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A sequence of soils derived from tuffaceous rocks was sampled on the H. J. Andrews Experimental Forest and adjacent U. S. Plywood land to determine some genetic and stability relationships. The less than two micron size fraction from each sample was characterized by x-ray diffraction with respect to crystalline clay minerals. Electron micrographs were made of selected clay samples to determine the morphology of kaolin minerals. Standard chemical and physical analyses were performed. The coarse-silt mineralogy of each sample was characterized by a randomized count of 300 grains.

Data from these procedures showed that kaolin minerals studied were not well crystallized. Mixtures of kaolinite and halloysite resulted in broad rather diffuse, moderately intense diffraction maxima with some degree of asymmetry. These diffraction maxima were characterized by a morphological sequence of kaolin minerals from

poorly crystallized platy forms of kaolinite to platy forms with rolled edges and eventually to fairly well crystallized tubular forms of halloysite.

The major part of the sand and silt fractions consisted of pseudomorphs of clay that were chemically reactive in cation exchange reactions and moisture retention. The number of pseudomorphs of clay decreased and measured clay increased upward in the soil profile from the C or R horizons into the B horizons. Pseudomorphs of montmorillonite were more resistant to disruption than pseudomorphs of kaolin.

Chemical weathering resulted in alteration of plagioclase and ferromagnesian minerals to heulandite-clinoptilolite and ultimately to pseudomorphs of montmorillonite. Physical and biological disruptive forces reduced these pseudomorphs to discrete montmorillonite. Intergrade minerals formed in well drained weathered horizons as a result of hydroxy interlayering of smectites. Alternate wetting and drying coupled with more intense weathering resulted in the transformation of halloysite to kaolinite upward in the soil.

Slipout and Budworm soils had clay fractions that were mainly expanding type clays and buff tuff and McKenzie River soils had clay fractions that were mainly non-expanding clays. As a result, Slipout and Budworm soils were more prone to slope failure than buff tuff and McKenzie River soils.

Free iron oxide content of all four soils increased in the same order as relative slope stability. There was no apparent relationship between stability and exchangeable ions.

More of the total clay occurred as silt and sand-size pseudomorphs of clay in McKenzie River and buff tuff soils than in Slipout and Budworm soils. These pseudomorphs functioned as individual primary particles increasing the solid-to-solid contact in the soil framework causing McKenzie River and buff tuff soils to have additional resistance to shear failure.

# Genetic and Stability Relationships of Four Western Cascade Soils

by

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Finally, out of context but nevertheless appropriate, " thanks be unto GOD".

TABLE OF CONTENTS


INTRODUCTION 1

DESCRIPTION OF STUDY AREA 4

- Location 4
- Geology 4
- Soils 7
- Climate and Vegetation 8

EXPERIMENTAL METHODS 10

- Sampling Method 10
- Laboratory Methods 11
  - Chemical Analyses 11
  - Physical Analyses 14
  - Soil Fractionation 14
  - Selective Dissolution Treatment 15
  - X-ray Diffraction Analysis 16
  - Silt Mineralogy 18
  - Electron Micrographs 20

RESULTS 21

- Clay Mineralogy 21
  - X-ray Diffraction 21
  - Electron Micrographs 33
- Coarse-Silt Mineralogy 36
- Physical Properties 46
- Chemical Properties 50
- Some Physical and Chemical Relationships 54

DISCUSSION 59

- Genetic Relationships 59
  - Electron Micrographs 59
  - Coarse-Silt Mineralogy 65
  - Pedogenic Formation of Montmorillonite 74
  - Chloritization 78
  - Halloysite-Kaolinite Transformation 81
- Stability Relationships 82
  - Water 82
  - Clay Mineral Structures 91

TABLE OF CONTENTS (CONTINUED)


| | |
|---|---|
| Clay Type | 94 |
| Significance of Color | 99 |
| Free Iron Oxide | 100 |
| Exchangeable Ions | 103 |
| Pseudomorphs | 105 |
| CONCLUSIONS | 107 |
| Genetic Relationships | 107 |
| Stability Relationships | 109 |
| BIBLIOGRAPHY | 112 |
| APPENDIX | 123 |

## LIST OF FIGURES


| Figure | | Page |
|---|---|---|
| 1. | Map of sample sites on the H. J. Andrews Experimental Forest. | 12 |
| 2. | Map of sample sites on U. S. Plywood land. | 13 |
| 3. | X-ray diffraction patterns of clay samples from Slipout soil profile number 1. | 25 |
| 4. | X-ray diffraction patterns of clay samples from Budworm soil profile number 1. | 26 |
| 5. | X-ray diffraction patterns of clay samples from buff tuff soil profile number 1. | 27 |
| 6. | X-ray diffraction patterns of clay samples from McKenzie River soil profile number 1. | 28 |
| 7. | X-ray diffraction patterns of dispersible clay and clay mechanically dispersed from the silt fraction of selected horizons from McKenzie River soil profile number 3. | 31 |
| 8. | X-ray diffraction patterns of dispersible clay and clay mechanically dispersed from the silt fraction of selected horizons from buff tuff soil profile number 2. | 32 |
| 9. | X-ray diffraction patterns of clay mechanically dispersed from the silt fraction of horizons from McKenzie River soil profile number 1. | 34 |
| 10. | Electron micrograph of clay from the A11 horizon of buff tuff soil profile number 2. | 35 |
| 11. | Electron micrograph of clay from the B21 horizon of buff tuff soil profile number 2. | 35 |
| 12. | Electron micrographs of clay from the R horizon of buff tuff soil profile number 2. | 37 |
| 13. | Electron micrograph of clay from the A3 horizon of McKenzie River soil profile number 1. | 38 |

LIST OF FIGURES (CONTINUED)


| Figure | | Page |
|---|---|---|
| 14. | Electron micrograph of clay from the $B21_t$ horizon of McKenzie River soil profile number 1. | 38 |
| 15. | Electron micrographs of clay from the R horizon of McKenzie River soil profile number 1. | 39 |
| 16. | Picture of the coarse silt fraction from the B3 horizon of McKenzie River soil profile number 1. (plain light) | 44 |
| 17. | Picture of the coarse silt fraction from the B3 horizon of McKenzie River soil profile number 1. (x-nicols) | 44 |
| 18. | Picture of the coarse silt fraction from the A11 horizon of Budworm soil profile number 2. | 45 |
| 19. | Picture of the coarse silt fraction from the A11 horizon of Budworm soil profile number 2. | 45 |
| 20. | Plots of cation exchange capacity, 15-bar moisture, and dispersed clay as a function of depth in four selected soil profiles. | 56 |

# LIST OF TABLES


| Table | | Page |
|---|---|---|
| 1. | Clay mineral distribution of four Western Cascade soils. | 22 |
| 2. | Mineral composition of the coarse silt fraction of four Western Cascade soils. | 40 |
| 3. | Mechanical analysis of four Western Cascade soils and 15-atmosphere moisture of four selected profiles. | 47 |
| 4. | Selected chemical properties of four Western Cascade soils. | 51 |
| 5. | Elemental analysis of citrate-dithionite and boiling sodium hydroxide extracts of silt fractions of four selected soil profiles. | 55 |
| 6. | Relationship between dispersible clay, total clay, and cation exchange capacity. | 58 |

# GENETIC AND STABILITY RELATIONSHIPS OF FOUR WESTERN CASCADE SOILS

## INTRODUCTION

During the December 1964 storm, an exceedingly large amount of rain fell on much of the Douglas-fir region of the Western Cascades of Oregon. This resulted in record-breaking streamflow from many of the affected watersheds. This storm caused flood damage of major proportions to the forested region, especially in mountainous headwaters where channel scouring and mass soil movements were common. The H. J. Andrews Experimental Forest and adjacent timber lands are typical of much of the forested region that was affected by this storm. Fredriksen (1965) has described the characteristics of this storm and much of the resulting damage to the H. J. Andrews Experimental Forest. He believes that storms of this nature are not uncommon to this area.

Dyrness (1967) made a detailed study of mass soil movements occurring in the H. J. Andrews Experimental Forest as a result of this storm. He grouped these mass movements into ten classes on the basis of their morphological characteristics. At the site of each mass movement, he recorded type and specific characteristics of each movement, general characteristics of the area, and assessment of factors influencing each mass movement. Of these mass movements, 10.6 percent occurred in undisturbed areas comprising 84.6

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percent of the forest. Similarly, 89.4 percent of these events occurred in connection with roads or logging operations which covered 15.4 percent of the forest. Steepness of slope and frequency of slides were apparently positively correlated with only 17.0 percent of the events occurring on slopes below 45 percent. Mass movements were most common on northwest and west exposures and were almost nonexistent on south and southwest exposures. All but two mass movements occurred at elevations below 2,900 feet with the bulk of the events occurring in the elevation range between 2,000 and 2,900 feet.

Pope and Anderson (1960) have shown that soils derived from pyroclastic rocks in the Western Cascades are relatively unstable and prone to mass soil movement. Likewise, Dyrness (1967) points out that mass soil movements occur much more frequently in areas of tuff and breccia than in areas where the bedrock is basalt or andesite. Areas with tuff and/or breccia substratum were the site of 93.6 percent of these events, even though only 37.2 percent of the area consisted of these rocks. Moreover, 63.8 percent of the mass soil movements occurred on greenish tuffs and/or breccias which make up only 8.0 percent of the area, and 29.8 percent of the mass soil movements occurred on reddish tuffs and/or breccias which make up 29.2 percent of the area.

The purpose of this study was to characterize physical, chemical, and mineralogical properties of four Western Cascade soils in

order to obtain insight as to cause and effect relationships between kinds of soil and underlying bedrock and soil mass movements, to relate interhorizonal variation in clay mineral composition to pedogenic transformations, and to determine what clay minerals were responsible for broad, rather diffuse, moderately intense 7.5 Å diffraction maxima in clay fractions of McKenzie River soils.

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# DESCRIPTION OF THE STUDY AREA

## Location

The H. J. Andrews Experimental Forest (Berntsen and Rothacher, 1959), consisting of the entire 15,000 acre drainage of Lookout Creek, is located about 40 miles east of Springfield in the Willamette National Forest. The U. S. Plywood land, consisting of several hundred acres, is located in the same vicinity a few miles southwest of the Andrews in T 17 S, R 3 E, and R 4 E.

## Geology

Bedrock of the H. J. Andrews Experimental Forest and adjacent U. S. Plywood land is almost wholly volcanic in origin. According to Peck et al. (1964) the Volcanic Rocks of the High Cascade Range make up most of the higher ridges and side slopes of the H. J. Andrews Experimental Forest. The western portion of the H. J. Andrews Experimental Forest consists of interbedded flows and pyroclastic rocks of the Sardine Formation and the Little Butte Volcanic Series. The U. S. Plywood lands are predominantly interbedded flows and pyroclastic rocks of the Sardine Formation with small areas of the Little Butte Volcanic Series occurring in the extreme western portion.

The Little Butte Volcanic Series consists of Oligocene and early Miocene age massive beds of andesite and dacite tuff, presumedly deposited as ash flows, and less abundant flows and breccia of olivine basalt and andesite. Tuffaceous rocks are predominantly massive pumice lapilli tuff and vitric tuff, but they include waterlaid tuff, volcanic conglomerate, and welded tuff.

The Sardine Formation consists of late Miocene flows of hypersthene andesite, basaltic andesite, olivine basalt, dacite, and rhyodacite with less abundant flow breccia, tuff-breccia, lapilli tuff, tuff, and conglomerate.

The Volcanic Rocks of the High Cascades are predominantly Pliocene age flows of olivine basalt and olivine andesite. Less abundant are flows of dense basalt, basaltic andesite, and pyroxene andesite. Pyroclastic rocks are rare.

Thornbury (1965) outlines the major events leading to the evolution of the Cascade Range. The first episode of volcanism occurred during the Eocene with formation of interbedded layers of volcanic ash and basalt flows. Oligocene time saw growth of a folded mountain range. Associated with formation of Oligocene mountains was eruptive activity during which extensive andesitic lava flows and pyroclastic materials were spread over the region. Next came the great outpouring of Miocene lavas known as the Columbia River basalts. This was followed by Pliocene uplift, warping, and folding.

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Volcanic activity accompanying this uplift resulted in the formation of the imposing chain of volcanic mountains of the Cascade Range. Pleistocene mountain glaciers formed on the volcanic peaks and extended down valleys for many miles. The ensuing glactiation and subsequent erosion sculptured the Cascade Range as it exists today.

The upper parts of most major stream valleys contain glacial cirques. Associated slopes and stream valleys contain deposits of glacial debris. Erosion was characterized by rapid down-cutting of streams, channel scouring, headwall erosion, and sidewall erosion. The Western Cascades became maturely dissected with a well developed dendritic drainage pattern resulting in the formation of long narrow stream valleys separated by long acute ridges (Peck, 1964). Valley bottoms have elevations that range between 500 and 2,000 feet and intervening ridges have elevations that range between 3,000 and 5,000 feet. Bedrock in much of the area is covered with a thick layer of soft weathered pyroclastic material. Many features of the landscape suggest that earthflows, slumps, and slides have been important in shaping the landscape. Shallow scars, hummocky benches, step-like terraces, and poorly drained depressions are common. Much of the study area has geomorphic characteristics similar to those of the Looney Unit (Balster and Parsons, 1968).

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## Soils

The Soil Survey Report of the H. J. Andrews Experimental Forest (Stephens, 1964) shows the nature and distribution of the soils. For purposes of this investigation, only soils derived from pyroclastic rocks were selected. McKenzie River, Budworm, and Slipout soils were sampled from the H. J. Andrews Experimental Forest and two unnamed soils were sampled from lands of the U. S. Plywood Corporation.

McKenzie River soil is a member of the clayey, mixed, mesic family of Typic Haplohumults. Typically these are very deep well drained soils with dark brown, shotty, medium textured A horizons and dark brown, moderately fine to fine textured B horizons having thin patchy clay films. The solum is medium acid in the A horizon grading to strongly acid in the B horizon. These soils are derived from reddish tuffs and breccias of the Little Butte Volcanic Series and the Sardine Formation.

Slipout soils are members of the clayey, montmorillonitic, mesic family of Typic Argiaquolls and Typic Haplumbrepts. Typically these are very deep somewhat poorly drained soils with very dark grayish brown, medium textured A horizons and olive brown, fine textured, mottled B horizons. Thin patchy clay films occur in B horizons of some profiles. The solum is medium acid in the A

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horizon grading to very strongly acid in the B horizon. Gravel is common throughout the profile. These soils are derived from greenish tuffs and breccias of the Sardine Formation.

Budworm soils are members of the clayey, montmorillonitic, mesic family of Ultic and Typic Hapludalfs. Typically these are very deep moderately well drained soils with very dark brown, gravelly, medium textured A horizons and olive brown, fine textured B horizons with thin patchy clay films. The solum is strongly acid. These soils are derived from greenish tuffs and breccias of the Sardine Formation.

The unnamed soils, derived from the buff colored tuff of the Little Butte Volcanic Series, are members of the fine loamy, mixed, mesic family of Typic Hapludalfs and Typic Haplohumults. Typically these are deep well drained soils with dark brown, shotty, medium textured A horizons and dark brown, moderately fine textured B horizons with thin patchy clay films. The solum is medium acid in the A horizon and strongly acid in the B horizon.

## Climate and Vegetation

Rothacher, Dyrness, and Fredriksen (1967) give a rather complete description of climate and vegetation of the H. J. Andrews Experimental Forest. Briefly they state that temperature and precipitation are typical of a maritime climate with wet, relatively mild

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winters and dry cool summers. Precipitation for the ten year period from 1952 to 1962 averaged 94.17 inches annually. The average January temperature is 35.0° F and the average July temperature is 69.2° F. The mean annual temperature is 49.1° F and the frost free growing season is 97 days at nearby McKenzie Bridge Weather Bureau Station. Humidity is generally high during winter months, and ranges from 30 to 50 percent through much of the summer. Potential evapotranspiration trends are approximately reverse to seasonal trends in precipitation resulting in a precipitation deficit from mid-May to late September.

Vegetation is similar to that occurring in much of the Douglas-fir region. The dominant tree species is Pseudotsuga Menziessi[1] (Mirb) Franco mixed with varying amounts of Tsuga heterophylla (Raf) Sarg. Other conifers are Thuja plicata D. Don. and Taxus brevifolia Nutt. Common hardwood species are Acer macrophyllum Pursh., Cornus Nuttallii Aud., Castanopsis chrysophylla (Dougl.) A. DC., Acer circinatum Pursh., and Alnus oregona Nutt. Important shrubs and herbs are Berberis nervosa Pursh., Rhododendron macrophyllum G. Don., Polystichum munitum (Kaulf.) Presl., and Xerophyllum tenax (Pursh.) Nutt.

[1] Nomenclature is after Peck (1961).

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# EXPERIMENTAL METHODS

## Sampling Method

Sampling sites were located on the H. J. Andrews Experimental Forest by using the Soil Survey Report (Stephens, 1964). One profile site from each of the McKenzie River, Budworm, and Slipout soils was located as close as possible to the type location for that soil. The remaining two sampling sites were located in representative areas of each soil. Dr. Chester T. Youngberg assisted in location of sampling sites for the unnamed soils on U. S. Plywood land. Samples from these four soils were given the following sample designation for purposes of identification in the discussion that followed in this manuscript:

McKenzie River samples

M1 soil profile number 1
M2 soil profile number 2
M3 soil profile number 3

Slipout samples

S1 soil profile number 1
S2 soil profile number 2
S3 soil profile number 3

Budworm samples

B1 soil profile number 1
B2 soil profile number 2
B3 soil profile number 3

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buff tuff samples

P1 soil profile number 1
P2 soil profile number 2

These soil series names have no status.

Figure 1 shows the location of sampling sites on the H. J. Andrews Experimental Forest and Figure 2 shows the location of sampling sites on U. S. Plywood land.

Pits were exposed at each of these sampling sites and profiles were described in standard terminology of the U. S. Department of Agriculture (1951). Profile samples were collected, by horizon, from each pit.

## Laboratory Methods

### Chemical Analyses

Standard chemical analyses were performed by the Pacific Northwest Forest and Range Experiment Station, Forestry Sciences Laboratory, Corvallis, Oregon. Cation exchange capacity was determined by saturating the exchange complex with magnesium acetate, displacing with sodium acetate and determination of magnesium in the extract on a Beckman Model B flame photometer. Similarly, exchangeable calcium, potassium, magnesium, and sodium were determined on an ammonium acetate extract. Free iron oxide was determined according to Olson (1965). Organic matter was

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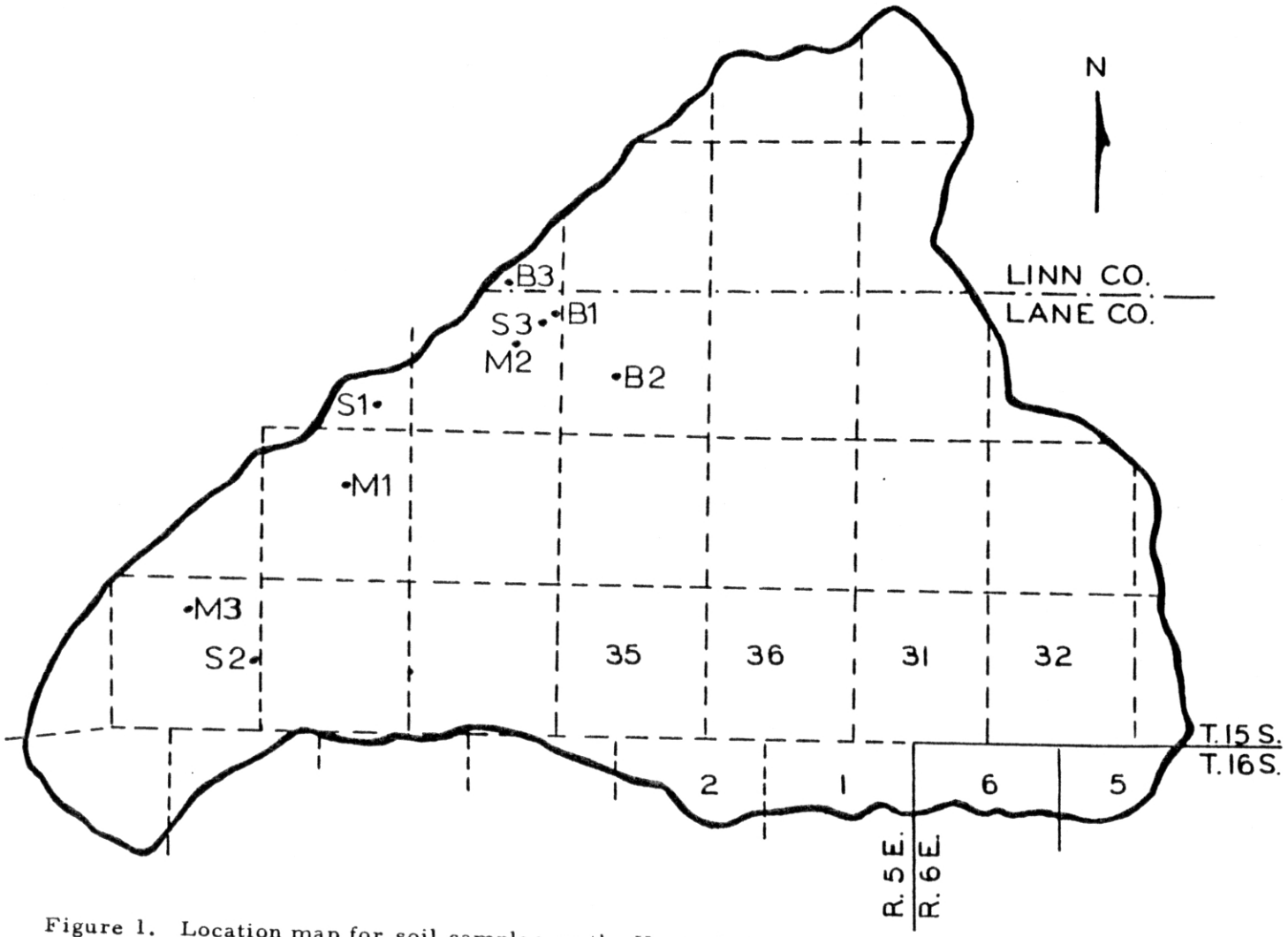


Figure 1. Location map for soil samples on the H. J. Andrews Experimental Forest.



