome	6250	82.2
NewFound Gap	5000	90.59
Big Cove	2620	59.99
Mt. LeConte	6350	85.86
Cataloochee Ranch	4800	62.47

Critchfield (1966) documented a rain intensity world record of 30.8 inches of rain in 4.5 hours in north central Pennsylvania in the flood of July 18, 1942. The maximum known intensities for the Great Smoky Mountain National Park are listed in Table 9. The maximum intensities Hurricane Camille of August, 1969, provided an unusual opportunity to study the effects of cata-

TABLE 9. Maximum Intensities (inches) for 5, 15 and 60 minutes for the GSMNP (after Frederick, et al., 1977)

minutes	5	15	60
2 year maximum:	0.44	0.85	1.5
100 year maximum:	0.78	1.67	3.25

strophicerosional forces (Woodruff, 1971). Some areas were more affected by Camille than others although in close proximity-why? Williams and Guy (1973) suggest the reason lies in uneven distribution of rainfall. Huff (1967) indicates that heavy storms with multicellular patterns of precipitation cause variable intensities and quantities near the storm centers (See Table 10). On the night of August 19-20, 1969, the central part of Virginia received 28 inches of rain from the remnants of Hurricane Camille (Williams and Guy, 1973).

TABLE 10. Maximum precipitation intensities in the Tennessee River Basin (1941-1979), TVA Precipitation Annual (1979).

YEAR	RAIN(INCHES)	DURATION	STATION
1968	5.5	1 HOUR	F1
1968	8.8	3 HOURS	F1
1968	11.13	6 HOURS	F1
1968	11.13	12 HOURS	F1
1964	13.10	24 HOURS	286

F1: Kimbrough Cemetery, Elevation 1020 feet, in Tennessee River water-shed, northwest Alabama.

286: Rosman #2, Elevation 2230 feet, in the French Broad water shed, near South/North Carolina, Georgia junction.

194/293 intense rainfalls, were recorded at rain gauges above 4000 feet; Clingmans Dome accounted for 122/194 of the 4000+ elevational rains (Bogucki, 1972). Of the 341 intense rainfalls recorded in and around the GSMNP from 1937 to 1968, 2 debris slide events have documented by study: August 4-5, 1938 Webb Mountain event, and the September 1, 1951 Mount LeConte event. Other slides must certainly have occurred, but have not been acknowledged. The low number of slides with corresponding precipitation data make it difficult to determine a precipitation threshold. The reliability of a threshold value depends on the completeness of the data base (Crozier, 1986). Due to the lack of rain gauge density and instrumentation to quantify debris slide

thresholds, precipitation data for locally intense storms is typically unavailable (Schwartz and Helfert, 1969).

The rainfall data can be approached in two different ways: 1) Although a recording station was not present on Anakeesta Ridge at any time, precipitation can be approximated by using the nearest recording stations, Mount LeConte and Newfound Gap, 2) the rest of the stations can be used to get a handle on duration, intensity and aerial extent of precipitation events in the higher elevations of the GSMNP. Only 4 recording gauges are positioned above 1218 m (4000 ft); high elevation precipitation information therefore is quite limited. Precipitation gauge information may be especially limiting if storms are localized.

More than 50 percent of the total storm rainfall depth occurs in 25 percent of the storm period; usually more than half of the total depth of rainfall is delivered as burst rainfall. Rainfall bursts occur most frequently in the first quarter of the storm period (Farmer and Fletcher, 1972).

Upland forests receive more precipitation, lose less water through evaporation, thus maintain wetter soils than lowland areas (Eschner and Patric, 1982).

The climatic parameter which will provide the most accurate indication of the triggering threshold is the maximum one day rainfall (Crozier, 1986). Slope-ripening will lower the threshold in accordance to the time elapsed since the last event (Crozier, 1986). Slope-ripening processes include freeze-thaw, oxidation, root rot, tree throw and weathering in general.

Bogucki (1970) has extrapolated a mean recurrence interval of 3 years for the unglaciated Appalachian Highlands: 11 slide-generating storms have occurred during 1938-1970. He feels that theoretically any area has a characteristic recurrence interval for storms of debris slide capability. Scott (1972), based on events in this century, determined that the recurrence interval

of relatively localized slide producing storms for the Blue Ridge Province is 3-4 years. Schneider (1973) feels that

it is not possible to arrive at a definite recurrence interval for slide-producing storms: first, there are limited precipitation data for these remote areas and the time span of frequency of occurrence data is too short to be considered accurate. Second, the rate and amount of precipitation necessary to cause debris slides probably varies considerably because of the numerous factors that may be involved and therefore is not known with any degree of certainty. The critical values for precipitation appear to vary from one site to another, depending on other factors.

July 13, 1984 a debris slide temporarily closed U.S. highway 441; 3.8 inches of rain was recorded for a 24 hour period at the Gatlinburg precipitation station. Dilworth (1983) indicates that developmental episodes of the slide tracks were observed in March 1975 and September/October, 1975; on consecutive days, March 13 and 14, rainfall exceeded 1.7 inches for a 24 hour period; on September 23 and 24, the 24 hour rainfall was 2.4 and 1.7 inches in the Big Cove area; on October 17/18 the 24 hour rainfall was 0.9 and 2.1 inches at the Clingmans Dome station. Eschner and Patric (1982) indicated that the debris avalanche storm threshold for most Appalachian soils is 125 mm (4.9 in) in 24 hours. The 4 inch threshold was reached or exceeded 14 times since 1953. Neary, et al. (1986) feel that a heavy rainfall of more than 125mm (4.9 in) in 3 days is required.

In their study of alluvial fans in central Virginia, Kochel and Johnson (1984) have noted recurrence intervals of 3,000 to 6,000 years based upon radiocarbon data. Due to lengthy recurrence intervals and the slow progress of many processes, which exceed the life-span of the researcher (Dietrich, et al., 1982), researchers are forced to develop conceptual models. Computer modeling would be useful if enough pertinent data could be collected and incorporated into the model.

CHAPTER VI

ROCK AND SLOPE ANALYSIS

Direct Shear Testing

Samples of the Anakeesta Formation from the study area were subjected to direct-shear testing in order to determine if values for cohesion (C) and the internal friction angle (ϕ). Factors which affect shear strength include confining pressure, rock type, nature of discontinuity, degree of weathering and pore water pressure. Due to limitations of testing equipment, pore water pressure could not be simulated in the laboratory; this parameter, therefore, was not incorporated into this study.

