Soil mineralogy used to distinguish solifluction deposits formed under a periglacial environment on the Boulder Batholith, Jefferson County, Montana
by Janette Louise Young Black

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
Using soil mineralogy and relative mineral stabilities, soils formed on stable sites were compared to soils formed on features believed to have undergone mass movement. The mass movement features studied are termed solifluction terraces. The terraces are gently sloping and extend into an arcuate, convex-downslope, steep, rocky front. The presence of Early to Late Wisconsinan glacial deposits in close proximity and at similar elevations to the study area terraces, coupled with the lack of glacial features within the study area, provides evidence that a periglacial environment existed in the area. This study indicates that the terraces were formed by periglacial processes during the Pleistocene. Specifically, the terraces are thought to be formed by solifluction, used here to indicate the slow movement of water-saturated material from higher to lower ground over a frozen substrate in a periglacial environment.

A study of the mineralogical changes in the soils within the study area was made in order to substantiate the solifluction hypothesis. An analysis of the degree of weathering and the distribution of the minerals within the soils found striking differences between the soils of the solifluction terraces and those found on the stable sites. Clay mineralogy analysis demonstrates sharp and erratic changes in the distribution of minerals within the solifluction terrace profile, which contrast sharply with the gradual changes in clay mineral distribution exhibited by the stable profiles. In addition, coarse-size minerals in the stable sites show a gradual decrease in weathering with depth whereas the solifluction soils contain a mixture of fresh and highly weathered minerals throughout the profile. These differences are likely the result of frost heaving and downslope motion in the solifluction terrace soils and support the concept that the origin of the terraces are attributable to the mass movement process of solifluction.

The study demonstrates the utility of integrating soil mineral analysis with geomorphology and that an analysis of mineral texture and distribution within a soil can provide valuable information to distinguish stable landforms from those formed by mass movement processes.

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APPROVAL

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Janette Louise Young Black

This thesis has been read by each member of the thesis committee and has been found to be satisfactory regarding content, English usage, format, citations, bibliographic style, and consistency, and is ready for submission to the College of Graduate Studies.

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Date May 14, 1984

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ABSTRACT

Using soil mineralogy and relative mineral stabilities, soils formed on stable sites were compared to soils formed on features believed to have undergone mass movement. The mass movement features studied are termed solifluction terraces. The terraces are gently sloping and extend into an arcuate, convex-downslope, steep, rocky front. The presence of Early to Late Wisconsinan glacial deposits in close proximity and at similar elevations to the study area terraces, coupled with the lack of glacial features within the study area, provides evidence that a periglacial environment existed in the area. This study indicates that the terraces were formed by periglacial processes during the Pleistocene. Specifically, the terraces are thought to be formed by solifluction, used here to indicate the slow movement of water-saturated material from higher to lower ground over a frozen substrate in a periglacial environment.

A study of the mineralogical changes in the soils within the study area was made in order to substantiate the solifluction hypothesis. An analysis of the degree of weathering and the distribution of the minerals within the soils found striking differences between the soils of the solifluction terraces and those found on the stable sites. Clay mineralogy analysis demonstrates sharp and erratic changes in the distribution of minerals within the solifluction terrace profile, which contrast sharply with the gradual changes in clay mineral distribution exhibited by the stable profiles. In addition, coarse-size minerals in the stable sites show a gradual decrease in weathering with depth whereas the solifluction soils contain a mixture of fresh and highly weathered minerals throughout the profile. These differences are likely the result of frost heaving and downslope motion in the solifluction terrace soils and support the concept that the origin of the terraces are attributable to the mass movement process of solifluction.

The study demonstrates the utility of integrating soil mineral analysis with geomorphology and that an analysis of mineral texture and distribution within a soil can provide valuable information to distinguish stable landforms from those formed by mass movement processes.

CHAPTER 1

INTRODUCTION

Purpose of Study

The purpose of this study is to substantiate the geomorphological interpretation of certain landforms on the uplands of the west central Boulder batholith in southwestern Montana. Using concepts from soil genesis and relative mineral stabilities, soils formed on stable sites were compared to soils formed on features believed to have undergone mass movement. A second objective is to demonstrate the utility of soil mineral analysis to aid in distinguishing stable landforms from those which have undergone mass movement.

The mass movement features studied are termed solifluction terraces. The terrace surfaces are gently sloping and extend into an arcuate, convex-downslope, steep rocky front. This study suggests that the terraces were formed by periglacial processes during the Pleistocene Epoch. Specifically, the terraces are thought to be formed by solifluction, used here to indicate the slow movement of water-saturated material from higher to lower ground over a frozen substrate in a periglacial environment.

Soil mineralogy provides the ideal bridge between geology and soil science. It is especially important for the understanding of rock weathering and mass wasting processes. Soils formed on stable sites exhibit a different mineral distribution profile than soils which have undergone mass movement. Mineral texture and distribution in soils found on solifluction terraces are compared to soils found on stable areas within the study area in order to test the solifluction hypothesis.

Because a concentrated study of typical examples would yield more information than a cursory examination of solifluction terraces in a more regional study, a detailed investigation of three representative examples was undertaken in order to determine their origin. The investigation was designed to test the periglacial hypothesis for the origin of the terraces. The principal objectives were: 1) to describe the terraces in detail, 2) to determine the processes responsible for terrace formation by using the mineral distribution pattern within the soil profile, and 3) to provide data and interpretations to aid in the identification and study of mass movement landforms in other areas.

Location

The study area is located in the west-central portion of the Boulder batholith in southwestern Montana. The study area is 8 km west of Boulder, Montana in the southern part of the Boulder Mountains. The area, referred to as Galena Park, is between 1950 and 2000 meters in elevation and is within the Deerlodge National Forest. The study is concentrated in the northwest portion of T5N, R5W and the southwest portion of T6N and R5W. It is accessible by logging roads that extend southward from U.S. Highway 91 (Figure 1).

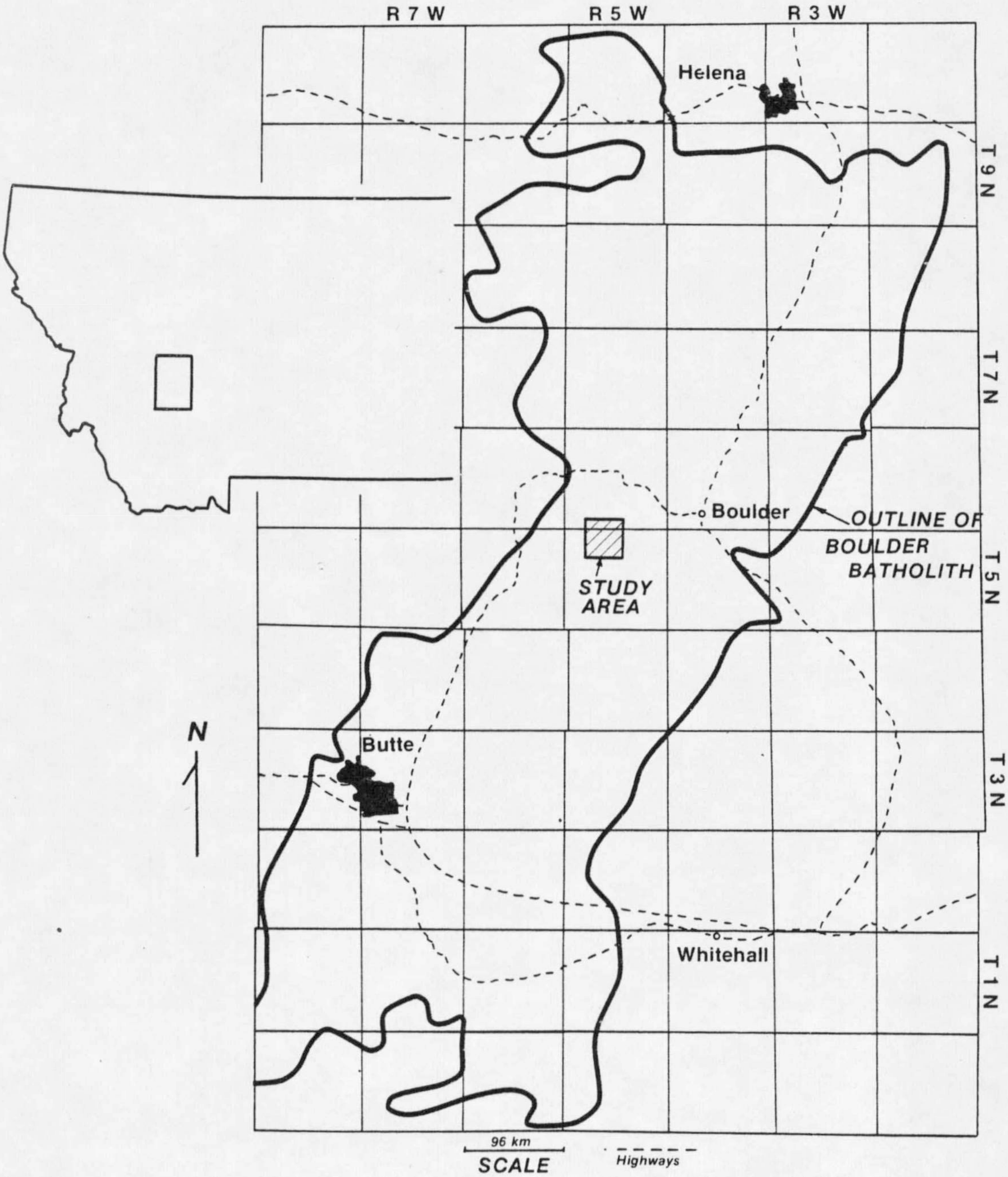


FIGURE 1. Index map showing location of study area and extent of the Boulder batholith (modified from Veseth, 1981)

Topography

The study area lies in the upland region between the Elk Park valley and the Boulder valley. Elk Park is a broad valley located above the Butte Basin on the uplifted block of the Continental Fault. The Boulder valley is a broad intermontane valley trending north-south between the Boulder Mountains and the Elkhorn Mountains. The origin of this valley is not completely understood. Although the Boulder River was larger during the Pleistocene, it is questionable if it could have formed the broad valley. Possibly the valley is formed by a combination of downfaulting and stream erosion (Becraft and others, 1963). Several tributaries to the river show evidence of rejuvenation, with knickpoints, sharp youthful valleys near the river, and sparse terrace deposits (Becraft and others, 1963). The geomorphic expression of the upland region between these two valleys is influenced by the base level established by these drainages. Periglacial processes also appear to have been a significant landshaping force seen in the present landscape.

The Boulder Mountains, on which the study area is located, have no sharply outlined peaks, nor do its hills constitute a well-defined range. The region has moderate relief, with smoothly rounded ridges rising about 300 meters above the major stream valleys. The maximum relief is 1200 meters. There are a few mountains over 2400 meters with most between 2100 and 2450 meters. To the north of the study area, a large ice sheet covered the northern Boulder Mountains during early

Wisconsinan time and subsequently caused nearly all the modern streams in the area to reoccupy glaciated valleys (Ruppel, 1962).

Galena Park is characterized by broad, east-west trending ridges with a series of gulleys and benches extending off both sides to minor tributaries. The benches have tor piles formed by weathering and erosion along joint planes which leave cores of subrounded boulders, usually 6 to 18 meters high, that project above the general land surface (Becraft and others, 1963; Sahinen, 1950). North Boulder Creek, a tributary to the Boulder River, drains the gently rolling uplands of the study area.

Jointing

Physical and chemical weathering is facilitated along joints, and these areas are preferentially weathered. Chemical weathering is enhanced by concentration of water and increased surface area along the joints. Joints also aid the process of frost and ice wedging in periglacial and glacial climates (Thornbury, 1969). In the study area, joint spacing generally falls into two categories, one group with spacing ranging from 0.5 to 2 meters, and another ranging from 4 to 13 meters. As a result, the granite blocks vary in size and shape. Where joints are widely spaced, large, subangular blocks are formed. A finer joint pattern results in increased weathering with the rock often being completely decayed.

In the northern Boulder Mountains, Ruppel (1963) measured nearly 500 joints and mapped two prominant sets, one that trends almost due east and dips steeply north and one that trends north and most commonly dips steeply west. In addition, Ruppel (1963) noted two less prominent, nearly vertical sets that trend N35E and N35W. Smedes (1966) measured 283 joints and found similar trends with the north-south vertical joint set the most pronounced. In the study area, 83 joints were recorded and the dominant joint pattern trends N25W with less prominant sets at N5W and N35W (Figure 2). Smedes (1966) states that the joints are not related to the emplacement or cooling of the batholith, but formed later as a result of regional stresses. Whatever their origin, the joints are the principal controlling feature in the formation of gulleys on the Boulder batholith as evidenced by its strikingly rectilinear drainage patterns (Ruppel, 1963; Klepper and others, 1957).

Climate and Vegetation

The present climate in the vicinity of the southern Boulder Mountains is semiarid with rainfall averaging 38-51 cm per year (Miller and others, 1962). The mean annual temperature is 5.4°C with an average January temperature of -6.9°C and an average July temperature of 16.7°C (Cordell, 1960). The total annual snowfall averages 144 cm. Frost occurs 150 to 180 days of the year. The average depth of penetration of frost is 168 cm (Miller and others, 1962).

Although the annual precipitation is low, the climate is favorable for moderately good soil moisture content due to the occurrence and type of precipitation. Three-quarters of the moisture falls during the months when the ground is not frozen, thus good infiltration is possible. The cool temperatures during the spring and summer also aid

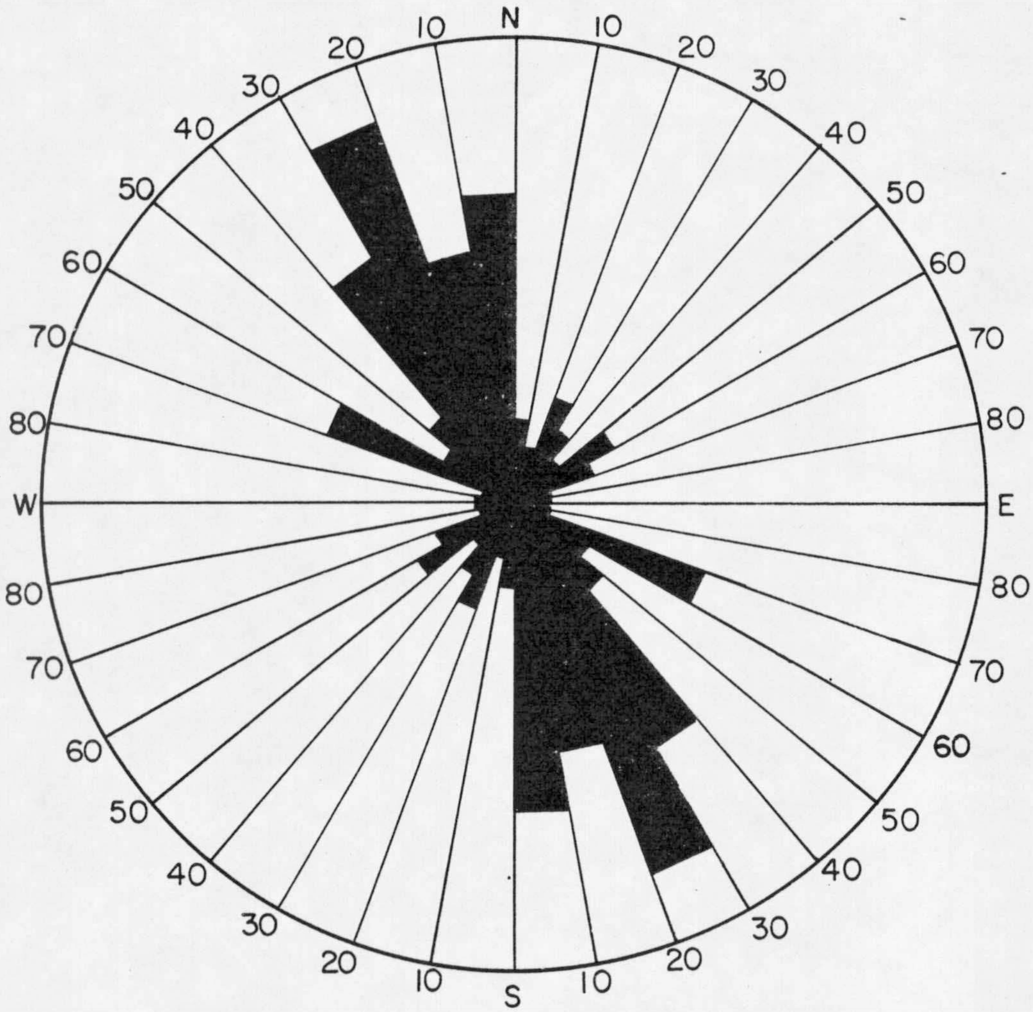


FIGURE 2. Rose diagram of dominant joint patterns in the study area.

in reducing moisture loss. Since evapotranspiration is low during this period, much of the moisture is effective in weathering and in moving the weathering products in the soil.

The study area is largely forested with Douglas fir (Pseudotsuga menziesii) and lodgepole pine (Pinus contorta) with undergrowth characterized by a mat of pine grass (Calamagrostis rubescens) in which twinflower (Linnaea borealis) is common. Kinnikinnick (Arctostaphylos uvi-ursi), bluebunch wheatgrass (Agropyron spicatum) and grouse whortleberry (Vaccinium scoparium) are also common. Meadow species are mainly bluebunch wheatgrass (agropyron spicatum), Idaho fescue (Festuca idahoensis), and big sagebrush (Artemisia tridentata).

Field and Laboratory Methods

Site Selection

A broad survey of the Boulder batholith was conducted to examine geomorphological provinces and to select typical landforms of the Boulder batholith for detailed study. From this preliminary survey four physiographic provinces were recognized: 1) Dissected fan-pediments and low to moderate grassy hills, 2) Steep main canyon walls, 3) Rolling uplands with a wooded and rugged terraine, and 4) Mountain peaks. The detailed study concentrated on landforms in the rolling uplands province. Within this province, solifluction terraces are found.

At Galena Park, the broad east-west trending ridges have a series of alternating gulleys and benches extending off the ridges on both the north and south sides at right angles to the ridges. The gulleys converge away from the ridges to form minor tributaries of the major stream valleys. Within the gulleys, step like features interrupt an otherwise smooth gradient to the local base level. The terraces are arcuate in shape, convex downslope, and extend completely across the gulley. These features appear to be similar to the stone-banked terraces described by Ruppel (1962) to the north of the study area and by Antevs (1932) who first introduced the term stone-banked terrace to avoid a genetic connotation in his study on Mount Washington, New Hampshire. Evidence from this study, however, supports the process of solifluction for the formation of the terraces in the study area. Hence the term solifluction terrace is used in this study. A literature review of periglacial landforms suggests that they are also similar to solifluction and gelifluction terraces described by Washburn (1956), Benedict (1970), and Small (1970).

Field Methods

Topographic maps, aerial photographs and surveys conducted with a Brunton compass and pacing were used to determine the size, shape, slope and elevation of the solifluction terraces. Joints were measured on tor piles located on the benches to determine the regional jointing pattern and its relationship to the orientation of the gulleys. In general, the gulleys trend in the direction of the regional joint pattern.

In order to characterize the soils formed on the solifluction terraces and compare them to soils of stable sites, twenty-one sample sites were examined (Figure 3). The sites were selected using

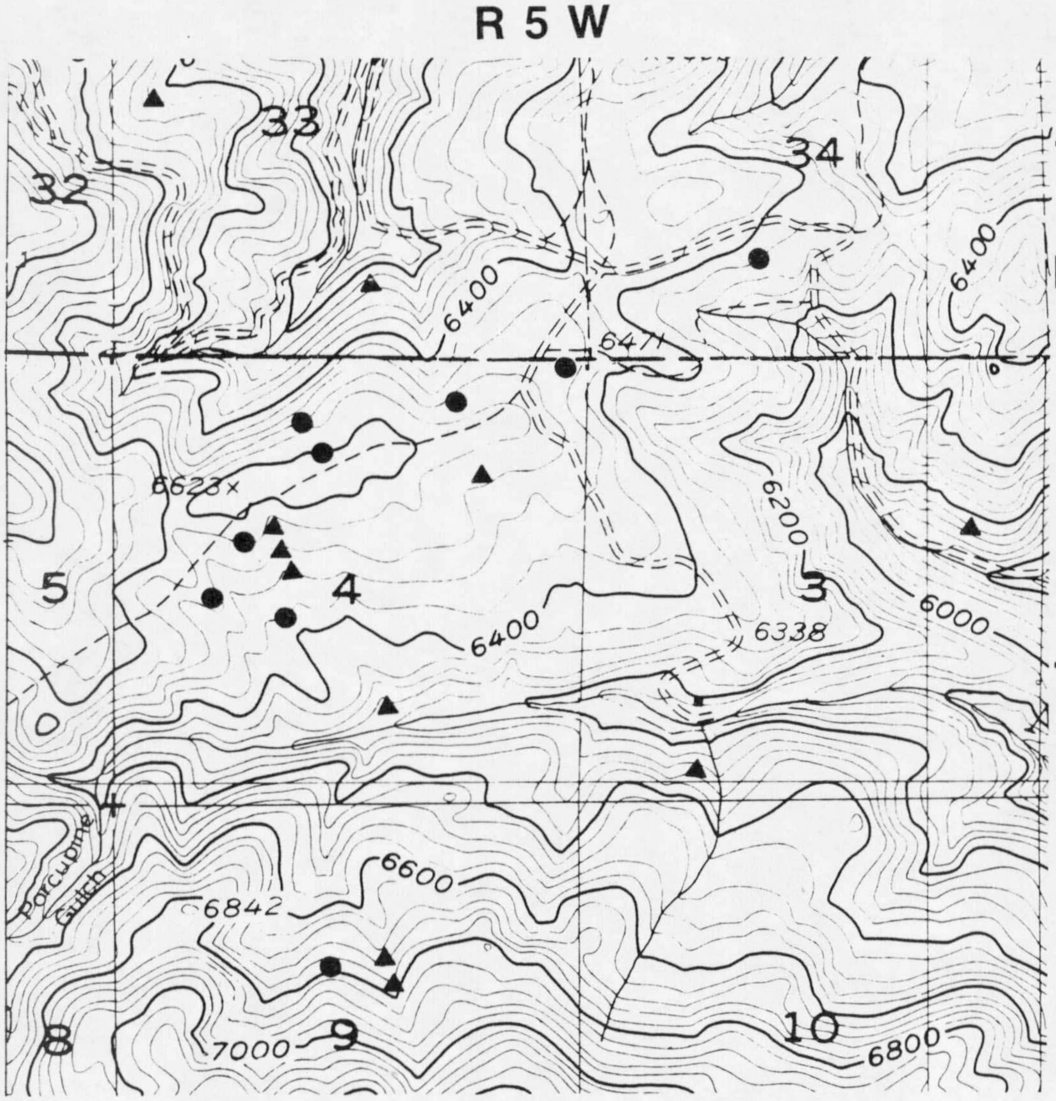


FIGURE 3. Topographic map of stable and solifluction terrace soil site locations.

geomorphology, vegetation, and U.S. Forest Service land-type data in order to locate stable areas as well as areas that have undergone mass movement. At each sample site, soil pits were excavated by hand to a depth of 60 cm or to bedrock. The soil profiles were described by Paul McDaniel, a Montana State University Plant and Soil Science graduate student, and modified based on additional observations made by the author. Samples were collected at 10 cm depth intervals for laboratory analysis. For each soil horizon the following properties were measured by procedures of the Soil Survey Staff (1951): thickness; moist color (Munsell Color Chart); texture; structure type, class, and grade; moist and wet consistence; effervescence (0.1 normal HCL); pH (Hellige-Truog field pH kit); and percent by volume coarse fragments >20 mm. Slope and aspect of the sites were determined with a clinometer. Dominant vegetation was identified and rockiness and stoniness classes (Soil Survey Staff, 1951) were estimated within a radius of 25 meters from the soil profile. Complete profile descriptions of each site are found in Appendix A.