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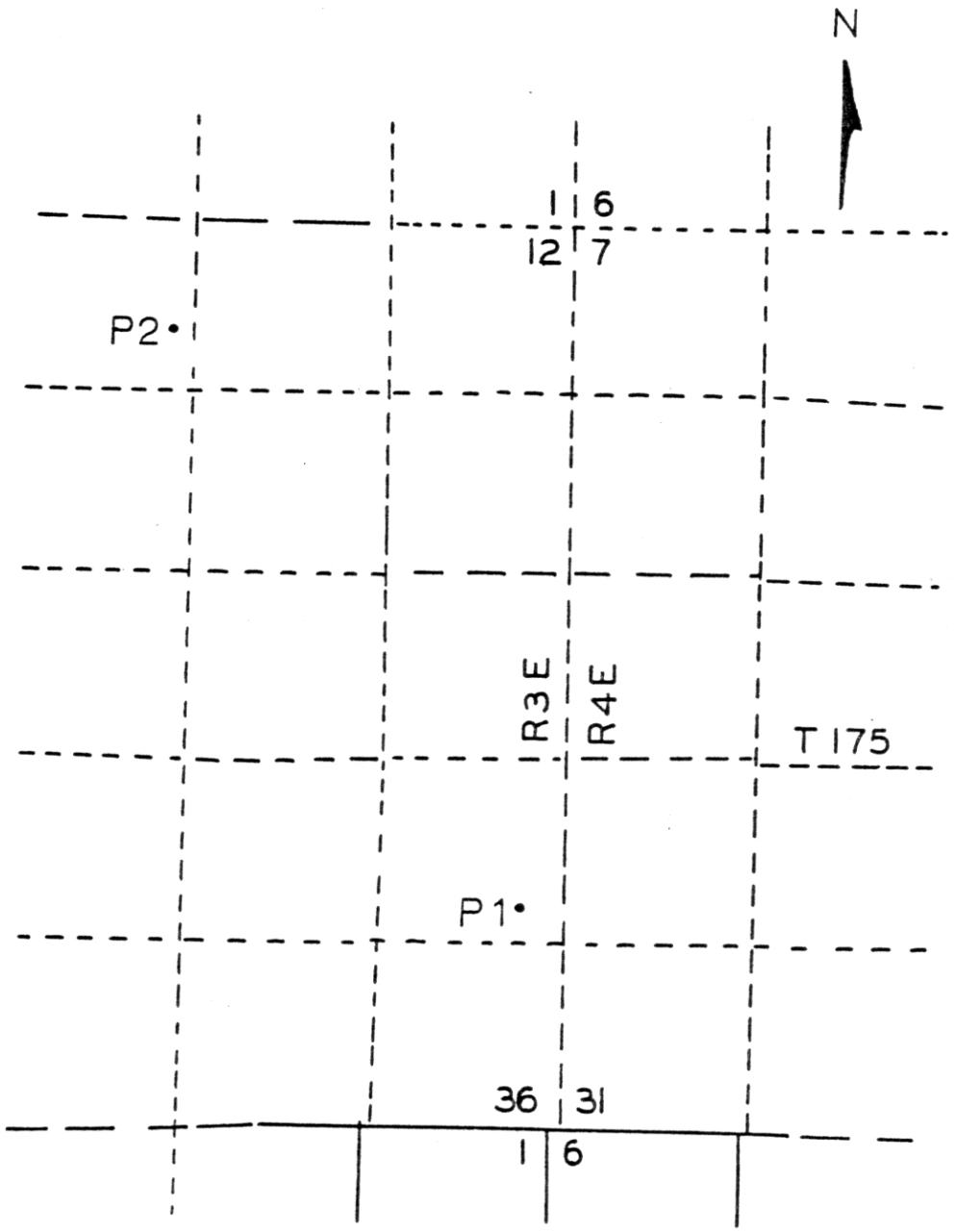


Figure 2. Location map for soil samples on U. S. plywood land.

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determined after Alban and Kellogg (1959). pH determinations were made with a Beckman model N pH meter.

Physical Analyses

Particle size distributions were determined by the Pacific Northwest Forest and Range Experiment Station, Forestry Sciences Laboratory, Corvallis, Oregon using the Bouyoucos hydrometer method (Forest Soils Committee, 1953). Fifteen-atmosphere water was determined in the Oregon State University soil physics laboratory using a pressure membrane apparatus.

Soil Fractionation

Soil samples were air-dried, crushed and passed through a 2 mm sieve. Appropriate sample sizes (sufficient to yield approximately ten grams of silt) were placed in 600 ml beakers and treated with 35 percent $H_2O_2$ for decomposition of organic matter and dissolution of $MnO_2$ (Jackson, 1956, pp. 35, 36). NaOAc washings were followed by four to six washings with distilled water to remove excess NaOAc and disperse the sample. Sand was separated from silt and clay by wet sieving through a 270 mesh sieve using dilute $Na_2CO_3$ solution (1 g. of $Na_2CO_3$ per 9 l. of distilled water) and distilled water. The supernatant, containing silt and clay, was transferred into 250 ml centrifuge tubes and centrifuged at 750

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R.P.M. for 3.9 minutes using an international centrifuge size 1 type SB with head number 242. The supernatant suspension, containing less than two micron clay, was decanted into a 2000 ml erlenmeyer flask. Resuspension, centrifugation, and decantation were continued until the supernatant was fairly clear (Jackson, 1956, pp. 101-141). The clay suspension was flocculated by addition of solid NaCl, allowed to stand for several hours, so the suspended material could settle out, and the clear supernatant solution was siphoned off and discarded. Dried sand and silt were stored in envelopes and sodium saturated clay was stored in glass jars and retained for subsequent x-ray examination.

Selective Dissolution Treatment

A clay sample of approximately 5 gms was placed in a 250 ml centrifuge tube and treated for removal of free iron oxide (Jackson, 1956, p. 55). Following this treatment, the sample was washed several times to remove excess NaCl and disperse the sample. The suspension was brought up to a volume of about 30 ml and quick frozen in a mixture of dry ice and acetone. Frozen samples were placed in a freeze-dry unit and left until dry. Dry samples were then placed in an oven at 105° C for 12 hours to obtain a beginning oven-dry weight. A 500 mg sub-sample was given a boiling 0.5 N KOH treatment for selective dissolution of amorphous material using a

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solid to solution ratio of 100 mg clay to 150 ml 0.5 N KOH. This was followed by a second iron removal (Jackson, 1956, p. 55). The remainder of the sequence consisted of a second KOH dissolution followed by an iron removal treatment. Clay samples were then washed five times in 1.0 M $(NH_4)_2CO_3$ and dried in an oven at 65° C for 12 hours to obtain an ending oven-dry weight. These clay samples were then saved for x-ray diffraction analysis and examination with the electron microscope.

X-ray Diffraction Analysis

Clay samples were prepared for x-ray identification by washing separate sub-samples of approximately 50 mg of clay twice in normal chloride salts of calcium and potassium. This was followed by two washings in distilled water to remove excess salt solution. Following the last washing, a small amount of sediment (sufficient to form a thin film on a petrographic slide) was removed from a vertical section in the centrifuge tube and smeared on a petrographic slide with a micro spatula (Theisen and Harward, 1962).

Duplicate slides of calcium saturated clay were placed in an oven at 65° C for one hour. One set was stored in a desiccator over a saturated solution of $Mg(NO_3)_2 \cdot 6H_2O$ with a relative humidity of

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54 percent for 24 hours. X-ray diffraction analysis was done with air to the goniometer passed through a saturated solution of $Mg(No_3)_2 \cdot 6H_2O$ to control relative humidity at about 54 percent. After analysis, these slides were solvated in glycerol. Treatment was similar to the procedure of Brown and Farrow (1956) except the slides were placed over liquid glycerol in a desiccator, heated to 105° C for three hours, and then cooled and equilibrated at room temperature for 12 hours. The remaining set was solvated in ethylene glycol. Treatment was similar to the procedure of Kunze (1955) except slides were placed over liquid ethylene glycol in a desiccator, heated to 65° C for three hours, and then cooled and equilibrated at room temperature for 12 hours. Desiccators were evacuated before being placed in the oven. X-ray diffraction patterns were obtained for each solvated sample without controlled atmosphere to the goniometer.

Slides of potassium saturated clay were dried in the oven at 105° C for at least six hours and were stored in a desiccator over drierite. X-ray diffraction analysis was done on each slide with air to the goniometer passed through two columns of drierite. Following analyses, the slides were placed over a saturated solution of $Mg(NO_3)_2 \cdot 6H_2O$ in a desiccator and allowed to rehydrate for at least 12 hours. X-ray diffraction analysis was carried out in a controlled atmosphere of about 54 percent relative humidity. These

slides were subsequently given two successive heat treatments, one at 300° C for three hours followed by one at 550° C for three hours. Each heat treatment was followed by x-ray diffraction analysis in a dry atmosphere.

X-ray diffraction patterns were obtained by using a Philips Norelco x-ray diffractometer equipped with a Geiger-Mueller tube and a Minneapolis-Honeywell recorder. The x-ray generator was operated at 50 KV and 35 MA using a copper tube and a nickel filter. Divergent and scatter slits of 1/4° and a 0.006 inch receiving slit were used for scanning from 3° to 14° 2θ.

Silt Mineralogy

A 2 gm sub-sample of the silt fraction was wet-sieved through a 325 mesh sieve retaining coarse silt approximately 40 to 50 microns in size. Free iron oxides were removed and the samples were dried (Jackson, 1956, p. 55). Sub-samples were obtained by coning and quartering. Silt grains were mounted in Permount® on a petrographic slide and cured for seven days at room temperature. This was followed by seven days of curing in an oven at 75° C. This method prevented the formation of small air bubbles on the slides. Mineralogical data were obtained by a randomized count of 300 grains using a petrographic microscope.

Preliminary examination of the silt fraction revealed that much

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of the silt consisted of undispersed pseudomorphs of clay. In an attempt to determine the cementing agent, 500 mg sub-samples of silt were given a sequence of treatments consisting of an iron removal (Jackson, 1956, p. 55), a sonification treatment, a selective dissolution treatment in boiling 0.5 N NaOH (Jackson, 1956, p. 524-525), and a final sonification treatment. The sonification treatment consisted of exposing silt samples to disruptive vibrations of 100,000 cycles/second for seven minutes with a Bronwill Biosonik$^{R}$ BP-111 ultrasonic system. Dispersion occurred following the sodium hydroxide dissolution treatment but not following iron removal.

Iron was determined on citrate-dithionite extracts using a Perkin-Elmer atomic absorption spectrophotometer model 303. Colorimetric determinations were made on sodium hydroxide dissolution extracts for silicon (U. S. Department of Agriculture, 1954) and aluminum (Hsu, 1963).

Prior to receiving the above sequence of treatments, selected samples of silt were crushed in a mortar and dispersed in a mixture of distilled water and dilute $Na_2CO_3$. Clay samples were collected by centrifugation and decantation. Slide preparation and x-ray diffraction analysis were the same as already described except the iron removal treatment and the boiling 0.5 N KOH dissolution treatment were omitted.

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Electron Micrographs

Sub-samples of clay, pretreated for removal of free iron and amorphous material, were prepared by washing about 50 mgs of clay twice in dilute $Na_2CO_3$. This was followed by two washings in distilled water to remove excess $Na_2CO_3$. Samples were dispersed in about 150 ml of distilled water in a 250 ml plastic bottle and placed on a shaker for two hours.

Specimens were prepared for examination with the electron microscope by placing a single drop of suspension on a parlodion film supported on a 400 mesh copper grid and allowing it to evaporate to dryness. Specimens were shadowed with platinum at an angle of 20 degrees. Photographs were obtained by use of a Philips EM 200 Electron microscope.

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# RESULTS

## Clay Mineralogy

### X-ray Diffraction

Clay samples pretreated for removal of free iron oxides and boiling 0.5 N NaOH soluble components were used in the x-ray identification procedure. Results are summarized in Table 1.

Smectites were the predominant crystalline clay minerals present in the inorganic colloidal fraction of horizon samples from profiles of the Slipout and Budworm soils. Smectites were less abundant in clay samples from horizons of buff tuff and McKenzie River soils. Their occurrence was indicated primarily by expansion of 14 Å components upon solvation with glycerol or ethylene glycol and subsequent collapse upon potassium saturation and heating (Figures 3, 4, 5, and 6).

Hydroxy interlayered smectites occurred in most samples. Their presence was indicated by incomplete collapse of some of the 14 Å component to positions intermediate between 10 Å and 14 Å. Generally, the relative amount of this interlayered component increased upward in the soil profile indicating that hydroxy interlayering was more intense near the surface. These intergradient species were more evident in samples from buff tuff and McKenzie River soils

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Table 1 (Continued)

| Sample | Horizon | Relative Amounts[2] Kaolin | Mica | Chlorite-Chloritic Intergrade | Smectite | Zeolite | Amorphous % |
|---|---|---|---|---|---|---|---|
| Budworm | | | | | | | |
| B3-2 | A12 | 1 | 2 | 2 | 6 | 1 | 33.50 |
| B3-3 | $B2_t$ | 1 | 1 | 2 | 6 | 1 | 18.23 |
| B3-5 | C | 1 | 1 | 2 | 6 | 1 | 26.16 |
| B3-6 | R | 1 | 1 | 2 | 6 | 1 | 23.14 |
| buff tuff | | | | | | | |
| P1-1 | A11 | 5 | 2 | 3 | 2 | 1 | 52.95 |
| P1-2 | A12 | 6 | 1 | 2 | 2 | 1 | 46.88 |
| P1-3 | B1 | 5 | 1 | 2 | 4 | 1 | 46.85 |
| P1-4 | II $B21_t$ | 4 | 1 | 2 | 5 | 1 | 50.61 |
| P1-5 | III $B22_t$ | 4 | 1 | 2 | 5 | 1 | 46.66 |
| P1-6 | III $B3_t$ | 4 | 2 | 2 | 5 | 1 | 53.16 |
| P2-1 | A11 | 5 | 1 | 4 | 2 | 1 | 43.61 |
| P2-4 | $B21_t$ | 6 | 2 | 2 | 2 | 1 | 37.45 |
| P2-6 | R | 6 | 2 | 2 | 2 | 1 | 49.52 |
| McKenzie River | | | | | | | |
| M1-1 | A11 | 5 | 2 | 3 | 2 | 1 | 60.25 |
| M1-2 | A12 | 5 | 2 | 4 | 2 | 1 | 54.68 |
| M1-3 | A3 | 5 | 2 | 3 | 2 | 1 | 57.84 |
| M1-4 | B1 | 5 | 2 | 3 | 3 | 1 | 54.69 |
| M1-5 | $B21_t$ | 5 | 2 | 2 | 3 | 1 | 44.39 |
| M1-6 | $B22_t$ | 5 | 2 | 2 | 2 | 1 | 44.96 |
| M1-7 | B3 | 6 | 1 | 1 | 1 | 1 | 49.37 |
| M1-8 | R | 6 | 2 | 1 | 2 | 1 | 49.28 |



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Table 1. (Continued)

| Sample | Horizon | Relative Amounts[2] Kaolin | Mica | Chlorite-Chloritic Intergrade | Smectite[1] | Zeolite | Amorphous[3] % |
|---|---|---|---|---|---|---|---|
| McKenzie River | | | | | | | |
| M2-1 | A11 | 5 | 2 | 4 | 2 | 1 | 56.63 |
| M2-4 | $B21_t$ | 5 | 2 | 2 | 3 | 1 | 59.11 |
| M2-6 | B3 | 6 | 2 | 2 | 3 | 2 | 60.35 |
| M3-1 | A11 | 5 | 2 | 3 | 2 | 1 | nd |
| M3-2 | A12 | 5 | 2 | 3 | 3 | 2 | 41.79 |
| M3-4 | $B21_t$ | 6 | 2 | 2 | 2 | 2 | 41.56 |
| M3-6 | R | 5 | 2 | 1 | 4 | 2 | 62.29 |

[1] Montmorillonite with trace amounts of biedillite

[2] Expressed as relative quantities of crystalline clay based on peak area in the following progression: 1) none; 2) trace, 0-10%; 3) minor, 10-25%; 4) intermediate, 25-50%; 5) major, 50-75%; 6) dominant, 75-100%

[3] Based on differential dissolution weight loss



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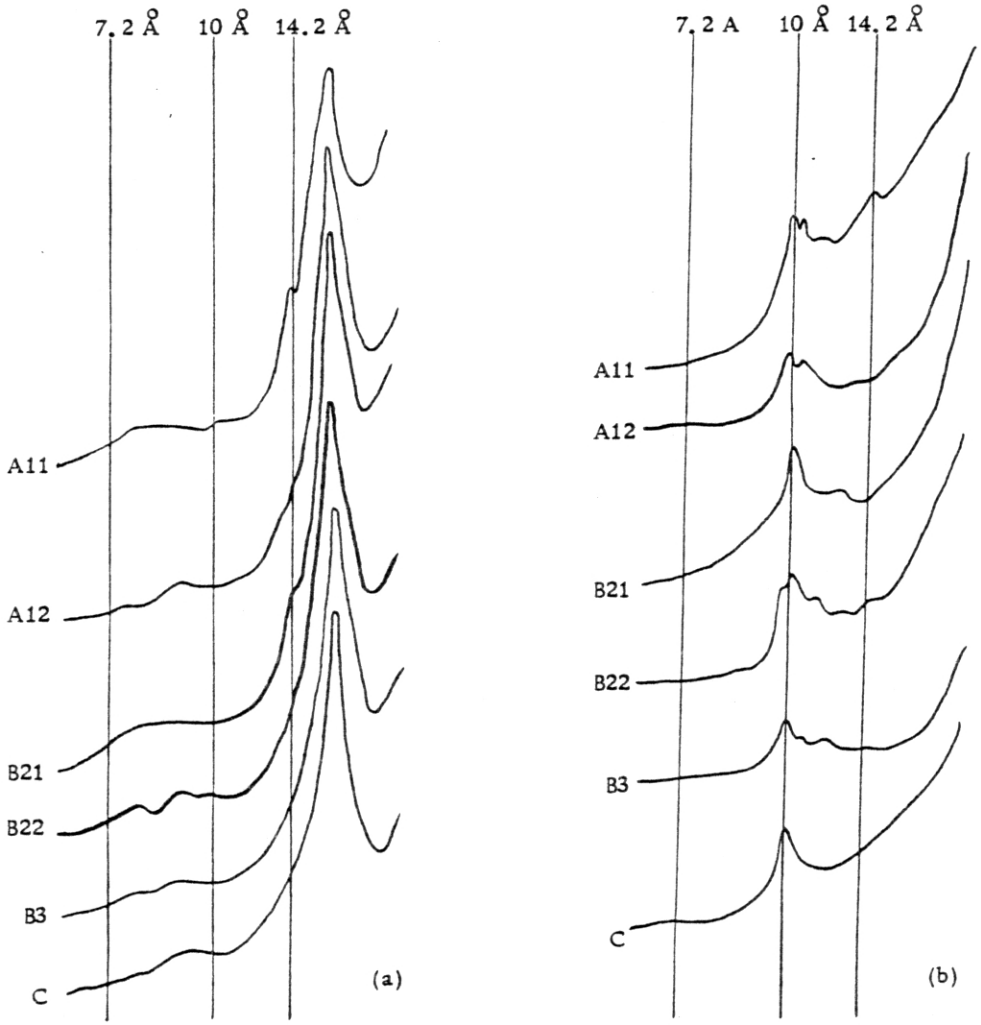


Figure 3. Smoothed traces of diffraction patterns of clay samples from Slipout soil profile number 1, (a) calcium saturated glycerol solvated samples, (b) potassium saturated samples heated to 550° C.

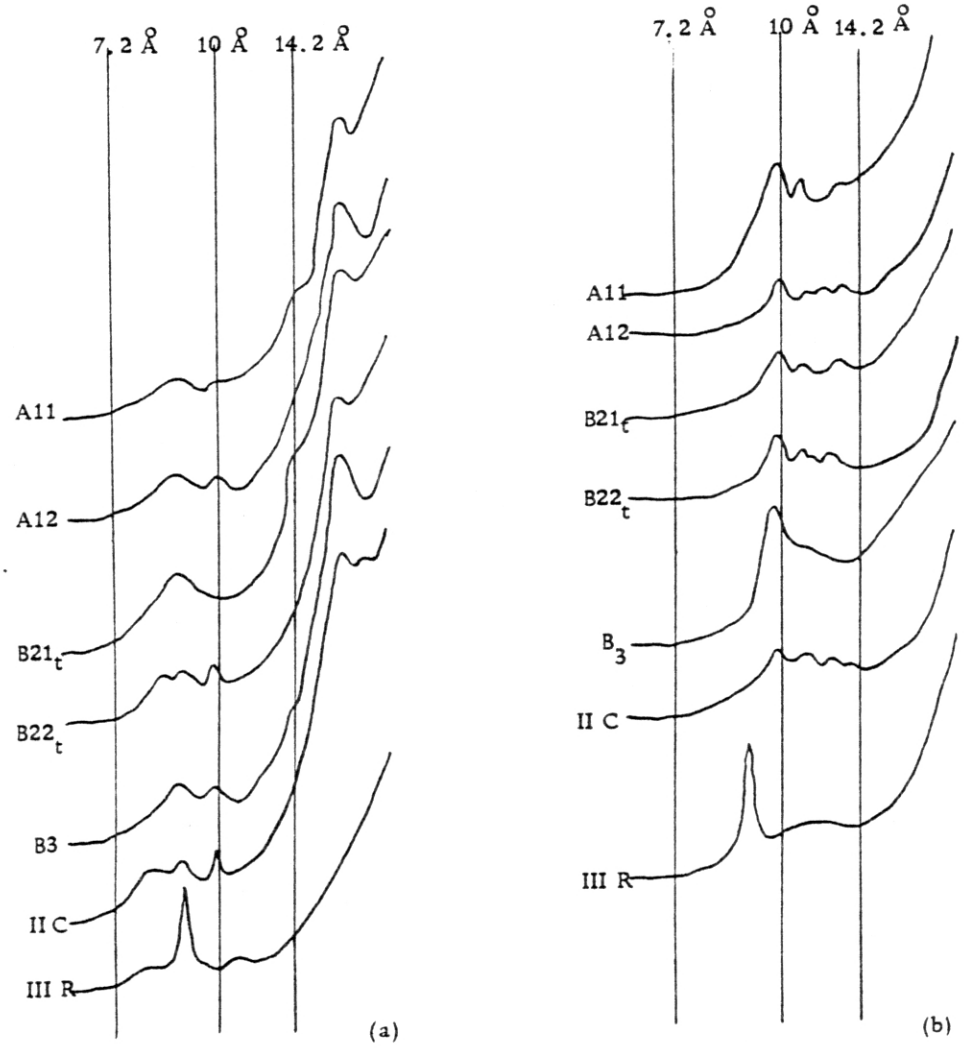


Figure 4. Smoothed traces of diffraction patterns of clay samples from Budworm soil profile number 1, (a) calcium saturated glycerol solvated samples, (b) potassium saturated samples heated to 550° C.