A discontinuity, as defined by Hoek and Bray (1981) is a distinct, physical separation within the rock mass. The discontinuities of importance in this study include slaty cleavage, bedding, and joints. Slide scar investigation reveals that failure planes occur not only along the soil-rock interface, but also below this interface within the highly oxidized Anakeesta Formation, along bedding/cleavage and joint release surfaces. Direct shear tests were performed along the slaty cleavage/bedding of 22 samples of Anakeesta phyllite.

The specific experimental procedure for determining the cohesion and internal friction angle is detailed within the Civil Engineering, Departmental Report #GT-87-1 as completed by Torbett and Ryan (1986) and is included as Appendix A. The general procedure involved cutting samples of Anakeesta Phyllite so that the samples would fit the direct shear testing device and that the cleavage planes were oriented parallel to the direction of shear (see Figure 23). Preliminary direct shear tests indicated that weathered Anakeesta Phyllite has an internal friction angle of 25 and a cohesion of 75 pounds per square inch. This testing differed from the Torbett and Ryan (1986) study in that the shearing plane was intact and was stressed until failure occurred during the shear run. Several normal stress conditions were tested with a variety of peak shear stresses

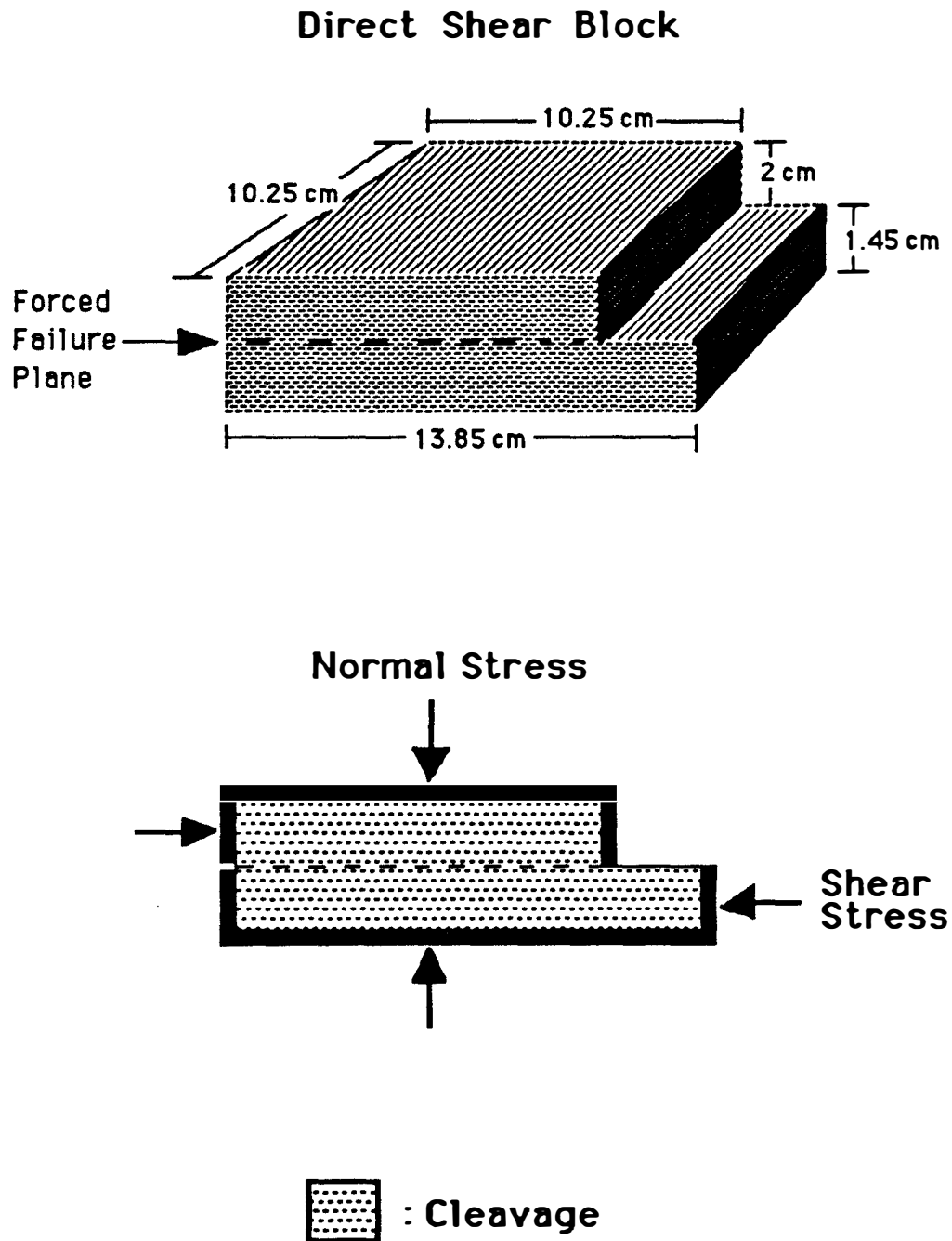


FIGURE 23. Direct shear block dimensions and stress configuration for direct shear testing.

(see Appendix B). Plotting normal stress versus maximum shear stress for each phyllite specimen enabled internal friction angles and cohesion to be determined (see Figure 24a). As interpreted from Figure 24a, the internal friction angle and cohesion for Anakeesta phyllite was determined to be -52.9° and 141 PSI (20,304 PSF), respectively. When the shear data from the less weathered samples 1, 2, 6, 11, 14 and 15 (Appendix B) were not used in the regression (see Figure 24b) then the internal friction angle and cohesion are 58.2° and 42.6 PSI (6134 PSF), respectively. These samples were pulled because they possessed a high shear strength to normal stress ratio; the failure planes were less weathered than the remaining samples. A friction angle of 58.2° falls on the high side of the range of $30-40^\circ$ as listed by Hoek and Bray (1981). It is possible that the sheared specimens were forced to fail along less weathered zones, and this pushed up the internal friction angle.