Lab Procedures

Due to the cost and time involved in laboratory analysis, only six soil sites, three located on solifluction terraces, and three located on stable areas, were analyzed in detail. At each of these sites a sample from each soil horizon was submitted to laboratory analysis.

Samples were air dried and hand sieved with a 2 mm sieve for coarse fragment separation. The coarse fragment contents were estimated on a volume basis with measurements of dry total sample and water displacement volumes of the coarse fragment fraction. Thin sections were made of fresh parent rock, and the coarse fractions of the A, B and C horizons. These were examined and described using a petrographic microscope to observe textural and percentage changes with depth in the soil profile.

The fine fractions were then treated to selectively remove calcium carbonate, organic matter, and exchangable cations using techniques discussed by Jackson (1956). The samples were then separated into the size fractions described in Table 1.

Table 1. Particle size classification.

2000–1000 microns	very coarse sand
1000–500	coarse sand
500–250	medium sand
250–100	fine sand
100–50	very fine sand
50–20	coarse silt
20–5	medium silt
5–2	fine silt
2–0.2	coarse clay
<0.2	medium and fine clay

Each horizon of the six profiles was analyzed for clay mineralogy. Samples of the coarse clay (2.0–0.2 microns) and medium to fine clays (<0.2 microns) from the A, B, and C horizons were prepared for X-ray analysis using methods described by Jackson (1956). X-ray analysis methods, using a Cu target and Ni filter, and interpretation procedures were from Whittig (1965). The relative abundance of alumino-silicate minerals in the clay fraction was determined by a semiquantitative method comparing the areas under the diffraction peaks (Klages and

Hopper, 1982). In this method, area weighting factors are established for various clay minerals. The peak areas are multiplied by the weighting factor, and the respective percentages of each clay mineral calculated.

Previous Investigations

Studies of Boulder Batholith

Several classic studies on the geology of the Boulder batholith and vicinity have been performed. Notable among these are the work of Barrell (1907), Weed (1912), Knopf (1913, 1957, 1963) and Billingsley (1915). Atwood (1916) was the first to describe the physiographic history of the Boulder batholith region. Corry (1931) studied the effect of the Continental fault on the landforms near Butte, Montana. Alden (1953) described the relationship between the volcanic and plutonic rocks as well as the Tertiary and glacial history as part of a regional study.

The mineralogy of the Butte quartz monzonite has been the subject of many studies as it makes up over 70% of the exposed batholith. Becraft (1955) set up a field classification for the Butte quartz monzonite based on slight differences in grain size, color, texture, and composition. This classification scheme is followed in most of the work on the Boulder batholith. The igneous petrology of the Boulder batholith has been described in the northern Elkhorn Mountains by Smedes (1966) and in the southern Elkhorn Mountains by Klepper and others (1957). To the north, Ruppel (1963) mapped and described the petrology of the batholithic and volcanic rocks found in the northern Boulder Mountains.

Glaciation on the Boulder batholith has been described and mapped by several workers. Ruppel (1962) mapped an early Wisconsinan ice sheet in the northern Boulder Mountains. Glacial features in the Elkhorn Mountains, northeast of the study area, have been mapped and described by Klepper and others (1957) and Smedes (1966). Preliminary mapping of the Elk Park valley, south-west of the study area, found evidence of Late Wisconsinan glaciation on the east side of the Elk Park valley (Smedes and others, 1962).

Studies of Periglacial Environments

The periglacial environment was first defined by Lozinski (1909) to designate the climate and climatically controlled features adjacent to Pleistocene ice sheets. Later this definition was extended to include any zone adjacent to glacial ice today or during any phase of the Pleistocene. The diagnostic and necessary criterion is a climate characterized by intense frost action and snow free ground for part of the year (Washburn, 1980).

Embelton and King (1968) studied frozen ground phenomena, periglacial mass wasting, and weathering processes. They reported that freeze-thaw action is the most important process of rock weathering in the periglacial environment. Washburn (1956) provided important information on the mechanism of frost creep as the result of detailed measurements made in northeastern Greenland. Högbom (1914) discussed the occurrence and significance of perenially frozen ground and a wide range of frost-induced phenomena was described and analyzed in detail. A regional summary of occurrences of both modern and "fossil" earth-flow phenomena was included in the study. In 1925, Kessler presented a comprehensive summary of periglacial phenomena. Studies on freeze-thaw action were undertaken by Lewis (1939) and McCabe (1939). Smith (1949) studied the distinctive landforms produced by intense frost action.

Solifluction action is an important mass wasting process in periglacial and permafrost environments. Early important contributions on solifluction and related phenomena were made by Eakin (1916), Ekblaw (1918) and Capps (1910). Solifluction associated with frozen substrate, including seasonally frozen ground as well as permafrost, was studied and defined as gelifluction by Baulig (1956). Solifluction lobes are widespread and the processes of solifluction have been observed for more than 80 years, however, it is only in the last 25 years that detailed quantitative studies on solifluction have appeared. Williams (1966) investigated soil movements resulting from solifluction in the discontinuous permafrost zone. Price (1973) observed solifluction lobes on mountain slopes in the Yukon territory. Kerfoot and Mackay (1972) recorded downslope movement of 0.35-1.4 cm per year in a solifluction lobe in the southern part of the continuous permafrost zone.

Stabilized solifluction terraces of Illinoian and Wisconsinan age are widespread in unglaciated areas. Studies by Péwé (1965, 1974) in Alaska and the Yukon territory found solifluction deposits underlying extensive loess deposits. Benedict (1970) studied rates and processes of downslope mass movement on turf-banked terraces formed during Late Pleistocene in the Colorado Rockies. Antevs (1932) studied stone-banked terraces on Mt. Washington, New Hampshire. In other studies, these features are termed antiplanation terraces (Eakin, 1916), solifluction benches (Russel, 1933), and sorted steps (Washburn, 1956).

Periglacial features in the Boulder batholith region have been described in the Elkhorn Mountains east of the study area (Klepper and others, 1957). Stone nets, stone stripes, and other features formed by frost action in a periglacial environment have been observed in the region, especially in the vicinity of Elk Peak. Ruppel (1962) described stone-banked terraces to the north of the study area.

Studies of Rock Weathering

The periglacial features studied in this report are composed primarily of weathering products of granite and closely related igneous rocks. Important studies on granite weathering have been conducted by Merrill, 1913; Reiche, 1950; Ruxton and Berry, 1957; Wahrhaftig, 1965; Ollier, 1969; and Carroll, 1970. Weathering of the Boulder batholith has been studied by Hood, 1963; Huang, 1973; and Darnell, 1974. Important to the process of granite weathering and grussification is the decomposition of biotite and feldspar. Grussification is used here to describe the weathering of granitic rock in place to form loosely consolidated material similar in composition to the parent rock. In studies conducted by Walker, 1949; Jackson and Sherman, 1953; Wahrhaftig, 1965; Nettleton and others, 1968; and Bustin and Mathews, 1979, the alteration and expansion of biotite was considered the primary agent in the grussification process. The weathering of feldspars has been studied by Graham, 1941; Fredrickson, 1951; Todd, 1968; and Wilson, 1975.

Goldich (1938) investigated chemical changes occurring in coarse-grained minerals during weathering. He proposed a mineral stability sequence based on relative resistance of minerals to weathering. Jackson and Sherman (1953) found differences in the stability of minerals with decreasing particle size and established a mineral stability sequence for clay-sized minerals. Other studies have been conducted to gain an understanding of the relationship between weathering processes and rock minerals. For example, Reiche (1950) developed a weathering potential index based on chemical analysis of minerals. Jackson and others (1948) studied the depths of mineral occurrence in the soil profile. Humbert and Marshall (1943) studied the depth function of quartz and feldspars in soils formed on diabase and granite.

CHAPTER 2

REGIONAL GEOLOGIC HISTORY

Tectonic and Regional Setting

The geologic history of the Boulder batholith region can be divided into three stages (Klepper and others, 1957). During the first stage, from Late Precambrian to Middle Mesozoic, the region lay near the edge of a passive continental margin. Early predominantly clastic sedimentation gave way to increasing carbonate deposition through time. During Late Mesozoic to Early Cenozoic, the region experienced a period of deformation and igneous activity. From Middle Cenozoic to the present, the area has undergone regional uplift and erosion.

Late Mesozoic-Early Cenozoic

Of importance to this study is the period of deformation and igneous activity. It is important to realize that tectonic, volcanic, and plutonic events are interrelated temporally and spacially with the emplacement of the Boulder batholith. Major folding, thrusting, and volcanism started about the same time, though not always at the same places (Robinson and others, 1968). Field evidence indicates that major thrusting began approximately 80 m.y.b.p. and recurred intermittently well into Maestrichtian time (Smedes and others, 1973). The earliest folding or tilting occurred before the Elkhorn Mountain Volcanics were extruded, as the volcanics rest unconformably on the

Niobrara Formation (Late Cretaceous) in at least two places (Klepper and others, 1957). Additional evidence that folding or tilting occurred prior to 79 m.y.b.p. is found in the Whitehall area where Elkhorn Mountain Volcanics lie on Madison Group (Mississippian) with an angular discordance of 15° to 20° (Alexander, 1955).

Volcanism began with the deposition of the locally tuffaceous Slim Slam Formation approximately 85 m.y. ago (Robinson and others, 1968) and climaxed during Early Campanian time (77-79 m.y. ago) when the region was buried under the calc-alkalic rocks of the Elkhorn Mountain Volcanics. Major volcanism ceased approximately 73 m.y. ago in Late Campanian time (Robinson and others, 1968) and was not renewed until the Lowland Creek Volcanics 50 m.y. ago. Severe folding and faulting took place after volcanic activity ceased, as volcanic rocks are as strongly deformed as the underlying sedimentary rocks (Klepper and others, 1957).

Emplacement of the Boulder batholith. The Boulder batholith was emplaced during the Late Cretaceous and Early Tertiary, approximately 68 to 78 million years ago. The batholithic rocks intrude rocks ranging in age from Precambrian to Late Cretaceous. Along most of its northern edge, the batholith is in contact with Paleozoic and Mesozoic sedimentary strata that generally dips southward (Knopf, 1963; Smedes, 1966). In most places, the eastern margin is in steep contact with the Upper Cretaceous Elkhorn Mountain Volcanics as a result of intrusion along a north-northeast trending fault (Klepper and others, 1957). To the west and northwest, the batholith passes beneath remnants of a relatively flat-lying roof of Elkhorn Mountain Volcanics. The southern part of the batholith cuts complexly folded and faulted Mesozoic, Paleozoic, and Proterozoic sedimentary rocks and Archean metamorphic rocks. The youngest rocks intruded by the batholith are the Elkhorn Mountain Volcanics, largely of Campanian age (Tilling and others, 1968; Klepper and others, 1957).

The method of emplacement of the Boulder batholith is still a much debated issue. Knopf (1957) suggested that the intruding magma made room for itself by partly shouldering aside the surrounding rocks, resulting in complex folding and imbricate, high-angle thrusting. Based on their analysis of the relations between the batholith and the folds in the intruded rocks, Klepper, Robinson, and Smedes (1971a, 1974) contend that the batholith rose passively into areas of minimum compression. They acknowledge that the border relations indicate forceful injection, but they qualify this statement by adding that the larger structural features suggest that the emplacement of the batholith was more of a result than a cause of regional structure.

Extent and composition of the Boulder batholith. The Boulder batholith lies east of a major salient of the Idaho batholith and west of the Montana Disturbed Belt (Figure 4). The batholith is about 120 km long and 50 km wide and extends from Helena, Montana to 32 km south of Butte, Montana (see Figure 1). The areal extent of the batholith is about 5700 km^2. The batholith is a NNE-trending calc-alkalic composite body ranging in composition from gabbro to alaskite. The plutons that comprise the Boulder batholith were intruded generally in order of increasing silica content. Most of the plutons can be assigned to one of four groups; mafic rocks, granodiorite, Butte quartz monzonite, and

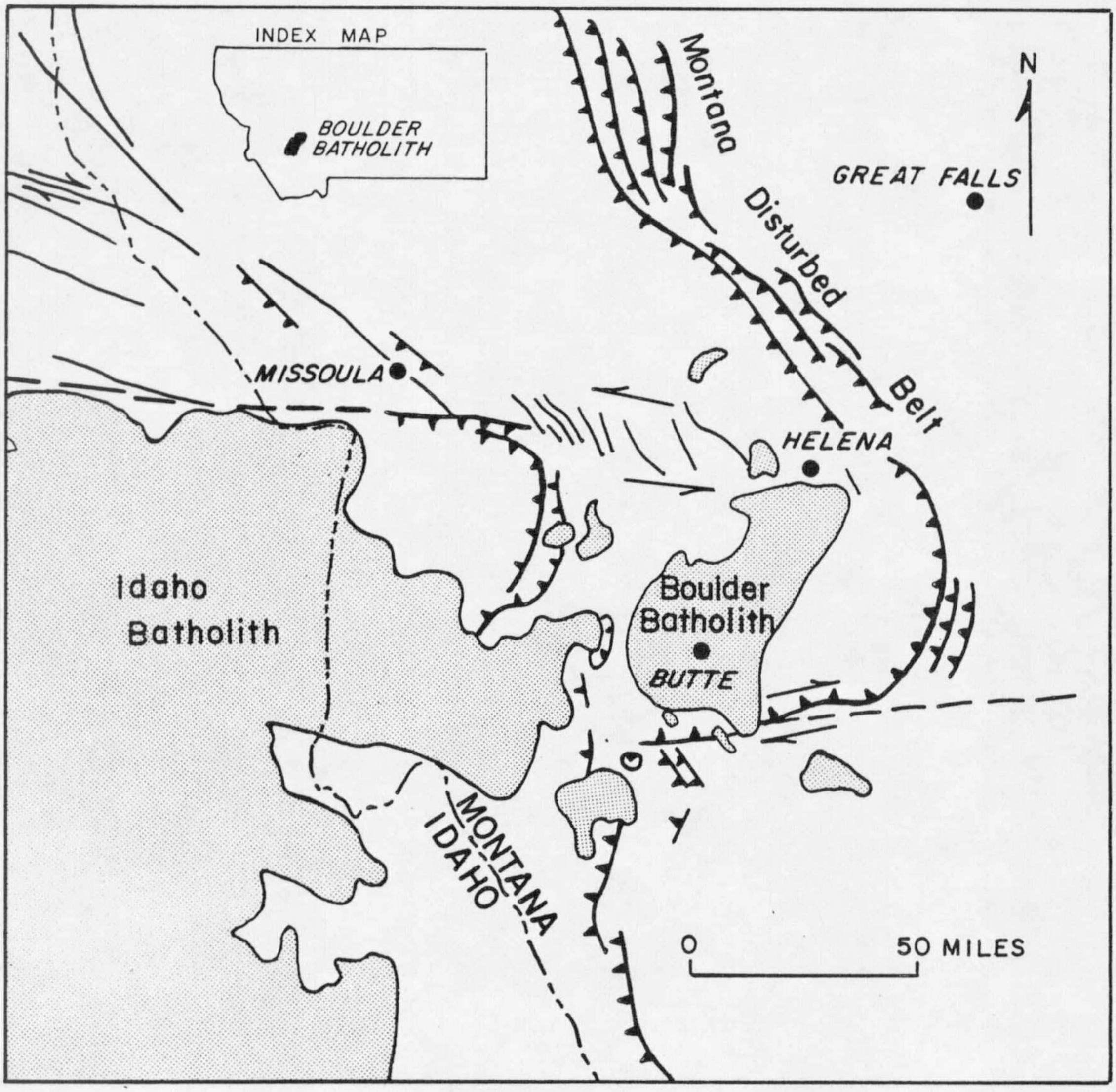


FIGURE 4. Regional and tectonic setting of the Boulder batholith. (modified from Smedes and others, 1973)

leucocratic rocks (Doe and others, 1968). The bulk of the batholith is quartz monzonite and granodiorite with over 70% of the exposed batholith composed of Butte quartz monzonite.

Preliminary mapping of the western part of the Boulder quadrangle during the course of this study and by Becraft and Pinckney (1961) indicates that the study area is underlain principally by coarse-grained Butte quartz monzonite. The rocks comprising the Butte quartz monzonite range in composition from 20-48% plagioclase, 15-45% potassium feldspar, 15-40% quartz, less than 1% to 12% biotite, and 1-3% magnetite, sphene, zircon, apatite, and chlorite (Becraft and others, 1963). The rocks range in texture from equigranular to distinctly porphyritic, and some contain phenocrysts of potassium feldspars as long as 3 cm (Klepper and others, 1957). Textural differences are found in the Butte quartz monzonite near the margins of the batholith, but the rock is relatively uniform in the central portions of the batholith (Becraft and others, 1963; Ruppel, 1963).

Early Cenozoic-Present

A period of crustal uplift and erosion began after the intrusion of the batholith. During the Tertiary, the area underwent two periods of erosion, each followed by a period of volcanism (Ruppel, 1963). The first period of erosion occurred during the Late Paleocene and Eocene time, carving the area into a mountainous terrain with mature relief. By as late as Early Oligocene, erosion had exposed the batholithic rocks, as tuffs containing Lower Oligocene fossils are deposited directly on them (Perry, 1962; Becraft and others, 1963). The quartz latite phase of the Lowland Creek Formation accumulated in the valleys and depressions of a maturely dissected mountainous terrain (Klepper and others, 1957). This was followed by a tectonic phase in which faulting and tilting resulted in erosion of a broad, rhyolite-covered landscape (Ruppel, 1963). During the erosional periods that preceded the Tertiary volcanics, a drainage system ancestral to the present system was established (Klepper and others, 1957; Ruppel, 1963). Since the end of Tertiary time, erosion has stripped away part of the Lowland Creek Volcanics and formed the present topography consisting of low mountains that have been locally modified by glaciation.

Quaternary Glaciation

During Quaternary glaciations, individual valley glaciers were widespread throughout the Rocky Mountain region and local icecaps existed in a number of mountain ranges (Richmond, 1965). In general, five Pleistocene glaciations, separated by interglaciations, are recognized. The last glaciation, termed the Pinedale, is subdivided into three stades, or minor advances, separated by brief interstades. The previous glaciation, known as Bull Lake, is subdivided into two advances, separated by minor interstades (Table 2).

During the Pleistocene, glaciation was important locally in the Boulder batholith region. Extensive glaciation occurred in the northern Boulder Mountains, where an ice sheet covered the area in Early Wisconsinan time (Ruppel, 1962). Glacial deposits and erosion features occur locally to the north, east and south of the study area (Becraft and others, 1963; Smedes and others, 1962; Knopf, 1963;

Table 2. Correlation of the Glaciations of the Rocky Mountains with those of the Midcontinent Region (modified from Richmond, 1965).

Approximate age B.P.	ROCKY MOUNTAINS Richmond, 1965			Aproximate age B.P.	Midcontinent Region After Frye & Willman (1960)		
	Nonglaciation	Gannett Peak Stade			RECENT		
800		Interstade					
900		Temple Lake Stade					
4.000	Altithermal interval			5.000			
6.500	Pinedale	Late stade			WISCONSINAN STAGE	VALDERAN SUBSTAGE	
10.000		Interstade		11.000		TWO CREEKAN SUBSTAGE	
12.000	Glaciation	Middle stade		12.500		WOODFORDIAN SUBSTAGE	several advances
		Interstade					
		Early stade		22.000		FARMDALIAN SUBSTAGE	
25.000	Interglaciation			28.000		ALTONIAN SUBSTAGE	
32.000	Bull Lake	Late stade	2nd episode				
45.000			Nonglacial interval				
			1st episode	50.000 to 70.000 estimated			
	Glaciation	Nonglacial interval			SANGAMONIAN STAGE		
		Early stade					
	Interglaciation						
	Sacagawea Ridge Glaciation				ILLINOIAN STAGE		
	Interglaciation				YARMOUTHIAN STAGE		
	Cedar Ridge Glaciation				KANSAN STAGE		
	Interglaciation				AFTONIAN STAGE		
	Washakie Point Glaciation				NEBRASKAN STAGE		

Pinckney and Becraft, 1961). There is little evidence of glaciation in the southern Boulder Mountains where the study area is located. Glaciation and periglacial processes occurring in the southern Boulder Mountains are discussed in greater detail in Chapter 3.

CHAPTER 3

GLACIATION AND PERIGLACIAL PROCESSES ON THE BOULDER BATHOLITH

This chapter discusses the glacial events and periglacial processes that affected the study area. It includes descriptions of glacial and periglacial features and a detailed description of the solifluction terraces found in the study area.

Glaciation

The earliest glaciation in the region occurred 37 km east of the study area in the Elkhorn Mountains during the Early Pleistocene (Klepper and others, 1957; Ruppel, 1962). Three other glaciations have been recognized in the Elkhorn Mountains (Ruppel, 1962) with the youngest glaciation occurring in Late Wisconsinan (Ruppel, 1962; Klepper and others, 1957) (Table 3).

Features characteristic of glaciation are common in the higher regions of the Elkhorn Mountains. They include headwall cirques, roches moutonnées, rock steps, and glacial grooves and striations on bedrock. Deposits clearly attributable to glaciation are found in the Elkhorn Mountains and include glacial outwash deposits now exposed as terraces along major river valleys. In addition, terminal, lateral and medial moraines, and kame terraces are found.

Ten kilometers southwest of the study area, small glacial deposits have been mapped, but not dated (Smedes and others, 1962).

Table 3. Correlation of local and regional glaciation. (Modified from Richmond, 1965)

WIND RIVER MOUNTAINS Richmond, 1964	WASHAKIE POINT GLACIATION	CEDAR RIDGE GLACIATION	SACAGAWEA RIDGE GLACIATION	BULL LAKE GLACIATION		PINEDALE GLACIATION			NEOGLACIATION	
				EARLY STADE	LATE STADE	EARLY	MIDDLE	LATE	TEMPLE LAKE STADE	GANNETT PEAK STADE
ELKHORN MOUNTAINS Klepper and others, 1957	WEATHERED MORAINES			CIRQUES MORAINES KAME TERRACES				CIRQUES MORAINES OUTWASH TILL		
NORTHERN BOULDER MOUNTAINS Ruppel, 1962				ICE SHEET CIRQUES MORAINES TILL OUTWASH						
ELK PARK Smedes and others, 1962								MORAINES OUTWASH TILL		
BULL MOUNTAINS Ruppel, 1962								COARSE BOULDER DEPOSITS		
GLACIER NATIONAL PARK Horberg, 1954				EARLY WISCONSINAN MOUNTAIN DRIFT		LATE WISCONSINAN MOUNTAIN DRIFT				
Richmond, 1960	EARLY PLEISTOCENE	EARLY OR MIDDLE PLEISTOCENE	MIDDLE PLEISTOCENE	BULL LAKE GLACIATION		PINEDALE GLACIATION			TEMPLE LAKE ADVANCE	HISTORIC ADVANCE
				EARLY	LATE	EARLY	MIDDLE	LATE		

Observations of these deposits made during the course of this study suggest a Late Wisconsinan age as evidenced by the freshness of the morphology and the lack of significant weathering. The deposits in this area consist of well formed to slightly dissected moraines, outwash, and till. Nineteen kilometers south of the study area, glaciation is indicated by coarse boulder deposits on Bull Mountain. These are thought to be Late Wisconsinan in age (Ruppel, 1962).