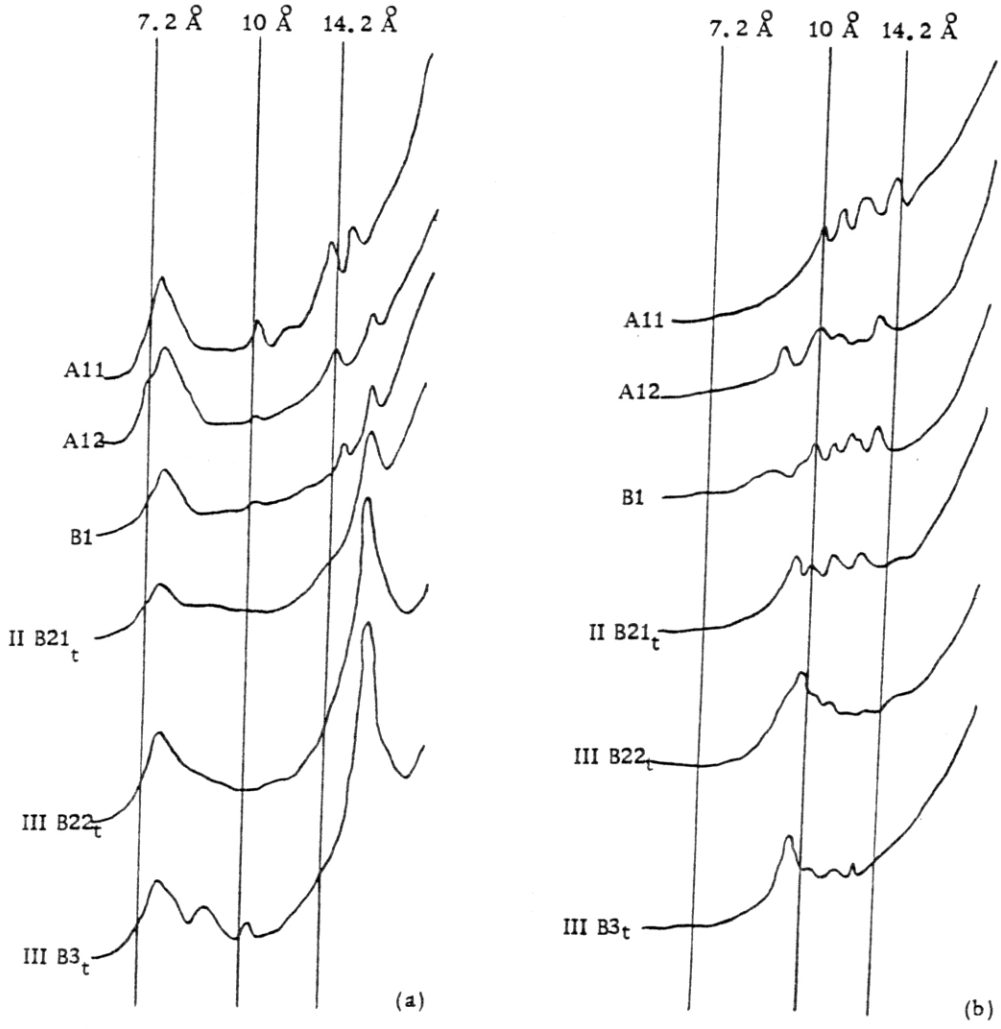


Figure 5. Smoothed traces of diffraction patterns of clay samples from buff tuff soil profile number 1, (a) calcium saturated ethylene glycol solvated samples, (b) potassium saturated samples heated to 550° C.

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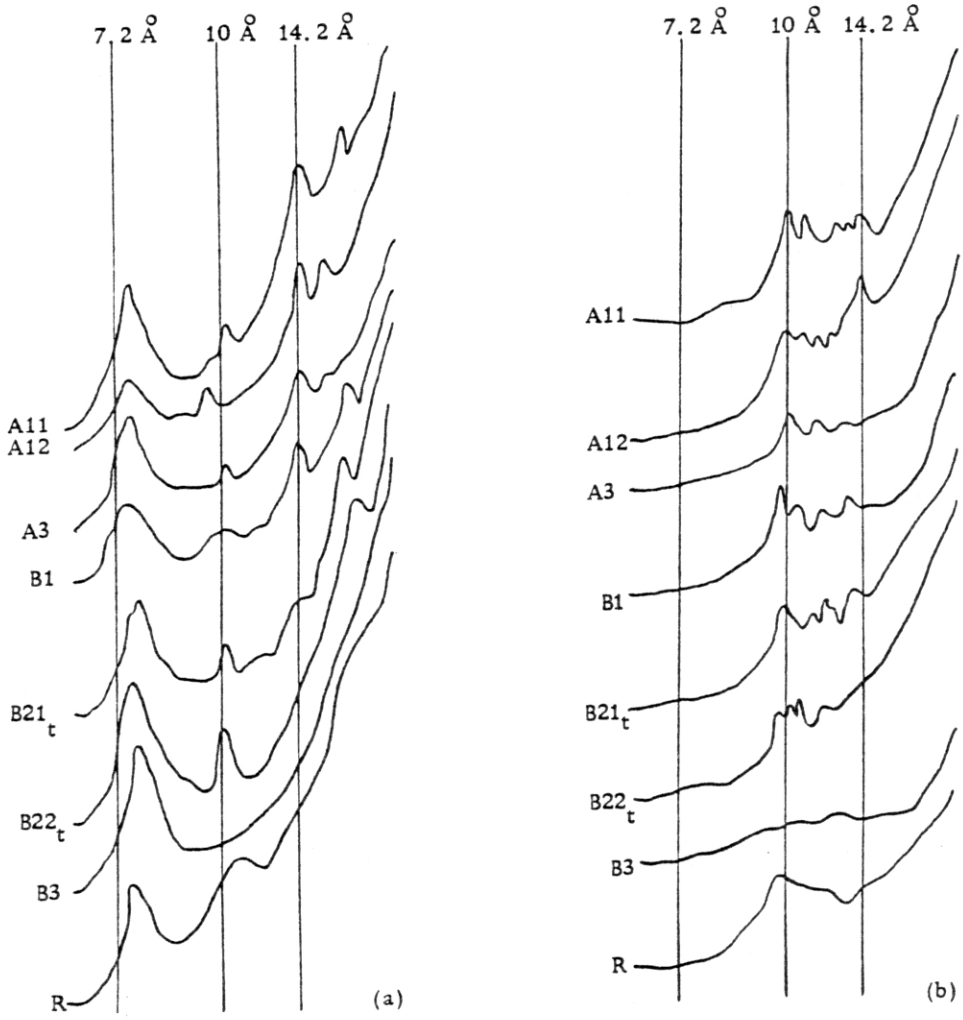


Figure 6. Smoothed traces of diffraction patterns of clay samples from McKenzie River soil profile number 1, (a) calcium saturated ethylene glycol solvated samples, (b) potassium saturated samples heated to 550° C.

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than they were in Budworm and Slipout soils.

Small 14.2 Å peaks persisted in the surface horizons of Slipout, buff tuff, and McKenzie River soils indicating the presence of chlorite.

Kaolin minerals dominated the clay fraction of horizon samples from profiles of buff tuff and McKenzie River soils but were not detected in clay samples from Budworm and Slipout soils. Their presence was indicated by strong rather broad 7.5 Å diffraction maxima (Figures 5 and 6).

A residual 10 Å peak showed that mica was present in most horizon samples of Budworm and McKenzie River soils but occurred in only a few horizon samples from Slipout and buff tuff soils.

The intense 8.9 Å peak (Figure 4) indicated that heulandite-clinoptilolite was the only crystalline component present in the clay fraction of the greenish tuff-breccia underlying Budworm soil profile number 1.

Vermiculite was probably present in samples from McKenzie River and buff tuff soils but detection was not possible because mica lines overlap those from potassium saturated samples and chlorite-like lines overlap those from solvated samples.

The sand and silt fractions of most of the soils under investigation contained pseudomorphs of clay that were not dispersed following treatment with cold sodium carbonate. Clay samples were

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mechanically dispersed from these pseudomorphs so that x-ray diffraction patterns of clay mechanically dispersed from the silt fraction could be compared with x-ray diffraction patterns of dispersible clay from the same horizon. To obtain mechanically dispersed clay, silt-size pseudomorphs of clay were crushed in a mortar and dispersed in dilute sodium carbonate. Clay was recovered by centrifugation and decantation.

X-ray diffraction patterns of clay samples from the A12, B21$_t$, and R horizons of McKenzie River soil profile number 3 showed that dispersible kaolin increased slightly toward the surface and dispersible montmorillonite was almost constant with depth (Figure 7a). By comparison, kaolin mechanically dispersed from the silt fraction of these same horizons remained almost constant with depth but montmorillonite mechanically dispersed from the silt fraction decreased sharply toward the surface (Figure 7b). Hydroxy interlayering of montmorillonite was also less evident in clay samples mechanically separated from the silt fraction than it was in dispersible clay samples.

X-ray diffraction patterns of clay samples from the A11, B21$_t$ and R horizons of buff tuff soil profile number 2 showed that dispersible kaolin decreased slightly toward the surface and dispersible montmorillonite increased toward the surface (Figure 8a). X-ray diffraction patterns of corresponding clay samples mechanically

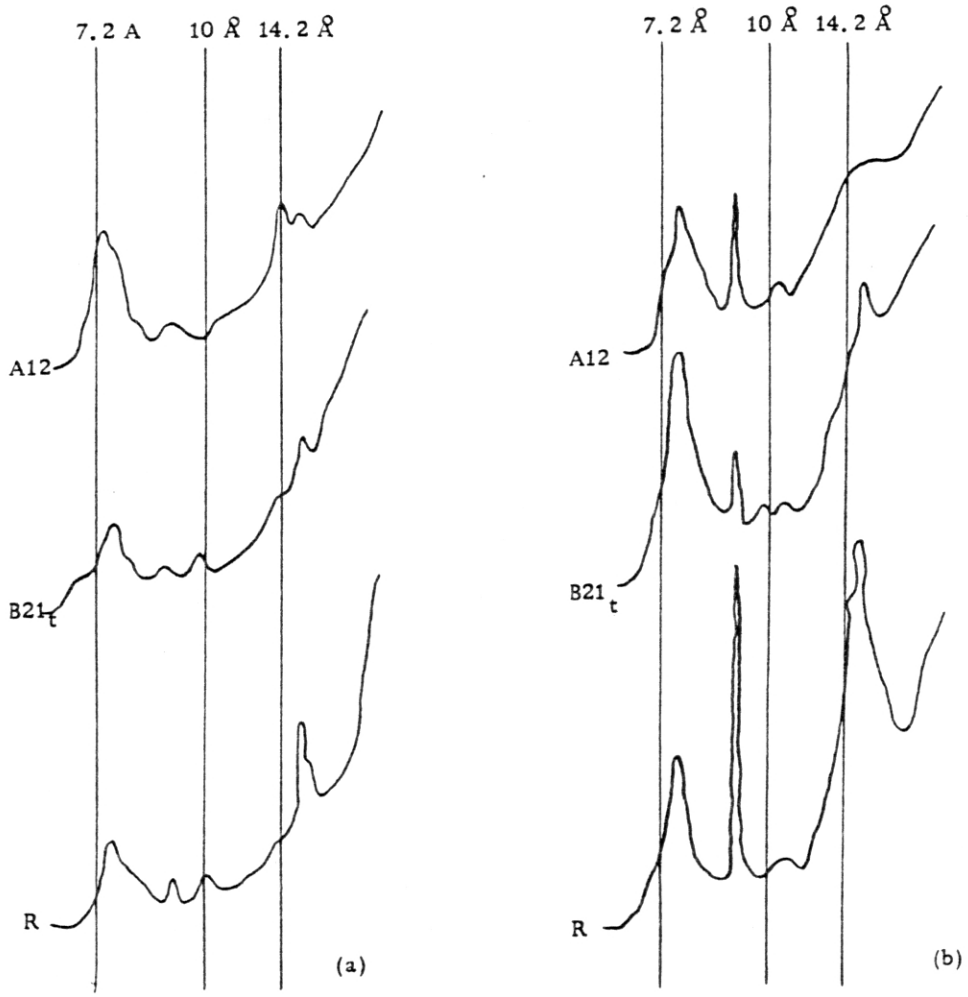


Figure 7. Smoothed traces of diffraction patterns of calcium saturated ethylene glycol solvated clay samples from McKenzie River soil profile number 3, (a) dispersible clay, (b) clay mechanically dispersed from the silt fraction.

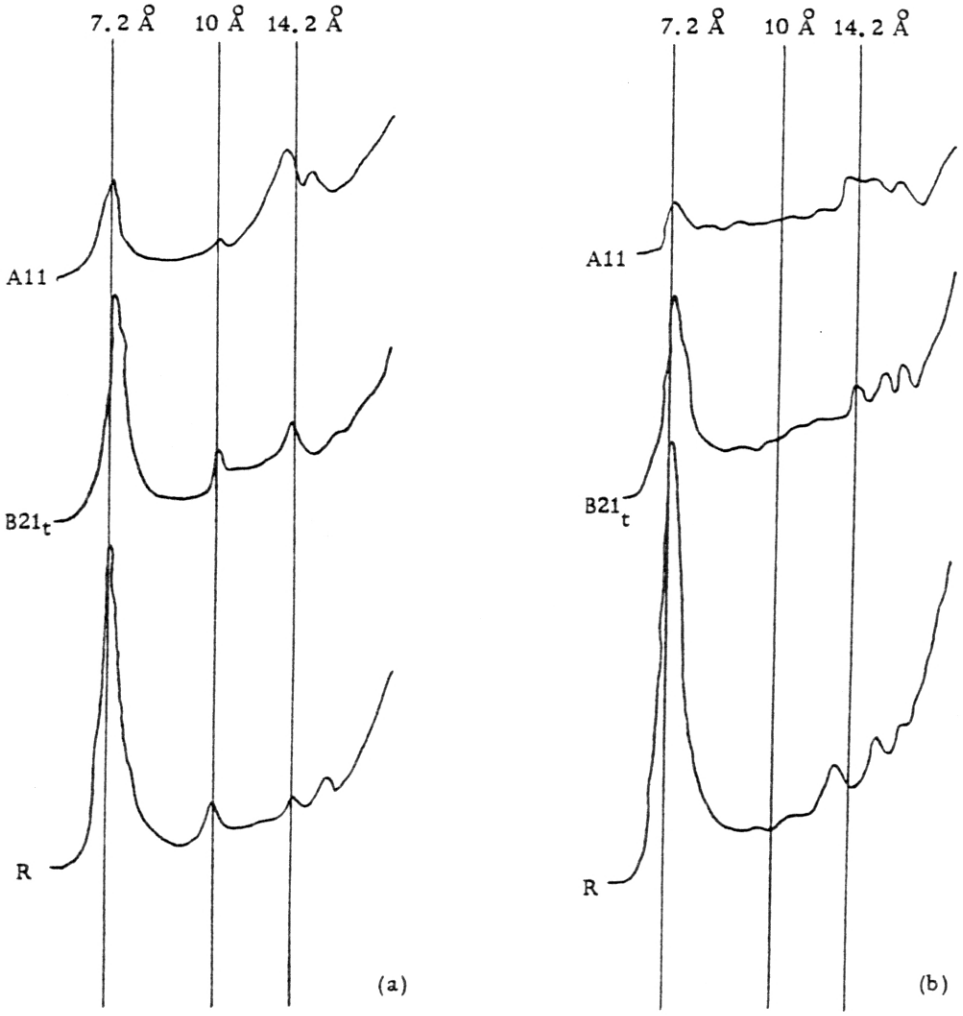


Figure 8. Smoothed traces of diffraction patterns of calcium saturated ethylene glycol solvated clay samples from buff tuff soil profile number 2, (a) dispersible clay, (b) clay mechanically dispersed from the silt fraction.

dispersed from the silt fraction showed that kaolin decreased sharply toward the surface and montmorillonite decreased slightly in the same direction (Figure 8b).

Similarly, diffraction patterns of clay samples from McKenzie River soil profile number 1 (Figure 6) showed that dispersible kaolin and dispersible montmorillonite both increased toward the surface, whereas kaolin and montmorillonite mechanically dispersed from the silt fraction by grinding both decreased upward (Figure 9).

Electron Micrographs

Kaolinite and halloysite are common constituents of clay fractions of many soils in Oregon. If either kaolinite or halloysite is present in a sample, there is no difficulty differentiating between them. However, when both minerals are present in a mixture, a problem of identification develops. The purpose of these electron micrographs was to determine if kaolinite and/or halloysite were present in clay components which exhibited broad, rather diffuse, moderately intense 7.5 Å diffraction bands.

Clay samples from the A11 and B2$_1$ horizons of buff tuff soil profile number 2 contain mainly kaolinite, but some halloysite was also present (Figures 10 and 11). Kaolinite occurred as small lath-like forms, the edges of which are slightly curled, and as larger platy forms with irregular but distinct well-defined edges. Halloysite

Figure 10. Electron micrograph of kaolinite and halloysite from the A11 horizon of buff tuff soil profile number 2. Scale is 1 micron.

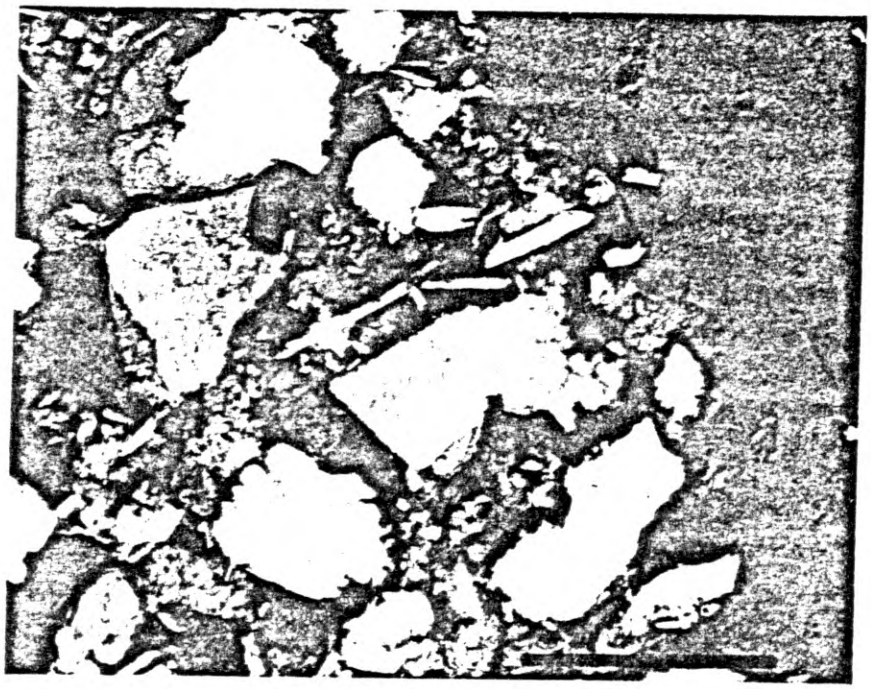

Figure 11. Electron micrograph of kaolinite and halloysite from the $B21_t$ horizon of buff tuff soil profile number 2. Scale is 1 micron.

Figure 12. Electron micrographs of halloysite and kaolinite from the tuff-breccia (R horizon) underlying buff tuff soil profile number 2. Scale is 1 micron.

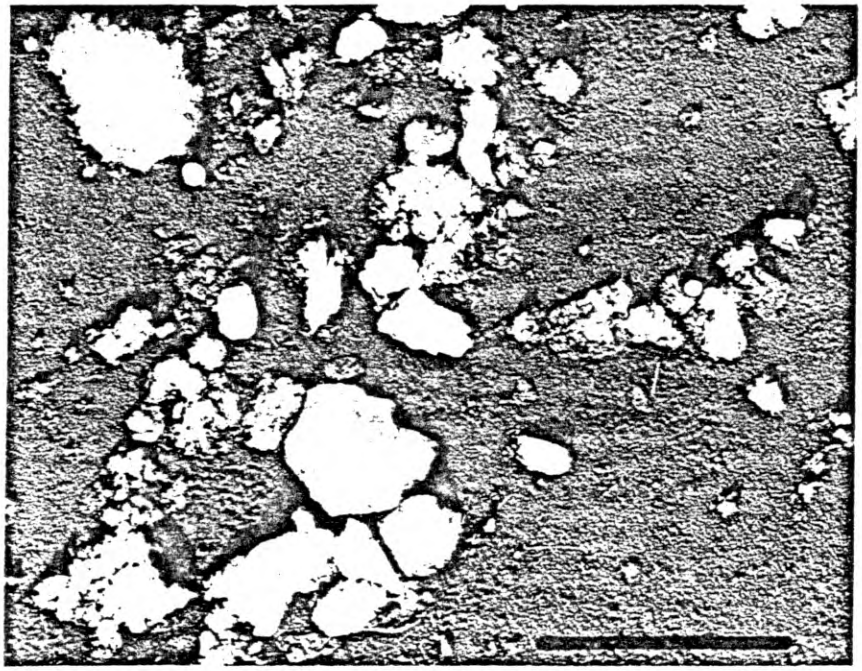

Figure 13. Electron micrograph of kaolinite from the A3 horizon of McKenzie River soil profile number 1. Scale is 1 micron.