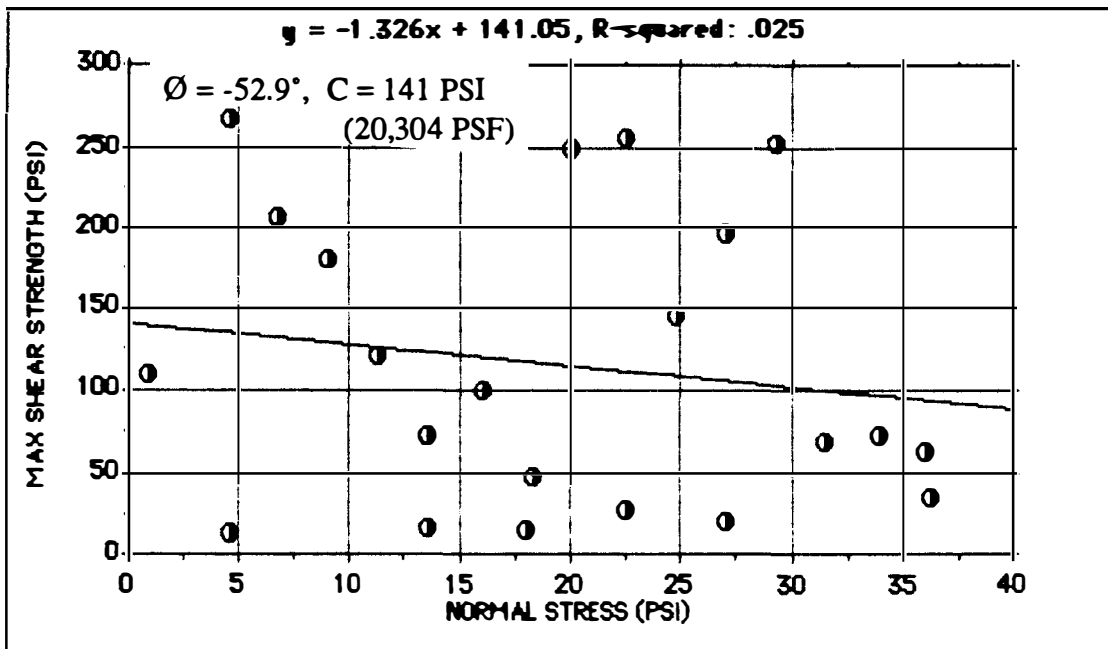
Hawley (1981) feels that it is not practical in natural slopes to assign meaningful values of C and phi because

for most natural slopes even the best sampling and testing operation would lead to a scatter of values of c and phi which would be more than sufficient to span the range between should have failed and stable. The more thorough the investigation, the greater is the range likely to be... .

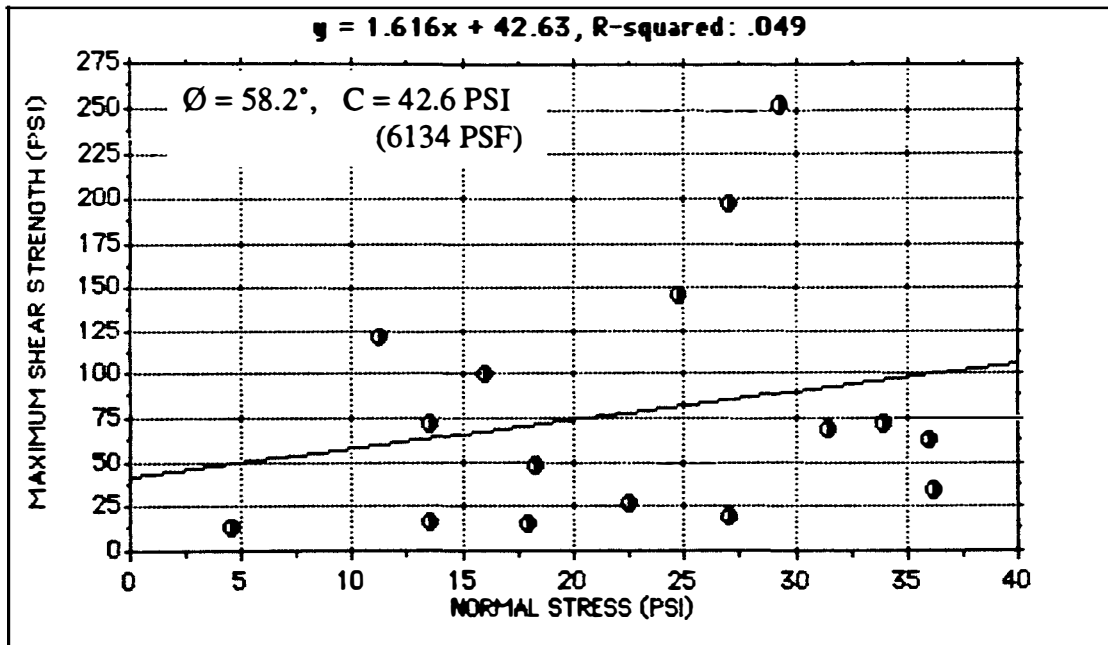
Although the regressions in figures 24a and 24b resulted in low R^2 values (0.025 and 0.049), by experimentally determining the cohesion and internal angle of friction, the FS can be determined with more confidence as compared to assuming a value, or pulling a value from a table (Torbett and Ryan, 1986). When pulling a value (cohesion or internal friction angle) from a table or assuming a value, the specific nature of the slope is not being taken into account.

Wedge Failure Analysis

Debris slides here are actually modified wedge failure sites, a structural coincidence of discontinuities which include joints, cleavage, bedding and faults. Based on the slope morphol-



a. Simple regression.



b. Without the unweathered phyllite sample data.

FIGURE 24. Simple Regression of direct shear data.

ogy and failure plane orientation, the Anakeesta Ridge slope failures appear to have initiated as wedge failures and then progressively developed into true debris slides along the way as evidenced by the debris fan consistency and morphology. The wedge symmetry is depicted by Figure 25; note how the two distinct failure planes interact to form the wedge shape. Moore (1986) has noted in the Blue Ridge that a wedge failure axis will occasionally develop in fault gouge or clay soils washed into fractures. Goethite, hematite and other iron oxides intermixed with randomly oriented phyllite chips were manifested within the northwesterly dipping joints along Anakeesta Ridge.

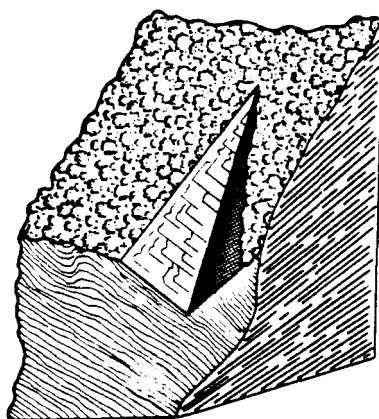


FIGURE 25. Schematic block diagram depicting wedge failure morphology (adapted from Moore, 1986).