The most extensive glaciation occurred in the northern Boulder Mountains where an ice sheet over 300 m thick covered approximately 325 km^2 (Ruppel, 1962). The southern limit was 8 km north of the study area, north of the Boulder River. The lower limits of the ice sheet reached elevations of approximately 2,100 m in the south and elevations of 1,800-2,000 m in the east, north and west. The glaciation occurred during Early Wisconsinan and included three phases; an initial valley-glacier phase, an intermediate mountain-ice sheet phase, and a final phase in which the mountain ice sheet rapidly thinned and the glaciers retreated.

Features characteristic of glaciation are common in the northern Boulder Mountains. They include smoothed and subdued cirques, prominent glacial troughs, till, and outwash, as well as lateral, medial, and terminal moraines. The poorly defined cirques and broad glacial theaters at the head of most major stream valleys were probably formed by alpine glaciers early in the glacial stage, and smoothed during the mountain-ice sheet phase. They were apparently not reshaped by alpine glaciers during the rapid recessional phase of glaciation. The cirques contrast strikingly with the well-defined glacial troughs that form the present stream valleys. The present depths and forms of many of the valleys, and the distribution of glacial debris and erosional features indicate that the glacial ice was locally more than 300 m thick during the ice sheet phase of glaciation (Ruppel, 1962).

Thus in the northern part of the Boulder batholith there appears to have been an Early Pleistocene glaciation in the Elkhorn Mountains, an Early Wisconsinan glaciation in the Elkhorn Mountains and the northern Boulder Mountains, and a Late Wisconsinan glaciation in the Elkhorn Mountains, on Bull Mountain and east of the Elk Park valley (Figure 5).

Periglacial Environment in Southern Boulder Mountains

The periglacial environment is defined as an area lying near the margins of glacial ice (Lozinski, 1909). That the environment adjacent to glacial ice has special characteristics has been recognized since systematic study of the Quaternary began. The periglacial environment is characterized by low temperatures, abundant meltwater from snow and ice, seasonally or permanently frozen ground at depth, and lack of vegetative cover. Intense freeze-thaw action is the most important weathering process operating in the periglacial environment (Embelton and King, 1968). It is the chief agent responsible for accelerated mechanical weathering, crude sorting of weathered materials in the soil, and mass movement of detritus on slopes. The alternation of freeze-thaw cycles disturbs the weathered debris overlying frozen ground or solid bedrock and breaks it down into finer sized particles. However, for effective freeze-thaw action moisture must also be available to enter the pore spaces of the rock and regolith in order to

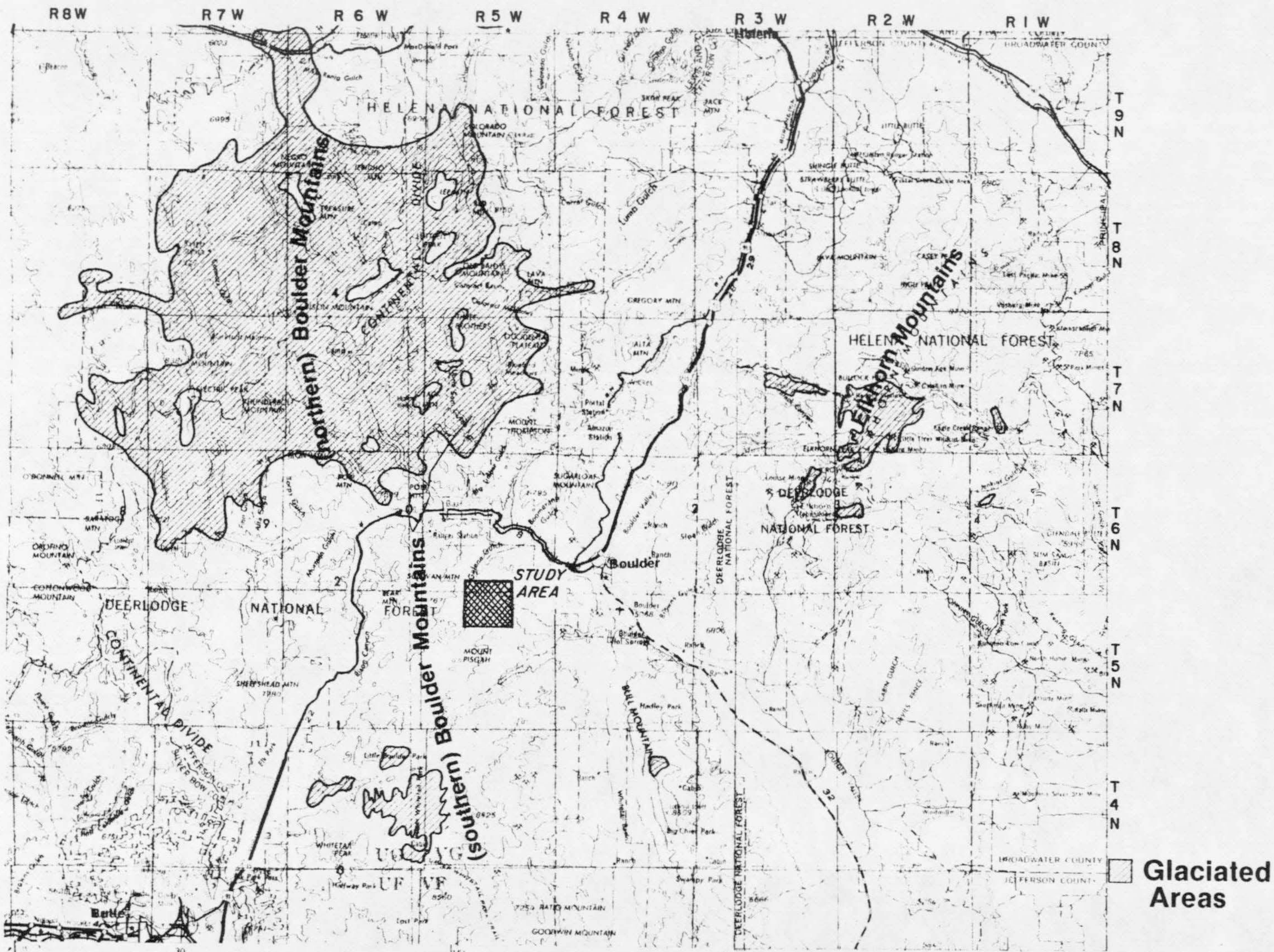


Figure 5. Location and extent of glaciation in the study area.

fracture and lubricate the material. In a periglacial environment, meltwater from snow or ice is available to saturate the unfrozen ground. In addition, frozen ground at depth prevents downward percolation of moisture in the thawed surface layers and facilitates movement downslope.

Studies of ice wedges (Black, 1976) and non-sorted polygons (Mears, 1981) indicated that the periglacial zone extends as far as 650 km south of continental glaciers. Denny (1951) studied relicts of solifluction deposits developed more than 100 km from the outer limits of glaciation. The present study area is 37 km from the site of Early Pleistocene glaciation and 8 km from the Early Wisconsinan ice sheet in the northern Boulder Mountains. Late Wisconsinan glaciation occurred 22 km northeast and 10 km south of the study area.

Although glaciation occurred proximal to the study area, fieldwork conducted during this study found no evidence of glaciation in the southern Boulder Mountains. The rounded cirques, well-defined glacial troughs, and widespread glacial deposits characteristic of the northern Boulder Mountains are lacking farther south. These observations are supported by mapping done by Becraft and Pinckney, (1961) and Ruppel (1962). Although the study area is at approximately the same elevation as nearby glaciated areas, glaciation was absent and periglacial processes were likely a major landshaping force during the Pleistocene.

Solifluction

The periglacial environment is particularly favorable for the phenomena of solifluction. Solifluction was defined by Andersson

(1906) as "slow gravitative transfer of weathered material that is saturated with water, from higher to lower ground." The process of solifluction is known to occur outside the periglacial environment and without a frozen substrate. However, solifluction is enhanced when the upper layer thaws and saturates the surface material and a frozen, impermeable layer occurs at depth. Solifluction includes two main processes of mass movement; 1) the flow of water-soaked debris resulting from seasonal thawing of the active layer, and 2) frost creep caused by alternate freeze and thaw of slope deposits (Embelton and King, 1968). In the first case, an excess of water reduces the shear strength of the material and relatively rapid flowage occurs. In the second case, frost heave produces planes of weakness within the material and reduces cohesion between the particles. During freezing of frost-susceptible material, ice crystals grow normal to the cooling surface and displace particles in this direction. On thawing, the particles resettle in a direction controlled by gravity. Thus if the cooling surface is inclined, the particles will always resettle slightly downslope from their original position (Embelton and King, 1968).

Solifluction deposits have recognizable geomorphological features. The deposits tend to be closely related to local rock types, for it is rare for material to travel more than three kilometers by solifluction (Embelton and King, 1968). Angular fragments are characteristic, and the thickness of deposits are variable. Solifluction deposits are usually better sorted than mudflows, with boulders and coarse debris occupying frontal and lateral positions. The sorting is the result of frost segregation and the "tongue" movement of the solifluction deposit.

Solifluction Terraces

In the study area, solifluction terraces were found in a region which was not covered by glacial ice during the Wisconsinan, but was exposed to the severe climate of the periglacial environment that is conducive to solifluction. These solifluction terraces are generally elongate downslope and have arcuate, convex-downslope fronts (see Figure 6). They merge into the general slope or are covered by additional terraces in the upslope direction. The terrace surfaces (treads) are gently sloping (2° to 4°) and are bounded by a steep, rocky front (risers) sloping 11° to 26°. The length of the risers range from 8 to 29 m; with widths between 20 and 40 m. The majority are between 29 and 34 m wide (Figure 7). Tread lengths are indeterminate because they merge into the slope uphill or are covered. A few can be traced for several tens of meters upslope. Concentration of boulders occurs on the riser front, where the boulders are often packed tightly together. Behind the risers the treads are built of scattered debris and blocks on the surface, with finer material underneath. These solifluction terraces probably formed during the Wisconsinan glaciation of nearby areas when the upland province of the southern Boulder Mountains was exposed to a periglacial environment.

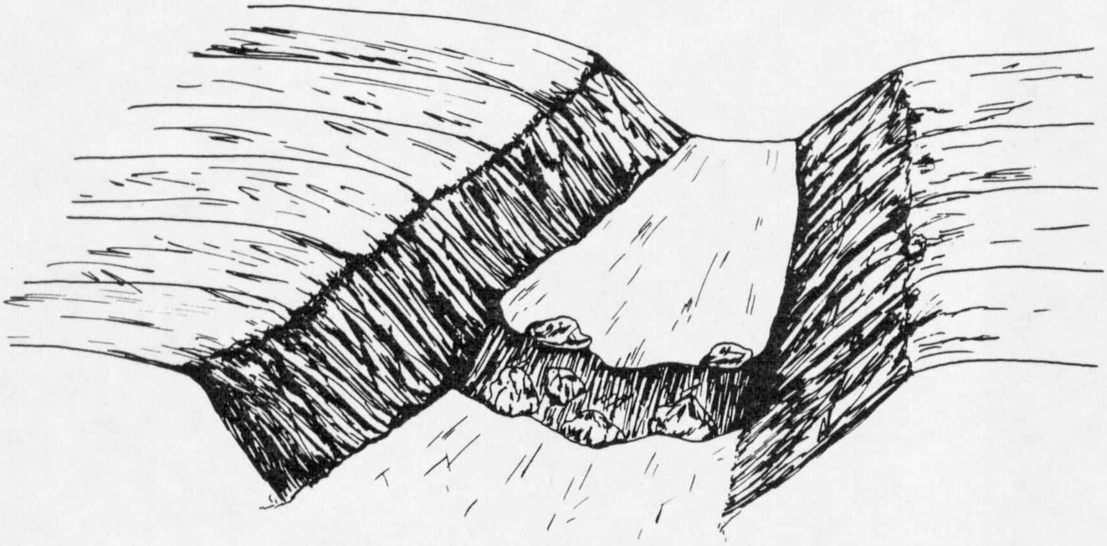
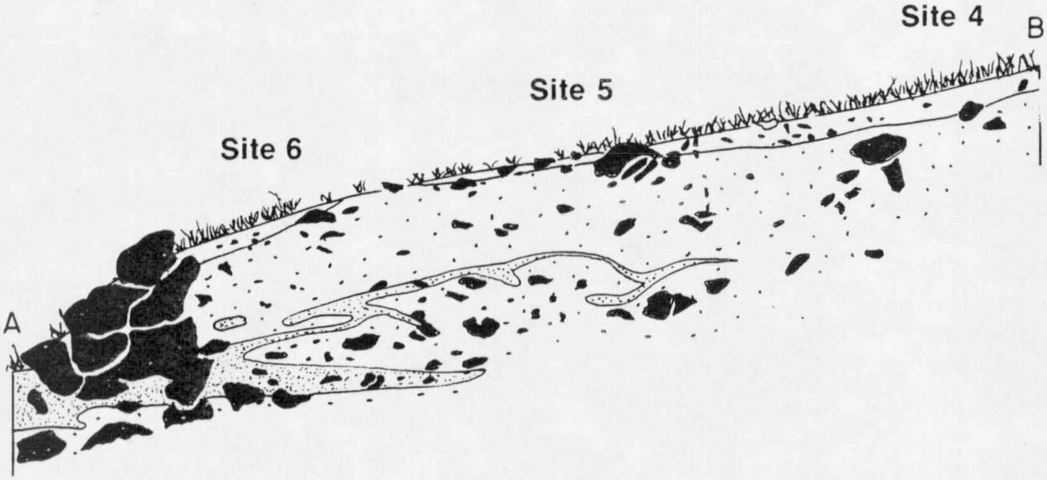


FIGURE 6. Diagram of a solifluction terrace



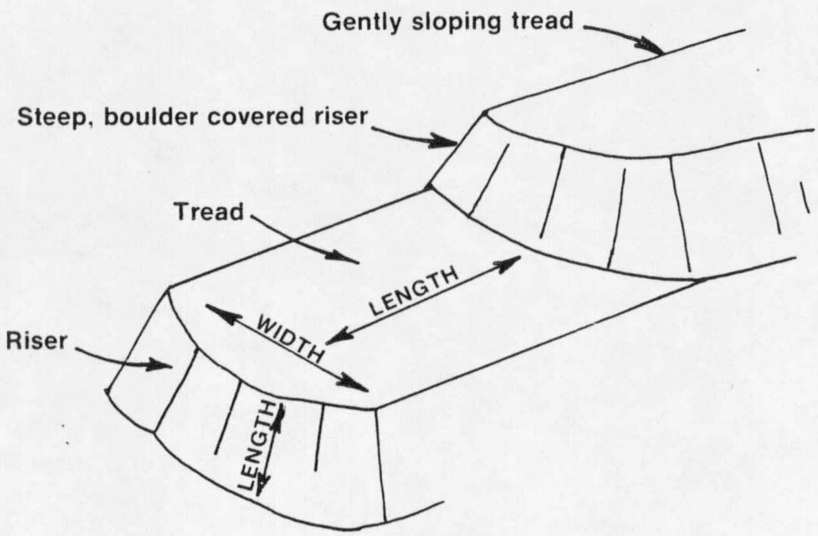


FIGURE 7. Longitudinal diagram of a solifluction terrace with schematic showing location of measurements.
(Modified from Benedict, 1970)

CHAPTER 4

ROCK WEATHERING AND MINERAL STABILITY SEQUENCES IN SOILS FORMED ON THE BOULDER BATHOLITH

In this chapter, basic concepts of rock weathering are briefly discussed, followed by a discussion of mineral stability sequences in the soil profile. The mineralogy of the parent rock and the soils formed on stable areas must be known in order to understand the changes from parent rock to soil. Soils formed on stable sites will exhibit a different mineral stability sequence than soils that have undergone mass movement. In order to test the mass movement hypothesis, the soil profiles of solifluction terraces were compared to those found on stable areas. The soil profiles were compared by a textural and mineralogical analysis of the coarse and clay-size minerals and by examining particle size distibution within the soil.

Rock Weathering

The transition from fresh bedrock to regolith involves numerous chemical and physical weathering processes. These processes are, in part, controlled by the physical, mineralogical, and chemical properties of the rock itself. Granitic rock, such as that found in the Boulder batholith, is formed under tremendous pressures and high temperatures and consequently is in a state of disequilibrium when exposed to conditions at the earth's surface. The overriding driving force for weathering processes is a continual readjustment the rock must make toward thermodynamic equilibrium dictated by the new environment at the earth's surface. As Ollier (1969) points out, it is not necessary to assume that the cooled magma at depth ever achieved a true thermodynamic equilibrium, only that this rock, when brought to the earth's surface, is less in equilibrium with surface conditions than its potential weathering products.

Reiche (1950) stated that physical weathering is brought about by expansion resulting from 1) unloading, which leads to the development of large scale fractures and joints in the bedrock, 2) crystal growth of salts and ice, a process which includes freeze-thaw action, and 3) organic activity, including plant and animal activities. `Physical weathering contributes to chemical weathering by increasing the surface area exposed to chemical reactions.

The chief chemical weathering processes are 1) hydration, 2) hydrolysis, 3) oxidation, 4) carbonation, and 5) solution. Chemical weathering processes may create relatively more stable new mineral phases, or leave a residue when other constituents are removed. Chemical weathering becomes more important than physical weathering as particle size decreases and surface area increases. The results of chemical weathering can be measured by the amounts and types of residual minerals and what newly formed material is found in the weathered residuum.

Mineral Stability Sequences

The mineralogical changes reflecting various stages or degrees of weathering are numerous. The distribution of mineral types and particle size within the soil profile yields information on the age and stability of the solum. Because of their chemical composition, bonding and temperature of formation, minerals weather at different rates. A weathering sequence based on the relative resistance to weathering of minerals has been established (Goldich 1938). His sequence is for coarse-grained minerals and resembles Bowen's reaction series (Table 4), reflecting the increasing instability of minerals with increasing departure from their temperature of formation. This

Table 4. Bowen's reaction series compared to Goldich (1938) mineral stability sequence.

Bowen			Goldich
Olivine		Least Stable	Olivine
	Calcic Plagioclase		Ca-feldspar (anorthite)
Augite			Pyroxene
	Calc-alkalic Plagioclase		Amphibole
Hornblende	Alkali-calcic Plagioclase		Na-feldspar (albite)
	Alkalic Plagioclase		
Biotite			Biotite
			K-feldspar (orthoclase)
			Muscovite
Potash Feldspar			Clay minerals
Muscovite			
Quartz			Quartz
			Al-oxides
		Most Stable	Fe-oxides

sequence does not imply that one mineral in the sequence will weather to a lower mineral in the sequence. For example, olivine does not weather to augite, but rather that under given weathering conditions, olivine will be completely disintegrated by weathering before augite. Jackson and Sherman (1953) established a weathering sequence for clay-size minerals (Table 5) and found differences from the coarse-grained mineral sequences, primarily due to the larger surface area exposed in clay-size minerals. The increased surface area hastens the weathering of minerals which, in coarser sizes, are relatively more stable. The presence of certain clay-size minerals can be used as indicators of the weathering intensity to which a particular soil has been subjected. In the study area biotite, quartz, illite, smectite, and kaolinite are found in the clay size fraction and are underlined in Table 5.

The degree of alteration and weathering of a mineral in a soil is greatest at the surface and decreases in intensity downward to the unaltered parent rock. Therefore, given enough time, only extremely stable minerals and secondary products of less stable minerals will be found near the surface of the soil profile. Figure 8 shows the relationship of coarse-size minerals and their alteration products of different stabilities within a stable soil profile. On the left are the individual minerals of the parent rock and their alteration products with M1 being the most stable and M5 the least stable. In the center and to the right, the approximate mineral composition of each horizon is shown in a generalized stable soil profile within the study area. The parent rock is Butte quartz monzonite, and its constituent minerals are shown along the horizontal axis of Figure 8 progressing from most stable on the left to least stable on the right according to Goldich's stability sequence. In the A horizon for example, quartz is the predominate unaltered mineral with significant amounts of partially weathered orthoclase and biotite, and minor amounts of highly weathered

Table 5. Weathering sequence of thirteen stages for clay-size minerals (modified from Jackson and Sherman, 1953).

Weathering Stage	Dominant clay-size mineral species	
		Least Stable
1	Gypsum (halite and other salts)	
2	Calcite (dolomite, aragonite, apatite)	
3	Olivine (pyroxenes, amphiboles)	
4	Biotite (glauconite, chlorite, nontronite)	
5	Albite (anorthoclase, stilbite, orthoclase, microcline)	
6	Quartz (cristobalite, opal)	
7	Illite (muscovite, sericite)	
8	Interstratified 2:1 layer silicates and vermiculite	
9	Smectite (and the members of the smectite group)	
10	Kaolinite (halloysite, disordered kaolinite)	
11	Gibbsite (boehmite, allophane, etc.)	
12	Hematite (goethite and other iron oxides)	
13	Zircon (illmenite, rutile, tourmaline and others)	
		Most Stable

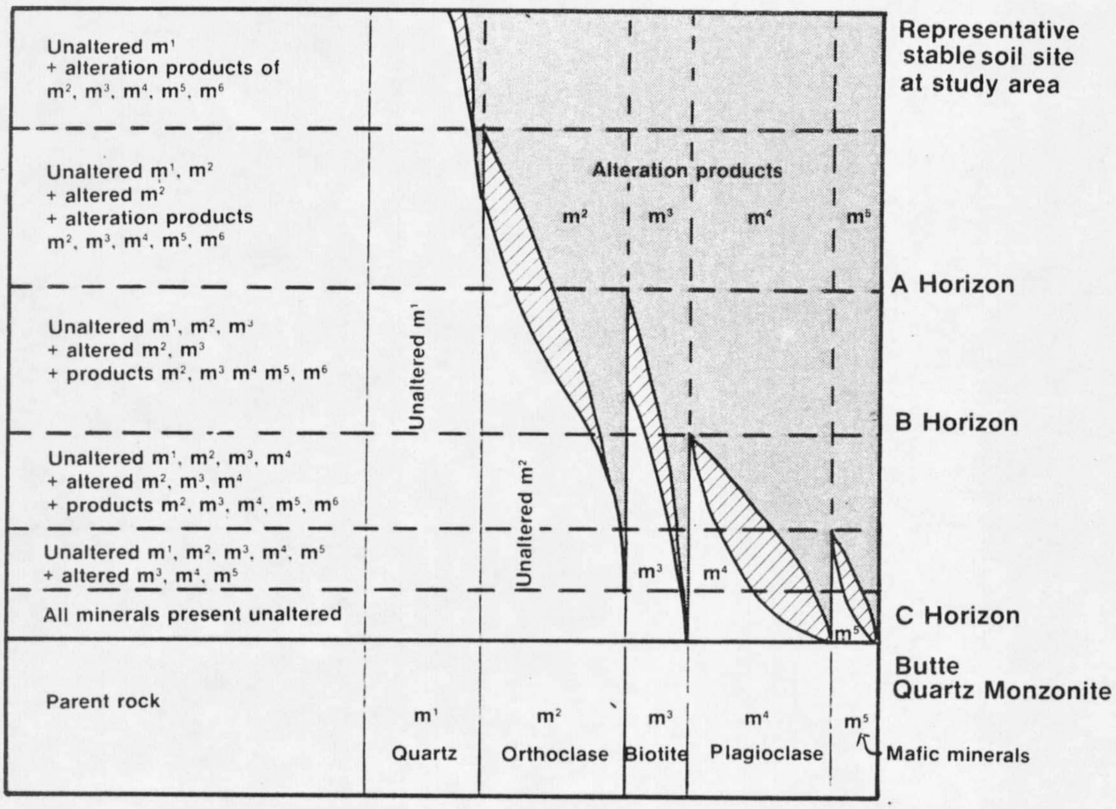


Figure 8. Relationship of minerals and alteration products of different stabilities in a soil profile (Modified from Jackson, K. C., 1970)

- unweathered minerals
- partially decomposed minerals
- weathering products of each mineral

plagioclase. Mafic minerals are typically absent. By contrast, at the base of the soil profile, adjacent to the parent rock, the entire spectrum from the most to the least stable minerals of the parent rock are found.