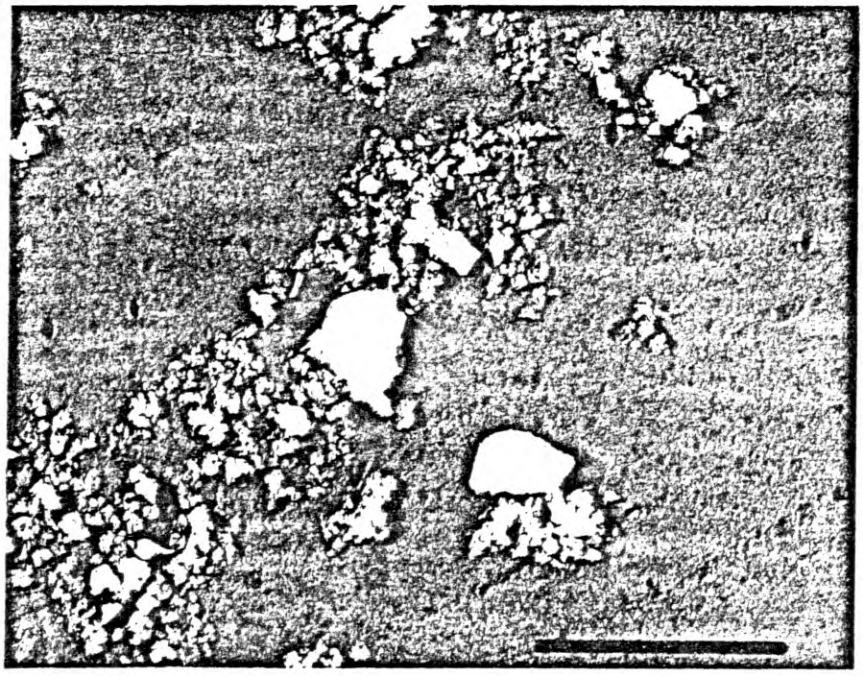

Figure 14. Electron micrograph of kaolinite and halloyside from the $B21_t$ horizon of McKenzie River soil profile number 1. Scale is 1 micron.

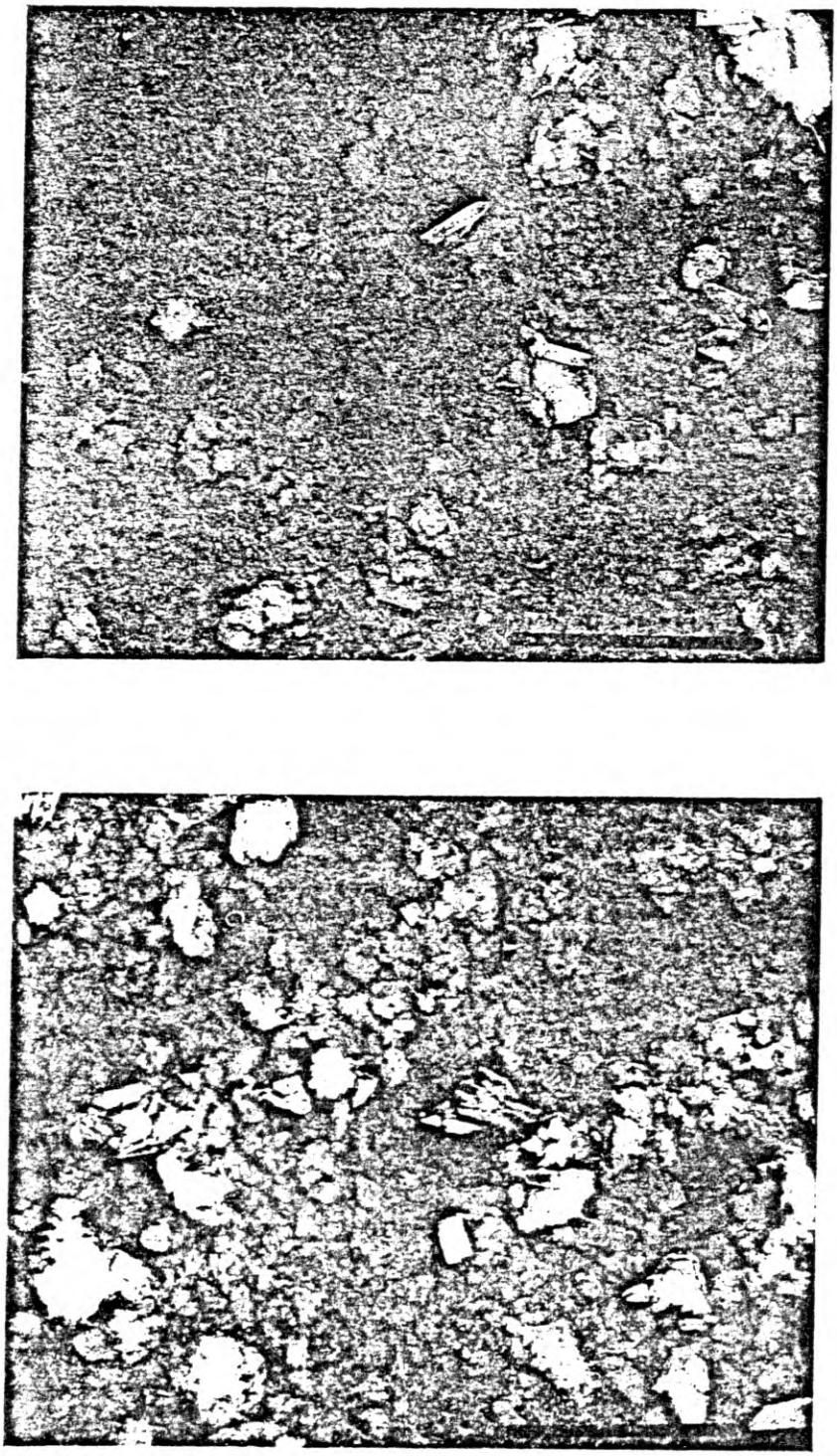

Figure 15. Electron micrographs of kaolinite and halloysite from the reddish tuff-breccia (R horizon) underlying McKenzie River soil profile number 1. Scale is 1 micron.

Table 2. Mineral composition of coarse silt (40μ to 50μ) fraction of four Western Cascade soils. (Percent)

| Sample | Horizon | Aggregates | Plagioclase | Biotite | Quartz | Amphibole-Pyroxene | Opaque | Glass | Heulandite-Clinoptilotite | Plant Opal |
|---|---|---|---|---|---|---|---|---|---|---|
| Slipout | | | | | | | | | | |
| S1-1 | A11 | 38.1 | 31.6 | 7.7 | 2.9 | 2.0 | 7.7 | 5.2 | 1.9 | 2.9 |
| S1-2 | A12 | 58.6 | 28.3 | 4.9 | 1.0 | 1.3 | 3.6 | 1.0 | 0.7 | 0.7 |
| S1-3 | B21 | 67.2 | 22.3 | 4.9 | 1.0 | 0.0 | 3.6 | 0.0 | 1.0 | 0.0 |
| S1-4 | B22 | 78.9 | 11.4 | 3.7 | 0.7 | 0.7 | 1.7 | 0.0 | 3.0 | 0.0 |
| S1-5 | B3 | 65.5 | 23.3 | 3.4 | 1.0 | 0.3 | 3.0 | 0.0 | 3.4 | 0.0 |
| S1-6 | C | 67.5 | 18.7 | 3.0 | 1.0 | 0.3 | 2.3 | 0.0 | 7.2 | 0.0 |
| S2-1 | A11 | 17.3 | 48.3 | 4.3 | 1.7 | 2.6 | 7.3 | 5.3 | 9.7 | 3.3 |
| S2-3 | II $B21_t$ | 29.0 | 42.9 | 4.3 | 2.0 | 2.6 | 5.6 | 1.3 | 11.2 | 1.0 |
| S2-5 | III B3 | 45.9 | 31.3 | 5.2 | 1.3 | 1.9 | 3.3 | 1.3 | 9.8 | 0.0 |
| S3-1 | A11 | 42.0 | 21.0 | 5.3 | 1.6 | 1.6 | 13.1 | 7.9 | 3.9 | 3.6 |
| S3-3 | B21 | 73.7 | 13.5 | 3.2 | 1.1 | 0.7 | 2.1 | 1.1 | 4.3 | 0.4 |
| S3-7 | C | 82.7 | 7.4 | 3.2 | 0.7 | 0.7 | 0.7 | 0.0 | 4.6 | 0.0 |
| Budworm | | | | | | | | | | |
| B1-1 | A11 | 18.5 | 50.3 | 2.3 | 1.0 | 5.6 | 5.3 | 6.6 | 7.3 | 3.0 |
| B1-2 | A12 | 14.8 | 52.4 | 4.7 | 1.3 | 9.7 | 6.7 | 2.7 | 7.1 | 0.7 |
| B1-3 | $B21_t$ | 22.5 | 52.1 | 2.6 | 1.1 | 6.0 | 4.9 | 0.4 | 10.6 | 0.0 |
| B1-4 | $B22_t$ | 23.2 | 50.7 | 3.0 | 0.7 | 4.6 | 6.6 | 0.0 | 11.3 | 0.0 |
| B1-5 | B3 | 42.9 | 37.5 | 1.4 | 1.4 | 3.2 | 3.9 | 1.8 | 7.9 | 0.0 |
| B1-6 | II C | 82.1 | 10.3 | 1.4 | 0.3 | 0.0 | 0.7 | 0.0 | 5.2 | 0.0 |
| B1-7 | III R | 57.8 | 6.0 | 0.7 | 0.3 | 0.3 | 0.7 | 0.0 | 34.2 | 0.0 |
| B2-1 | A11 | 55.2 | 18.9 | 2.3 | 1.7 | 1.7 | 7.0 | 6.3 | 3.6 | 3.3 |
| B2-3 | $B2_t$ | 44.7 | 35.0 | 2.3 | 2.3 | 2.0 | 5.7 | 2.3 | 3.0 | 2.7 |
| B2-5 | B32 | 91.8 | 4.9 | 0.3 | 0.0 | 1.3 | 1.0 | 0.0 | 0.7 | 0.0 |
| B2-6 | R | 76.4 | 7.6 | 0.3 | 0.0 | 0.3 | 0.3 | 0.0 | 15.0 | 0.0 |

Table 2. (Continued)

| Sample | Horizon | Aggregates | Plagioclase | Biotite | Quartz | Amphibole-Pyroxene | Opaque | Glass | Heulandite-Clinoptilotite | Plant Opal |
|---|---|---|---|---|---|---|---|---|---|---|
| Budworm | | | | | | | | | | |
| B3-2 | A12 | 64.8 | 18.9 | 3.3 | 1.3 | 2.6 | 5.5 | 1.3 | 2.3 | 0.0 |
| B3-3 | $B2_t$ | 61.8 | 21.9 | 0.7 | 2.0 | 6.3 | 2.7 | 0.3 | 4.3 | 0.0 |
| B3-5 | C | 68.6 | 18.8 | 1.7 | 0.7 | 4.8 | 2.1 | 0.0 | 3.4 | 0.0 |
| B3-6 | R | 75.7 | 11.3 | 0.3 | 1.0 | 5.7 | 3.0 | 0.0 | 3.0 | 0.0 |
| buff tuff | | | | | | | | | | |
| P1-1 | A11 | 55.9 | 15.4 | 1.0 | 11.4 | 8.0 | 6.0 | 1.3 | 0.0 | 1.0 |
| P1-2 | A12 | 56.7 | 18.5 | 0.3 | 10.1 | 6.7 | 7.7 | 0.0 | 0.0 | 0.0 |
| P1-3 | B1 | 60.7 | 20.3 | 1.0 | 11.3 | 4.0 | 2.7 | 0.0 | 0.0 | 0.0 |
| P1-4 | II $B21_t$ | 67.0 | 18.3 | 0.3 | 10.0 | 2.3 | 2.0 | 0.0 | 0.0 | 0.0 |
| P1-5 | III $B22_t$ | 66.7 | 25.3 | 0.3 | 6.0 | 1.0 | 0.7 | 0.0 | 0.0 | 0.0 |
| P1-6 | III $B23_t$ | 55.0 | 35.0 | 0.3 | 5.3 | 2.7 | 1.7 | 0.0 | 0.0 | 0.0 |
| P2-1 | A11 | 8.6 | 78.9 | 1.0 | 4.0 | 2.3 | 1.7 | 2.0 | 0.0 | 1.7 |
| P2-4 | $B21_t$ | 5.4 | 74.4 | 1.4 | 14.8 | 1.7 | 2.4 | 0.0 | 0.0 | 0.0 |
| P2-6 | R | 6.7 | 60.0 | 0.0 | 29.7 | 1.0 | 2.7 | 0.0 | 0.0 | 0.0 |
| McKenzie River | | | | | | | | | | |
| M1-1 | A11 | 60.1 | 13.1 | 12.7 | 4.8 | 1.7 | 1.8 | 3.5 | 0.0 | 2.2 |
| M1-2 | A12 | 60.3 | 12.7 | 11.4 | 4.4 | 1.8 | 3.1 | 4.4 | 0.4 | 1.7 |
| M1-3 | A3 | 74.1 | 6.7 | 9.6 | 3.3 | 0.4 | 2.1 | 1.7 | 1.3 | 0.8 |
| M1-4 | B1 | 84.6 | 2.2 | 7.0 | 2.6 | 0.0 | 1.8 | 0.0 | 1.3 | 0.4 |
| M1-5 | $B21_t$ | 88.9 | 1.8 | 7.1 | 1.3 | 0.4 | 0.0 | 0.0 | 0.4 | 0.0 |
| M1-6 | $B22_t$ | 89.7 | 0.7 | 8.6 | 0.7 | 0.0 | 0.0 | 0.0 | 0.3 | 0.0 |
| M1-7 | B3 | 90.8 | 0.0 | 6.9 | 1.0 | 0.0 | 0.3 | 0.0 | 1.0 | 0.0 |
| M1-8 | R | 93.3 | 0.3 | 4.0 | 1.0 | 0.0 | 0.3 | 0.0 | 1.0 | 0.0 |

Table 2. (Continued)

| Sample | Horizon | Aggregates | Plagioclase | Biotite | Quartz | Amphibole-Pyroxene | Opaque | Glass | Heulandite-Clinoptilotite | Plant Opal |
|---|---|---|---|---|---|---|---|---|---|---|
| McKenzie River | | | | | | | | | | |
| M2-1 | A11 | 50.2 | 18.6 | 5.7 | 4.0 | 3.6 | 8.3 | 5.0 | 0.0 | 4.7 |
| M2-4 | $B21_t$ | 74.7 | 5.9 | 6.5 | 3.1 | 3.1 | 3.4 | 1.5 | 1.5 | 0.3 |
| M2-6 | B3 | 88.9 | 3.3 | 2.3 | 0.7 | 2.6 | 1.0 | 0.3 | 1.0 | 0.0 |
| M3-2 | A12 | 71.4 | 8.3 | 1.8 | 0.7 | 1.0 | 8.7 | 4.0 | 2.3 | 2.0 |
| M3-4 | $B21_t$ | 88.9 | 0.3 | 1.3 | 1.0 | 0.0 | 2.0 | 0.3 | 5.3 | 1.0 |
| M3-6 | R | 90.3 | 0.0 | 0.7 | 0.7 | 0.0 | 0.7 | 0.0 | 7.7 | 0.0 |

fraction consisted of pseudomorphs of clay after plagioclase, ferromagnesian minerals, devitrified glass, and volcanic ash (Figure 16). These pseudomorphs were weakly birefringent (Figure 17) and some displayed faint pleochroism. Examination of the fine silt and sand in reflected light revealed that pseudomorphs of clay also persist in the fine silt and various sand fractions.

Labradorite and bytownite were the dominant plagioclase minerals present. Magnetite, limonite, and hematite were grouped as opaque minerals. Hornblende, actinolite, augite, and hypersthene were grouped as amphibole-pyroxene. It was not possible to distinguish between zeolites optically, but clinoptilolite and heulandite were identified by means of x-ray diffraction.

All soils showed a marked increase in unweathered plagioclase, amphiboles, and pyroxenes near the surface. These occurred as transparent, euhedral and subhedral crystals, some of which were fractured with ragged terminations. Many were rimmed with glass (Figure 18 and Figure 19). This was accompanied by the presence of volcanic glass which further indicates a surficial deposit of volcanic ash.

Phytoliths of Douglas-fir occurred in surface horizons but seldom extended below 24 inches in depth. Quartz maintained a small but fairly constant percentage of the total composition of the coarse-silt fraction. Heulandite and clinoptilolite decreased toward the

Figure 16. Pseudomorphs of plagioclase, ferromagnesian minerals, devitrified glass, and volcanic ash. (plain light)

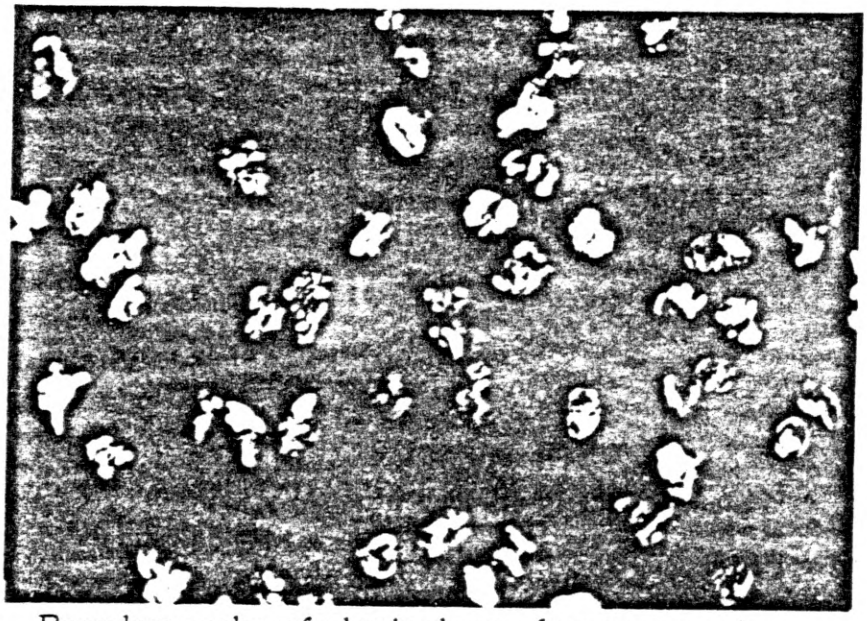

Figure 17. Pseudomorphs of plagioclase, ferromagnesian minerals, devitrified glass, and volcanic ash. (x-nicols)

surface, probably as a result of alteration to pseudomorphs of clay. Opaque minerals exhibited variations in amount with depth.

Several weathering states of plagioclase were represented from fresh, euhedral and subhedral, glass-clear crystals, to intermediate states showing various degrees of etching, and finally to completely weathered pseudomorphs of clay. Zeolitic alteration of plagioclase and glass shards probably occurred subsequent to deposition when circulating hot water caused complete replacement by zeolites and aggregates of green clay (Peck et al., 1964). This was accompanied by the formation of chalcedonic quartz. Pyroxenes, amphiboles, and biotite were replaced almost completely by green clay with little or no change in volume.

## Physical Properties

Clay content increased with depth from the A horizons into the B2 horizons in all profiles except Slipout soil profile number 2 and Budworm soil profile number 3. The transition from B horizons to R horizons resulted in sharp decreases in clay content which were accompanied by increases in sand content (Table 3). There were no additional consistent trends in the sand and silt fractions except in profiles of McKenzie River soils where silt content remained almost constant or increased slightly with depth and sand content tended to decrease with depth.