The slope factor of safety (FS) can be determined utilizing failure cohesion plane values, the internal angle of friction, slope height, and slope orientation relationships. Due to the nature of the slide, soil-rock interface cohesion is not a consideration in this landslide problem. Intercleavage rock cohesion, however, is critical. In this procedure, discontinuities were plotted on a Schmidt stereonet as outlined by Hoek and Bray (1981), so that a range of kinematically possible failure modes would be indicated. Juxtaposed on these data are slope orientation and dip (See

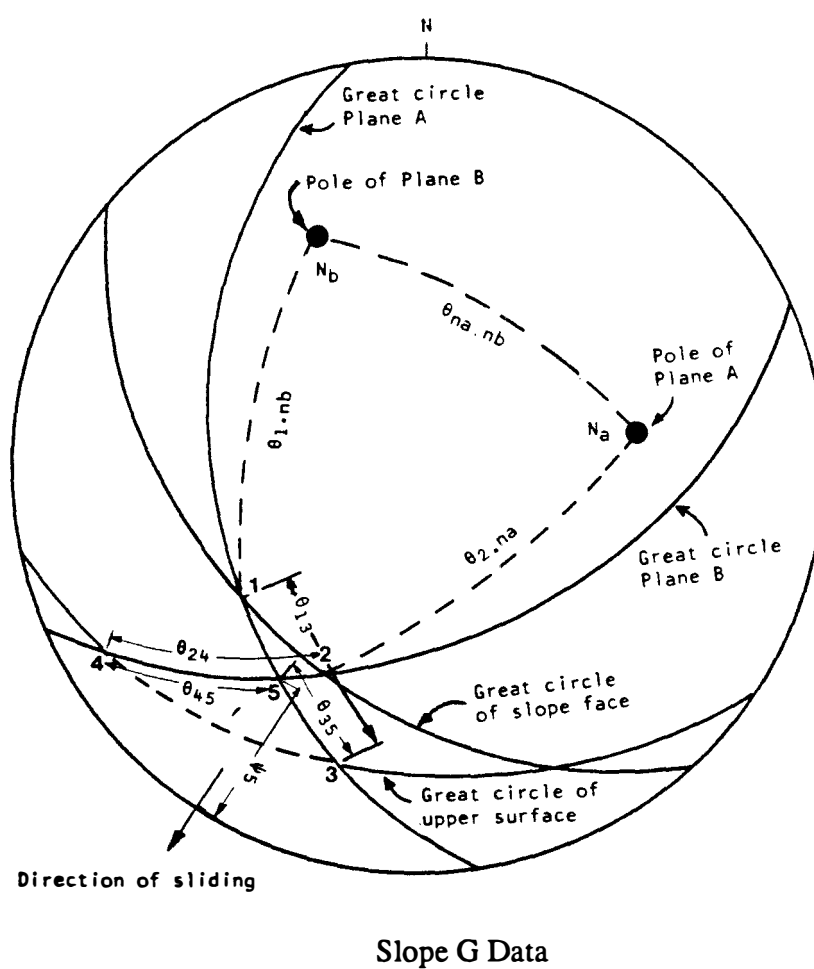


FIGURE 26. Graphic solution for a wedge failure
(adapted from Hoek and Bray, 1981)

Figure 26). Geometric relationships were utilized from stereonet data, with the experimentally determined parameters of cohesion and internal friction angle and the slope height, to obtain the slope factor of safety. The most complete structural data are available for slopes A and G (see Figure 5, page 10). These data have been evaluated according to the wedge failure analysis as detailed by Hoek and Bray (1981). A sample calculation is outlined in Appendix C.

“Back analysis,” the evaluation of a slope after the mass wasting event has taken place is a useful exercise in that the approximate rock cohesion can be determined. This procedure is based upon the limiting equilibrium evaluation of wedge failures as treated by Hoek and Bray (1981). A BASIC computer program that solves for cohesion is listed in Appendix D.

TABLE 11. Wedge failure analysis data

Wedge Analysis Data for Slide G					
Phi A	Phi B	Coh A	Coh B	H	FS
58.2	58.2	6134	6134	375	2.53
Back-calculation:					
58.2	58.2	4217	3225	375	1.00*
35	35	2839	2171	375	1.00*
Wedge Analysis Data for Slide A					
58.2	58.2	6134	6134	531	1.19
Back-calculation:					
58.2	58.2	6490	4992	531	1.00*
35	35	4118	3168	531	1.00*

A/B: Planes A and B

Phi: Internal friction angle, in degrees

Coh: Cohesion, in pounds/ft² = pounds/in² × 144

H: Slope height, feet

FS: Factor of safety

* Back-calculation of cohesion, assuming FS = 1 and adjusting the cohesion for each plane assuming cohesion approaches 0 as the failure plane approaches 90.

CHAPTER VII

GEOMORPHIC SIGNIFICANCE OF DEBRIS SLIDES

"Most of a landscape consists of curved, sloping surfaces, largely shaped by mass-wasting (Bloom, 1978)." Landscapes change in response to energy supplied from a variety of sources. The Appalachian mountains are undergoing active erosion at the present time and avalanching and sliding of debris during torrential rains is an important erosion mechanism in the mountains (Hack and Goodlett, 1960). Debris slides are dynamic areas that are subject to ongoing natural disturbance (Feldkamp, 1984). Regional rates of net erosion are crude measures of the energy conditions attending geomorphic processes. The mountain environment is a high energy environment (Slaymaker, et al., 1972). The vigor of erosion, and frequency of larger erosional events, is rarely so extensive and uniformly productive of high erosion rates as in the mountains. The rate of slope weakening as a result of erosion, overburden release and weathering may mean that in some rock types, slope adjustment by landsliding is an ongoing, and perhaps cyclic process, separated by long periods of apparent stability (Crozier, 1986). Crozier (1986) looks toward the history of the slope and its environment: tectonic uplift, fluvial downcutting, weathering, climatic change and vegetational change. Tectonic uplifting (Crozier 1986) increases potential gravitational energy available to drive processes and subjects material to severe high altitude atmospheric conditions. The latter creates local climatic patterns which propagate orographic precipitation and low evapotranspiration, thus ensuring high runoff and high fluvial activity. Periodic landsliding may maintain non-equilibrium conditions on steep slopes (Hupp, 1983b). Hack and Goodlett (1960) feel that debris avalanching plays an important role in the overall adjustment between slope processes, mountain form and vegetation. Landslides, then, are a product of their environment.