In areas that have undergone mass movement, this sequence is disturbed and a mixture of minerals of different stabilities will be found throughout the soil profile. Mass wasting processes such as soil creep, earth flow, and mudflow disturbs the surface layer, exposes new material at the surface and brings new material up from depth. Additional processes favored by a periglacial environment often cause vertical motion in the soil minerals. These processes include frost heaving and solifluction. These processes are discussed in the previous chapter.

Particle Size Distribution

For soils within the study area, graphs of particle size distribution within each soil profile were made (Appendix B). These graphs show the percentage of different size fractions within each soil horizon for every site. Two characteristics of the soils are readily apparent. The first is that the sand size fraction makes up approximately fifty percent of the soil. This reflects the original igneous parent rock texture with crystals that generally range in size from 0.5 to 3 millimeters. Secondly, the particle size distribution is much more uniform with depth on stable sites than on solifluction terraces. Percentages of sand, silt, and clay size particles do not vary greatly between horizons in stable site profiles. On the solifluction terraces, however, the percentages of most particle sizes show wide variation between the A, B, and C horizons. The variation within the particle sizes in each horizon within the solifluction sites is indicative of churning in the soils, possibly through frost heaving, as well as movement of material through solifluction.

Parent Material and Soil Mineralogy

In order to understand the mineralogy of the soils, the mineralogy of the parent material must be known. The parent rock in the study area is predominantly a medium- to coarse-grained, gray, quartz monzonite. Based on thin section analysis the parent rock in the study area is composed of 20-25% plagioclase, ranging in composition from oligoclase to andesine, 40-45% orthoclase, 20-25% quartz, 5-10% biotite, and 0-5% hornblende. In the specimens studied, the ratio of orthoclase to plagioclase is approximately 2:1.

The unaltered rock is light gray, having a hypidiomorphic-granular texture with crystals ranging in size from 1.5 to 2.5 mm. The plagioclase crystals are subhedral to euhedral and are often zoned with a core of andesine and a rim of oligoclase. Albite twinning is also occasionally seen. Orthoclase crystals range from small anhedral grains to occasional large, anhedral to euhedral crystals. Microperthite and Carlsbad twinning is observed. Quartz is anhedral and usually occurs as groundmass or irregular interstitial masses. The quartz exhibits strain shadows and undulose extinction. Biotite occurs as shiny, black euhedral to anhedral plates.

Stable Sites

Observations on the stable sites were made of the coarse size minerals in the C horizon and compared to the coarse size minerals found in the A horizon. Ideally, a change in the relative percentage of the stable minerals would be seen with change in depth within the soil horizon. For example, in the surface horizons where the weathering environment is the most intense, quartz would be expected to have a higher concentration relative to the less stable coarse-size minerals. Observations indicate that weathering has not proceeded to such an extent that significant changes in the percentage of individual coarse grained minerals with depth can be seen (Table 6). However, observations of the textural changes in the minerals relative to depth on stable sites reveal important differences.

C horizon. Petrographic observations using thin sections and microscopic observations of the coarse fraction of the C horizon show little visual evidence of weathering (Figure 9). The biotite is shiny, black with irregular edges. Even in this relatively unweathered sample, iron oxide stain can be seen on the feldspars. Plagioclase and orthoclase occur as subhedral crystals with hairline fractures. Slight sericitization of the plagioclase crystals is common. Quartz remains anhedral to subhedral and contains some microfractures.

A horizon. In the A horizon weathering has caused the disintegration of the coarse size minerals (Figure 9). In general there appears to be a slight decrease in grain size. The biotite flakes have curled edges and have become a golden brown. Plagioclase

Table 6. Distribution of coarse-size minerals within stable site soil profiles (>100 microns).*

Site	Horizon	Depth cm	Quartz	Orthoclase	Biotite	Plagioclase
1	A1/A2	0–10	M	M	L	L
	B2	30–40	M	H	L	L
	C	45–55	M	M	M	L
2	A2	15–25	M	H	L	L
	B2t	30–40	M	M	L	L
	C	75–85	M	M	L	M
3	A1	0–10	M	H	L	L
	A2	15–25	M	H	L	L
	B2	30–40	M	H	M	M
	C1/C2	45–55	M	M	M	L

* H = high (35-60%), M = medium (20-35%), L = low (5-20%)

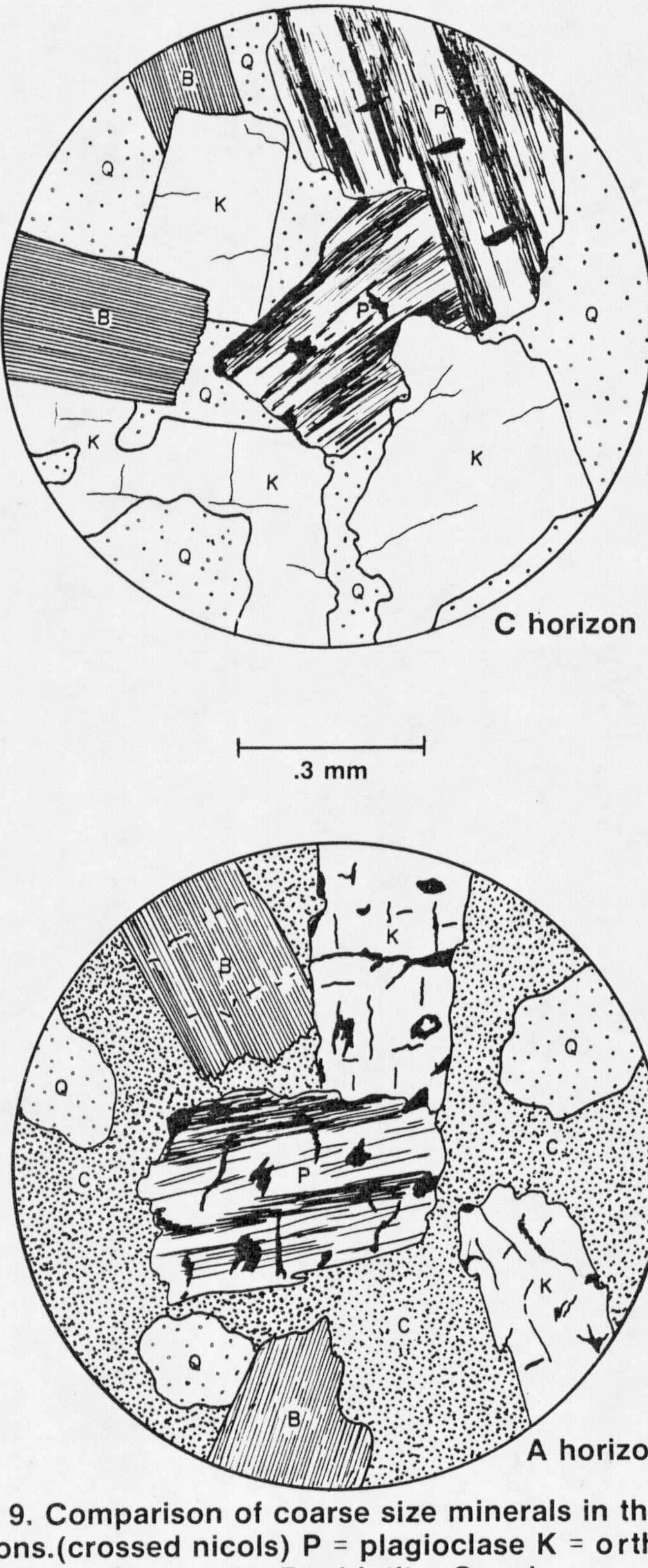


FIGURE 9. Comparison of coarse size minerals in the A and C horizons.(crossed nicols) P = plagioclase K = orthoclase Q = quartz, B = biotite, C = clay

minerals show increased fracturing and sericitization, especially along the fractures. Only slight evidence of chemical weathering is seen on orthoclase, which is cloudy due to argillization.

Solifluction Terrace Sites

On solifluction sites the coarse sized fraction contains a mixture of fresh and unaltered minerals at every depth. This supports the observations first made in the field on the solifluction soil sites, where fragments of unweathered rock were located in the A and B horizons and discrete units of clay and weathered material were found adjacent to relatively unweathered material in the C horizon.

In general textural changes caused by weathering within the coarse fraction in the soil profile decreases with depth on the stable sites. By contrast, both weathered and fresh coarse-size minerals are found distributed throughout the soil profile.

These observations indicate that the soils on the solifluction terraces have undergone turbation. This lends support to the hypothesis that the solifluction terraces were formed by mass movement processes. Further observations were made in the clay size fraction which is discussed in the following section.

Clay Mineralogy

Mineralogy of the clay size fraction was determined using x-ray diffraction techniques described by Whittig (1965). X-ray diffraction peaks were used to determine the clay mineralogy, the amount of clay minerals, and the distribution of the clay minerals within the different soil profiles. This information was used to compare soil profiles formed on stable areas to soils formed on solifluction terraces.

The minerals present in the clay size fraction of the soils include quartz, illite, smectite, kaolinite, and vermiculite. Quartz, identified by a strong peak at 3.34A and occasionally by a less intense peak at 4.26A, was present in all the samples (Figure 10). A 10A reflection is characteristic of the illite minerals, but a weak 002 reflection occurring with the peak suggests it may also partially represent fine particle size biotite (Hood, 1963). Kaolinite was identified as a peak at 7.15A on Mg-saturated, ethylene glycol-solvated samples. A reflection at 14.4A on the Mg-saturated air dried diffraction pattern shifted to the 17A region on treatment with ethylene glycol. This expansion is characteristic of the smectite minerals. Upon saturation of the sample with potassium and heating to 500°C for two hours, the 14A peak collapsed to 10A, confirming it's identification as smectite.

Semiquantitative estimates of the various clay minerals in each sample were made by measuring the areas beneath the diffraction peaks. Areas were adjusted using weighting factors established by Klages and Hopper (1982) to give estimated relative mineral abundance in percentages. These percentages were obtained using the equation:

$$\text{Relative Clay Mineral Abundance} = \frac{\text{Area of diffraction peak of individual clay minerals } (2\text{-}.2\mu) \div \text{ weighting factor}}{\text{Areas of diffraction peaks of clay minerals } (2\text{-}.2\mu)}$$

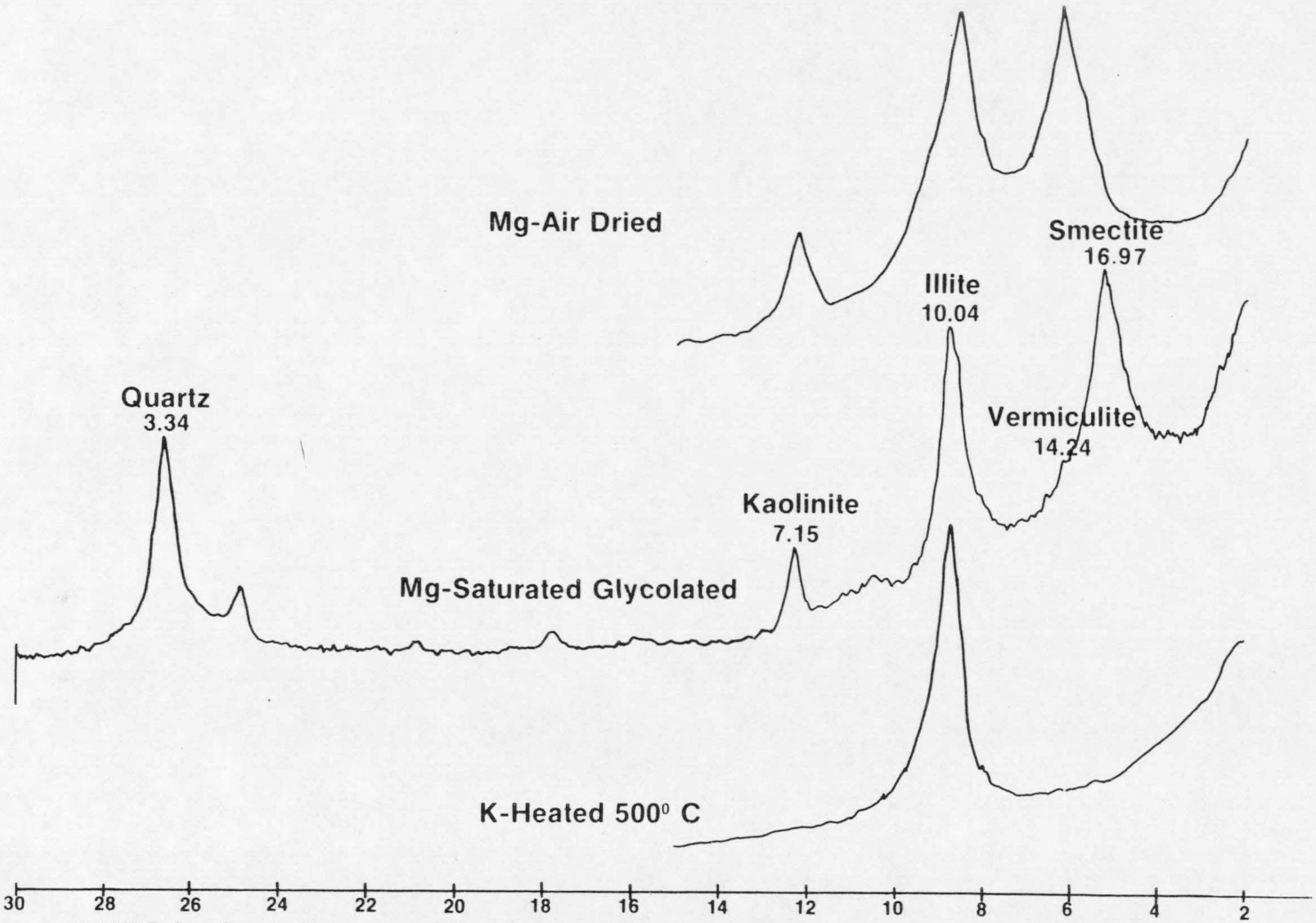


FIGURE 10. Example of x-ray diffraction pattern for clay size (2-.2 microns) material

Table 7 gives the range of percentages of clay minerals found in soil profiles developed on stable sites, and Appendix C contains the relative clay mineral abundance for each soil profile. Location of the stable sites relative to the solifluction terrace sites is shown in Figure 11.

Table 7. Range of relative clay mineral abundance found in soil profiles on stable sites (2-.2 microns).

Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Vermiculite %
A1/A2	0-25	6.0- 9.5%	57.0-80.0%	4.0-19.0%	10.0-15.0%	0-2.0%
B2/B2t	30-40	7.0-17.0%	41.0-81.0%	4.0-35.0%	9.0-13.0%	trace
C	45-85	9.0-18.0%	45.0-71.0%	5.0-26.0%	13.5-16.0%	trace-2%

Stable Sites

Relative Clay Mineral Abundance. In general the clay size fraction of the stable sites profile is characterized by an abundance of illite. There is a tendency for illite to increase gradually from the parent rock to the surface. Smectite and quartz are the next most abundant clay-size minerals. Both minerals gradually decrease in relative abundance from the parent rock to the surface (Figure 12). Kaolinite also shows a steady decrease in relative abundance from the parent rock to the surface. However, in site 2, where a strong argillic horizon is developed on the stable site, kaolinite increases in the B horizon while illite decreases in relative abundance (Figure 13). Vermiculite is a minor clay-size mineral which tends to decrease from the parent material to the surface.

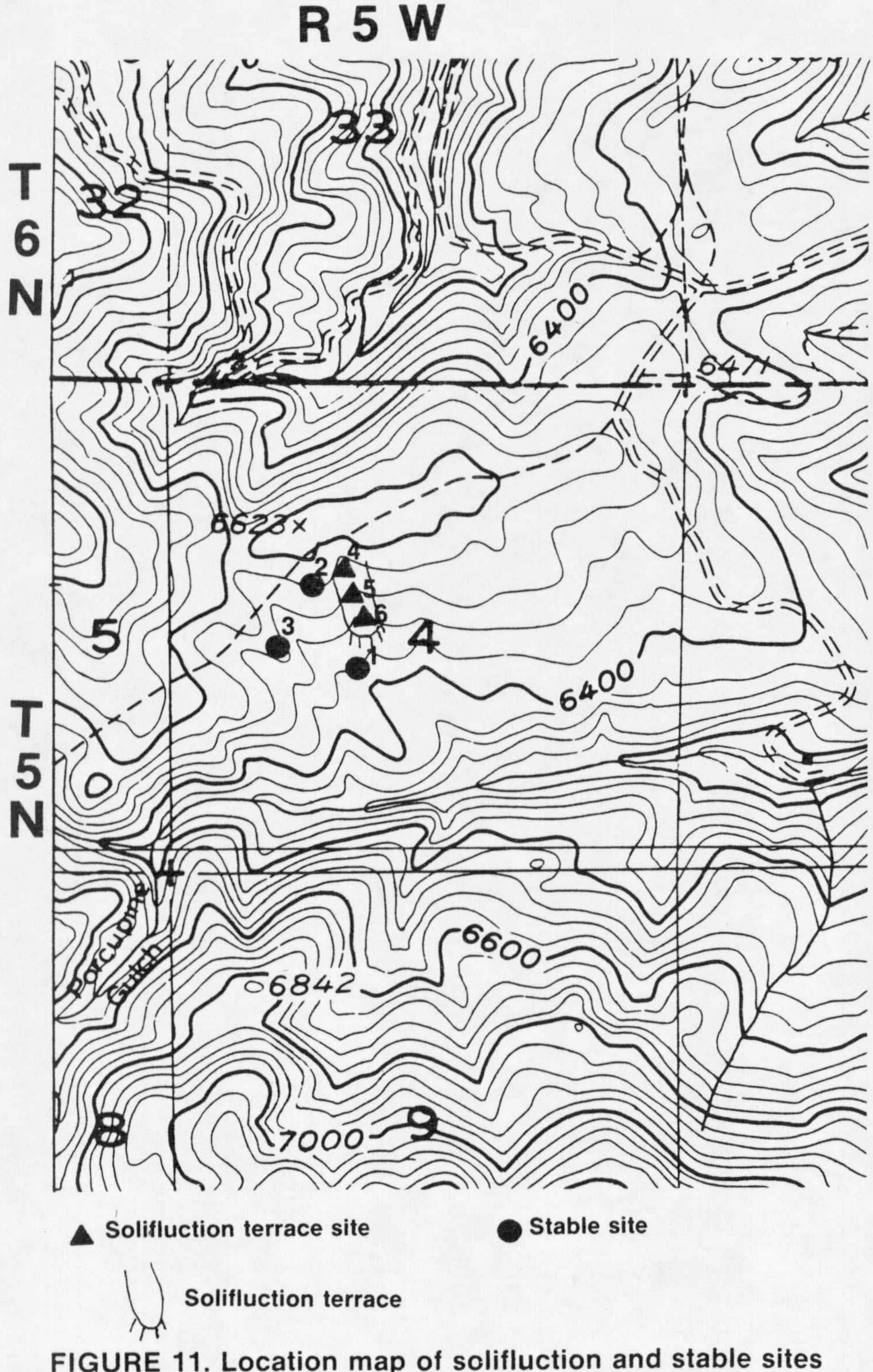


FIGURE 11. Location map of solifluction and stable sites subjected to detailed mineralogic analysis.

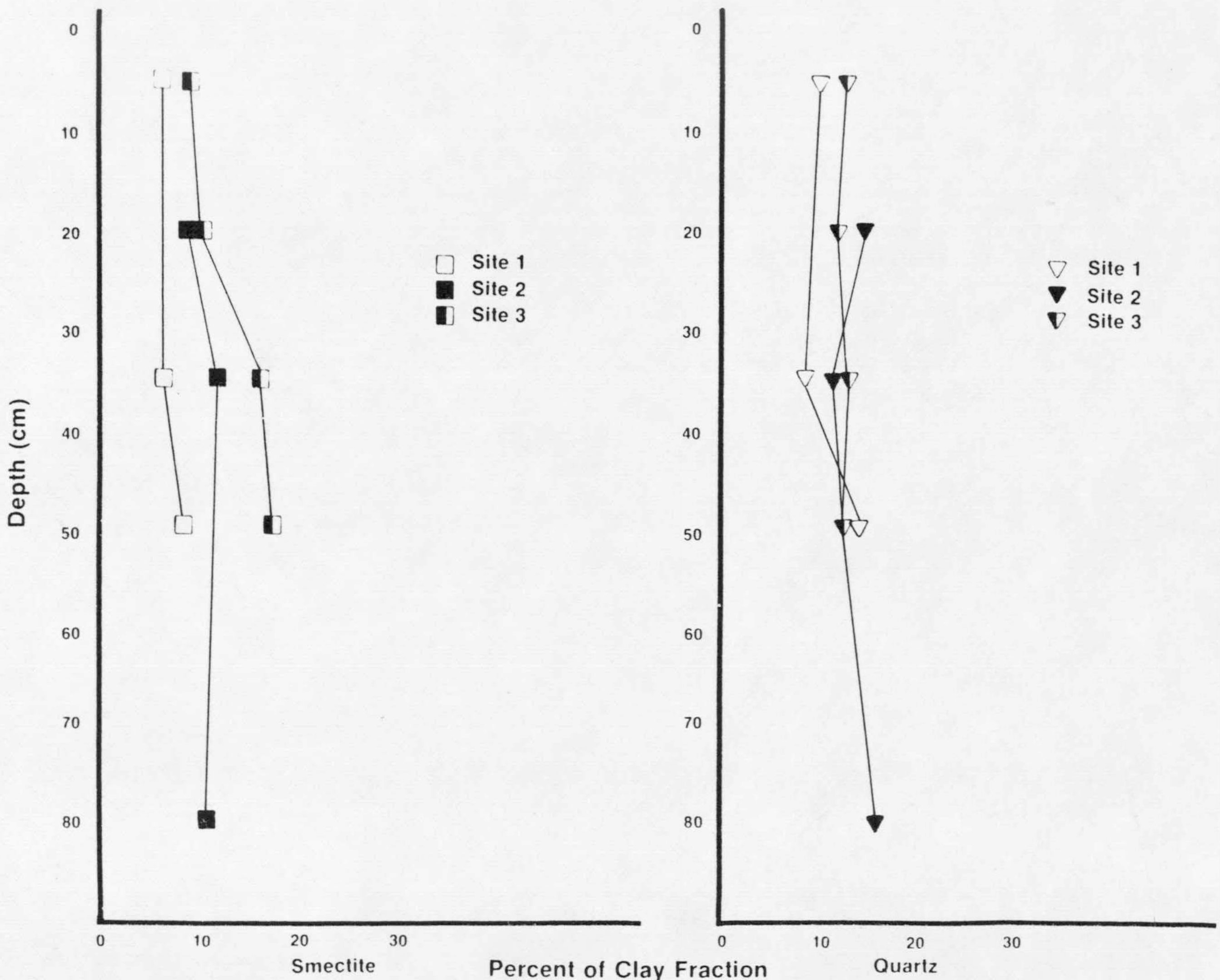


Figure 12. Change in relative abundance of smectite and quartz with depth on stable sites.