Table 3. Mechanical analysis of four Western Cascade soils and 15-atmosphere moisture of four selected profiles.

| Sample | Horizon | Sand % | Silt % | Clay % | Textural Class | Field Texture | 15 Bar Moisture % |
|---|---|---|---|---|---|---|---|
| Slipout | | | | | | | |
| S1-1 | A11 | 28.31 | 42.77 | 28.93 | Clay loam | Loam | 18.11 |
| S1-2 | A12 | 22.10 | 32.88 | 45.55 | Clay | Clay loam | 20.83 |
| S1-3 | B21 | 20.83 | 30.70 | 48.48 | Clay | Silty clay | 21.92 |
| S1-4 | B22 | 27.38 | 35.64 | 36.99 | Clay loam | Silty clay | 23.55 |
| S1-5 | B3 | 32.57 | 39.44 | 28.00 | Clay loam | Silty clay | 21.73 |
| S1-6 | C | 24.62 | 39.54 | 35.85 | Clay loam | Silty clay loam | 22.46 |
| S2-1 | A11 | 36.06 | 34.02 | 29.92 | Clay loam | Loam | nd |
| S2-3 | II $B21_t$ | 37.10 | 35.14 | 27.77 | Clay loam | Clay loam | nd |
| S2-5 | III B3 | 24.73 | 30.95 | 44.33 | Clay | Clay | nd |
| S3-1 | A11 | 27.49 | 38.17 | 34.10 | Clay loam | Loam | nd |
| S3-3 | B21 | 19.65 | 39.20 | 41.15 | Clay | Clay loam | nd |
| S3-7 | C | 16.40 | 42.69 | 40.91 | Clay | Clay | nd |
| Budworm | | | | | | | |
| B1-1 | A11 | 36.40 | 30.46 | 33.05 | Clay loam | Loam | 15.34 |
| B1-2 | A12 | 35.25 | 28.93 | 35.82 | Clay loam | Clay loam | 15.32 |
| B1-3 | $B21_t$ | 24.43 | 22.59 | 52.98 | Clay | Clay | 20.18 |
| B1-4 | $B22_t$ | 23.16 | 24.00 | 52.85 | Clay | Clay | 20.45 |
| B1-5 | B3 | 19.94 | 28.16 | 51.90 | Clay | Clay | 21.02 |
| B1-6 | II C | 24.50 | 43.93 | 31.57 | Clay loam | Silty clay loam | 21.46 |
| B1-7 | III R | 57.02 | 28.47 | 14.51 | Sandy loam | nd | 17.10 |
| B2-1 | A11 | 37.24 | 34.44 | 28.33 | Clay loam | Loam | nd |
| B2-3 | $B2_t$ | 24.01 | 36.90 | 39.09 | Clay loam | Clay | nd |
| B2-5 | B32 | 34.02 | 45.17 | 20.82 | Loam | Loam | nd |
| B2-6 | R | 34.70 | 48.08 | 17.23 | Loam | nd | nd |

Table 3. (Continued)

| Sample | Horizon | Sand % | Silt % | Clay % | Textural Class | Field Texture | 15-Bar Moisture % |
|---|---|---|---|---|---|---|---|
| Budworm | | | | | | | |
| B3-2 | A12 | 29.75 | 35.53 | 34.73 | Clay loam | Loam | nd |
| B3-3 | $B2_t$ | 31.23 | 35.79 | 32.99 | Clay loam | Clay loam | nd |
| B3-5 | C | 35.25 | 45.74 | 19.01 | Loam | Loam | nd |
| B3-6 | R | 55.33 | 35.63 | 9.05 | Sandy loam | nd | nd |
| buff tuff | | | | | | | |
| P-1 | A11 | 28.16 | 49.62 | 22.22 | Loam | Loam | 13.90 |
| P1-2 | A12 | 31.63 | 44.77 | 23.61 | Loam | Silt loam | 13.99 |
| P1-3 | B1 | 23.02 | 51.46 | 25.53 | Silt loam | Silt loam | 16.57 |
| P1-4 | II $B21_t$ | 27.21 | 47.36 | 25.43 | Loam | Clay loam | 19.45 |
| P1-5 | III $B22_t$ | 23.42 | 45.37 | 31.21 | Clay loam | Silty clay loam | 21.30 |
| P1-6 | III $B3_t$ | 27.55 | 40.38 | 32.08 | Clay loam | Silty clay loam | 21.62 |
| P2-1 | A11 | 46.86 | 32.27 | 20.88 | Loam | Loam | nd |
| P2-4 | $B21_t$ | 43.33 | 29.23 | 27.45 | Clay loam | Loam | nd |
| P2-6 | R | 49.95 | 27.10 | 22.95 | Loam | nd | nd |
| McKenzie River | | | | | | | |
| M1-1 | A11 | 30.69 | 43.02 | 26.30 | Loam | Loam | 18.54 |
| M1-2 | A12 | 27.67 | 43.05 | 29.28 | Clay loam | Loam | 18.48 |
| M1-3 | A3 | 25.82 | 44.84 | 29.35 | Clay loam | Silty clay loam | 19.49 |
| M1-4 | B1 | 16.56 | 49.14 | 34.31 | Clay loam | Silty clay loam | 23.23 |
| M1-5 | $B21_t$ | 12.90 | 46.46 | 40.65 | Clay | Silty clay | 26.73 |
| M1-6 | $B22_t$ | 12.97 | 45.94 | 41.09 | Clay | Silty clay | 26.40 |
| M1-7 | B3 | 14.16 | 51.82 | 34.02 | Silt loam | Clay loam | 29.31 |
| M1-8 | R | nd | nd | nd | nd | nd | nd |

Table 3. (Continued)

| Sample | Horizon | Sand % | Silt % | Clay % | Textural Class | Field Texture | 15-Bar Moisture % |
|---|---|---|---|---|---|---|---|
| McKenzie River | | | | | | | |
| M2-1 | A11 | 32.47 | 38.61 | 28.93 | Clay loam | Loam | nd |
| M2-4 | $B21_t$ | 16.61 | 40.49 | 42.91 | Clay | Silty clay loam | nd |
| M2-6 | B3 | 16.27 | 39.94 | 43.80 | Clay | Silty clay loam | nd |
| M3-2 | A12 | 20.53 | 45.62 | 33.86 | Clay loam | Loam | nd |
| M3-4 | $B21_t$ | 10.10 | 49.66 | 40.24 | Clay | Silty clay loam | nd |
| M3-6 | R | 20.68 | 50.61 | 28.71 | Silt loam | nd | nd |

Fifteen-bar moisture bore no particular relationship to clay content but showed a consistent increase with depth except in the transition from the II C horizon to the III R horizon of Budworm soil profile number 1 where it dropped from 21.46 percent to 17.10 percent.

## Chemical Properties

Free iron oxide content of the whole soil increased with depth from the A horizons to the B2 horizons of McKenzie River and buff tuff soils and decreased regularly with depth in Slipout and Budworm soils (Table 4). McKenzie River soils were much higher in free iron than either Slipout or Budworm soils while the buff tuff soils had intermediate values.

Cation exchange capacities fluctuated slightly, but with the exception of buff tuff soil profile number 2, increased slightly with depth. Many cation exchange capacity values were not reasonable when compared with clay percentages. For example, the III R horizon of Budworm soil profile number 1 contained 14.51 percent clay and had a cation exchange capacity of 33.84 meq/100 gms of soil. The R horizon of Budworm soil profile number 3 contained 9.05 percent clay and had a cation exchange capacity of 48.16 meq/100 gms of soil. This would imply that the clay samples in these two horizons had cation exchange capacities of 233.22 meq/100 gms and

T 4. Selected chemical properties of four Western Cascade soils.

| Sample | Horizon | pH | Organic matter % | C. E. C. Meq/100 gms | Ca | Mg | K | Na | Base saturation % | Free iron oxides % $Fe_2O_3$ |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Meq/100 gms | | | | | |
| Slipout | | | | | | | | | | |
| S1-1 | A11 | 5.6 | 6.53 | 40.17 | 7.75 | 2.04 | 0.70 | 0.20 | 26.61 | 2.69 |
| S1-2 | A12 | 5.3 | 1.55 | 44.69 | 4.90 | 2.68 | 0.38 | 0.14 | 18.12 | 2.00 |
| S1-3 | B21 | 4.9 | 0.77 | 42.76 | 2.57 | 2.10 | 0.42 | 0.13 | 12.21 | 1.74 |
| S1-4 | B22 | 4.8 | 0.64 | 40.21 | 1.87 | 1.76 | 0.40 | 0.11 | 10.30 | 1.23 |
| S1-5 | B3 | 5.0 | 0.26 | 36.51 | 3.06 | 2.06 | 0.38 | 0.13 | 15.42 | 1.81 |
| S1-6 | C | 5.0 | 0.41 | 42.46 | 4.47 | 2.58 | 0.49 | 0.16 | 18.13 | 1.31 |
| S2-1 | A11 | 5.7 | 4.03 | 32.89 | 14.49 | 3.67 | 0.97 | 0.13 | 58.56 | 1.37 |
| S2-3 | $IIB21_t$ | 5.7 | 0.44 | 25.69 | 11.70 | 3.01 | 0.88 | 0.17 | 61.35 | 1.44 |
| S2-5 | IIIB3 | 5.5 | 0.25 | 39.54 | 19.04 | 4.94 | 0.65 | 0.31 | 63.08 | 1.86 |
| S3-1 | A11 | 5.4 | 6.85 | 41.46 | 7.00 | 2.48 | 0.86 | 0.13 | 25.25 | 3.16 |
| S3-3 | B21 | 5.5 | 0.73 | 40.74 | 14.78 | 7.42 | 0.56 | 0.14 | 56.91 | 2.57 |
| S3-7 | C | 5.2 | 0.24 | 50.23 | 21.08 | 10.01 | 0.50 | 0.22 | 63.33 | 0.92 |
| Budworm | | | | | | | | | | |
| B1-1 | A11 | 5.3 | 3.41 | 33.76 | 8.04 | 2.94 | 0.73 | 0.16 | 35.16 | 2.01 |
| B1-2 | A12 | 5.6 | 1.50 | 33.73 | 12.81 | 4.18 | 0.75 | 0.17 | 53.10 | 2.04 |
| B1-3 | $B21_t$ | 5.4 | 1.09 | 41.57 | 16.24 | 7.15 | 0.87 | 0.21 | 58.86 | 1.86 |
| B1-4 | $B22_t$ | 5.4 | 1.08 | 40.60 | 14.97 | 6.82 | 0.96 | 0.22 | 56.58 | 1.99 |
| B1-5 | B3 | 5.4 | 0.76 | 44.88 | 15.82 | 7.44 | 0.99 | 0.21 | 54.50 | 1.37 |
| B1-6 | IIC | 5.2 | 1.10 | 44.72 | 16.23 | 7.62 | 0.98 | 0.29 | 56.17 | 0.85 |
| B1-7 | IIIR | nd | 0.08 | 33.84 | 22.42 | 6.40 | 8.89 | nd | 111.44 | 0.87 |
| B2-1 | A11 | 5.5 | 5.47 | 43.88 | 16.67 | 4.24 | 1.24 | 0.15 | 50.82 | 2.16 |
| B2-3 | $B2_t$ | 5.3 | 2.05 | 55.01 | 16.98 | 7.50 | 0.93 | 0.19 | 46.54 | 1.86 |
| B2-5 | B32 | 5.3 | 0.55 | 64.83 | 30.63 | 9.68 | 1.60 | 0.35 | 65.19 | 1.01 |
| B2-6 | R | nd | 0.24 | 61.10 | 31.72 | 10.01 | 1.52 | 0.58 | 71.73 | 0.91 |

Table 4. (Continued)

| Sample | Horizon | pH | Organic matter % | C.E.C. Meq/100 gms | Ca | Mg | K | Na | Base saturation % | Free iron oxides % $Fe_2O_3$ |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Meq/100 gms | | | | | |
| Budworm | | | | | | | | | | |
| B3-2 | A12 | 5.3 | 1.58 | 52.27 | 15.89 | 8.01 | 0.72 | 0.14 | 47.37 | 1.72 |
| B3-3 | $B2_t$ | 5.2 | 0.89 | 55.51 | 12.80 | 8.30 | 0.36 | 0.20 | 39.02 | 1.56 |
| B3-5 | C | 5.1 | 0.26 | 51.98 | 14.71 | 9.58 | 0.29 | 0.21 | 47.69 | 1.43 |
| B3-6 | R | nd | 0.10 | 48.16 | 15.60 | 8.53 | 0.32 | 0.23 | 51.25 | 1.24 |
| buff tuff | | | | | | | | | | |
| P1-1 | A11 | 5.5 | 3.02 | 29.64 | 7.47 | 3.59 | 0.96 | 0.10 | 40.89 | 2.92 |
| P1-2 | A12 | 5.5 | 1.35 | 29.93 | 9.56 | 5.55 | 0,55 | 0.20 | 52.89 | 2.69 |
| P1-3 | B1 | 5.1 | 0.75 | 30.45 | 7.35 | 8.36 | 0.13 | 0.30 | 53.00 | 2.68 |
| P1-4 | $IIB21_t$ | 5.1 | 0.57 | 31.02 | 8.47 | 9.93 | 0.11 | 0.37 | 60.86 | 2.94 |
| P1-5 | $IIIB22_t$ | 5.3 | 0.31 | 36.19 | 11.11 | 14.03 | 0.03 | 0.40 | 70.65 | 2.68 |
| P1-6 | $IIIB3_t$ | 5.4 | 0.25 | 39.21 | 12.53 | 15.39 | 0.03 | 0.46 | 72.46 | 2.57 |
| P2-1 | A11 | 6.0 | 5.93 | 25.08 | 5.54 | 0.72 | 0.63 | 0.08 | 27.79 | 2.34 |
| P2-4 | $B21_t$ | 5.4 | 0.36 | 10.97 | 1.25 | 0.42 | 0.21 | 0.09 | 17.96 | 2.58 |
| P2-6 | R | 5.1 | 0.15 | 8.57 | 1.66 | 0.42 | 0.21 | 0.11 | 28.00 | 2.13 |
| McKenzie River | | | | | | | | | | |
| M1-1 | A11 | 6.0 | 4.61 | 34.29 | 7.37 | 1.51 | 1.05 | 0.11 | 29.28 | 4.21 |
| M1-2 | A12 | 5.9 | 3.42 | 33.05 | 7.59 | 2.28 | 0.94 | 0.10 | 33.01 | 4.44 |
| M1-3 | A3 | 5.9 | 1.65 | 33.46 | 6.52 | 3.03 | 0.66 | 0.12 | 30.87 | 4.52 |
| M1-4 | B1 | 5.9 | 1.02 | 37.74 | 9.44 | 5.13 | 0.48 | 0.10 | 40.14 | 4.89 |
| M1-5 | $B21_t$ | 5.6 | 0.28 | 39.91 | 6.72 | 4.61 | 0.49 | 0.09 | 29.84 | 5.24 |
| M1-6 | $B22_t$ | 5.6 | 0.44 | 39.89 | 4.13 | 3.75 | 0.54 | 0.08 | 21.31 | 5.53 |
| M1-7 | B3 | 5.4 | 0.34 | 52.34 | 5.36 | 7.09 | 0.85 | 0.10 | 25.60 | 3.96 |
| M1-8 | R | nd | nd | nd | nd | nd | nd | nd | nd | nd |

Table 4. (Continued)

| Sample | Horizon | pH | Organic matter % | C. E. C. Meq 100 gms | Ca | Mg | K | Na | Base saturation % | Free iron oxides % $Fe_2O_3$ |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Meq/100 gms | | | | | |
| McKenzie River | | | | | | | | | | |
| M2-1 | A11 | 5.5 | 7.14 | 34.17 | 5.29 | 0.97 | 0.89 | 0.15 | 21.36 | 3.96 |
| M2-4 | $B21_t$ | 5.7 | 0.65 | 38.67 | 5.67 | 3.98 | 1.08 | 0.14 | 21.11 | 4.34 |
| M2-6 | B3 | 5.6 | 0.50 | 39.27 | 5.43 | 5.04 | 0.41 | 0.15 | 28.09 | 4.50 |
| M3-2 | A12 | 5.4 | 4.75 | 30.21 | 4.60 | 1.58 | 0.72 | 0.10 | 23.17 | 5.93 |
| M3-4 | $B21_t$ | 5.3 | 1.01 | 32.80 | 4.04 | 3.04 | 0.72 | 0.12 | 24.15 | 6.13 |
| M3-6 | R | nd | 0.37 | 32.56 | 3.69 | 3.41 | 0.48 | 0.27 | 24.11 | 6.02 |

532.15 meq/100 gms respectively.

Base saturation of Slipout soil profile number 1 decreased with depth. This was accompanied by a parallel decrease in pH. The remaining soil profiles showed a slight tendency for base saturation to increase with depth without a corresponding increase in pH. Calcium and magnesium were the dominant exchangeable bases.

Organic matter showed a regular decrease with depth.

Citrate-dithionite and sodium hydroxide extracts of silt fractions contained mostly silicon and aluminum. Iron was present in much smaller amounts (Table 5). More silicon and aluminum were present in extracts from Slipout and McKenzie River soils than buff tuff and Budworm soils. Extracts from McKenzie River soil profile number 1 contained considerably more iron than the three other soils indicating that McKenzie River soils are more strongly weathered.

## Some Physical and Chemical Relationships

The major part of the coarse-silt fraction of most of the soils under investigation consisted of pseudomorphs of clay. This was also true of the fine silt and much of the sand. Comparison of measured clay percentage, cation exchange capacity, and 15-bar moisture is often helpful in detecting incomplete dispersibility such as this.

Plots of clay, cation exchange capacity, and 15-bar moisture for buff tuff soil profile number 1 parallel each other (Figure 20b).

Table 5. Elemental analysis of citrate-dithionite and boiling sodium hydroxide extracts of silt reactions of five selected soil profiles.

| Sample | Horizon | Elemental Analysis (O. D. Weight Basis) $SiO_2$ % | $Al_2O_3$ % | $Fe_2O_3$ % | Sample | Horizon | Elemental Analysis (O. D. Weight Basis) $SiO_2$ % | $Al_2O_3$ % | $Fe_2O_3$ % |
|---|---|---|---|---|---|---|---|---|---|
| Slipout | | | | | buff tuff | | | | |
| S1-1 | A11 | 12.48 | 9.07 | 1.79 | P1-1 | A11 | 10.06 | 7.45 | 1.79 |
| S1-2 | A12 | 12.88 | 10.13 | 1.14 | P1-2 | A12 | 10.19 | 8.09 | 1.79 |
| S1-3 | B21 | 13.27 | 9.90 | 0.89 | P1-3 | B1 | 9.72 | 7.45 | 1.40 |
| S1-4 | B22 | 14.77 | 12.10 | 0.73 | P1-4 | $IIB21_t$ | 10.19 | 8.01 | 1.32 |
| S1-5 | B3 | 15.11 | 11.30 | 1.04 | P1-5 | $IIIB22_t$ | 7.40 | 4.80 | 1.32 |
| S1-6 | C | 15.62 | 9.98 | 1.94 | P1-6 | $IIIB3_t$ | 7.06 | 4.57 | 1.04 |
| Budworm | | | | | McKenzie River | | | | |
| B1-1 | A11 | 7.70 | 4.08 | 1.79 | M1-1 | A11 | 11.64 | 9.45 | 3.89 |
| B1-2 | A12 | 7.49 | 4.65 | 1.79 | M1-2 | A12 | 11.34 | 9.45 | 3.68 |
| B1-3 | $B21_t$ | 6.21 | 3.40 | 2.11 | M1-3 | A3 | 11.64 | 10.17 | 3.50 |
| B1-4 | $B22_t$ | 7.66 | 4.91 | 1.94 | M1-4 | B1 | 11.34 | 9.07 | 3.14 |
| B1-5 | B3 | 10.06 | 6.43 | 1.49 | M1-5 | $B21_t$ | 12.11 | 9.37 | 3.38 |
| B1-6 | IIC | 12.75 | 7.79 | 0.43 | M1-6 | $B22_t$ | 12.48 | 9.56 | 3.50 |
| B1-7 | IIIR | 15.19 | 5.07 | 0.43 | M1-7 | B3 | 12.48 | 8.43 | 1.63 |

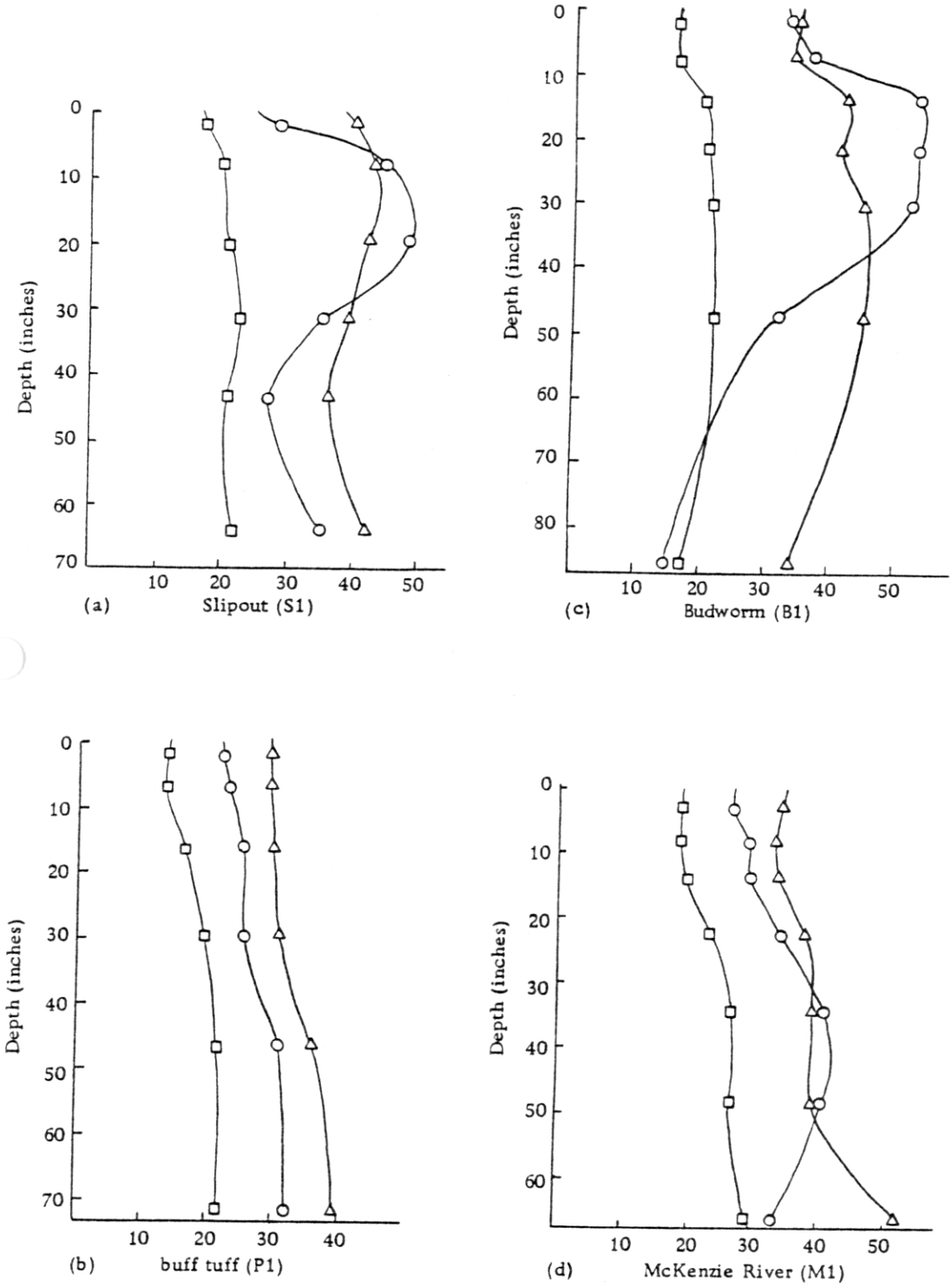


Figure 20. 15-atmosphere water (□—□%), cation exchange capacity (△—△meq/100 gms), and dispersed clay (○—○%) as a function of depth in four Western Cascade soils.

Similar plots of clay and cation exchange capacity for Slipout, Budworm, and McKenzie River soils did not parallel each other but plots of cation exchange capacity and 15-bar moisture were almost parallel (Figure 20a, c, d).

Table 6 shows the relationship betweem measured clay, total clay, and cation exchange capacity. Total clay is the higher of either measurable clay, or 2.5 times the 15-bar moisture percentage, if one-half or more of the control section has a 15-bar moisture to clay ratio of 0.6 or more (U. S. Soil Conservation Service, 1967).