Erosion of the Appalachians has been historically attributed to running water (Scott, 1972). The bulk of the sediment leaves the mountains by alluvial processes. How do the majority of slope materials reach the nearest stream of sufficient load bearing capabilities? The steep forested slopes of the Blue Ridge (Scott, 1972) are generally lacking perennial stream channels and rarely experience overland flow, however they have many hollows. As these areas supply a large amount of the sediment to the eventual depositional area, the oceans, and in the absence of overland flow, with the negligible effect of ice, and wind, mass movement is the only realistic transportation process. Although Scott (1972) is correct in that there is little or no overland flow, piping acts as a soil and fine particle removing agent. In a transect of Anakeesta Ridge, several sampling sites revealed voids that ranged from 4 to 15 cm (1.6 to 5.9 inches) between the vegetational mat and the bedrock. Annual erosion (Slaymaker, et al., 1972) is affected by fewer and more intense events, so the role of rare events becomes greater.

The importance of landsliding as a geomorphic agent is gaining more credibility as ancient events are being recognized (Clark, 1987). In recent literature (Hart and Scaramella, 1987; Schultz, 1986) large-scale, massive, wedge and sliding slope failures have been revealed. The use of aerial photography has been useful in identifying these mass movements. Approximate volumes of prehistoric slides in the eastern United States range in volume of material from approximately 3 million to 1 billion cubic meters (Southworth, 1987). Schultz's (1986) recognition of giant ancient mass movement sites in the southern Appalachians makes one wonder how many, smaller scale mass movements are evading detection? Slide scars are difficult to identify due to vegetational camouflage. Many mass movement features are escaping detection due to this cover. After surveying several debris slide locals in the southeast, Scott (1972) concluded that slide scars in general heal relatively quickly, so they are difficult to find after a number of years has passed. An exception to this are the slide scars of Mount LeConte, which are still prominent from the 1951 event. Due to inaccessibility and the healing of slide scars, there exists an incomplete record of slides (Scott, 1972).

Bell (1983) considers landsliding as one of the most effective and widespread mechanisms by which landscape is developed. Debris slide/flow events are a major erosional process in the forested, eastern mountains of the United States (Eschner and Patric, 1982). "Although debris avalanching in the Appalachian Mountains is a rare phenomenon in human history (100-1,000+ year return period), it is a major and frequent geomorphic process influencing soil formation" in addition to an important landscape forming process (Neary et al., 1986). Field study of selected areas should convince even the most skeptical observer that mass-movement must be recognized as one of the major geomorphic processes (Sharpe, 1968).

In a transect of Anakeesta Ridge, all 7 major drainage ways, F2-F8 exhibit evidence of mass movement activity (See Figure 18, page 37). Scott (1972), and Schneider (1973) find that debris avalanching is common in the Blue Ridge Mountains. Scott (1972) estimates at least 1700 slides have occurred over the last century alone. In one of central Virginia's worst natural disasters on record, erosion resulted mainly from debris avalanches and channel scour with nearly half of all the storm-related sediment coming from down-slope trending strips on the hillsides (Williams and Guy, 1973). Mills, et al. (1987) agree with Scott (1972) that most hillslope erosion in the Blue Ridge is accomplished by debris slides. Schneider (1973) feels that the storms that produce the slides were probably common events in the last 100 years in the Appalachians south of the glacial border. Hack and Goodlett (1960), Bogucki (1970) and Koch (1974) all have earmarked the debris slide as a major agent in valley formation: slide events, as extraordinary erosional events, initiate first order valleys or modify existing valleys; the importance of the process depends upon the frequency of debris slide producing storms.

The Appalachian Highlands may be rising relative to the Atlantic Coast at rates up to 6 mm/year, and because these rates exceed geologic time averages, the movements are episodic or oscillatory (Brown and Oliver, 1976). The long-term denudation rate in the Appalachians is approximately 40 mm/1000 years (Hack, 1980). Bloom (1978) concludes that "steep valley walls

or cliffs retreat at rates of 0.1-3.0 mm/year in a variety of subarctic, desert, mountainous, rain forest and savanna environments on diverse lithologies." If a debris slide removes 30 cm (11.8 inches), than one debris slide is equal to approximately 100 years of slow process denudation.

It is widely believed (Wolman and Miller, 1960) that "the infrequent events of immense magnitude are most effective in the progressive denudation of the earth's surface . . . analysis of the transport of sediment by various media indicate that a large portion of the work is performed by events of moderate magnitude which recur relatively frequently rather than by rare events of unusual magnitude (Wolman and Miller, 1960)." "The relative importance in geomorphic processes of extreme or catastrophic events and more frequent events of smaller magnitude can be measured in terms of the relative amounts of work done on the landscape...in many basins 90 percent of the sediment is removed by storm discharges which recur at least once every 5 years...where stresses generated by frequent events are incompetent to transport available materials, less frequent ones of greater magnitude are obviously required (Wolman and Miller, 1960)." Debris avalanching is a major contributor to long-term erosion rates (Neary et al, 1986).

In Strahler's (1952) concept of steady state (concept of graded slope) the slope forms and the mantling debris are in a state of continuous adjustment as dictated by open system components such as relief, climate, bedrock type and weathering index, slope azimuth, vegetation to name a few. Sliding phenomena (Zaruba and Mencl, 1982) is among the more significant exogenic denudation processes. In dynamic equilibrium, landforms are adjusted to the bedrock underlying them and the processes acting upon them (Hack, 1980). To geomorphologists (Crozier, 1986) instability represents a

geologically short-lived condition in which slopes tend to be reduced in mass, height or angle as a result of some perturbation of geological or environmental conditions...large-scale slope movements...rapidly destroy the conditions necessary for its operation...landslide activity is a self-annihilating process of landform adjustment which tends to give way in time to conditions where form and process take on a more stable and long-lasting relationship.

As debris slide magnitude storms are more likely to occur in the southeastern United States, their importance as a process must be appreciated. Ryan and Clark (1988) have found debris sliding to be a significant transportational agent along Anakeesta Ridge.

The compound slide scars of Anakeesta Ridge are at various stages according to the "Pacific Northwest model" for the origin and evolution of bedrock hollows (See figure 27). The older scars such as A and B are at an accumulation stage in the mid and lower track sections and are depicted by stage "B" in the model. In the scar head area, however, all of the slopes are at stage "A." Although scar G has been recently (1984) evacuated, it is beginning to undergo accumulation; the scar is at stage "B." White (1949) proposed a similar sequence of events leading to a debris slide: The cycle begins with an exposure of bedrock, followed by a plant succession, the plants deplete soluble minerals and organic matter, the plants deteriorate, at which time, the cycle begins once again.