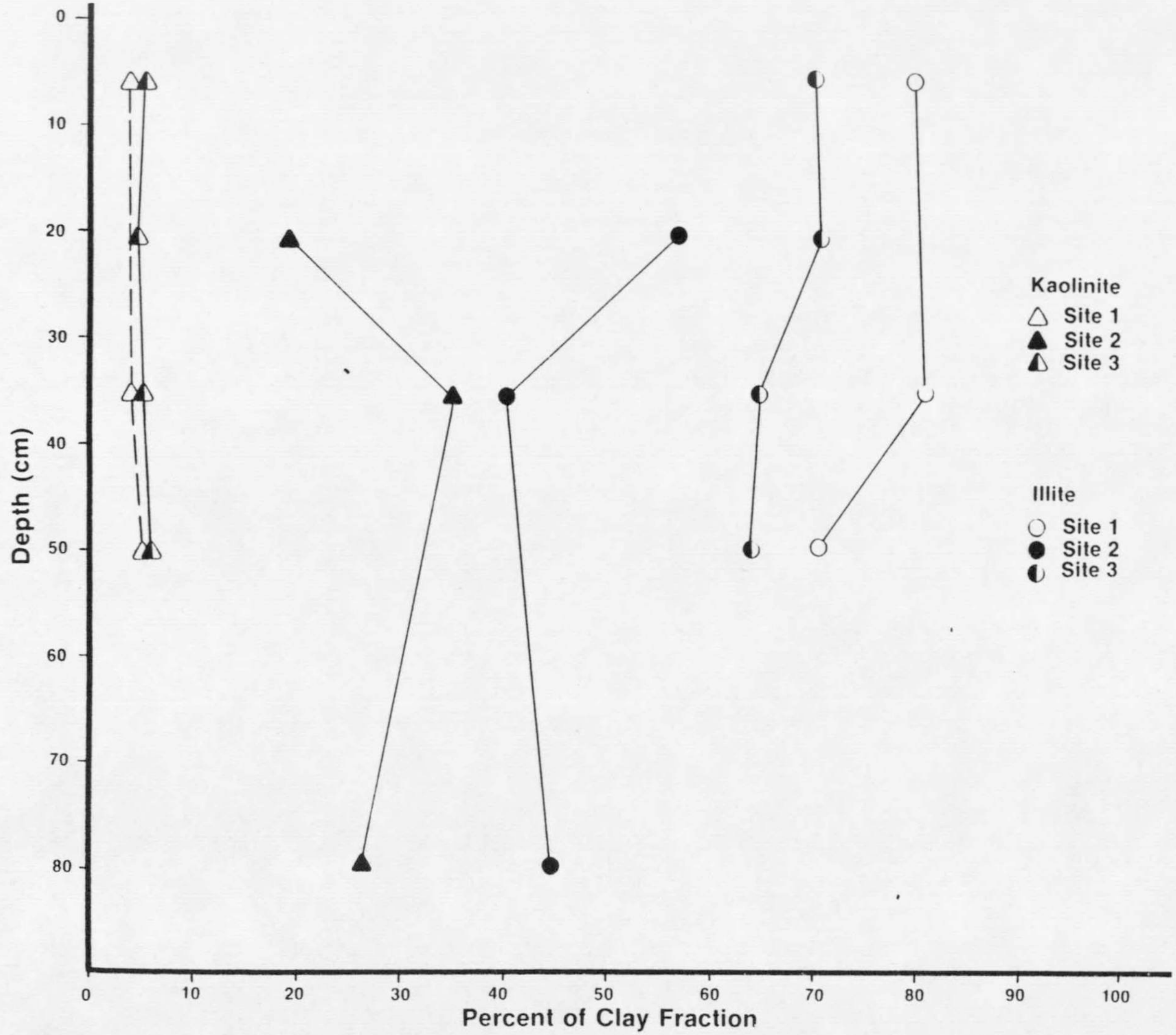


Figure 13. Change in relative abundance of kaolinite and illite with depth on stable sites.

Absolute Abundance. Absolute abundance is the measurement of the percent clay-size mineral of the whole rock versus relative abundance which measures the percent clay-size mineral relative to the other clay-size minerals present. It is calculated using the relative abundance in relation to the percent clay-size fraction. Illite gradually increases from the parent rock to the top of the B horizon and then decreases into the A horizon. Smectite follows the same pattern. Quartz and kaolinite gradually decrease from the parent rock to the surface (Figure 14). In site 2, there is a sharp increase in kaolinite in the argillic horizon as mentioned earlier, but there is no corresponding decrease in illite. These relationships are illustrated and the data included in Appendix D.

Mineral Assemblages. Illite is the dominant mineral in all horizons in all of the stable soil sites. Clay mineral assemblages for almost all of the horizons in order of decreasing amounts are; illite-quartz-smectite-kaolinite (Table 8). A variation of this occurs in site 3 where the B and C horizons have an illite-smectite-quartz-kaolinite clay mineral assemblage. This change is probably the result of illuviation of clay from the A horizon to the B horizon. Illuviation is the process of translocation of clay-size particles from the A horizon, the zone of leaching, to the B horizon, the zone of accumulation. Site 2 varies by having a large amount of kaolinite throughout the soil profile so that the clay mineral assemblage changes to illite-kaolinite-quartz-smectite.

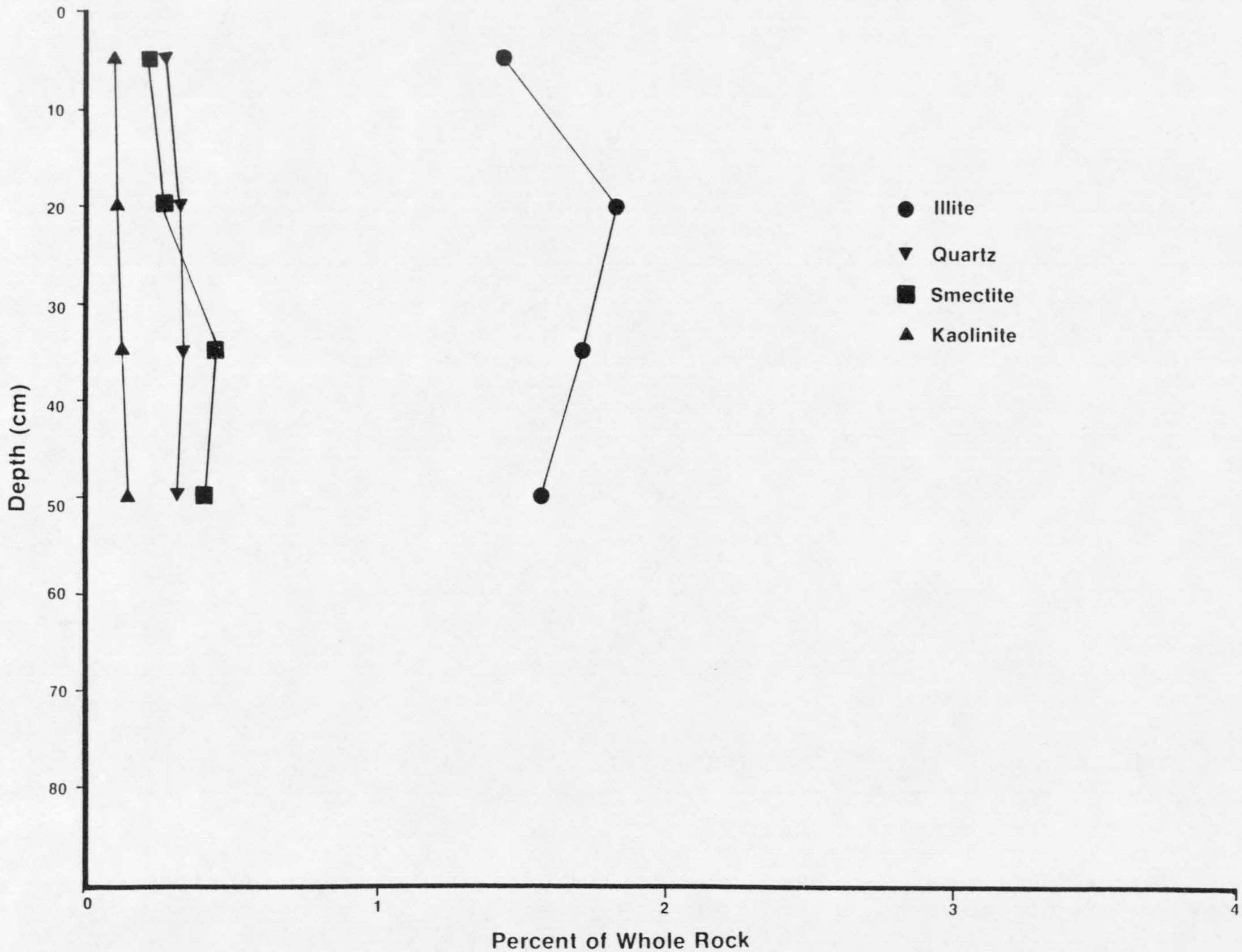


Figure 14. Absolute abundance of clay minerals on stable Site 3.

Table 8. Clay mineral assemblages of stable sites from most to least abundant.

Site	Horizon	Clay Mineral Assemblage
1	A	Illite-Quartz-Smectite-Kaolinite
	B	Illite-Quartz-Smectite-Kaolinite
	C	Illite-Quartz-Smectite-Kaolinite
2	A	Illite-Kaolinite-Quartz-Smectite
	B	Illite-Kaolinite-Quartz-Smectite
	C	Illite-Kaolinite-Quartz-Smectite
3	A	Illite-Quartz-Smectite-Kaolinite
	B	Illite-Smectite-Quartz-Kaolinite
	C	Illite-Smectite-Quartz-Kaolinite

The significance of the data mentioned above with respect to factors of soil formation and mineral stability are included in a later discussion.

Solifluction Terrace Sites

Relative Clay Mineral Abundance. On the solifluction terraces, illite is still the most abundant mineral, however in terms of relative percentages it is less than the illite formed on stable sites (Table 9). Instead of gradually increasing from the parent rock to the surface, as in the stable sites, illite stays essentially the same from the C horizon to the B horizon and increases sharply in the A horizon (Figure 15). In general, kaolinite is the next most abundant mineral found in the solifluction terrace soils, rather than smectite as in the stable sites profiles. Kaolinite tends to increase just slightly from the C to the B horizon and then sharply decreases in the A horizon. In the A and C horizons the relative abundance of smectite is similar to that found in the stable sites. However, instead of a gradual decrease from the parent rock to the surface, there is an increase of smectite

Table 9. Range of relative clay mineral abundance found in soil profiles on solifluction terrace sites (2-.2 microns).

Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Vermiculite %
A1/A2	0-25	4.0- 7.0%	68.0-76.0%	7.0-15.0%	10.0-13.0%	trace
B2t	30-55	10.0-29.0%	42.0-54.0%	18.0-26.0%	11.0-14.0%	trace
C	45-85	9.0-17.0%	51.0-57.0%	15.0-27.0%	10.0-12.0%	0-3.0%

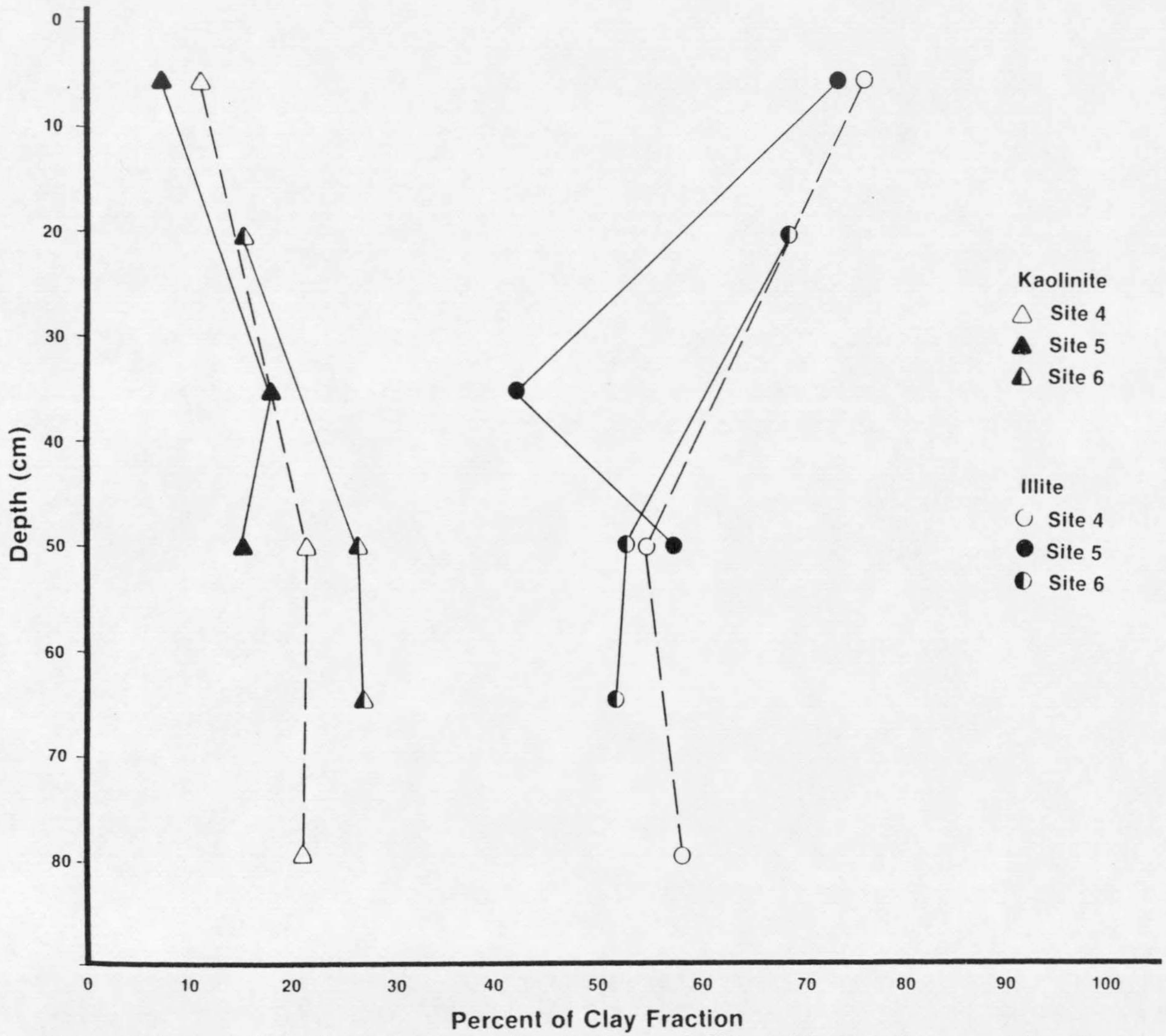


Figure 15. Change in relative abundance of kaolinite and illite with depth on solifluction terrace sites.

in the B horizon (Figure 16). Changes in relative abundance of quartz in the solifluction soil profile are slight, with no real pattern emerging. There is a tendency for the relative abundance of quartz to be almost equal between the B and C horizons, but the A horizon may have slightly more or less quartz relative to these horizons. Vermiculite remains a minor mineral and tends to decrease from the parent material to the surface.

Absolute Clay Mineral Abundance. A study of absolute abundances, the percent clay of the whole rock, demonstrates the greater variability in the distribution of the minerals on a solifluction terrace.

In general, sites located at the head and toe (sites 4 and 6) of the solifluction terrace exhibit similar mineral distribution and abundances and will be discussed separately from the unique profile located in the center of the terrace. In sites 4 and 6, illite changes just slightly from the C to the B horizon. There is a marked twofold increase from the B horizon to the A horizon. Kaolinite is the next most abundant clay mineral found in these sites and it tends to decrease from the parent material to the surface with slight variations occurring in the B horizon. Quartz is distributed fairly uniformly in the site at the head of the solifluction terrace. At the toe of the solifluction terrace, quartz increases from the B to the A horizon. In both sites, smectite gradually decreases from the parent rock to the surface with just a slight increase in the B horizon.

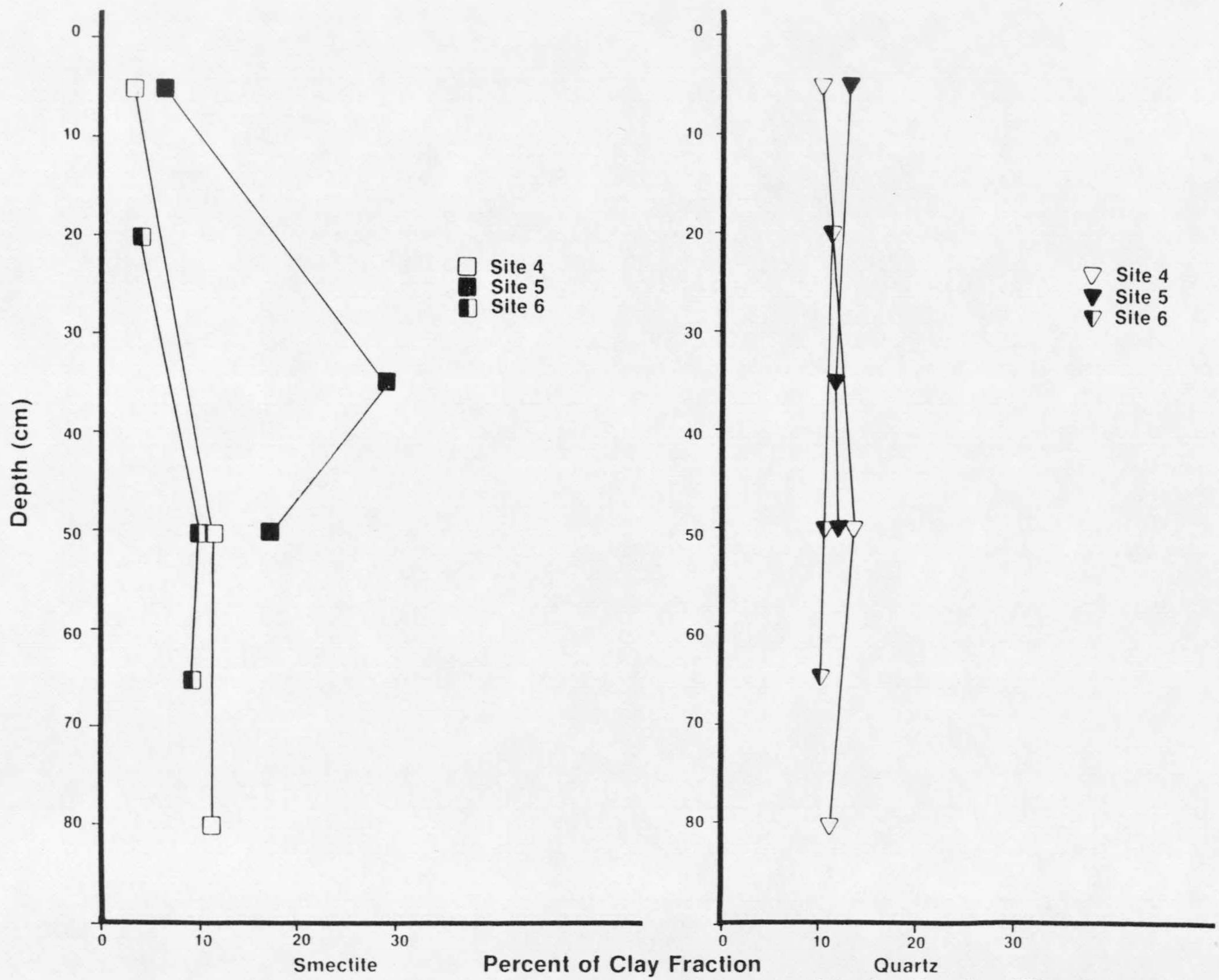


Figure 16. Change in relative abundance of smectite and quartz with depth of solifluction terrace sites.

The soil profile located in the center of the solifluction terrace is unique in many aspects. The distribution and abundance of the clay minerals exhibit rapid changes with respect to depth in the profile. Illite doubles in absolute abundance from the C to the B horizon and triples from the C to A horizon (Figure 17). Whereas kaolinite was the next most abundant mineral in the other solifluction soil sites, smectite is more abundant and exhibits a different distribution pattern. Smectite increases fivefold from the C horizon to the B horizon and decreases equally rapidly from the B to the A horizon. Kaolinite repeats a similar pattern with a threefold increase from the C horizon to the B horizon and a threefold decrease from the B to the A horizon. Quartz increases from the C to the B horizon but exhibits only a slight decrease to the A horizon.

Mineral Assemblages. Illite dominates all the horizons in the soil profiles formed on the solifluction terraces (Table 10). Mineral assemblages of illite-kaolinite-quartz-smectite are found in the head and the toe of the solifluction sites. In the center of the terrace the B and C horizon have clay mineralogy suites of illite-smectite-kaolinite-quartz. The smectite increase is much more pronounced in the B horizon than in the C horizon. In the A horizon, an illite-quartz-kaolinite-smectite mineral assemblage is present. These variations may be the result of influxes of new material and/or mixing of material within the soil profile.

T-test. A t-test was run to check the probability that the differences in the smectite, kaolinite, illite and quartz content of

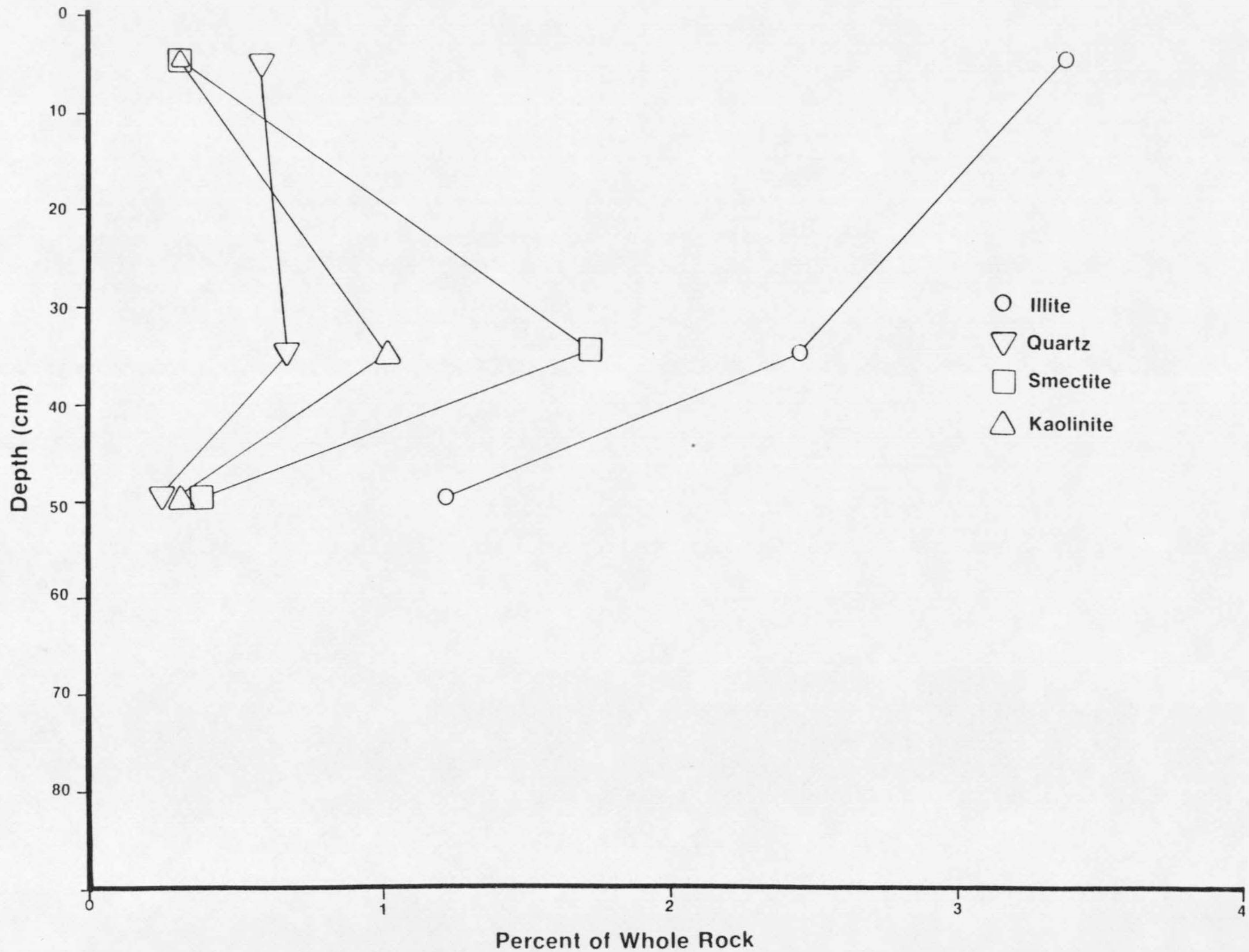


Figure 17. Absolute abundance of clay minerals on solifluction terrace Site 5.