Table 6. Relationship between dispersible clay, total clay, and cation exchange capacity.

| Sample | Horizon | C. E. C. Meq/100 gms | Dispersible clay % | Total clay [1] % | Clay / Total clay | Sample | Horizon | C. E. C. Meq/100 gms | Dispersible clay % | Total clay [1] % | Clay / Total clay |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Slipout | | | | | | buff tuff | | | | | |
| S1-1 | A11 | 40.17 | 28.93 | 45.28 | 0.64 | P1-1 | A11 | 29.64 | 22.22 | 34.75 | 0.64 |
| S1-2 | A12 | 44.69 | 45.02 | 45.02 | 1.00 | P1-2 | A12 | 29.93 | 23.61 | 34.98 | 0.68 |
| S1-3 | B21 | 42.76 | 48.48 | 48.48 | 1.00 | P1-3 | B1 | 30.45 | 25.53 | 41.43 | 0.62 |
| S1-4 | B22 | 40.21 | 36.99 | 58.88 | 0.63 | P1-4 | $IIB21_t$ | 31.02 | 25.43 | 48.63 | 0.52 |
| S1-5 | B3 | 36.51 | 28.00 | 54.33 | 0.52 | P1-5 | $IIIB22_t$ | 36.19 | 31.21 | 53.85 | 0.59 |
| S1-6 | C | 42.46 | 35.85 | 56.13 | 0.64 | P1-6 | $IIIB3_t$ | 29.21 | 32.08 | 54.05 | 0.59 |
| Budworm | | | | | | McKenzie River | | | | | |
| B1-1 | A11 | 33.76 | 33.05 | 33.05 | 1.00 | M1-1 | A11 | 34.29 | 26.30 | 46.35 | 0.57 |
| B1-2 | A12 | 33.73 | 35.82 | 35.82 | 1.00 | M1-2 | A12 | 33.05 | 29.28 | 46.20 | 0.63 |
| B1-3 | $B21_t$ | 41.57 | 52.98 | 52.98 | 1.00 | M1-3 | A3 | 33.46 | 29.35 | 48.73 | 0.60 |
| B1-4 | $B22_t$ | 40.60 | 52.85 | 52.85 | 1.00 | M1-4 | B1 | 37.74 | 34.31 | 58.08 | 0.59 |
| B1-5 | B3 | 44.88 | 51.90 | 51.90 | 1.00 | M1-5 | $B21_t$ | 39.91 | 40.65 | 66.83 | 0.61 |
| B1-6 | IIC | 44.72 | 31.57 | 53.65 | 0.59 | M1-6 | $B22_t$ | 39.89 | 41.90 | 66.00 | 0.63 |
| B1-7 | IIIR | 33.84 | 14.51 | 42.75 | 0.34 | M1-7 | B3 | 52.34 | 34.02 | 73.28 | 0.46 |

[1] According to method of U. S. Soil Conservation Service

# DISCUSSION

## Genetic Relationships

### Electron Micrographs

Kaolinite and halloysite are common constituents of clay fractions of many Ultisols and Alfisols in Oregon. Both minerals have essentially the same chemical composition and structure but differ in degree of stacking order of alumina-silica sheets. Kaolinite unit layers are held rigidly together by hydrogen bonds resulting in a 7.15 Å basal spacing. Halloysite, on the other hand, contains a layer of water molecules bonded between each unit layer resulting in a hydrated basal spacing of 10.1 Å. Dehydration yields a 7.2 Å to 7.6 Å basal spacing (Brindley, 1961).

Singleton (1966) has shown that broad moderately intense peaks with basal spacings between 7.3 Å and 7.4 Å are common in certain Haplohumults in Oregon. Similar results were obtained in the course of this study. Figures 6a and 8a illustrate the kinds of diffraction patterns obtained from ethylene glycol solvated, calcium saturated clay samples from selected horizons of two representative soil profiles. Diffraction tracings of buff tuff soil profile number 2 samples yielded sharp rather intense bands with basal spacings of 7.25 Å. Peaks were almost uniformly symmetrical but intensity

increased slightly with depth. Tracings of McKenzie River soil profile number 1 samples showed broad, rather diffuse, moderately intense peaks with basal spacings between 7.49 Å and 7.56 Å. Peaks were almost uniformly asymmetrical and intensity increased slightly with depth.

Bramao et al. (1952) showed that no soil kaolinite had basal spacings as small as those of pure kaolinite and none had well developed interference maxima. On the basis of x-ray diffraction patterns, they distinguished between kaolin minerals as follows: (1) strong rather sharp 7.25 Å basal spacings, (2) broad basal plane reflections ranging between 7.3 Å and 7.45 Å with rather low intensities, and (3) broad relatively weak basal spacings ranging from 7.25 Å to 7.45 Å. Group 1 they called kaolinite, group 2 halloysite, and group 3 an indefinite group of randomly organized kaolinite or mixture of kaolinite and halloysite not necessarily a transitional form between group 1 and group 2.

According to Brindley (1961), if either kaolinite or halloysite is present in a sample, there is no difficulty differentiating between them. However, when both minerals are present in a mixture, a problem of identification develops. Kaolinite is likely to be identified and halloysite is likely to be overlooked in a cursory x-ray examination. Brindley and Comer (1956) reported that at 1:1 mixture of kaolinite and halloysite produced a diffraction band that was

indistinguishable from the pattern produced by partially disordered kaolinite. Similarly, Brindley, Souza Santos, and Souza Santos (1963) showed that additions of at least 20 percent of well-formed kaolinite to well-formed halloysite produced an x-ray diffraction pattern comparable to poorly crystallized halloysite. Addition of 40 percent of well-formed kaolinite completely submerged the halloysite resulting in a pattern appearing like one from disordered kaolinite. Consequently, x-ray diffraction is not entirely satisfactory in distinguishing between kaolinite and halloysite. The purpose of the electron micrographs was to determine if kaolinite and/or halloysite were present in clay components which exhibited broad, rather diffuse, moderately intense 7.5 Å diffraction bands.

From the application of the electron microscope to the study of clay minerals, Bates, Hildebrand, and Swineford (1950) have shown that halloysite consists of well-developed lath-like tubes in contrast to the well-known pseudohexagonal plates of kaolinite. Both minerals, however, depart from these ideal characteristic forms. Kaolinite may occur in euhedral crystals elongated in one direction, subhedral to anhedral particles of small to medium size, groups of overlapping oriented plates, and pseudohexagonal plates with ragged edges and perforations suggestive of solution. Similarly, halloysite exists in the form of hollow tubes, many of which are split longitudinally or have collapsed to form laths or ribbons. Splitting and

fairly well-developed tubular forms of halloysite (Figure 15).

These microscopic observations showed that clay samples from these two profiles contained kaolin minerals with a variety of morphological forms from fairly well-crystallized platy forms of kaolinite to well-developed tubular forms of halloysite with a gradation of forms in between.

From electron microscope studies by Bates and Comer (1959), it appears that a morphological sequence of kaolin minerals exists from well-defined platy forms of kaolite through less well-developed platy and lath-like forms to curled and rolled forms, and eventually to fully rolled tubular forms of halloysite. Brindley, Souza Santos, and Souza Santos (1963) and Souza Santos, Souza Santos, and Brindley (1964) showed that x-ray diffraction patterns run more or less parallel with this morphological sequence. In a subsequent study, Brindley and Souza Santos (1966) distinguished four stages in this sequence grading one into the other as follows:

A. Well formed platy particles, capable of a high degree of preferential basal plane orientation, with well-ordered layer sequences.

B. Platy particles often less well developed than those in A but capable of considerable preferred orientation, with considerable layer sequence disorder mainly nb/3 displacements.

C. Partially rolled and curved forms, with much less tendency towards preferred orientation, and with greater structural disorder than in case B, but less than in D.

D. Rolled and tubular forms, with no tendency towards preferred basal plane orientation and with a highly disordered layer sequence.

They called groups A and B kaolinite and groups C and D halloysite.

Electron micrographs by Bramao et al. (1952) showed that soil kaolin minerals have less mineral perfection and smaller crystal size than kaolin minerals from naturally occurring geologic deposits. Small particle size and poor crystallinity of soil kaolins were the result of fluctuating environmental conditions. Soils are subject to cycles of alternate wetting and drying and corresponding cycles of cooling and warming. Consequently, a kaolin mineral may begin to synthesize in one part of a cycle and have its growth interrupted or stopped by another part of the cycle. The next cycle of crystal growth may occur in a warmer environment or in a more hydrous slightly different chemical environment. Where kaolinite and halloysite can crystallize undisturbed in an ideal environment, they can attain their characteristic euhedral form. However, where weathering conditions are dynamic and reflect the interplay of many variables, as in soil, perfect crystal formation would not be expected to take place.

Results showed that soil kaolins studied here were not well crystallized. X-ray diffraction patterns and electron micrographs of clay samples from buff tuff soil profile number 2 showed that in spite of this fact, intense diffraction maxima with a high degree of symmetry were obtained from relatively pure samples of either

kaolinite or halloysite. On the other hand, x-ray diffraction patterns and electron micrographs of clay samples from McKenzie River soil profile number 1 indicated that mixtures of kaolinite and halloysite resulted in broad rather diffuse, moderately intense diffraction maxima with some degree of asymmetry. These diffraction maxima were characterized by a morphological sequence of kaolin minerals from poorly crystallized platy and lath-like forms of kaolinite to platy forms with curled and rolled edges and eventually to fairly well crystallized tubular forms of halloysite.

Coarse-Silt Mineralogy

Results of the coarse-silt mineralogy showed that tuffaceous rocks underlying the four soils in this investigation were strongly altered and weathered prior to profile development. Consequently, the parent material of these soils contain few weatherable primary minerals. Fresh unweathered plagioclase, amphiboles, and pyroxenes are abundant in surface horizons as a result of recent volcanic ash falls, but these decrease sharply with depth in most profiles. The coarse-silt fractions of all profiles except buff tuff soil profile number 2 were dominated by pseudomorphs of clay. Examination of the fine silt and sand in reflected light revealed that pseudomorphs of clay also persisted in the fine silt, very fine sand, fine sand, medium sand, coarse sand, and in some cases into the very coarse sand.

Incomplete dispersibility, such as this, can usually be detected by comparing measured clay percentages with other soil properties such as cation exchange capacity, 15-atmosphere mositure, and surface area.

Plots of clay, cation exchange capacity, and 15-bar moisture (Figure 20b) for buff tuff soil profile number 1 parallel each other. This would be expected because Table 2 showed that aggregate percentages of this soil were almost constant with depth. Similar plots (Figure 20a, c, d) of clay, cation exchange capacity, and 15-bar moisture for Slipout, Budworm, and McKenzie River soils showed that clay increased more abruptly than cation exchange capacity in the B2 horizons and decreased more abruptly than cation exchange capacity below the B2 horizons. Plots of cation exchange capacity and 15-bar moisture were almost parallel for all four soils.

These data indicated that the clay-size fraction was not solely responsible for the cation exchange capacities of these soils. Sand and silt-size pseudomorphs of clay apparently participate in cation exchange reactions. Ogg and Hendrick (1920) showed that sand and silt, as well as clay, could adsorb ammonia from a solution of ammonium sulfate. Mitchell and Muir (1937) and Brown (1951) confirmed this observation when they demonstrated that sand and silt fractions of soil participated in cation exchange reactions. A series of subsequent studies by McConaghy and McAleese (1957), McAleese

and McConaghy (1958), and McAleese and Mitchell (1958a, 1958b) showed that pseudo-aggregates were responsible for a major part of the cation exchange capacity of basaltic soils in northern Ireland.

Similarly, one could expect moisture retention characteristics to be affected by these pseudomorphs of clay. Lund (1959) has shown that 15-bar moisture is related to clay content. Abrol and Khosla (1966) showed that 15-bar moisture was also related to surface area. They reported almost a linear relationship but cautioned that this could be expected to hold true only for soils of similar mineralogy.

Texture is one of the criteria used as family differentiae in the Comprehensive System of Soil Classification. Because some soils do not disperse well, clay content used is the higher of either measured clay, or 2.5 times the 15-bar moisture percentage, if one-half or more of the control section has a 15-bar moisture to clay ratio of 0.6 or more (U. S. Soil Conservation Service, 1967). Applying this method of estimating clay to Slipout, Budworm, buff tuff, and McKenzie River soils yielded total clay values for these soils as shown in Table 6. These data show that total clay and cation exchange capacity bear a good relationship to each other whereas dispersible clay does not bear much relationship to either cation exchange capacity or total clay. For example, in the McKenzie River soil profile, cation exchange capacity and total clay both increased gradually with depth down through the $B2_t$ horizon, below which both

increased rather sharply. By comparison, dispersible clay had a rapid increase with depth down through the $B2_t$ horizon followed by a rather abrupt decrease in the B3 horizon. This suggests that mechanical analysis is not always a good indicator of amount of phyllosilicates or related physico-chemical properties such as moisture holding capacity, hydraulic conductivity, and cation exchange capacity. Many statements can be made about a soil if texture is known. Such statements, however, are valid only if the soil can be separated into a reactive clay fraction and a relatively inert silt and sand fraction.

Soil scientists have known for many years that certain soils get finer textured as a result of prolonged rubbing between the thumb and forefinger in the process of determining soil texture in the field. Similarly, particle size distribution of these same soils depends on the intensity of the dispersion treatment in the laboratory. This is particularly true of soils derived from basic volcanic parent materials.

Table 4 showed that field textures and laboratory mechanical analyses did not always agree. On the basis of particle size distribution, the A3 and B1 horizons of McKenzie River soil profile number 1 are clay loam. They were called silty clay loam in the field. Similarly, the III $B22_t$ and III $B3_t$ horizons of buff tuff soil profile number 1 and the B22 and B3 horizons of Slipout soil profile number 1

were called clay loam. They were designated silty clay loam and silty clay respectively in the field. Apparently the disruptive force of rubbing these soils between the thumb and forefinger resulted in a decrease of sand and silt-size pseudomorphs of clay and a corresponding increase in the clay fraction. The B21 horizon of Slipout soil profile number 3 and the $B21_t$ horizon of McKenzie River soil profile number 3 were called clay in the laboratory. In the field, these samples were called clay loam and silty clay loam respectively. With additional rubbing, these samples would have become finer textured in the field.

Field textures of these soils depend on the amount of rubbing that the soil material is exposed to during the process of texture determination. In other words, if a soil sample were worked for a short period of time, a small amount of sand and silt-size pseudomorphs of clay would be disrupted and the apparent texture would be relatively coarse. On the other hand, if rubbing were continued for a longer period of time, more sand and silt-size pseudomorphs of clay would be separated into discrete particles and the apparent texture would be finer. Mechanical analyses can be carried out in such a manner that soil samples are exposed to a standard dispersion treatment resulting in consistent reproducible results. To obtain equally consistent results in the field, each soil sample would have to be rubbed between the thumb and forefinger for precisely the same

amount of time, at the same moisture content, and with precisely the same amount of normal pressure as each subsequent soil sample. Such precision cannot be expected in the field. However, in soils that contain incompletely dispersible material, textures obtained in the field by prolonged rubbing may be more meaningful than those obtained in the laboratory using standard dispersion methods.

In an attempt to bring about the dispersion of silt-size pseudomorphs of clay, sub-samples of the silt fraction were given a sequence of treatments consisting of a citrate-dithionite iron removal, a sonification treatment, a selective dissolution treatment in boiling sodium hydroxide, and a final sonification treatment. Table 5 shows the results of elemental analyses of these extracts for iron, aluminum, and silicon reported as oxides of these metals. There was no apparent relationship between amounts of these metals and content of pseudomorphs to clay. However, dispersion of silt aggregates occurred following the boiling sodium hydroxide dissolution and sonification treatments but not after the iron removal indicating that aggregate stability was due to aluminum and/or silicon.

Flach et al. (1969) pointed out that silica is not considered an effective cementing agent in soils of humid regions. However, silica cementation of soil would occur in humid regions if conditions were such that silica could precipitate. Volcanic rocks of the Western Cascades contain abundant feldspar, ferromagnesian minerals, and

volcanic glass. Weathering of these rocks would produce large amounts of amorphous silica. Drying would cause the soil solution to become saturated with respect to silica resulting in incipient cementation.

Deshpande, Greenland, and Quirk (1964, 1968) reported that removal of iron oxide had no effect on aggregate stability or particle size distribution. They were of the opinion that hydrous oxides of aluminum were more important to aggregate stability than oxides of iron. Similarly, Saini, MacLean, and Doyle (1966) showed that aluminum was more highly correlated with aggregation than either organic matter or iron. They concluded that more than one cementing agent was probably responsible for aggregate stability but aluminum was the most important.

Table 2 showed that the number of pseudomorphs decreased upward in the soil profiles of all four soils. This was accompanied by a corresponding increase in measured clay probably as a result of destruction of these pseudomorphs. Flach, Cady, and Nettleton (1968) suggested that disruption of pseudomorphs is brought about by alternate wetting and drying, shrinking and swelling, and mechanical disturbance of root movement and soil fauna. St. Arnaud and Whiteside (1963) likewise showed that mechanical disintegration of rocks, minerals, and rock fragments was at a maximum in surface horizons and decreased downward. They suggested that

physical breakdown occurred as a result of frost action. Duff, Webley and Scott (1963) reported that several 2-ketogluconic acid-producing bacteria actively decomposed soil, rock samples, and most common silicate minerals releasing iron, aluminum, and silicon. Henderson and Duff (1963) demonstrated that certain fungi have the ability to decompose naturally occurring silicate minerals, rocks, and mineral matter of soils resulting in release of metallic ions and silica. These processes all increase in intensity toward the surface and would result in destruction of pseudomorphs of clay as well as primary minerals, rocks, and rock fragments.

Flach, Cady, and Nettleton (1968) suggested the term pedoplasmation, the formation of soil plasma, for the process that results in the transformation from saprolite to soil B horizons. Pedoplasmation is an important soil forming process in many soils formed on strongly weathered parent materials. According to Flach, Cady, and Nettleton (1968), it may be the most important process causing formation of soil horizons from weathered rock. It has a strong effect on clay distribution and is responsible for formation of many argillic horizons.

X-ray diffraction patterns obtained during the course of this study suggested that pedoplasmation may also affect clay mineral composition. Figure 7a shows diffraction patterns of dispersible clay from the A12, $B21_t$, and R horizons of McKenzie River soil

profile number 3. Figure 7b shows diffraction patterns of clay mechanically dispersed from the silt fraction of these same horizons. On the assumption that peak intensity is roughly proportional to concentration, it is apparent that dispersible kaolin increased toward the surface whereas kaolin as pseudomorphs remained about constant. Dispersible montmorillonite was almost constant with depth but montmorillonite mechanically dispersed from pseudomorphs decreased sharply toward the surface. Diffraction patterns of similar clay samples from the A11, $B21_t$, and R horizons of buff tuff soil profile number 2 showed almost the same depth trend. Figure 8a showed that dispersible montmorillonite increased toward the surface and Figure 8b showed that montmorillonite, mechanically dispersed from pseudomorphs, decreased toward the surface. Kaolin decreased toward the surface in both samples but the decrease was sharper in the case of pseudomorphic kaolin. Mica occurred in all dispersible clay samples, but there was only a slight indication that it occurred as psuedomorphs in the B horizon. Similarly, diffraction patterns of clay samples from McKenzie River soil profile number 1, Figures 6 and 9, showed that dispersible montmorillonite and montmorillonite mechanically dispersed from pseudomorphs had an inverse depth relationship. Dispersible montmorillonite increased toward the surface and montmorillonite as pseudomorphs decreased toward the surface. Kaolin mechanically dispersed from

pseudomorphs decreased toward the surface and mica was present only in dispersible clay samples.

It is apparent that destruction of silt-size pseudomorphs of kaolin took place rather easily and dispersible kaolin occurred in almost equal amounts in all horizons. On the other hand, silt-size pseudomorphs of montmorillonite were more resistant to destruction. As a consequence, dispersible montmorillonite had a vertical distribution pattern that was related to the intensity of the disruptive forces. It was abundant in surface horizons where disruptive processes were intense but gradually decreased in amount with depth so that very little dispersible montmorillonite occurred in lower horizons where destruction of pseudomorphs was less severe. In these lower horizons, the major part of the total montmorillonite was contained in stable silt-size pseudomorphs of clay.

## Pedogenic Formation of Montmorillonite

Clay mineral composition of all three McKenzie River soil profiles is quite similar. Consequently, discussion of pedogenic clay transformations is based on one soil profile of the McKenzie River soil series. Similarly, only one representative soil profile of Budworm, Slipout, and buff tuff soils is discussed in detail.

Montmorillonite is the predominant crystalline clay mineral present in clay samples from profiles of Budworm and Slipout soils.

Diffraction maxima of clay samples from Budworm soil profile number 1 (Figure 4) are not nearly as sharp and intense as corresponding diffraction maxima of clay samples from Slipout soil profile number 1 (Figure 3). Table 3 shows that Budworm soil profile number 1 contains more clay than Slipout soil profile number 1, but comparison of diffraction patterns from these two soil profiles indicate that montmorillonite in the latter apparently has better crystallinity.

Peck et al. (1964) suggested that alteration of plagioclase to zeolites took place shortly after deposition as a result of hydrothermal solutions circulating through nonwelded tuffs. The clay fraction of one of these semi-consolidated tuffs is essentially all heulandite-clinoptilolite (Figure 4, III R horizon). Figure 4 also shows that the transition from semi-consolidated tuff to soil, much of the heulandite-clinoptilolite readily altered to montmorillonite.

Mason and Sand (1960) reported the intimate association of clinoptilolite and montmorillonite in beds of clay in Patagonia. Bramlette and Posnjak (1933) showed that vitric pyroclastic rocks altered to montmorillonite with clinoptilolite as an intermediate stage in the alteration process. Ames, Sand, and Goldich (1958) reported that tuff and volcanic ash, under the influence of hot spring water, altered to clinoptilolite and ultimately to hectorite or montmorillonite depending on the mineral composition of the water. Kerr and Cameron (1936) reported that volcanic ash was altered to montmorillonite with

heulandite as an intermediate alteration product.