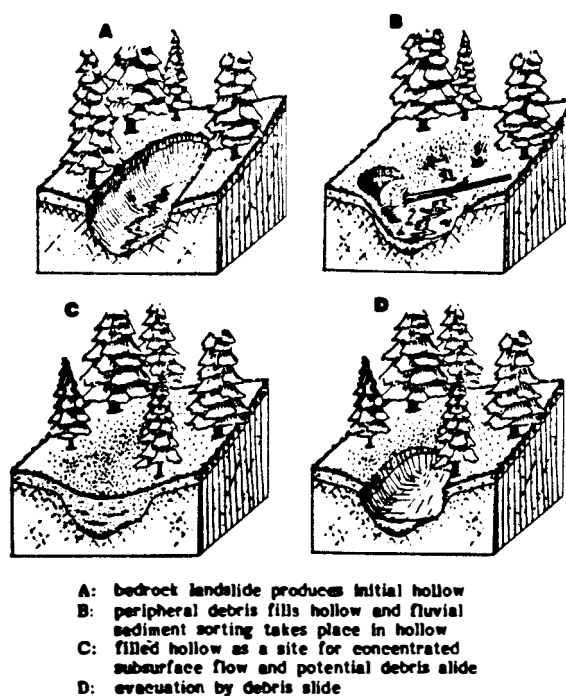


FIGURE 27. The Pacific Northwest model for the origin and evolution of bedrock hollows (after Dietrich et al., 1982).

CHAPTER VIII

SUMMARY AND CONCLUSIONS

The purpose of this work has been to enhance the knowledge of slope processes as they operate on the landscape within the Great Smoky Mountain National Park; to identify, date, and classify debris slides along Anakeesta Ridge for an historical perspective; and to evaluate the engineering properties of Anakeesta phyllite in terms of the shear strength parameters, internal friction angle and cohesion such that a factor of safety may be calculated.

It was determined that 1273 storms (1 or more inches of rain per 24 hour period) occurred during the period of 1951-1987 in the general area of the Great Smoky Mountain National Park. This amount of moisture very nearly guarantees that the highland slopes will be in a moist condition. Moisture information on three debris slides for which the specific slope failure dates are known indicate that an excess of 3 inches of rain within 24 hours triggered one slide; another slide was triggered by two or more consecutive days in which the precipitation averaged 1.5 inches of rain per 24 hour period; the September 1, 1951 debris slides were triggered by a cloudburst of at least 4 inches of rain in less than 6 hours.

By monitoring slope retreat along selected slide scars and comparing slope retreat with precipitation events, it was determined that a series of <2 inch rainfalls had at least as much of an erosional impact as a single 2 or 3 inch rain. It is possible, however, that some of the <2 inch rainfalls were of greater intensity than the 2 or 3 inch rainfalls. Recent slide scar retreat has been nonexistent in some areas, primarily side-slopes, whereas slope retreat has been as high as 37.2 cm over a 7 month period in the scar head areas. Erosion primarily occurs at the scar head because slope angles are steepest at the scar head, toe support is lacking, and tree throw is highly operative due to shallow rooting systems and exposure to wind.

The principal triggering factor for debris slides are high precipitation events, perhaps in the form of a cloudburst, coupled with a ripened (weathered) slope. Tree levering may also play a factor. Local intensity records are all but impossible to obtain unless a storm happens to center over a precipitation gauge.

As to the question of the significance of debris slides as landscape modifiers, this study has revealed that a high percentage of the draws about Anakeesta Ridge are associated with debris fans. That debris slides are major geomorphic agents is not the question, but how often on the average, does the ridge undergo a “flushing” event. Debris fans along the ridge, range in approximate minimum dates of origin from 1749 to 1971 according to dendrochronological data. Because the time between the mass movement event and plant growth is variable, only a minimum age may be estimated from dendrochronological data. Tree coring data yields dates which vary from 6 to 87 years between events and averages 13.8 years. I feel that this frequency represents a combination of slope ripening and the time between precipitation events: material must be available for debris slides/flows to take place and there is more than enough precipitation on a regular basis for debris slides/flows to take place.

Anakeesta Ridge is moving toward a stage of accumulation in the midslope sections-it is getting “ripe” for a flushing event, especially along slide G where several rotting log dams are storing up slaked Anakeesta phyllite. The logs have been in place on slide G since 1984; log dams have been in place at the distal end of slide track A since 1951. Log jams act as an effective debris dam until that time when the logs are structurally weakened by rotting. The logs are then part of the regolith and contribute to vegetal debris in next slide event.

Sequential aerial imagery for Anakeesta Ridge indicates that slope failure is initiated in the mid-slope region, with the upper-slope placed in a position of instability for future failure. 81 percent of the Anakeesta Ridge debris slides/flows originate in drainage depressions. After the initial event, porewater pressure does not initially play a significant role because the slope has

made the undrained to drained transition. The slope will regain its undrained status once the slope becomes revegetated and weathered material accumulates.

The compound slide scars on Anakeesta Ridge have increased in aerial extent and volume (as estimated from scar head area and depth) from 4,300 m² and 1790 m³ in 1953 to 128,000 m² and 84,100 m³ in 1987. The scar head and upper slide track areas are primary debris volume contributors.

Many times it is assumed that the failure plane occurs at the rock-soil interface. In this study, this was not the case: the failure plane occurred below the rock-soil interface within the weathered Anakeesta phyllite along cleavage, joint and bedding planes. Cleavage is the preferred southeast dipping failure plane and joints are the preferred northwest dipping failure planes. The chute morphology is characterized by wedge failure planes as formed by the intersection of cleavage/bedding and joints. Weathering is extended beneath the slope surface due to the pervasiveness of joints, cleavage and bedding.

In this study all of the profiled slopes (A-G) are concave in profile with the upper slope segments averaging 43.4°, the mid-slope segments averaging 33.6° and the lower slope segments averaging 25.6°.

Due to the coincidence of discontinuities and the weak mechanical properties of weathered phyllite, the Anakeesta Formation is a prime candidate for debris slides. Direct shear testing of Anakeesta phyllite yielded an internal friction angle (ϕ) of 58.2° and a cohesion (c) of 6134 pounds/foot². The discontinuity interaction of cleavage/bedding and joints, produce the characteristic wedge failures on the compound slide scars of Anakeesta Ridge. When the experimentally determined internal friction angle and cohesion data are utilized in a FS wedge failure calculation, a FS range of 1.19 to 2.53 was generated. If the FS values had been extremely low, I would have suspected that the test samples were more weathered than insitu slope material. However, the FS values, exceed one; this leads me to believe that the samples were forced to fail

along a less weathered failure plane, thereby yielding high cohesion and internal friction values. Test specimens may have failed along zones *not* of the least resistance, whereas natural events would have failed along the weak zones of a rock.