Table 10. Clay mineral assemblages of solifluction terrace sites from most to least abundant.

Site	Horizon	Clay Mineral Assemblage
4	A	Illite-Kaolinite-Quartz-Smectite
	B	Illite-Kaolinite-Quartz-Smectite
	C	Illite-Kaolinite-Quartz-Smectite
5	A	Illite-Quartz-Kaolinite-Smectite
	B	Illite-Smectite-Kaolinite-Quartz
	C	Illite-Smectite-Kaolinite-Quartz
3	A	Illite-Kaolinite-Quartz-Smectite
	B	Illite-Kaolinite-Quartz-Smectite
	C	Illite-Kaolinite-Quartz-Smectite

the stable and the solifluction sites are significant. The test was made by comparing the average relative abundance of each clay mineral in each horizon of the stable sites to the average relative abundance of the same mineral in the same soil horizon in the solifluction sites.

For example, the average relative abundance of smectite in the B horizon of the stable sites was compared to the average relative abundance of smectite in the B horizon of the solifluction sites. The t-test was also applied using the average absolute abundance values of each clay-size mineral in each horizon of the stable and solifluction sites. The results were that the probability of the clay mineral contents being from different populations ranges from 80 to 90 percent. Illite had higher t-test results in the A horizon whereas smectite and kaolinite exhibited higher probability levels in the B horizon. As a general rule, if the probability is 95% or more the differences are considered real, and if the probability is between 95% and 80% there may be real differences but a statistically significant difference is not proven. Therefore the results show a general trend of differences between the stable and solifluction terrace sites. In future studies it is recommended that the number of sites subjected to detailed mineralogical analysis be increased in order to obtain statistically significant results.

Discussion

Illite

In stable sites, illite gradually increases from the parent material to the top of the B horizon and then decreases from the B to the A horizon (Figure 18). In the solifluction sites, the absolute abundance of illite changes little between the B and C horizon, and then increases by a large amount from the B horizon to the A horizon. Studies performed on weathering of granitic rock attribute high illite contents to the weathering of biotite and/or potassium feldspar in dry environments (Veseth, 1981). This agrees with the high percentage of illite in soils found on the Boulder batholith which contain biotite and potassium feldspar and where leaching is diminished because of the dry climate. The distribution of illite in the stable sites soil profiles indicates that biotite in the parent material is weathering to illite. Biotite increases with depth in the stable soil profile and illite content increases from the parent rock to the B horizon. The decrease of illite in the A horizon is likely the result of the instability of the mineral coupled with increased leaching in the A horizon. On solifluction sites, a marked increase in illite content occurs in the A horizon. As seen in the stable sites, illite does not survive near the surface, therefore it is likely that on the solifluction terrace there is an influx of illite-bearing material. Two likely mechanisms are downslope movement either by runoff or by mass movement processes and frost heaving bringing illite from depth.

Quartz

In general, quartz decreases from the parent material to the surface in soil formed on stable sites (Figure 19). Careful observations of changes in the relative abundance of quartz with depth suggest a slightly higher concentration in the A horizon relative to

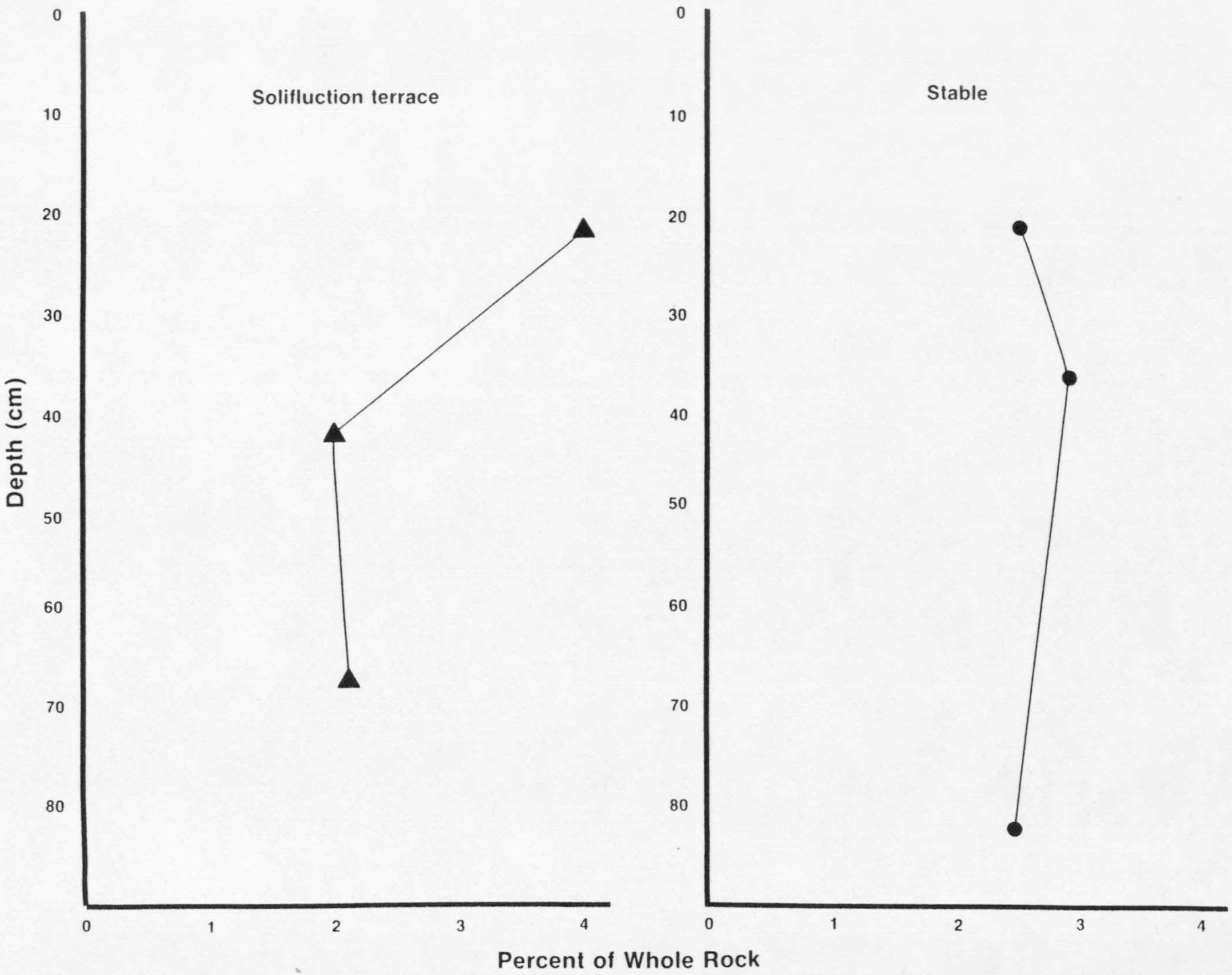


Figure 18. Comparison of the absolute abundances of illite in stable and solifluction sites.

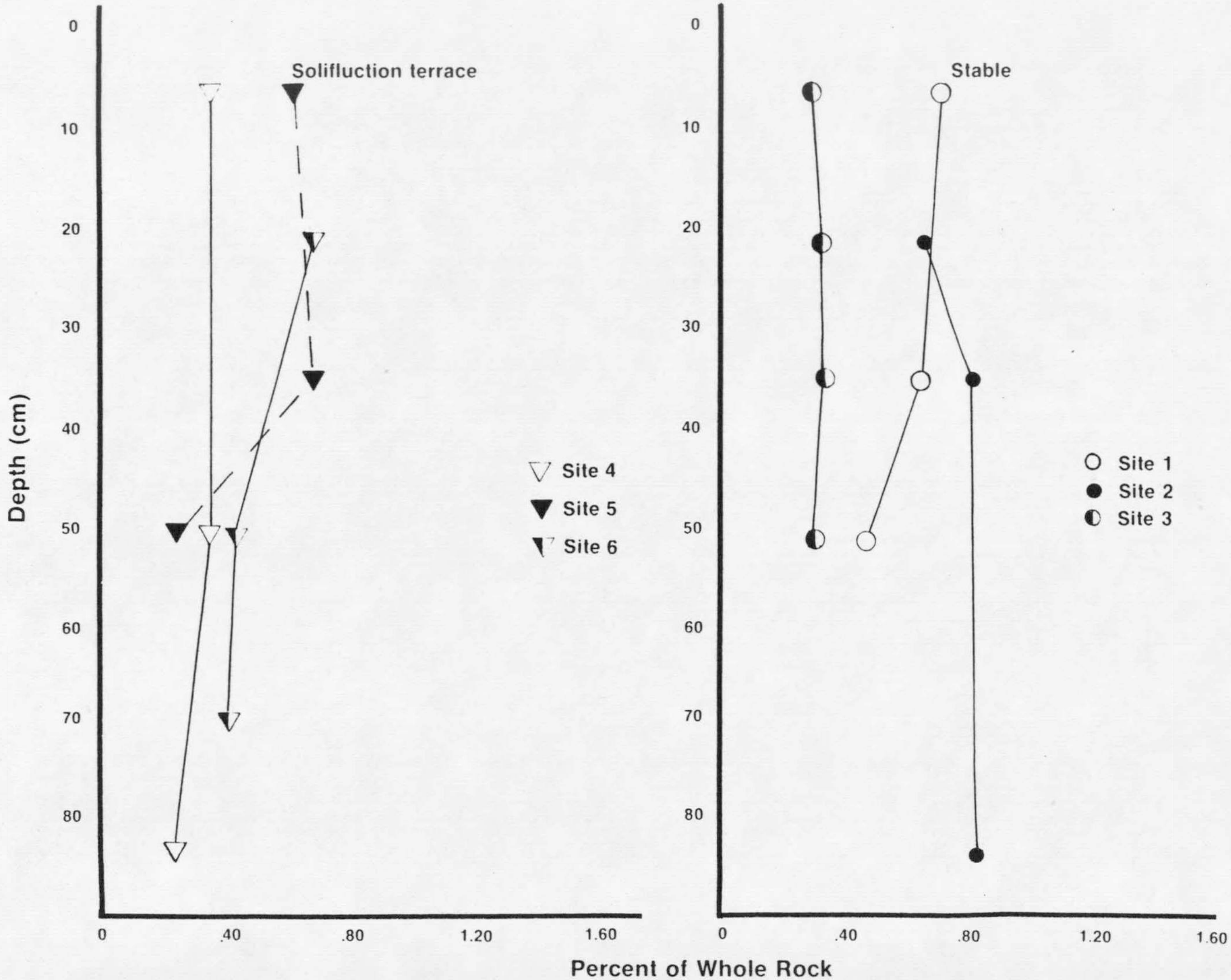


Figure 19. Comparison of the absolute abundances of quartz in stable and solifluction sites.

the B horizon. This may be attributable to the illuviation of clay minerals from the A horizon leaving the more resistant quartz minerals behind. The diminished relative abundance of quartz within the B horizon is probably the result of the concentration of clay-size particles by illuviation. On the solifluction terrace, each soil profile has a different distribution of quartz. At the head of the terrace the B horizon contains more quartz than the C horizon, suggesting an influx of quartz-bearing material. In the center of the solifluction terrace, the B horizon contains more quartz than the A horizon and they both have greater absolute abundance of quartz than the C horizon. This indicates a mixing of material from the base of the profile bringing quartz to the surface. There may also be an influx of material from upslope. At the toe of the solifluction terrace, quartz is found in greatest absolute abundance at the surface which suggests that material is being moved downslope into this area.

Smectite and Kaolinite

In stable sites, smectite generally decreases from the parent material to the surface (Figure 20). There is a tendency for a slight increase in the B horizon, reflecting illuviation of smectite from the A horizon. Soil profiles at the head and toe of the solifluction terrace show a similar smectite distribution. However, in the center of the solifluction terrace, smectite greatly increases in the B horizon. Kaolinite is a small fraction of the clay minerals present in the stable sites and, in general, decreases in abundance from the parent material to the surface (Figure 21). However, in

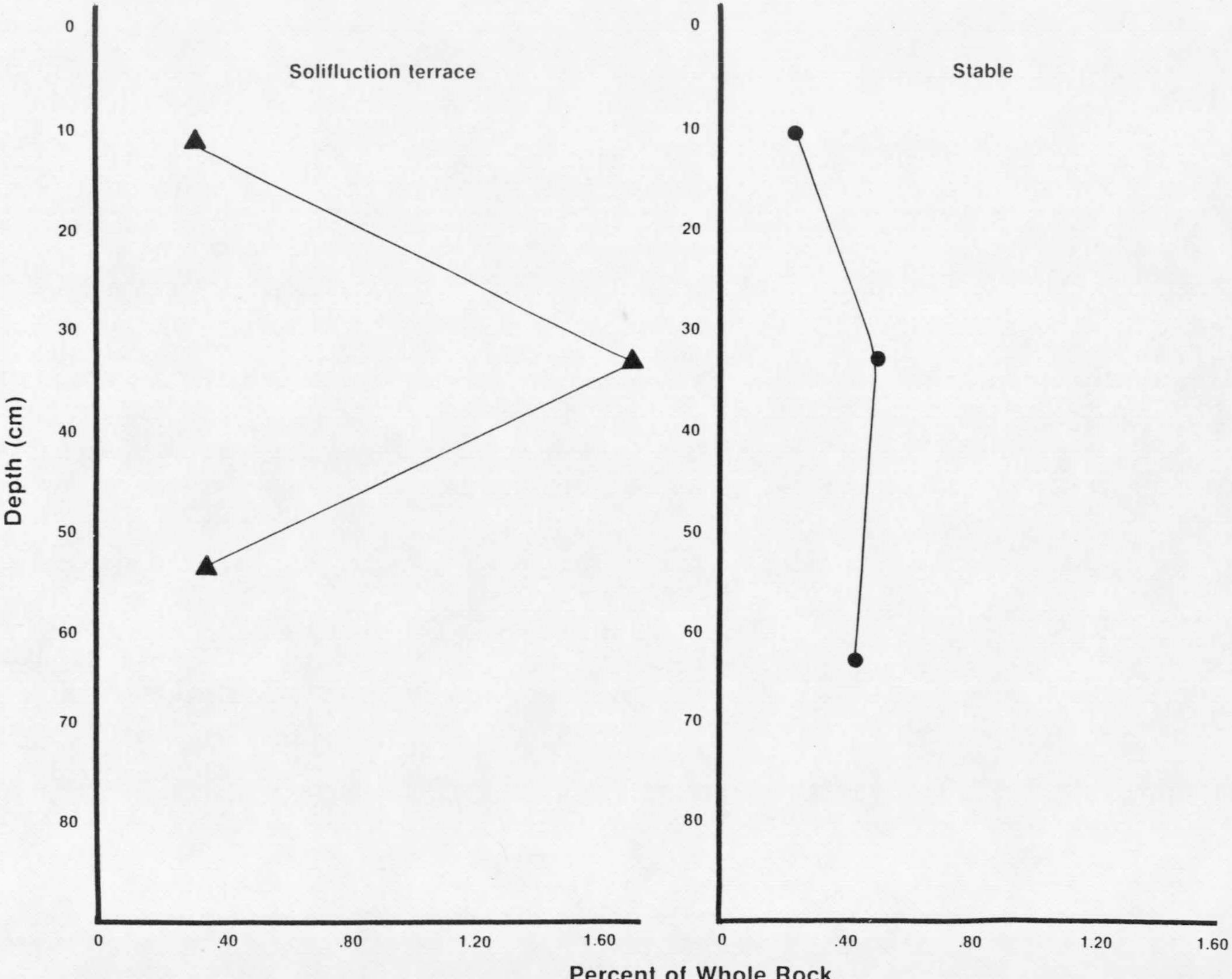


Figure 20. Comparison of the average absolute abundances of smectite in stable sites 1 and 3 and solifluction site 5.

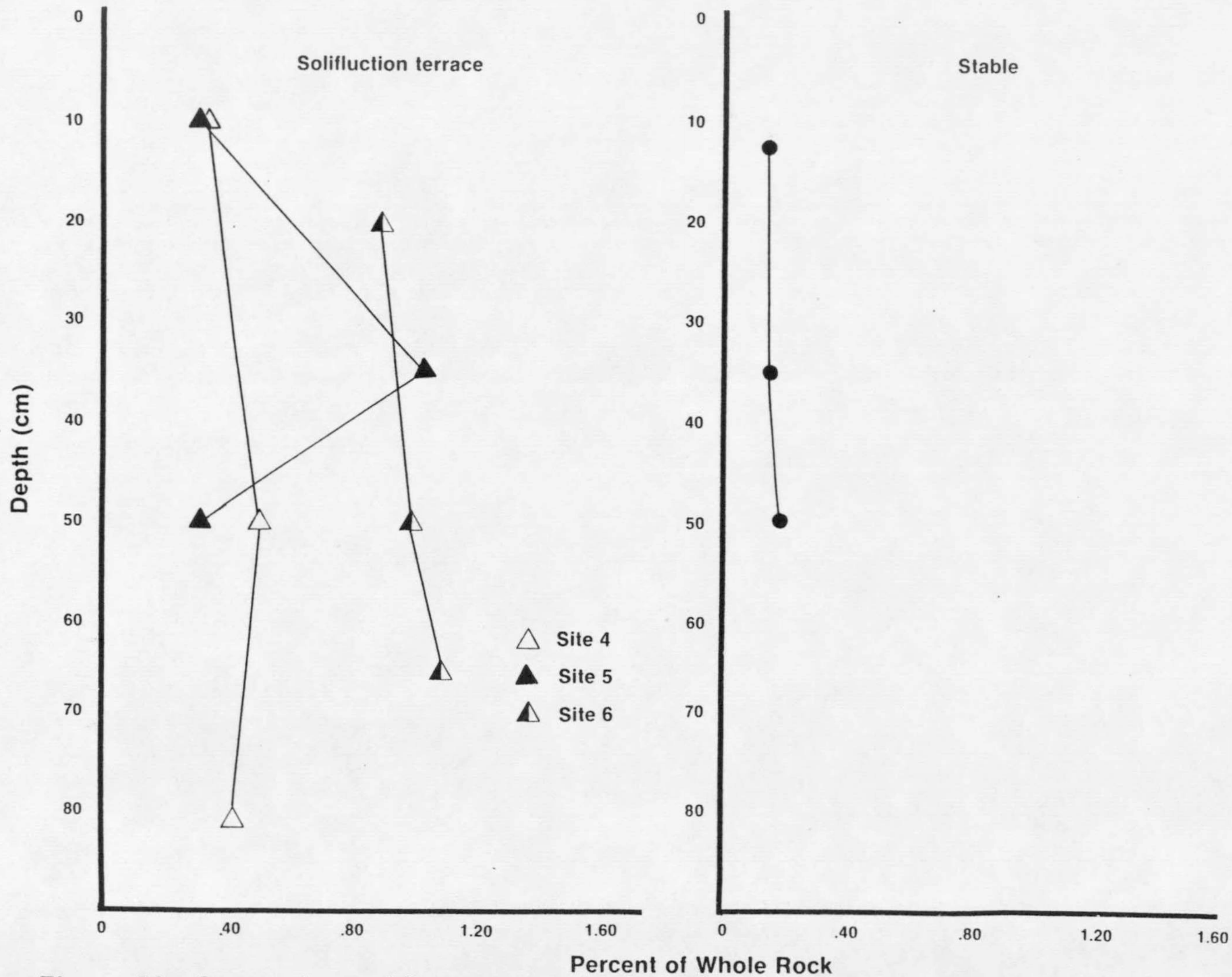


Figure 21. Comparison of the average absolute abundances of kaolinite in stable sites 1 and 3 and solifluction sites.

stable site 2, kaolinite is more prevalent, especially in the B horizon. This profile also contained several prominent clay skins in the B horizon. Clay skins are coatings of oriented clay particles along channels through which clay-bearing water percolates downward. Clay skins are indicative of illuviation of clay minerals in a stable environment over a prolonged period. Stable site 2 is the most developed of the stable sites with an argillic horizon and a soil thickness greater than 150 cm. The time necessary to develop this profile may also have been sufficient for kaolinite to have formed and be concentrated in the B horizon.

Although kaolinite is the least abundant clay mineral in most of the stable sites, it is the second most abundant clay mineral in the solifluction terrace. In soil profiles at the head and toe of the solifluction terrace, kaolinite gradually decreases from the C to A horizon. The soil site in the center of the solifluction terrace exhibits the greatest change with a threefold increase from the C to the B horizon and a threefold decrease from the B to the A horizon. This pattern is similar to that exhibited by smectite in this soil profile.

In summary there is an increase in the kaolinite and smectite content in the B horizon on both the solufluction terrace and stable sites. The magnitude of the increase of these clay minerals is much greater on the solifluction terrace. In a granitic terrane, illuviation is an important mechanism in the concentration of clay minerals in a B horizon, as weathering of clay minerals in place is usually not sufficient to form an argillic horizon. A long period of illuviation in a stable environment is shown in the stable sites by the presence of clay skins. It is significant to note that in the solifluction terrace, clay skins are absent.

Mass movement in the solifluction terrace is the suggested mechanism responsible for both the magnitude of the increase of smectite and kaolinite and the absence of clay skins. Motion of the soil, supplies fresh minerals to the A horizon from which kaolinite and smectite are weathered. Weathering and leaching of the secondary minerals would result in a high concentration in the B horizon. Soil motion is also responsible for the destruction of clay skins.

Summary

Evidence of mass movement occurring on the solifluction terraces is found in the soil mineralogy. On stable sites, the coarse fraction shows moderately intense weathering at the surface which becomes gradually less severe with depth. The change in clay mineral content and particle size distribution in the soil profile is also gradual with depth.

On the solifluction sites however, fresh and weathered coarse-size materials are found together throughout the profile. Discrete units of clay and highly altered material are found at the base of the soil profile adjacent to the relatively fresh parent rock. Fragments of unweathered parent rock are found mixed in with the weathered upper horizons of the solifluction terrace soil profiles. In addition, the changes in clay mineral content within the profile are abrupt and striking, and in sharp contrast to the gradual changes seen in the stable sites. The particle size distribution also shows wide variation between soil horizons. These differences are attributable to frost heaving and downslope motion and are illustrative of the mass movement origin of the solifluction terraces.