Heulandite-clinoptilolite (Deer, Howie, and Zussman, 1963) is a group of zeolites with structures made up of a silica-alumina framework that is considered a hydrated equivalent of the feldspars. The net negative charge in the crystal lattice is balanced by cations, generally calcium, sodium, and potassium. The framework contains channels and relatively large interconnecting cavities in which cations and water molecules are loosely bound. These can be removed or replaced quite easily, usually without destroying the framework (Hay, 1966). Structure and composition relationships of the heulandite-clinoptilolite group are a matter of debate. Mumpton (1960) defined clinoptilolite as a high-silica member with a silica-alumina ratio of 4.25 to 5.0. He defined heulandite with a corresponding silica-alumina ratio ratio of 2.75 to 3.25. Mason and Sand (1960) maintained that the difference was that calcium predominated in heulandite and sodium and potassium predominated in clinoptilolite. Heulandite treated with potassium chloride had thermal stability of clinoptilolite and clinoptilolite treated with calcium chloride behaved like heulandite after heating (Hay, 1966). As part of the standard characterization treatments in this study, clay specimens were treated with potassium chloride prior to heat treatment. Heulandite-clinoptilolite in all clay samples except M3-6 was thermally stable. Obviously more work needs to be done before heulandite can be

distinguished from clinoptilolite.

In any event, these zeolites have silica-alumina ratios high enough and contain sufficient bases to produce environments that would lead to the formation of montmorillonite. In addition, sufficient weatherable ferromagnesian minerals occurred in surface horizons of these soils to produce solutions relatively rich in magnesium, iron, and calcium. In the wet part of the year, solutions of these cations might be too dilute to form montmorillonite, but as Keller (1964) points out, during the dry season, when potential evaporation exceeds precipitation, such solutions would become saturated with respect to magnesium, calcium, iron, sodium, and potassium and their combination with silica-alumina colloids would result in formation of montmorillonite. In view of the fact that Slipout soils are somewhat poorly drained and Budworm soils are moderately well drained, formation of montmorillonite could also be explained on the basis of impedded drainage. Downward movement of percolating water would be restricted. This would tent to concentrate metallic cations and produce a chemical system that would also favor montmorillonite. Barshad (1959) maintained that clay formation was more pronounced in poorly drained soils than it was in well drained soils. This would account for the fact that montmorillonite had better crystallinity in somewhat poorly drained Slipout soils than it had in moderately well drained Budworm soils.

The various aspects of the weathering process that resulted in the formation of montmorillonite can perhaps be represented by a generalized equation in the form:

(plagioclase + $H_2O$) → (heulandite-clinoptilolite + ions and amorphous intermediate products) → (pseudomorphs of montmorillonite + soluble salts) → (montmorillonite)

## Chloritization

Recently, numerous workers have shown that chloritic intergrades can be formed in the laboratory. Hsu and Rich (1960), working with the synthetic cation exchanger Dowex-50, showed that hydroxy-aluminum ions, with an Al-OH ratio of 1:2, became non-exchangeable on the resin and reduced its cation exchange capacity. Rich (1960) formed synthetic chloritic intergrades by treating vermiculite with aluminum chloride and adjusting the pH of the clay suspension to 4.5 with sodium hydroxide. Similarly, Sawhney (1960a) produced aluminum interlayers in montmorillonite as well as in vermiculite using aluminum chloride and adjusting the pH to 5.0 and aging for 15 days. Rich and Obenshain (1955) showed that easily collapsed dioctahedral vermiculite was made difficult to collapse by repeated saturation with aluminum followed by drying. Adjustment of the clay suspension to pH 7.0 increased resistance to collapse. Slaughter and Milne (1960) formed chlorite-like structures

from montmorillonite by thoroughly mixing a clay suspension with freshly precipitated hydroxides of either aluminum or magnesium followed by aging or dehydration. Carstea (1965, 1967, 1968), using a method similar to Slaughter and Milne (1960), was able to produce intergrades that closely resembled chlorite by adjusting the pH to 4.6 and aging for ten days. Fairly stable chloritic intergrades were also formed at a pH of 6.2

In view of the fact that chlorite-like structures form so easily from montmorillonite and vermiculite under laboratory conditions, it is not surprising that chloritic intergrades are common in nature. Soil clays showing intergradient properties occur typically in strongly acid Red-Yellow Podzolic soils in Virginia (Rich and Obenshain, 1955), strongly acid Gray Brown Podzolic soils in Indiana (Klages and White, 1957), several strongly acid soils in Connecticut (Sawhney, 1960b), medium acid Podzolic soils in Alberta (Pawluk, 1963), surface soils in Pennsylvania (Jeffries, Rolfe, and Kunze, 1953), and certain very strongly acid soils of Oregon (Singleton, 1966).

Soil reaction is one of the most important factors controlling hydroxy interlayer formation because charged hydroxy polymers occur in soils only under acid conditions. Weathering intensity is also an important factor. Jeffries, Rolfe, and Kunze (1953), Hathaway (1955), Rich and Obenshain (1955), and Singleton (1966) reported that chlorite-like minerals were more abundant in surface horizons

than in lower horizons because weathering was more intense near the soil surface. In addition, Sawhney (1960b) showed that interlayers had greater stability in well-drained soils than in poorly-drained soils. He reported that less weathering, associated with poor drainage, resulted in restricted hydroxy interlayer formation. Sawhney (1960b) also suggested that high amounts of organic matter tended to reduce interlayer formation by chelating aluminum.

The evidence indicates that intergrade minerals form in weathered soils when montmorillonite and vermiculite are present and environmental conditions are such that large amounts of aluminum and magnesium are in the weathering complex. Under these conditions, charged polymers either move into interlayer positions (Hsu and Bates, 1964) or hydroxide precipitates are adsorbed onto surfaces of expanded clays (Slaughter and Milne, 1960) forming chlorite-like minerals. If complete interlayers are formed, chlorite results.

More chloritic intergrades occurred in buff tuff and McKenzie River soils than in Budworm and Slipout soils. Chlorite occurred in the A1 horizons of all three McKenzie River soil profiles, Budworm soil profile number 2, and Slipout soil profile number 1. Fresh alumino-silicate minerals (Table 2) and exchangeable magnesium (Table 4) were abundant in all these soils. Soil reactions (Table 4) were comparable, ranging from medium acid in surface soils to very strongly acid in subsoils. Organic matter contents (Table 4)

observations of clay samples from the A11, $B21_t$, and R horizons of buff tuff soil profile number 2 indicate that kaolinite is the end product of clay mineral weathering in this soil profile. Apparently alternate wetting and drying coupled with more intense weathering in surface horizons has resulted in the transformation of halloysite to kaolinite.

The statements concerning transformation of halloysite to kaolinite in this soil are valid only if there is no discontinuity in the profile between the B horizon and the underlying tuff-breccia. Such a discontinuity is suggested by the sharp increase in quartz content between the $B21_t$ horizon and the R horizon (Table 2). However, comparison of particle size distribution (Table 3), cation exchange capacity, and exchangeable bases (Table 4) do not support this possibility.

## Stability Relationships

### Water

Shearing strength of cohesive soil is usually considered to depend on cohesion and internal friction of the soil (Terzaghi and Peck, 1967). Cohesion is that part of shearing resistance that is independent of applied pressure. It is the result of coulombic forces and van der Waal forces that exist between clay particles. Internal

friction consists of ordinary sliding friction and is equal to the normal pressure multiplied by a coefficient of friction. Thus the general equation for shearing strength is

$$S = C + (P-\mu w)\ \mathrm{TAN}\ \phi' \qquad (1)$$

where C is the cohesion, P is the normal pressure, $\mu w$ is the pore water pressure and $\phi'$ is the effective angle of internal friction.

Shear strength of cohesive soil depends on several factors, the most important of which is water. For purposes of this discussion, water can be placed in three categories as follows: (1) adsorbed water, (2) capillary water, and (3) free water.

There is a considerable body of opinion that contends that adsorbed water has some sort of definite rigid configuration but there is essentially no agreement as to the precise nature of this configuration (Grim, 1953). Hendricks and Jeffeson (1938) suggested that water was adsorbed on silicate clays as a layer of water molecules in the form of a hexagonal net bonded to the oxygen or hydroxals in the basal surfaces of the clay minerals. Stability of this configruation arose from the attraction between hydrogen atoms of the water molecules and neighboring oxygen atoms of silicate layers or oxygen atoms of water molecules in the net. Presence of the first molecular layer favored formation of a second layer and water structure was thus propagated away from the surface of the clay mineral.

According to this scheme, no hydrogens were available for bonding together the outer layer of water molecules from two adjacent clay particles. Consequently, a plane of relative weakness would exist at the junction of these two water envelopes. Hendricks, Nelson, and Alexander (1940) supported this view and presented endothermic peaks and corresponding basal spacings of montmorillonite as evidence of cation hydration and formation of a monolayer of water molecules in a hexagonal net. Mackenzie (1950) and Walker (1949) took exception to this view because it did not take into account the effect of adsorbed cations. They suggested that a definite space relationship existed because of the position of cations in the interlayer position and the frequency of their distribution. Cations in montmorillonite occupy spheres of vibration limited by the interlayer distance and as such probably interfere with formation of any regular rigid configuration of water molecules. In order to fit into any regular hexagonal net of water molecules, interlayer cations would have to be displaced from their exchange sites. Mackenzie (1950) felt that water sorption, at low moisture contents, depended primarily on exchangeable cations. Macey (1942) suggested that water molecules should be able to build up a structure capable of fitting onto planes of the layer lattice in such a way that certain dimensions were common to the lattice and the molecular layer of water. He pointed out that if the atomic structure of ice were superimposed on the Si-O

plane which forms the surface of clay minerals, certain oxygen atoms of the ice structure coincide exactly with oxygen atoms of the silicate lattice. Therefore, initially adsorbed water could have the structure of ice. More recently Barshad (1949) suggested another concept for the nature of adsorbed water. At low moisture contents, the monomolecular layer of water was discontinuous. Water molecules tend to form tetrahedra with the bases of linked silica tetrahedra. At higher moisture contents, water molecules seemed to form more closely packed layers consisting of hexagonal rings similar to the hexagonal rings of linked silica tetrahedra. At even higher states of hydration, centers of hexagonal rings of water molecules and centers of hexagonal oxygen rings of linked silica tetrahedra, not occupied with exchangeable cations, were filled with water molecules.

Because of large specific surface and high charge density of some clays, they fix water initially by polar adsorption. Unsatisfied charges within the mineral structure give rise to a field of potential energy that acts on hydrated ions in the aqueous medium surrounding the particles and on water molecules themselves because of their polar nature. Water molecules are oriented toward the solid surface and are immobilized to a certain degree. Opposed to this binding force is the kinetic energy of the cations and molecules of water. Water held by this electric field is quite strongly bound and does not possess the same physical properties as free water. It behaves more

or less like a rigid solid. Binding energy is greatest at the solid surface and decreases with increasing distance as an exponential function. This means that binding energy decreases with increasing thickness of water film around the particle. Ultimate thickness of the film depends upon water content of the system and total internal surface. Kinetic energy, of cations and water molecules, is at a maximum at an infinite distance from the solid surface and decreases with decreasing distance from the surface. Because of the kinetic energy of water molecules, adsorbed water has a rigid configuration for only a limited distance from the solid surface. Grim and Culbert (1945) and Ruiz (1962) conclude that this film is about 25 Å thick, which is equivalent to about ten molecular layers of water. As more water is adsorbed, film thicknesses in the order of 40 Å result, involving only partially immobilized water molecules. These films act as semirigid or plastic bodies and flow or deform under relatively small stress.

Grim and Culbert (1945) have shown that strength and other properties of molding sand and clay systems was due to the nonliquid nature of adsorbed water. Similarly, Grim (1948b) and Williamson (1948) have shown that this thin water film was responsible for the plastic properties of clays. Plasticity developed when just enough water was added to clay to satisfy the requirements of all available surfaces for water with a definite configuration plus a little more

water which developed little or no configuration of water molecules (Grim, 1953). If additional water were added, characteristic stickiness developed. When a certain critical film thickness was exceeded, water began to take on liquid properties and clay tended to lose some of its internal cohesion.

Capillary water can be considered liquid water that occurs in soil masses as a result of tension or negative pressure. It exists as continuous and tightly stretched films around soil particles and in pores. Forces binding these films to soil arise from surface tension of water and curvature of these water surfaces. Tension differences are directly related to meniscus curvature and water movement always takes place in response to unbalanced tensions. At equilibrium, the meniscus curvature is directly related to the pore pressure of the water as follows:

$$\mu w = -\frac{2\ Ts}{R} \qquad (2)$$

where $\mu w$ is the pore water pressure, Ts is the surface tension, and R is the radius of curvature of the meniscus. Ordinarily tension is higher in dry soils since, at lower water content, the remaining water is in smaller pores where the meniscus curvature, R, is smaller. However, moisture may flow from a relatively dry sandy soil into a wetter clay soil because clay soils usually have much smaller pores than sandy soils.

Total normal stress P at any point on a section through a saturated soil consists of the effective stress $\overline{P}$ and the neutral stress or pore water pressure $\mu w$. The equation $P = \overline{P} + \mu w$ is one of the most important in soil mechanics (Terzaghi and Peck, 1967). As an initially saturated soil dries and becomes partially saturated, tension increases in the pore water from zero to $-\mu w$. Tension increases with decreasing water content but total normal stress on a given section through the soil remains practically unaltered. Surface tension simultaneously produces an effective capillary pressure $Pk = -\mu w$, the negative value of the pore water pressure. As capillary pressure increases, shearing resistance of the soil increases following the equation:

$$S = C + (P-\mu w)\ TAN\phi' \qquad (1)$$

and with $\mu w = -Pk$,

$$S = C + (P-[-Pk])\ TAN\phi' \qquad (3)$$

and shear strength ultimately becomes

$$S = C + (P+Pk)\ TAN\phi' \qquad (4)$$

where C is the cohesion, P is the normal pressure, Pk is the capillary pressure, and $\phi'$ is the effective angle of internal friction. Tension, therefore, produces contact pressures in the soil which in turn produce frictional resistance to shearing.

Free water is that part of liquid water that moves downward in soils because of the force of gravity. Friction and viscous drag along sidewalls of pores and against mineral grains tends to retard its movement. It occurs temporarily in soils, following periods of heavy rainfall, but is responsible for most water-caused slope failures.

Slope failures occur most commonly during periods of heavy rainfall. This has led some to associate a decrease in shear strength with the lubricating action of water. Consequently, lubrication is one of the oldest explanations of water-caused slope failures (Erskine, 1965). This explanation is unacceptable for two reasons. First of all, water in contact with common minerals such as quartz acts as an anti-lubricant and not as a lubricant. The coefficient of static friction between smooth dry quartz surfaces is 0.17 to 0.20 as compared to 0.36 to 0.41 for wet surfaces (Terzaghi, 1960). Horn and Deere (1962) showed a similar relationship to exist between wet and dry surfaces of quartz, microcline, and calcite. They did, however, point out that the coefficient of static friction for muscovite, phlogopite, biotite, and chlorite decreased with addition of water. Secondly, only an extremely thin film of lubricant is required to produce full lubrication. Further addition of lubricant has no effect on the coefficient of static friction between surfaces (Hardy and Hardy, 1919). In support of this argument, Terzaghi (1960) pointed out that in humid regions most sediments contain far more water than was needed for

according to the equation

$$S = C + (P-hw)\ TAN\phi' \qquad (5)$$

where S is shearing resistance, C is cohesion of the soil, P is pressure per unit area due to the weight of the soil and water, h is the height of the piezometric surface, w is the unit weight of water, and $\phi'$ is the effective angle of sliding friction. The greater the term hw becomes, the greater is that part of the total weight of the overburden which is carried by the water. As soon as hw becomes equal to P, the overburden floats and failure will occur when gravitationally induced stress exceeds cohesive strength.

## Clay Mineral Structures

For purposes of this study, crystalline clay minerals were placed into two contrasting groups based on properties of expansion or non-expansion. Expanding clay minerals are smectite and vermiculite and non-expanding minerals are kaolinite, halloysite, micaceous clay minerals (illite), chlorite, and chloritic intergrades.

Following is a brief summary on atomic structure of some clay minerals, a subject that has been discussed in considerable detail by Grim (1953, 1962, 1965).

Kaolinite is a two-layer silicate consisting of a single silica tetrahedral sheet and a single octahedral sheet combined in a unit by

sharing of oxygens. There are few, if any, substitutions of cations within the structure. Consequently, the unit is electrically balanced. Unit thickness is about 7.15 Å. The plane between unit layers is a cleavage plane, but cleavage is not pronounced because adjacent units are oxygen surfaces oriented to hydroxyl surfaces held tightly together by hydrogen bonds resulting in properties of non-expansion. Well-crystallized kaolinite is composed of flakes with hexagonal outlines. Poorly crystallized kaolinite has a less distinct hexagonal shape and displays various degrees of disorder along the b-axis.

Halloysite has a structure similar to kaolinite, but successive silicate layers are displaced more or less randomly in both the a and b horizontal directions. The highly hydrated form consists of kaolinite layers separated from each other by a single molecular layer of water. This results in a basal spacing of about 10 Å. The less hydrated form does not possess the layer of water. Consequently, the basal spacing is about 7.3 Å. Halloysite typically has a tubular form but commonly these tubes collapse, split, or unroll.

Micaceous clay minerals differ from biotite by having less substitution of aluminum for silicon, a lower charge in the lattice, less interlayer potassium, and less regularity of stacking. Also, a small amount of water may be present between silicate layers. Aluminum is the dominant octahedral cation, but magnesium and iron may also be present. Micaceous minerals may be either dioctahedral or

trioctahedral. They have a basal spacing of about 10 or 11 Å. The source of charge deficiency is largely in the silica sheet and therefore close to the surface of the unit layer. The balancing cation between unit layers is chiefly potassium. Consequently, structural unit layers are relatively fixed in position so that polar ions cannot enter between them causing expansion.

Chlorite consists of alternate mica-like and brucite-like layers combined to form a unit. Mica layers are unbalanced by isomorphous substitution of aluminum for silicon in the tetrahedral layer. Magnesium of the brucite layer is partly replaced by aluminum or ferric iron to produce the excess positive charge necessary to balance the substitution of aluminum for silicon. Cations in the octahedral layer are quite variable, but the mineral is usually trioctahedral. Thickness of the mica-brucite unit is fixed at about 14 Å and expansion does not occur.

The Smectite group of clay minerals is characterized by a three-layer silicate structure consisting of units of two silica tetrahedral sheets and a central octahedral sheet held together by shared oxygens. There may be moderate substitution of aluminum for silicon in tetrahedral positions although octahedral substitution of magnesium and iron for aluminum is more common. The unbalanced charge is satisfied by cations between silicate layers. Balancing cations are rather loosely held and are exchangeable. Succssive silicate layers

are loosely held together and water or other polar liquids may enter between them. Thickness of the interlayer varies with the adsorbed cation and amount of water in the system. Consequently, the mineral has no fixed c-axis dimension and the spacing depends on the environment.

Vermiculite has structural characteristics of both chlorite and montmorillonite. It consists of silicate units like those in mica with a negative charge deficiency due largely to tetrahedral substitution of aluminum for silicon. Interlayer balancing cations are largely hydrated magnesium. Consequently, interlayer water is present and expansion occurs. However, unlike montmorillonite, expansion of vermiculite is limited to about 14 Å in water systems.

Chloritic intergrades are layer silicates formed by precipitation of hydroxy interlayers of aluminum and iron between silicate layers of montmorillonite or vermiculite. These ions are not easily exchangeable. Properties of expansion and thermal stability are intermediate between chlorite on the one hand and montmorillonite and vermiculite on the other.

## Clay Type

Smectites were the predominant crystalline clay minerals present in the clay fraction of horizon samples from profiles of the Slipout and Budworm soils. Smectites were less abundant in clay samples from buff tuff and McKenzie River soils. Kaolin minerals dominated

the clay fraction of horizon samples from profiles of buff tuff and McKenzie River soils but were not detected in clay samples from Budworm and Slipout soils. Chlorite was present in surface horizons of Slipout, buff tuff, and McKenzie River soils. Hydroxy interlayered smectites occurred in many samples but were more common in upper horizons of the sola than they were in lower horizons of the sola. These intergradient species and chlorite were more evident in clay samples from buff tuff and McKenzie River soils than they were in clay samples from Budworm and Slipout soils. Small amounts of mica were present in the clay samples from Budworm and McKenzie River soils but occurred in only a few horizon samples from Slipout and buff tuff soils (Table 1).

These data show that Slipout and Budworm soils have clay fractions that are dominated by expanding type clays and buff tuff and McKenzie River soils have clay fractions of mixed mineralogy that are dominated by non-expanding clays. As a consequence, Slipout and Budworm soils are more prone to slope failure during periods of heavy prolonged rainfall than buff tuff and McKenzie River soils.

Shear strength of soils containing significant amounts of clay (Table 3) is determined by the shear strength of the clay in the soil. Shear strength of the clay, in turn, depends on the type of clay. Hardy (1965) and Crawford (1964) showed that expanding type clays experienced a substantial reduction in shearing resistance as a result

of internal swelling. More specifically, Gibson (1953) showed that the angle of internal friction is almost zero for montmorillonite, ten to 15 degrees for illite, and 20 degrees for kaolinite. Residual shear strength of these same samples decreased from kaolinite through illite to montmorillonite. Kenney (1967) in a later study showed that hydrous mica had more strength than kaolinite but less strength than mica. Montmorillonite had the least resistance to shear failure. Studies by Trask and Close (1958) and Langston, Trask, and Pask (1958) showed that for a given water content, illite was stronger than kaolinite, and montmorillonite was stronger than illite. Trask (1959) later went on to show that when these clays were given free access to water, as in a natural environment, weakness had an inverse relationship to cation exchange capacity. Consequently, kaolin was stronger than illite and montmorillonite was weaker than illite.

Forces of hydration, at particle surfaces, are responsible for a portion of the swelling tendency in expanding clays. These forces, however, are operative over distances of only molecular order of magnitude. Osmotic forces of repulsion, on the other hand, are effective over much greater distances (van Olphen, 1963). They are responsible for essentially all swelling that takes place in expanding clays.