The back calculation for cohesion when setting the FS at 1, yielded the cohesion range of 2117 to 6490 PSF. This is close to the range of cohesion for weathered rock: 200-5000 PSF (Hoek and Bray, 1981). But in comparison to soft sedimentary rock, it does not make the range: 20,000-400,000 PSF. The lower values are, however, reasonable in magnitude due to the weathered nature of the Anakeesta phyllite.

Several slope-modifying processes are continuing to operate on the slide scars of Anakeesta Ridge. Transportational processes operating on Anakeesta Ridge include creep, overland flow (fluvial transport), debris sliding, freeze-thaw activity, faunalurbation, slaking, and bank slumping. Slope retreat is accomplished by sheet wash, mass movement and tree throw. Appreciable amounts of fine sediment are moved downslope by slopewash. Tree-throw continues to operate proximally to the Anakeesta Ridge slide scars. Tree-throw is present at every breach of a ridge crest in the study area and is common along the unfailed slopes of Anakeesta Ridge. In a transect of Anakeesta Ridge, 73 isolated tree-throw sites were observed. In terms of biogenic transport, tree-throw importance is clearly expressed by the microtopography created by decaying tree-throw mounds; long after the tree has been uprooted and decayed, the mound will remain. These mounds could be mistaken for "blowout" holes, especially if the tree has completely rotted away.

Areas of future slide scar development include the Chimneys; slide scars as observed from U.S. 441, have been developing in the most recent cycle since 1985. The chimneys are underlain by Anakeesta phyllite, are very steep, and have a northern (moisture retaining) exposure.

Anakeesta Ridge slope instability is the result of the coincidence of weathering, an accumulation of regolith, tree levering against shallow root networks, and high precipitation

events. Landslide potential is limited only by the availability of excess water, steep slopes and material.

High precipitation events and thus debris slides/flows cannot be prevented, however, hazardous areas can be delineated, and risks be assessed. This way, landslides will not be studied simply for forensic purposes. Periodic aerial reconnaissance of areas of interest may be sufficient to detect incipient slide development. The Anakeesta Ridge slides have developed through headward erosion that was easily tracked through aerial photographs; incipient slides are followed by additional, ongoing sliding.

A debris slide in a particular area is not a one-time event. It has been documented that portions of Anakeesta Ridge have failed in the past, and under the present climatic regime, will continue to do so in the future. The question is not if, but when, and what will be the associated debris volume.

Future studies of interest concerning debris slides and flows include: 1) Perform direct shear tests under a variety of pore pressure conditions, 2) Determine the range of rainfall intensities that are typically experienced in the Great Smoky Mountains National Park, 3) Perform a more extensive tree coring study to further delineate debris slide/flow frequency, 4) Develop a sediment budget for the slide scars by tracking material movement from the scar head to the debris fan, and determine the volume of material that is being stored by tree dams, 5) Determine the useful life span of a tree dam in the GSMNP, 6) Continue to monitor the expansion of the slide scars on Anakeesta Ridge.

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APPENDICES

APPENDIX A

Equipment and Procedure for Direct Shear Testing

EQUIPMENT:

- (1) Karol-Warner direct shear device with modified (lower) sample box--the lower box has been enlarged from 4" x 4" to 4" x 5.6" to accommodate a larger lower slab. This permits continual sample contact with upper sample (4" x 4"), during 0.9" maximum shear. Model # 595, Serial # 2351406. Unit includes x-gauge #25441, strain gauge # 25611s.
- (2) y-gauge (vertical)--Soil Test (0.0001") LC-3.
- (3) Watanabe WX1 000 X-Y Recorder, Serial #2020036, with LYDT X and Y pistons.
- (4) Boston Gear Radiotrol--D.C. Motor Speed Control.
- (5) 24 inch rock-cutting saw and slab saw (complements of Geology Department).
- (6) Planimeter: Numonics Corporation, Electronic Graphics Calculator, model #1224, computer interfaceable.

PROCEDURE:

- (1) Cut samples to the dimensions provided in figure 23 on page 73.
- (2) Determine the surface area over which normal force will be applied.
- (2) Place sample in direct shear device.
- (3) Position (normal) pressure bar over steel ball and brass cap with nuts so that pressure will be applied to sample and not to the outer shear ring via lower nuts.
- (4) Run direct shear unit in forward mode until x-displacement registers on x-recorder and x-gauge, then stop.
- (5) Position LYDT y-gauge over y-pedestal (band of tape on LYDT piston should be positioned at edge of sensor opening).

(6) Apply a minimum normal stress to sample in order to stabilize pressure bar (0.2 on pressure gauge).

(7) Position X-Y plotting pen to left-central area of graph paper (scale: 500X, 10Y).

Mark start of test.

(8) Zero the strain, x, and y gauges.

(9) Apply normal stress to sample.

(10) Apply shear stress to sample at motor-speed of 20 on radiotrol unit until the sample fails or until the limits of the load ring (2500 pounds) are reached. Monitor the load ring gauge and record the value at which failure occurred.

The equipment list and procedure was adapted from the Direct Shear Project Report, Civil Engineering 5920, Departmental Report #GT-87-1, that was submitted to Dr. Drumm by C. Allen Torbett and Patrick T. Ryan, Jr. on February 5, 1986.

APPENDIX B

Data from direct shear tests

Normal Stress (PSI)

Shear Strength (PSI)

1	▲	4.50	267.00
2	▲	9.00	180.00
3		27.00	197.10
4		31.50	69.00
5		36.00	62.70
6	▲	6.75	206.80
7		11.25	121.50
8		24.75	146.20
9		29.25	252.90
10		16.00	100.10
11	▲	20.00	249.20
12		34.00	72.50
13		18.30	47.43
14	▲	22.50	255.35
15	▲	.90	109.43
16		13.50	72.39
17		36.20	35.23
18		18.00	14.93
19		13.50	17.02
20		22.50	26.51
21		27.00	20.11
22		4.50	13.49

▲ Unusually high shear strength for the corresponding normal stress.