The most important criteria for establishing the mass movement origin of the terraces are: 1) the striking changes in the distribution of illite, smectite and kaolinite within the soil profile; 2) the mixing of weathered and unweathered minerals throughout the soil profile; and 3) the lack of clay skins in the argillic horizons on the solifluction terraces.

CHAPTER 5

CONCLUSIONS

Solifluction terraces on the uplands of the west central Boulder batholith are widespread and have a distinctive morphology. They occur in gulleys and vary greatly in size, ranging from 15 to 60 meters in length. They are elongate downslope with steep, arcuate, convex-downslope fronts covered with large boulders. Behind the steep boulder fronts, the terraces are relatively flat and contain a mixture of fine and coarse-grained material.

Terrace formation is likely the result of mass movement processes. The alignment of boulders in the downslope direction on the terrace tread and the sorting and separation of boulders from the rest of the deposit can be attributed to extensive frost heaving over a prolonged period. In addition, the presence of Early to Late Wisconsinan glacial deposits in close proximity and at similar elevations to the study area terraces, coupled with the lack of glacial features within the study area, provides evidence that a periglacial environment existed in the study area. It is therefore suggested that the specific mass movement process responsible for terrace formation is solifluction in a periglacial environment.

A study of the textural and mineralogical changes in the soils within the study area was made in order to substantiate the solifluction hypothesis. An analysis of the degree of weathering and the distribution of the minerals within the soils found striking differences between the soils of the solifluction terraces and those found on the stable sites. Clay mineralogy analysis demonstrates sharp and erratic changes in the distribution of minerals within the solifluction terrace profile, which contrast sharply with the gradual changes in clay mineral distribution exhibited by the stable profiles. In addition, coarse-size minerals in the stable sites show a gradual decrease in weathering with depth whereas the solifluction terrace soils contain a mixture of fresh and highly weathered minerals throughout the profile. These differences are likely the result of frost heaving and downslope motion in the solifluction terrace soils and serve to support the concept that the origins of the terraces are attributable to the mass movement process of solifluction.

In addition to demonstrating that the terraces in the study area originated through solifluction in a periglacial environment, a method of differentiating landforms that can be applied to other areas is presented. A comparison of soil mineralogy, weathering intensities, and soil structures in soils of landforms under study to soils formed in stable areas may show differences useful in determining their origin. The study demonstrates the utility of integrating soil mineral analysis with geomorphology and that an analysis of mineral texture and distribution within a soil can provide valuable information to distinguish stable land forms from those formed by mass movement processes.

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APPENDICES

APPENDIX A

SOIL PROFILE DESCRIPTIONS FOR PEDONS FORMED ON STABLE SITES AND UNSTABLE SITES

By Paul McDaniel and Janette Black

Site 1

CLASSIFICATION: Mollic Cryoboralf
LOCATION: stable site, NE 1/4, SW 1/4, Sec 4, T5N, R5W
ELEVATION: 6480 ft (1975 m)
PHYSIOGRAPHIC POSITION: simple, convex, rolling upland
SLOPE AND ASPECT: 22%, south-southwest
VEGETATION: Douglas fir, lodgepole pine, pinegrass
STONINESS: <10% large granite upcrop nearby

PEDON DESCRIPTION

01,02 0-4 cm. Needles, twigs, and decayed organic material.

A1 0-3 cm. Sandy loam; black (10 YR 2/1) moist; strong, fine; granular structure; soft, very friable, nonsticky and nonplastic; clear, smooth boundary, slightly acid (pH 6.5)

B2 3-53 cm. Gravelly, sandy loam; dark brown (10 YR 3/3) moist; moderate, fine to medium, subangular blocky structure; slightly hard, very friable, nonsticky and nonplastic; gradual wavy boundary; slightly acid (pH 6.5)

C 53+ cm. Gravelly sandy loam; dark yellowish brown (10 YR 4/6) moist; single grain to weak, medium, subangular blocky structure; loose, very friable, nonsticky and nonplastic; boundary not reached; neutral (pH 6.5-7.0)

REMARKS: Located on ridge west of solifluction terraces of this report. No evidence of transport. Amount of weathered material decreased with depth. Biotite increases with depth in profile.

* Dry colors in weathered granite are very subjective because of the salt and peppery appearance due to the coarse grain size and differing colors of the constituent minerals. Dry colors are given when available.

Site 2

CLASSIFICATION: Mollic Cryoboralf
LOCATION: stable site, SE 1/4, NW 1/4, Sec 4, T5N, R5W
ELEVATION: 6560 ft (1999 m)
PHYSIOGRAPHIC POSITION: simple, convex, rolling upland
SLOPE AND ASPECT: 9%, south-southeast
VEGETATION: Douglas fir, lodgepole pine, pinegrass
STONINESS: <10%

PEDON DESCRIPTION

01,02 — 0-5 cm. Needles, twigs, and decayed organic material.

A1 — 0-4 cm. Dark, gray brown (10 YR 4/2) sandy loam; black (10 YR 2/1) moist; strong, medium, granular structure; loose, soft, very friable, nonsticky and nonplastic; abrupt, smooth boundary; very stongly acid (pH 4.5).

A2 — 4-22 cm. Brown (10 YR 5/3), sandy clay loam; dark brown (10 YR 3/3) moist; strong, coarse, subangular blocky structure; slightly hard, very friable, slightly sticky and slightly plastic; gradual, wavy boundary; stongly acid (pH 5.0-5.5).

B2t — 22-85 cm. Yellowish brown (10 YR 5/6) gravelly, sandy, clay loam; yellowish brown (10 YR 5/4) moist; stong, very coarse, angular blocky structure; very hard, friable, sticky and plastic; diffuse, broken, variable boundary; medium acid (pH 5.5-6.0).

C — 50-85+ cm. Yellowish brown (10 YR 5/6) gravelly, sandy clay loam; yellowish brown (10 YR 5/4) moist; weak, medium, subangular blocky structure; loose, friable, slightly sticky and slightly plastic; boundary not reached; slightly acid (pH 6.5).

REMARKS: Located on ridge west of solifluction terraces of this report. No evidence of transport. Prominent, non continuous clay skins in the argillic horizon. Banding in B2t and C horizons runs horizontally for width of profile and indicates a stable, residual profile.

Site 3

CLASSIFICATION: Mollic Cryoboralf
LOCATION: stable site, NW 1/4, SW 1/4, Sec 4, T5N, R5W
ELEVATION: 6525 ft (1988 m)
PHYSIOGRAPHIC POSITION: simple, convex, rolling upland
SLOPE AND ASPECT: 7%, south-southeast
VEGETATION: Lodgepole pine, pinegrass
STONINESS: <10%

PEDON DESCRIPTION

O1,O2 — 0-3 cm. Needles, twigs, and decayed organic material.

A1 — 0-9 cm. Brown (10 YR 5/3) sandy loam; dark yellowish brown (10 YR 3/3) moist; strong, fine, granular structure; soft, very friable, nonsticky and nonplastic; clear, smooth boundary; very stongly acid (pH 4.5-5.0).

B2 — 9-40 cm. Yellowish brown (10 YR 5/4) gravelly, sandy loam; yellowish brown (10 YR 3/4) moist; strong, medium to coarse, subangular blocky structure; soft, very friable, nonsticky and nonplastic; gradual wavy boundary; strongly acid (pH 5.5).

C1 — 40-50 cm. Gravelly, sandy loam; dark, yellowish brown (10 YR 3/6) moist; strong, medium, subangular blocky structure; loose to slightly hard, friable, nonsticky and nonplastic; gradual wavy boundary; slightly acid (pH 6.0-6.5).

C2 — 50+ cm. Gravelly, sandy loam; dark yellowish brown (10 YR 3/4) moist; weak, medium, subangular blocky structure; loose, friable, nonsticky and nonplastic; boundary not reached; sightly acid (pH 6.0-6.5).

REMARKS: Site located on ridge west of other stable sites of this report. No evidence of transport. Few clay skins in the B2 horizon. Banding in B2 horison occuring at 35 cm and running width of profile indicates a stable, residual profile.

Site 4

CLASSIFICATION: Mollic Cryoboralf
LOCATION: solifluction site, SW 1/4, NW 1/4, Sec 4, T5N, R5W
ELEVATION: 6535 ft (1992 m)
PHYSIOGRAPHIC POSITION: simple, concave, rolling upland
SLOPE AND ASPECT: 5%, south
VEGETATION: Douglas fir, pinegrass, kinnikinnick
STONINESS: <10%, large boulder outcrop nearby

PEDON DESCRIPTION

01,02 0-3 cm. Needles, twigs, and decayed organic material.

A2 0-10 cm. Brown (10 YR 5/3) sandy loam; dark brown (10 YR 3/3) moist; strong, fine, granular structure; loose, soft, very friable, nonsticky and nonplastic; clear, smooth boundary; very strongly acid (pH 4.5).

B21 10-35 cm. Brown (10 YR 5/3) sandy loam; dark brown (10 YR 3/3), moist; strong, fine, granular structure; soft, very friable, nonsticky and nonplastic; clear, smooth boundary; strongly acid (pH 5.5).

B22t 35-52 cm. Brown (10 YR 4/4) sandy clay loam; dark brown (10 YR 3/3) moist; strong, medium, prismatic structure; slightly hard, friable, slightly sticky and plastic; clear wavy boundary; slightly acid (pH 6.5).

C 50+ cm. Gravelly sandy clay loam; dark yellowish brown (10 YR 3/6) moist; weak, medium, subangular blocky structure; loose, friable, slightly sticky and slightly plastic; boundary not reached; neutral (pH 6.5-7.0).

REMARKS: Site located on head of solifluction terrace. Evidence of movement; weathered and unweathered material found throughout profile, coarse fragments found throughout profile, clay pockets found in C horizon. Water table at 110 cm. Well expressed, continuous argillic horizon.

Site 5

CLASSIFICATION: Mollic Cryoboralf
LOCATION: solifluction site, SE 1/4, NW 1/4, Sec 4, T5N, R5W
ELEVATION: 6525 ft (1989 m)
PHYSIOGRAPHIC POSITION: simple, convex, rolling upland
SLOPE AND ASPECT: 10%, south-southwest
VEGETATION: Aspen, bunchgrass, kinnikinnick
STONINESS: 20%

PEDON DESCRIPTION

01,02 0-3 cm. Needles, twigs, and decayed organic material.

A1 0-10 cm. Very dark grayish brown (10 YR 3/2) sandy loam-sandy clay loam; very dark brown (10 YR 2/2) moist; moderate, medium, granular structure; soft, very friable, nonsticky and nonplastic; clear, smooth boundary; strongly acid (pH 5.5).

A2 10-18 cm. Dark gray brown (10 YR 4/2) sandy loam-sandy clay loam; very dark grayish brown (10 YR 3/2) moist; moderate, medium, granular structure; soft, very friable, nonsticky and nonplastic; gradual irregular boundary; strongly acid (pH 5.5).

B2t 18-58 cm. Gravelly sandy clay loam; dark brown (10 YR 4/3) moist; weak, medium angular blocky structure, slightly hard-hard friable, sticky and plastic; diffuse, broken boundary; slightly acid (pH 6.5).

C 50-90+ cm. Gravelly sandy loam; dark brown (10 YR 4/3) moist; single grain structures; loose, nonsticky and nonplastic; boundary not reached; neutral (pH 6.5-7.0).

REMARKS: Site located in center of solifluction terrace. Evidence of movement; weathered and unweathered material found throughout the profile, coarse fragments found throughout profile. Water table at 80 cm. Well expressed argillic horizon.

Site 6

CLASSIFICATION: Mollic Cryoboralf
LOCATION: solifluction site, SE 1/4, NW 1/4, Sec 4, T5N, R5W
ELEVATION: 6520 ft (1987 m)
PHYSIOGRAPHIC POSITION: simple, convex, rolling upland
SLOPE AND ASPECT: 10%, southeast
VEGETATION: Douglas fir, aspen, pinegrass
STONINESS: 50%

PEDON DESCRIPTION

01,02 0-3 cm. Needles, twigs, and decayed organic material.

A1 0-15 cm. Loam; black (10 YR 2/1) moist; moderate, fine, granular structure; soft, very friable, slightly sticky and slightly plastic; abrupt, smooth boundary; slightly acid (pH 6.5).

B2t 15-70 cm. Gravelly, sandy clay loam; dark brown (10 YR 3/3) moist; strong, medium prismatic structure; hard, friable, sticky and plastic; gradual, smooth boundary; slightly acid (pH 6.5).

C1 70-90+ cm. Gravelly sandy clay loam; dark yellowish brown (10 YR 3/6) moist; single grain to weak, medium subangular blocky structure; loose to hard, very friable, sticky and plastic; boundary not reached; neutral (pH 7.0).

REMARKS: Located at toe of solifluction terrace. Evidence of movement; four large coarse fragments located in top of the profile corresponding to the boundary between A1 and B2t horizons, weathered and unweathered material found throughout the profile. Water table at 90 cm. Not strongly expressed, but continuous and deep argillic horizon, thin clay skins.

APPENDIX B

TABLE OF PARTICLE SIZE DISTRIBUTION ON STABLE AND SOLIFLUCTION SITES

Table 11. Particle size data for stable site 1.

Size (microns)	A1/A2 0-10cm %	B2 30-40cm %	C 45-55cm %
2000-1000 (very coarse sand)	12.37	16.16	16.10
1000- 500 (coarse sand)	22.54	23.73	21.33
500- 250 (medium sand)	10.45	10.23	9.75
250- 100 (fine sand)	13.27	12.93	15.94
100- 50 (very fine sand)	9.42	8.71	9.11
50- 20 (coarse silt)	8.65	7.17	6.25
20- 5 (medium silt)	14.56	11.22	9.78
5- 2 (fine silt)	2.54	2.25	1.64
2- 0.2 (coarse clay)	2.96	3.09	3.25
0.2-0.08 (medium clay)	3.44	4.41	6.14
<0.08 (fine clay)	Nd	Nd	Nd
Greater than 2000	42.4	39.1	51.8
2000- 50 (total sand)	67.60	71.76	72.23
50- 2 (total silt)	25.74	20.64	17.67
<2 (total clay)	6.67	7.60	10.10
Textural class	gr.cosl	gr.cosl	Vgr.cosl

Nd = No data

Table 12. Particle size data for stable site 2.

Size (microns)	A2 15-25cm %	B2t 30-40cm %	C 75-85cm %
2000-1000 (very coarse sand)	13.02	9.45	9.05
1000- 500 (coarse sand)	21.69	17.60	18.74
500- 250 (medium sand)	9.43	7.98	8.50
250- 100 (fine sand)	11.21	9.94	11.87
100- 50 (very fine sand)	7.54	6.90	8.34
50- 20 (coarse silt)	8.76	6.99	7.35
20- 5 (medium silt)	16.48	16.52	15.04
5- 2 (fine silt)	3.28	4.64	4.11
2- 0.2 (coarse clay)	4.34	7.11	5.49
0.2-0.08 (medium clay)	4.77	9.83	10.42
<0.08 (fine clay)	Nd	Nd	Nd
Greater than 2000	33.3	36.4	46.1
2000- 50 (total sand)	62.90	51.86	56.50
50- 2 (total silt)	28.52	28.14	26.50
<2 (total clay)	8.58	20.00	17.00
Textural class	gr.cosl	gr.loam	gr.cosl

Nd = No data

Table 13. Particle size data for stable site 3.

Size (microns)	A1 0-10cm %	A2 15-25cm %	B2 30-40cm %	C1/C2 45-55cm %
2000-1000 (very coarse sand)	13.65	15.63	13.50	15.06
1000- 500 (coarse sand)	19.73	22.87	20.68	21.63
500- 250 (medium sand)	9.78	10.24	10.35	10.95
250- 100 (fine sand)	14.08	12.98	15.98	15.90
100- 50 (very fine sand)	8.56	8.97	10.02	8.67
50- 20 (coarse silt)	7.62	7.87	7.43	6.92
20- 5 (medium silt)	12.29	11.41	9.63	8.29
5- 2 (fine silt)	2.43	3.06	3.43	3.18
2- 0.2 (coarse clay)	2.04	2.41	2.61	2.43
0.2-0.08 (medium clay)	3.83	4.88	7.21	7.35
<0.08 (fine clay)	Nd	Nd	Nd	Nd
Greater than 2000	34.1	39.6	41.8	39.5
2000- 50 (total sand)	72.80	70.69	70.52	72.22
50- 2 (total silt)	22.33	22.35	20.49	18.38
<2 (total clay)	4.87	6.96	8.99	9.41
Textural class	gr.cosl	gr.cosl	gr.cosl	gr.cosl

Nd = No data

Table 14. Particle size data for solifluction site 4.

Size (microns)	A2 0-10cm %	B21 15-25cm %	B22t 45-55cm %	C 75-85cm %
2000-1000 (very coarse sand)	16.05	11.55	26.48	13.32
1000- 500 (coarse sand)	18.86	17.37	21.92	20.41
500- 250 (medium sand)	7.69	8.40	9.21	14.07
250- 100 (fine sand)	9.97	12.51	12.25	20.05
100- 50 (very fine sand)	7.65	8.70	6.55	8.35
50- 20 (coarse silt)	9.65	9.70	5.11	4.89
20- 5 (medium silt)	19.12	19.89	9.49	8.48
5- 2 (fine silt)	4.39	4.86	2.94	2.40
2- 0.2 (coarse clay)	3.27	4.02	2.41	2.10
0.2-0.08 (medium clay)	4.46	4.65	4.89	7.09
<0.08 (fine clay)	Nd	Nd	Nd	Nd
Greater than 2000	34.6	39.7	43.6	62.0
2000- 50 (total sand)	60.21	58.53	76.41	76.20
50- 2 (total silt)	33.15	34.44	17.54	15.77
<2 (total clay)	6.64	7.03	6.06	8.03
Textural class	gr.cosl	gr.cosl	gr.cosl	V.gr.cosl

Nd = No data

Table 15. Particle size data for solifluction site 5.

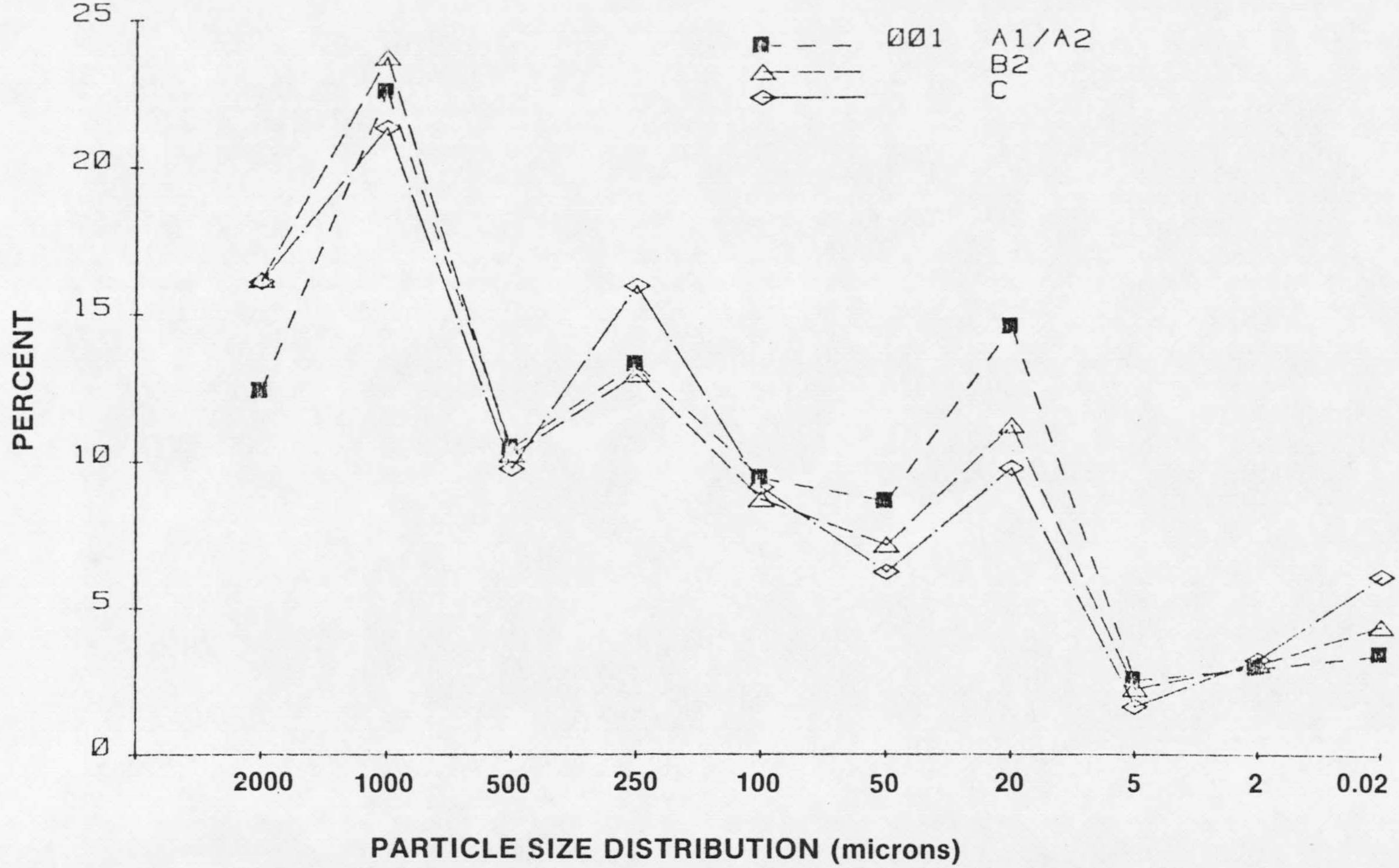
Size (microns)	A1 0–10cm %	B2t 30–40cm %	C 45–55cm %
2000–1000 (very coarse sand)	10.81	9.10	15.00
1000– 500 (coarse sand)	18.31	17.44	21.17
500– 250 (medium sand)	7.74	9.09	10.73
250– 100 (fine sand)	9.50	13.53	15.83
100– 50 (very fine sand)	7.30	9.16	9.08
50– 20 (coarse silt)	10.07	7.19	6.36
20– 5 (medium silt)	21.77	10.82	8.25
5– 2 (fine silt)	3.69	2.52	2.44
2– 0.2 (coarse clay)	4.69	5.93	2.18
0.2–0.08 (medium clay)	4.30	13.10	8.78
<0.08 (fine clay)	Nd	Nd	Nd
Greater than 2000	40.1	45.5	52.7
2000– 50 (total sand)	53.65	58.33	71.80
50– 2 (total silt)	35.53	20.53	17.04
<2 (total clay)	10.82	21.15	11.16
Textural class	gr.cosl	gr.scl	Vgr.cosl