Bolt and Miller (1955), Warkentin, Bolt, and Miller (1957), and Warkentin, and Schofield (1962) made theoretical calculations of

surface, Co is the ionic concentration in the bulk solution, R is the general gas constant, and T is the absolute temperature.

Engineers are concerned primarily with three specific factors in connection with swelling soils (Hardy, 1965). These are magnitude of swelling pressure, amount of volume increase that may occur, and the effect that these will have on shear strength of the soil. The conventional approach in soil mechanics is to relate shear strength to the effective stress of the soil in the following manner:

$$S = C + (P-\mu w)\ TAN\phi' \qquad (1)$$

where S is the shear strength, C is the cohesion, P is the normal pressure, $\mu w$ is the pore water pressure, and $\phi'$ is the effective angle of internal friction. The effective stress, $P - \mu w$, decreases with increasing pore water pressure. Peterson et al. (1960), Ringheim (1964), and Hardy, Brooker, and Curtis (1962) indicate that osmotic swelling pressure may have the same effect as pore water pressure in reducing effective stress and ultimately shear strength of soil. Hardy (1965), therefore, suggests that equation (1) be modified for swelling soils by including a separate term for swelling pressure as follows:

$$S = C + (P-\mu w-Ps)\ TAN\phi' \qquad (7)$$

where Ps is the magnitude of the osmotic swelling pressure.

Equation (1) would apply to all clay soils and equation (7) would apply to those soils that contain expanding type clays. Therefore, under the same environmental conditions, osmotic swelling pressure would induce shearing failure in an expanding clay but would have no effect on the shear strength of a non-expanding clay. Consequently, because Slipout and Budworm soils contained predominantly expanding type clay, they had less resistance to shear failure than buff tuff and McKenzie River soils.

## Significance of Color

The source of green color in sediments has been of interest for many years but rarely has it been defined in terms of mineralogy (Keller, 1953). It has occasionally been ascribed to chlorite or glauconite and often vaguely as due to ferrous iron. Keller (1953) studied representative green sedimentary rocks to identify the source of green color. Ferric oxide is red and commonly stains sediments red but silicates which contain ferric iron are green. He found the intensity of the green color increased with ferric iron content and not with ferrous iron content. Green-exciting vibrations result because of the bonding and position of the iron in relation to the silica tetrahedra. Coordination of iron in the silicate lattice is more important than valence in determining mineral color. Keller (1953) concluded that montmorillonite was responsible for the green color

of most non-marine sediments. Grim (1951) also showed that montmorillonite was a source of green color in sediments, and reducing conditions were not necessary. Red sediments, except in rare cases, are due to pigmentary red iron oxide or hydroxide. Strong oxidizing environments tend to oxidize any non-ferric iron added and to liberate ferrous iron from structures of iron-bearing minerals. Red sediments tend to persis because of the stable nature of ferric oxide.

## Free Iron Oxide

Table 4 shows free iron oxide content of the four Western Cascade soils under investigation. McKenzie River soils were much higher in free iron than either Slipout or Budworm soils. Buff tuff soils had intermediate values. Relative slope stability of these soils followed the same decreasing progression.

Large amounts of iron oxides in soils have long suggested their obvious correlation with associated physical properties. Consequently, iron has been recognized for a long time as a cementing agent in well-aggregated soils. Lutz (1936) showed a high positive correlation between free iron oxide and degree of aggregation and suggested that free iron acted as a cementing agent as well as a flocculating agent. Chesters, Attoe, and Allen (1957) and Kemper and Koch (1966) supported this hypothesis with similar results. The latter investigators went on to suggest, that because of similarities between iron

and aluminum, oxides of these two metals should aggregate soil particles to approximately the same extent. Experimental results, however, showed that higher contents of aluminum oxide were not necessarily associated with significantly higher aggregate stability. Kenney, Moum, and Berre (1967) conducted consolidation tests on various soil samples and demonstrated that resistance to deformation decreased in samples following treatment for iron removal. They concluded that iron compounds somehow acted as cementing agents contributing strength to the soil.

Most hardened laterites contain large amounts of iron oxide, many contain almost no aluminum oxide, but few if any contain no iron oxide (Sivarajasingham et al., 1962). For such an iron-rich network to form in soil, enrichment, recrystallization, and dehydration must occur (Alexander and Cady, 1962). Alternate wetting and drying are the main cause of movement and recrystallization. Goethite and hematite are deposited in crystalline form in channels, pores, and nodules forming a rigid hard framework. Concretions, on the other hand, do not necessarily contain more iron oxide than the surrounding soil, but their concentration near the soil surface suggest that alternate wetting and drying are essential to their formation (Flach et al., 1969).

Koth and Page (1946) showed from electron micrographs that stable soil aggregates were heavily coated with iron and aluminum

oxides. When these oxides were removed, aggregation was destroyed. Weldon and Hide (1942) suggested that both iron and aluminum oxide were responsible for aggregate formation. More recently, Greenland, Oades, and Sherwin (1968) presented evidence in the form of electron micrographs showing iron oxide present as small rounded particles and irregularly shaped aggregates primarily on edges of kaolinite flakes. There was no evidence that iron oxide occurred as a coating on clay surfaces. Deshpande, Greenland, and Quirk (1964, 1968) showed that removal of iron oxide had no effect on aggregate stability or particle size distribution and concluded that hydrous oxides of aluminum were more important as a soil cement than oxides of iron. Saini, MacLean, and Doyle (1966) found aluminum to be more highly correlated to aggregation than were iron or organic matter. They felt that more than one property was involved in aggregate formation but aluminum was the most important. Many workers in the field of clay mineralogy have also amply demonstrated that aluminum oxides effectively modify properties of clay minerals to such an extent that strength could be affected.

Silica cementation of soil is generally considered to be a feature of arid regions but could occur in humid regions if conditions were such that silica could precipitate. Volcanic rocks of the Western Cascades contain abundant feldspar, ferromagnesian minerals, and volcanic glass. Weathering of these rocks would produce large

amounts of amorphous silica. Drying could cause the soil solution to become saturated with respect to amorphous silica resulting in incipient cementation.

The fact that stability increased with iron oxide content suggested that iron may be an important cementing agent. Aluminum oxide content of these soils was not known but the foregoing discussion points out that hydrous oxides of aluminum are likely to be more effective cementing agents than oxides of iron. Silica could also be acting as a cementing agent but it is not presently considered very effective in humid regions.

Exchangeable Ions

It has been known for some time that cohesive soils change their physical properties not only as a function of clay type but also as a function of exchangeable ions (Kelley and Brown, 1925; Hissink, 1925). Vees and Winterkorn (1967) showed that shear strength of sodium montmorillonite was less than that of calcium montmorillonite, and aluminum montmorillonite had greater strength than calcium montmorillonite. Winterkorn and Moorman (1941) replaced naturally occurring hydrogen, calcium, and sodium with potassium, on the exchange complex of Putnam clay, and the angle of internal friction increased from 19 to 22 degrees. Sullivan and Graham (1940) saturated several different clay samples with a number of cations, many

of which are not common to soils, and showed that strength increased in the following progression: lithium, sodium, potassium, ammonium, rubidium, and cesium.

Adsorbed ions also have a pronounced effect on swelling properties of clays. Grim (1948a) showed that sodium montmorillonite swelled more than calcium montmorillonite. Winterkorn and Baver (1934) saturated Putnam clay with potassium, hydrogen, magnesium, calcium, sodium, and lithium and found swelling to increase in that order. In another similar study, Havlicek and Kaza (1961) showed that swelling pressures in montmorillonite decreased in the following progression: sodium, iron, magnesium, aluminum, calcium, hydrogen, and potassium.

The foregoing discussion has been based on studies in which the investigators have dealt with monoionic systems. Such is not the case with natural soils. Table 4 shows that exchangeable hydrogen and aluminum dominate the exchange complex in one Slipout profile, one profile derived from buff tuff, and all three McKenzie River profiles. On the basis of exchangeable ions, these soils containing mostly hydrogen and aluminum, should have the greatest shear strength. However, this is not the case. McKenzie River and buff tuff soils are relatively stable whereas Slipout soils are prone to slope failure. The remaining soil profiles have their exchange complexes dominated by calcium, magnesium, sodium, and potassium. Generally calcium

is the most abundant, followed by magnesium, and small amounts of potassium and sodium. The high potassium value for B1-7 sample seems to be an artifact. Three Budworm soil profiles, two Slipout soil profiles, and one buff tuff soil profile had about the same percent base saturation and contained similar kinds and amounts of exchangeable bases on their exchange complex. In spite of this, these soils showed very little similarity in their resistance to slope failure. Budworm and Slipout soils were by far the most susceptible to mass soil movements.

These data would seem to indicate that there is no relationship between exchangeable ions and stability. Natural soils are not monoionic. They are complex colloidal systems containing several adsorbed species, each of which independently cause physical properties of soils to change. This would tend to cause soils to have ionic-dependent strength properties intermediate between sodium-saturated soils on one hand and hydrogen or potassium-saturated soils on the other. In view of this fact, no precise correlation is to be expected between exchangeable ions and stability in natural soils.

## Pseudomorphs

Many soils contain incompletely dispersible aggregates and pseudomorphs. Because these materials do not disperse easily, the U. S. Soil Conservation Service (1967) estimates clay content by

using the higher of either measured clay, or 2.5 times the 15-bar moisture percentage, if one-half or more of the control section has a 15-bar moisture to clay ratio of 0.6 or more. Values for total clay, calculated by this method, and dispersible clay appear in Table 6. Ratios of these two values showed that a greater percentage of the total clay was dispersible in Budworm and Slipout soils than in buff tuff and McKenzie River soils. Of the total clay in McKenzie River soil, 58 percent occurred as dispersible clay. Corresponding values for buff tuff, Budworm, and Slipout soils were 61 percent, 93 percent, and 74 percent respectively.

In other words, more of the total clay occurred as silt and sand-size pseudomorphs of clay in buff tuff and McKenzie River soils than in Slipout and Budworm soils. These pseudomorphs would function as individual primary particles in the soil framework. This would tend to increase the solid-to-solid contact forming a more continuous stress-resistant framework. As a consequence, buff tuff and McKenzie River soils have greater resistance to shear failure than Slipout and Budworm soils.

pseudomorphs of clay that were chemically reactive in cation exchange reactions and moisture retention. These pseudomorphs of clay resisted being separated into discrete particles by standard laboratory dispersion methods but were disrupted by prolonged rubbing between the thumb and forefinger in the process of determining soil texture in the field. Consequently field textures of these soils did not always agree with laboratory mechanical analyses. The number of pseudomorphs of clay decreased and measured clay increased upward in the soil profile from the C or R horizons into the B horizons indicating that physical and biological disruption of pseudomorphs was an important process in the formation of argillic horizons. Pseudomorphs of montmorillonite were more resistant to disruption than pseudomorphs of kaolin.

Chemical weathering in these soils resulted in the alteration of plagioclase feldspars and ferromagnesian minerals to heulandite-clinoptilolite and ultimately to pseudomorphs of montmorillonite. Subsequent disruption by physical and biological processes reduced these pseudomorphs of clay to discrete montmorillonite.

Intergrade minerals formed in well drained weathered horizons of all four soils in this study where smectites were present and aluminum was abundant. Under these conditions, charged polymers either moved into interlayer positions or hydroxide precipitates were adsorbed onto surfaces of expanded clays forming chlorite-like minerals.

clays. During periods of heavy prolonged rainfall, Slipout and Budworm soils experienced a substantial reduction in shearing resistance as a result of osmotic swelling. Consequently, Slipout and Budworm soils were more prone to slope failure than buff tuff and McKenzie River soils.

Montmorillonite was the source of the green color in the greenish tuff-breccias and pigmentary red iron oxide or hydroxide was responsible for the color of the reddish tuff-breccias.

Free iron oxide content of all four soils increased in the same direction as relative slope stability suggesting that iron may be an important cementing agent in these soils. However, the literature points out that hydrous oxides of aluminum are likely to be more effective cementing agents than oxides of iron. Silica could also be acting as a cementing agent but it is not considered very effective in soils of humid regions. These comments should make it obvious to the reader of this manuscript that our knowledge of cementing agents in soils is very incomplete. Further investigation is very much needed.

There was no apparent relationship between exchangeable ions and relative stability. Natural soils are not monoionic. They are complex colloidal systems containing several adsorbed species, each of which independently causes physical properties of soils to change. Collectively these ionic species cause soils to have ionic-dependent strength properties intermediate between sodium-saturated soils on

the one hand and hydrogen or potassium-saturated soils on the other.

Ratios of measured clay to total clay showed that more of the total clay occurred as silt and sand-size pseudomorphs of clay in McKenzie River and buff tuff soils than in Slipout and Budworm soils. These pseudomorphs functioned as individual primary particles increasing the solid-to-solid contact in the soil framework. As a consequence, McKenzie River and buff tuff soils had additional resistance to shear failure.

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# APPENDIX

## REPRESENTATIVE SOIL PROFILE DESCRIPTIONS

### McKenzie River soil profile number 1

| Soil Profile | | |
|---|---|---|
| 01 | 3-0" | Decomposed leaves, twigs, and needles |
| A11 | 0-6" | Dark brown (7.5YR3/3) very gravelly loam, brown (7.5YR5/3) dry; strong medium, fine and very fine granular structure; slightly hard, friable, slightly sticky, slightly plastic; many roots; many medium and fine interstitial pores; 60% fine gravel and spherical shot; medium acid (pH 6.0); gradual smooth boundary. |
| A12 | 6-10" | Dark brown (7.5YR3/3) very gravelly loam, brown (7.5YR5/3) dry; strong medium, fine and very fine granular structure; slightly hard, friable, slightly sticky, slightly plastic; many roots; many fine and medium interstitial pores; 50% fine gravel and spherical shot; medium acid (pH 5.9); gradual smooth boundary. |
| A3 | 10-17" | Dark brown (7.5YR3/4) silty clay loam, brown 7.5YR5/4) dry; moderate medium and fine subangular blocky structure; slightly hard, friable, slightly sticky, plastic; many roots; many fine and medium interstitial pores; medium acid (pH 5.9); gradual smooth boundary. |
| B1 | 17-27" | Dark brown (7.5YR3/4) silty clay loam, brown (7.5YR 5/4) dry; weak medium subangular blocky structure breaking into moderate very fine subangular blocky structure; slightly hard, friable, sticky, plastic; common roots; many fine and very fine tubular pores; medium acid (pH 5.9); gradual smooth boundary. |
| $B21_t$ | 27-41" | Dark brown (7.5YR4/4) silty clay, reddish yellow (7.5YR6/6) dry; weak medium subangular blocky structure breaking into moderate fine and very fine subangular blocky structure; slightly hard, friable, very sticky, plastic; common roots; many fine tubular pores; thin patchy clay films in pores and on some ped faces; medium acid (pH 5.6); gradual smooth boundary. |
| $B22_t$ | 41-55" | Dark brown (7.5YR4/4) silty clay, reddish yellow (7.5YR6/6) dry; moderate medium subangular blocky structure breaking into moderate very fine subangular blocky structure; common roots; many fine tubular pores; thin patchy clay films on vertical and horizontal ped surfaces; medium acid (pH 5.6); clear wavy boundary. |
| B3 | 55-78" | Dark brown (9YR4/4) very gravelly clay loam, reddish yellow (9YR6/6) dry; weak medium subangular blocky structure breaking into weak very fine subangular blocky structure; few roots; many fine tubular pores; moderate patchy clay films on vertical and horizontal ped surfaces; 50% gravel-sized fragments of soft tuff and breccia; strongly acid (pH 5.4). |
| R | | Fragment of tuff-breccia taken from the B3 horizon for characterization purposes. |

Soil Profile — buff tuff soil profile number 1

A11 0-4" — Dark brown (10YR3/3) gravelly loam, brown (10YR5/3) dry; moderate medium and fine granular structure; slightly hard, friable, slightly sticky, nonplastic; many roots; many interstitial pores; 25% spherical shot and 10% gravel and stone-sized fragments of andesite; strongly acid (pH 5.5); clear wavy boundary.

A12 4-10" — Dark yellowish brown (9YR4/4) silt loam, light yellowish brown (9YR6/4) dry; moderate medium and fine subangular blocky structure; many roots; many fine tubular pores; 15% gravel-sized fragments of tuff and andesite; strongly acid (pH 5.5); clear wavy boundary.

B1 10-23" — Yellowish brown (9YR5/4) heavy silt loam, light yellowish brown (10YR6/4) dry; moderate medium and fine subangular blocky structure; slightly hard, friable, slightly sticky, slightly plastic; common roots; many fine and medium pores; 10% gravel-sized fragments of tuff and andesite; strongly acid (pH 5.1); clear wavy boundary.

$IIB21_t$ 23-27" — Dark brown (7.5YR4/4) clay loam, reddish yellow (7.5YR6/6) dry; moderate medium and fine subangular blocky structure; hard, friable, sticky, plastic; common roots; many fine and medium tubular pores; thin patchy clay films in pores and on some vertical and horizontal ped faces; 10% cobble-sized fragments of tuff; strongly acid (pH 5.1); gradual wavy boundary.

$IIIB22_t$ 37-57" — Dark brown (7.5YR4/4) silty clay loam, light brown (7.5YR6/5) dry; moderate medium and fine subangular blocky structure; hard, firm, sticky, plastic; common roots; many fine and medium tubular pores; moderate patchy clay films in pores and on vertical and horizontal ped faces; occasional gravel-sized fragment of tuff; strongly acid (pH 5.3); gradual wavy boundary.

$IIIB3_t$ 57-88" — Dark brown (7.5YR4/6) silty clay loam, light brown (7.5YR6/5) dry; moderate medium subangular blocky structure; hard, firm sticky, plastic; few roots; many fine and medium tubular pores; common splotches of $MnO_2$; 15% fragments of tuff; strongly acid (pH 5.4).

R — Fragment of tuff taken from the $IIIB3_t$ horizon for characterization purposes.

Soil Profile

Budworm soil profile number 1

| | | |
|---|---|---|
| 01 | 1/2-0" | Twigs, needles, and leaves. |
| A11 | 0-5" | Very dark brown (10YR2/2) gravelly loam, gray brown (10YR5/2) dry; moderate fine and very fine granular structure; slightly hard, friable, slightly sticky, slightly plastic; many roots; many fine interstitial pores; 30% fine gravel; strongly acid (pH 5.3); clear smooth boundary. |
| A12 | 5-10" | Very dark gray borwn (10YR3/2) gravelly clay loam, gray brown (10YR5/2) dry; weak fine and very fine subangular blocky structure; hard, friable, slightly sticky, plastic; many roots; many fine interstitial pores; 25% gravel; medium acid (pH 5.6); clear smooth boundary. |
| $R21_t$ | 10-17" | Olive brown (2.5YR4/3) clay, pale brown (10YR6/3) dry; moderate medium and fine subangular blocky structure; hard, firm, sticky, plastic; common roots; common very fine tubular pores; thin patchy clay skins on vertical and horizontal ped surfaces; 10% gravel; strongly acid (pH 5.4); gradual wavy boundary. |
| $B22_t$ | 17-26" | Olive brown (2.5YR4/3) clay, pale brown (10YR6/3) dry, moderate medium and fine subangular blocky structure; hard firm, sticky, plastic; common roots; common very fine tubular pores; moderate patchy clay films on vertical and horizontal ped surfaces; 15% soft breccia fragments; strongly acid (pH 5.4); gradual smooth boundary. |
| B3 | 26-34" | Olive brown (2.5YR4/3) clay, pale brown (10YR6/3) dry; weak medium and fine subangular blocky structure; slightly hard, firm, sticky, plastic; common roots; common very fine tubular pores; 20% soft fragments of breccia; strongly acid (pH 5.4); abrupt wavy boundary. |
| IIC | 34-60" | Light olive brown (2.5YR5/3) silty clay loam, light gray (2.4YR7/3) dry; weak medium subangular blocky structure; slightly hard, firm, sticky, plastic; few roots; common fine and very fine tubular pores; 25% soft fragments of breccia; strongly acid (pH 5.2). |
| IIIR | | Sample of breccia occurring 36 inches below the bottom of the IIC horizon. |

Soil Profile

Slipout soil profile number 1

| | | |
|---|---|---|
| A11 | 0-4" | Very dark gray brown (10YR3/2) loam, gray brown (10YR5/2) dry; moderate fine granular and medium very fine subangular blocky structure; slightly hard, friable, slightly sticky, slightly plastic; many roots; many fine and very fine interstitial pores; 15% fine gravel and spherical shot; medium acid (pH 5.6); clear wavy boundary. |
| A12 | 4-12" | Very dark gray brown (2.5YR3/2) clay loam, gray brown (10YR5/2) dry; weak medium subangular blocky structure breaking into moderate very fine subangular blocky structure; hard, friable, sticky, plastic; many roots; many very fine tubular pores; strongly acid (pH 5.3); clear wavy boundary. |
| B21 | 12-27" | Olive brown (2.5YR3.5/4) silty clay, light yellowish brown (2.5YR 6/4) dry; moderate medium fine and very fine subangular blocky structure; hard, firm, sticky, plastic; common roots; many very fine and common coarse and medium tubular pores; few soft breccia fragments; very strongly acid (pH 4.9); gradual wavy boundary. |
| B22 | 27-36" | Light olive brown (2.5YR5/3) silty clay, light yellowish brown (2.5 YR6/3) dry; weak medium fine and very fine subangular blocky structure; hard, firm, sticky, plastic; common roots; many very fine tubular pores; very strongly acid (pH 4.8); clear wavy boundary. |
| B3 | 36-51" | Olive (2.5YR5/3) silty clay, pale yellow (2.5YR7/4) dry; weak medium, fine and very fine subangular blocky structure; hard, friable, sticky, plastic; few roots; many fine and very fine tubular pores; common distinct mottles; very strongly acid (pH 5.0); clear wavy boundary. |
| C | 51-78" | Pale olive (2.5YR6/3) silty clay loam, very pale brown (10YR7/5) dry; weak medium, fine and very fine subangular blocky structure; hard, friable, slightly sticky, plastic; few roots; many fine and very fine tubular pores; common distinct mottles; common breccia fragments; very strongly acid (pH 5.0). |