APPENDIX C

Wedge Stability (Factor of Safety) Calculation Sheet

INPUT DATA	
$\psi_a = 47^\circ$ $\psi_b = 53^\circ$ $\psi_5 = 36^\circ$ $\theta_{na.nb} = 74^\circ$	$A = \frac{\cos\psi_a - \cos\psi_b \cdot \cos\theta_{na.nb}}{\sin\psi_5 \cdot \sin^2\theta_{na.nb}} = .9502654$ $B = \frac{\cos\psi_b - \cos\psi_a \cdot \cos\theta_{na.nb}}{\sin\psi_5 \cdot \sin^2\theta_{na.nb}} = .7619401$
$\theta_{24} = 42^\circ$ $\theta_{45} = 34^\circ$ $\theta_{2.na} = 79^\circ$	$X = \frac{\sin\theta_{24}}{\sin\theta_{45} \cdot \cos\theta_{2.na}} = 6.271201$
$\theta_{13} = 38^\circ$ $\theta_{35} = 18^\circ$ $\theta_{1.nb} = 72^\circ$	$Y = \frac{\sin\theta_{13}}{\sin\theta_{35} \cdot \cos\theta_{1.nb}} = 6.447294$
$\phi_A = 17.5^\circ$ $\phi_B = 17.5^\circ$ $\gamma = 160 \text{ lb/ft}^3$ $\gamma_w = 62.5 \text{ lb/ft}^3$ $c_A = 2000 \text{ lb/ft}^2$ $c_B = 2000 \text{ lb/ft}^2$ $H = 375 \text{ ft}$	$F = \frac{3c_A}{\gamma H} \cdot X + \frac{3c_B}{\gamma H} \cdot Y + \left(A - \frac{\gamma_w}{2\gamma} \cdot X\right) \tan\phi_A + \left(B - \frac{\gamma_w}{2\gamma} \cdot Y\right) \tan\phi_B$ F = FACTOR OF SAFETY = 1.004668

Data derived from the graphic solution of the wedge failure from slope G, Figure 26, page 77.

APPENDIX D

Basic Program for Determining Cohesion of Wedge failure planes

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10' PROGRAM : WEDGE 5 V.1.2
20' THIS PROGRAM DETERMINES THE COHESION (IN POUNDS/ SQUARE FOOT) ALONG FAILURE
30' OF THE WEDGE UNDER ANALYSIS. IT IS ASSUMED THAT COHESION
40' APPROACHES ZERO ALONG PLANE B (THE STEEPEST OF THE TWO
50' PLANES) AS PSIB APPROACHES 90 DEGREES AND IS A MAXIMUM ALONG
60' PLANE A AS PSIA APPROACHES 0 DEGREES.
70' CAB=TOTAL COHESION ALONG PLANES A,B.
80  PSIA=35
90  PSIB=64
100 PSI5=28
110 THEAB=79
120 THE24=32
130 THE45=19
140 THE2A=38
150 THE13=26
160 THE35=68
170 THE1B=11
180 PHIA=25
190 PHIB=25
200 GAM=160
210 H=40
220 FS=1
230 PRINT, "COHESION: SLOPE # 5, PLANES A & B"
240 PRINT, "FS = ";FS
250 PRINT, "PHIA= ";PHIA
260 PRINT, "PHIB= ";PHIB
270 PRINT, "H= ";H
280 PRINT, "PSIA= ";PSIA
290 PRINT, "PSIB= ";PSIB
300 PRINT, "PSI5= ";PSI5
310 PRINT, "THEAB= ";THEAB
320 PRINT, "THE24= ";THE24
330 PRINT, "THE45= ";THE45
340 PRINT, "THE2A= ";THE2A
350 PRINT, "THE13= ";THE13
360 PRINT, "THE35= ";THE35
370 PRINT, "THE1B= ";THE1B
380 PI=3.141593
390 '
400 PSIA=PSIA*PI/180
410 PSIB=PSIB*PI/180
420 PSI5=PSI5*PI/180
430 THEAB=THEAB*PI/180
440 THE24=THE24*PI/180
450 THE45=THE45*PI/180

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460 THE2A=THE2A*PI/180
470 THE13=THE13*PI/180
480 THE35=THE35*PI/180
490 THE1B=THE1B*PI/180
500 PHIA=PHIA*PI/180
510 PHIB=PHIB*PI/180
520 '
530 A=(COS(PSIA)-(COS(PSIB)*COS(THEAB)))/(SIN(PSI5)*(SIN(THEAB))^2)
540 B=(COS(PSIB)-(COS(PSIA)*COS(THEAB)))/(SIN(PSI5)*(SIN(THEAB))^2)
550 X=SIN(THE24)/(SIN(THE45)*COS(THE2A))
560 Y=SIN(THE13)/(SIN(THE35)*COS(THE1B))
570 CFACT=3/(GAM*H)
580 CAB=2*((FS-((A-32.2/GAM*X)*TAN(PHIA))-((B-32.2/GAM*Y)*TAN(PHIB)))/((CFACT*Y)))
590 PSIA=PSIA*180/PI
600 PSIB=PSIB*180/PI
610 PRINT,"COHESION OF PLANE A + PLANE B = ";CAB
620 IF PSIA>PSIB GOTO 730 ~
630 BFRAC=(90-PSIB)/90
640 AFRAC=(90-PSIA)/90
650 IF (AFRAC+BFRAC) > .9999 AND (AFRAC+BFRAC) < =1! GOTO 700
660 EXTRA=1-(AFRAC+BFRAC) ~
670 AFRAC=(AFRAC*EXTRA)+AFRAC
680 BFRAC=(BFRAC*EXTRA)+BFRAC
690 GOTO 650
700 COHA=AFRAC*CAB
710 COHB=BFRAC*CAB
720 GOTO 750
730 PRINT, "PSIB MUST EXCEED PSIA, CHECK YOUR DATA"
740 GOTO 770
750 PRINT,"WITH A FS=";FS;"      C ALONG PLANE A= ";COHA;"PSF"
760 PRINT, ,      "      C ALONG PLANE B= ";COHB;"PSF"
770 END

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This computer program was originally developed for the Direct Shear Project Report, Civil Engineering 5920, Departmental Report # GT-87-1 that was submitted to Dr. Drumm by C. Allen Torbett and Patrick T. Ryan, Jr. on February 5, 1986.

VITA

Patrick T. Ryan, Jr. taught High School science for four years in Northern Minnesota prior to receiving his B.S. in Geology at the University of Minnesota-Duluth. He entered the Masters program at the University of Tennessee-Knoxville in 1984. While completing his Masters Degree he worked as a teaching assistant for the Geology Department; as a research assistant for the Department of Civil Engineering; and on a full-time basis as the state-wide wildlife and boating safety education coordinator for the Tennessee Wildlife Resources Agency. He is currently employed by the Osakis School District in Minnesota. While going to school he and his wife Kathy were blessed with Tess, Maeve and Eve.