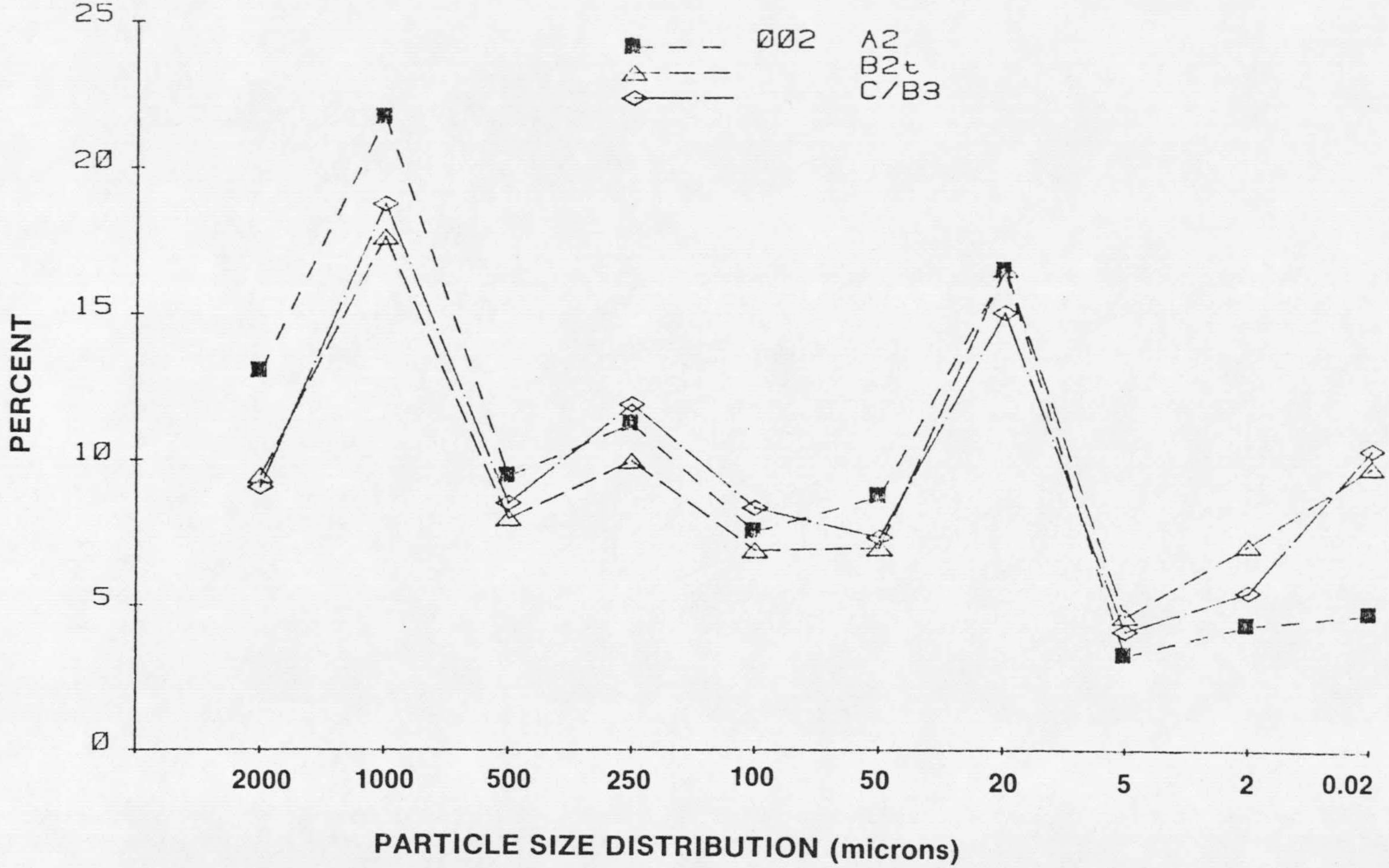
Nd = No data

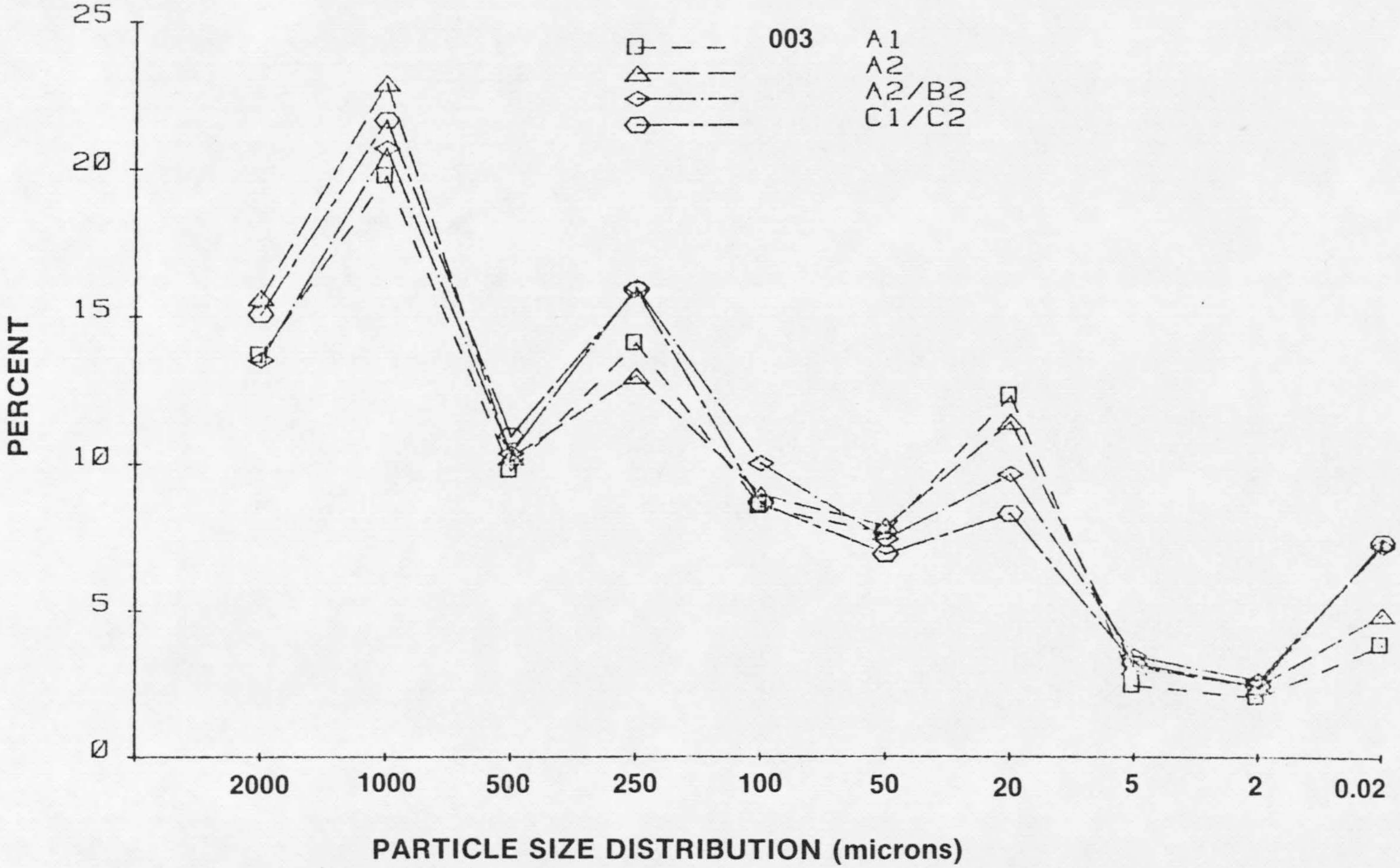
Table 16. Particle size data for solifluction site 6.

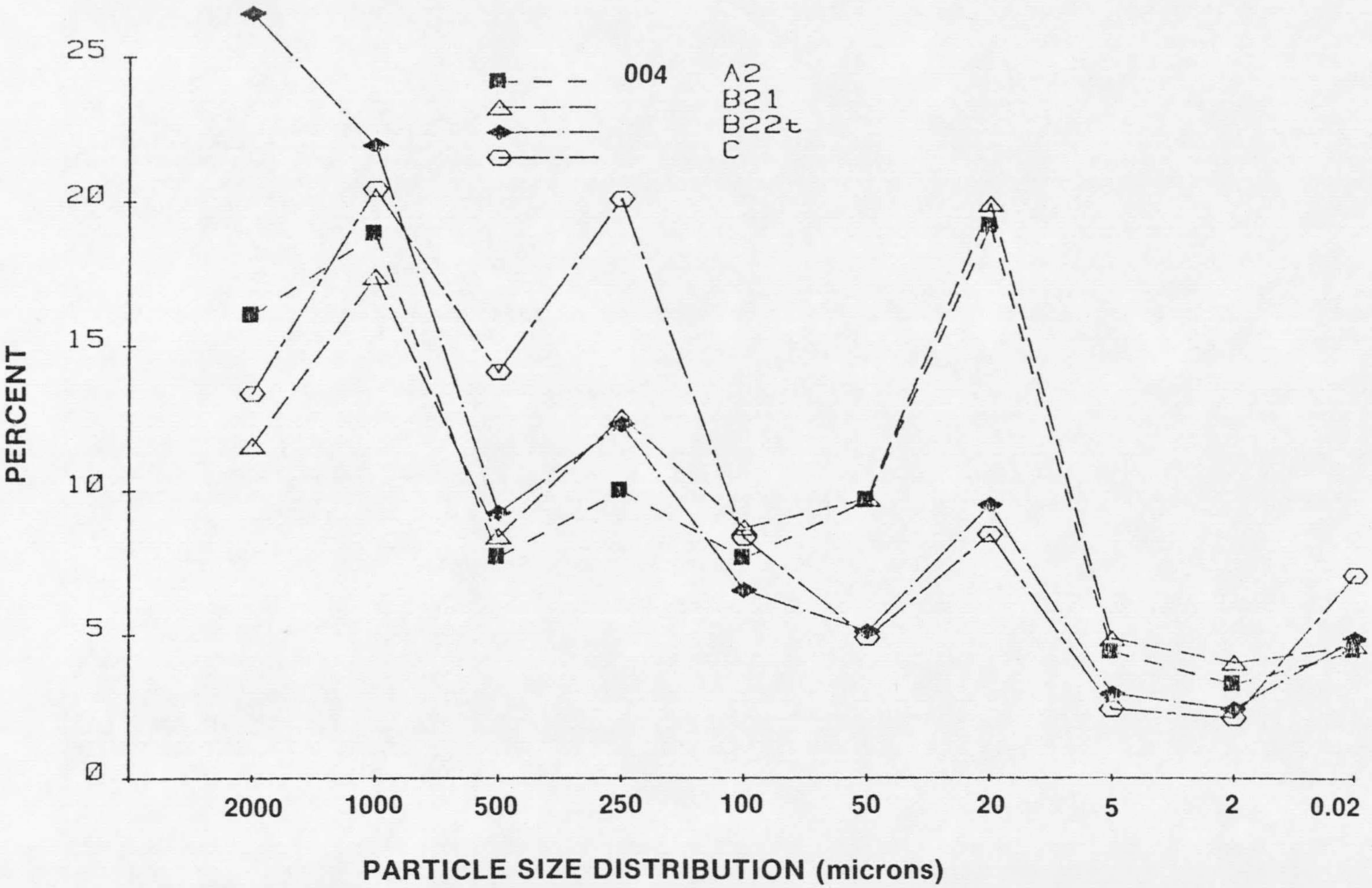
Size (microns)	A1 15-25cm %	B22t 45-55cm %	C 60-70cm %
2000-1000 (very coarse sand)	8.75	9.45	7.29
1000- 500 (coarse sand)	15.25	16.06	14.44
500- 250 (medium sand)	8.00	10.37	8.82
250- 100 (fine sand)	11.35	14.63	16.31
100- 50 (very fine sand)	8.50	9.21	10.94
50- 20 (coarse silt)	11.75	8.50	10.90
20- 5 (medium silt)	13.88	12.11	12.18
5- 2 (fine silt)	3.84	3.01	3.14
2- 0.2 (coarse clay)	5.98	3.86	4.14
0.2-0.08 (medium clay)	11.07	9.93	11.13
<0.08 (fine clay)	Nd	Nd	Nd
Greater than 2000	45.5	49.7	40.3
2000- 50 (total sand)	51.84	59.71	57.79
50- 2 (total silt)	29.47	23.62	26.22
<2 (total clay)	18.69	16.67	16.00
Textural class	gr.loam	gr.cosl	gr.sl

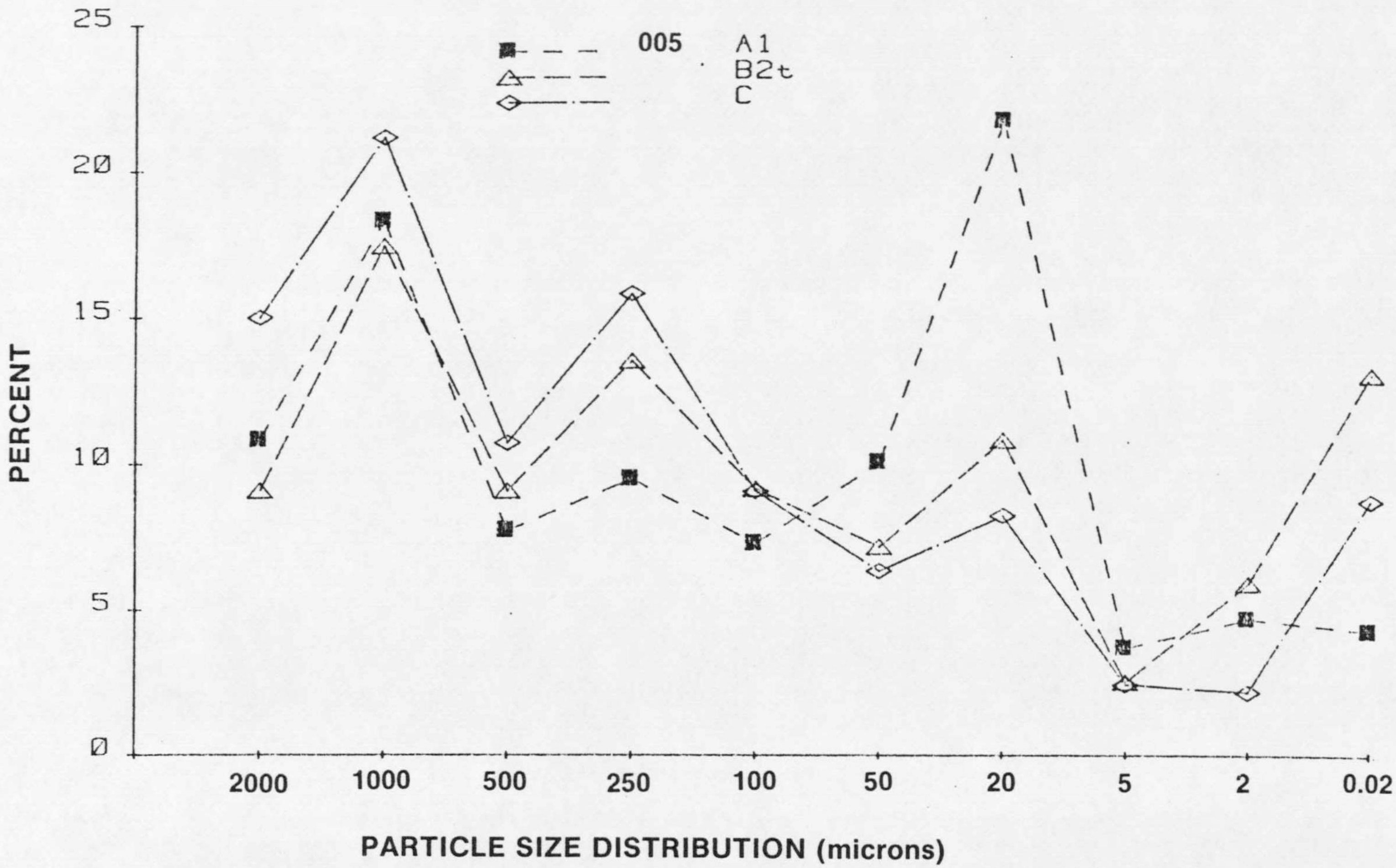
Nd = No data


PERCENT
25
20
15
10
5
Ø
ØØ1 A1/A2
B2
C
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)

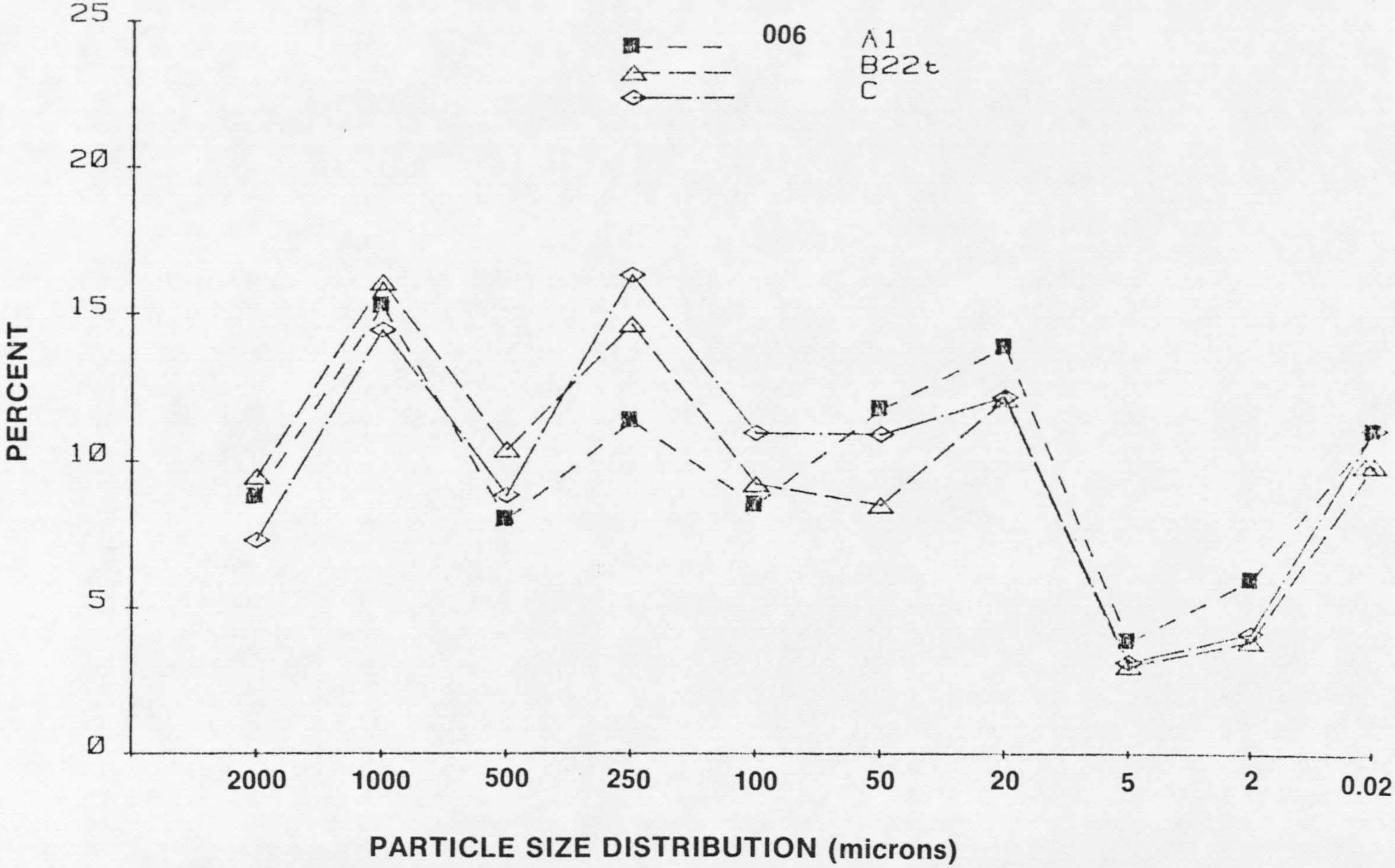

002 A2
B2t
C/B3
PERCENT
25
20
15
10
5
0
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)


003
A1
A2
A2/B2
C1/C2
PERCENT
25
20
15
10
5
Ø
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)


004
A2
B21
B22t
C
PERCENT
25
20
15
10
5
0
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)



005
A1
B2t
C
25
20
15
10
5
0
PERCENT
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)


006
A1
B22t
C
PERCENT
25
20
15
10
5
0
2000
1000
500
250
100
50
20
5
2
0.02
PARTICLE SIZE DISTRIBUTION (microns)

APPENDIX C

RELATIVE ABUNDANCE OF CLAY MINERALS WITH DEPTH ON STABLE AND SOLIFLUCTION TERRACE SITES

Table 17. Relative mineral abundance of clay minerals (2-.2 microns) on stable sites.

Site	Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Vermiculite %
1	A1/A2	0-10	6.0%	80.0%	4.0%	10.0%	-
	B2	30-40	7.0%	81.0%	4.0%	9.0%	trace
	C	45-55	9.0%	71.0%	5.0%	14.0%	1.0%
2	A2	15-25	9.0%	57.0%	19.0%	15.0%	-
	B2t	30-40	12.0%	41.0%	35.0%	12.0%	trace
	C	75-85	11.0%	45.0%	26.0%	16.0%	2.0%
3	A1	0-10	9.0%	70.0%	5.0%	13.0%	2.0%
	A2	15-25	10.0%	71.0%	4.0%	12.0%	2.0%
	B2	30-40	17.0%	66.0%	4.0%	13.0%	trace
	C1/C2	45-55	18.0%	64.0%	6.0%	13.0%	trace

Table 18. Relative mineral abundance of clay minerals (2-.2 microns) on solifluction terraces.

Site	Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Vermiculite %
4	A2	0-10	4.0%	76.0%	11.0%	10.0%	trace
	B22t	45-55	11.0%	54.0%	21.0%	14.0%	trace
	C	75-85	11.0%	57.0%	21.0%	11.0%	trace
5	A	0-10	7.0%	73.0%	7.0%	13.0%	trace
	B2t	30-40	29.0%	42.0%	18.0%	12.0%	-
	C	45-55	17.0%	57.0%	15.0%	12.0%	-
6	A1	15-25	4.0%	68.0%	15.0%	12.0%	trace
	B22t	45-55	10.0%	52.0%	26.0%	11.0%	trace
	C	60-70	9.0%	51.0%	27.0%	10.0%	2.0%

APPENDIX D

ABSOLUTE ABUNDANCE OF CLAY MINERALS BY HORIZON ON STABLE AND SOLIFLUCTION TERRACE SITES

Table 19. Absolute amounts of clay minerals (2-.2 microns) on stable sites.

Site	Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Total % (2-.2) clay
1	A1/A2	0-10	0.18%	2.35%	0.12%	0.30%	2.96%
	B2	30-40	0.20%	2.50%	0.13%	0.26%	3.09%
	C	45-55	0.28%	2.30%	0.18%	0.46%	3.25%
2	A2	15-25	0.40%	2.50%	0.82%	0.65%	4.34%
	B2t	30-40	0.87%	2.88%	2.50%	0.85%	7.11%
	C	75-85	0.59%	2.45%	1.44%	0.87%	5.49%
3	A1	0-10	0.19%	1.43%	0.10%	0.27%	2.04%
	A2	15-25	0.25%	1.82%	0.11%	0.31%	2.41%
	B2	30-40	0.44%	1.71%	0.12%	0.33%	2.61%
	C1C2	45-55	0.43%	1.56%	0.13%	0.30%	2.43%

Table 20. Absolute amounts of clay minerals (2–.2 microns) on solifluction terraces.

Site	Horizon	Depth cm	Smectite %	Illite %	Kaolinite %	Quartz %	Total % (2–.2) clay
4	A2	0–10	0.12%	2.47%	0.35%	0.33%	3.27%
	B22t	45–55	0.27%	1.30%	0.51%	0.33%	2.41%
	C	75–85	0.23%	1.20%	0.43%	0.23%	2.10%
5	A	0–10	0.32%	3.40%	0.33%	0.61%	4.69%
	B2t	30–40	1.73%	2.46%	1.04%	0.69%	5.93%
	C	45–55	0.36%	1.24%	0.32%	0.25%	2.18%
6	A1	15–25	0.25%	4.09%	0.90%	0.68%	5.98%
	B22t	45–55	0.40%	2.01%	1.02%	0.43%	3.86%
	C	60–70	0.37%	2.11%	1.10%	0.41%	4.14%