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Rock weathering pathways of the three major bedrock formations in the central western Cascade Mountains of Oregon are distinctive as a result of chemical and mineralogical differences in parent materials and variations in climate found along an elevation gradient. These influences are expressed in secondary mineralogy, cation retention, elemental distribution in soils and runoff waters and rates of soil development. Parent materials include welded ashflow rhyodacitic tuffs, unwelded airfall ash breccias, andesite and basalt flows and flow breccias.

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ROCK WEATHERING IN THE CENTRAL WESTERN CASCADES

by

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ABSTRACT

Rock weathering pathways of the three major bedrock formations in the central western Cascade Mountains of Oregon are distinctive as a result of chemical and mineralogical differences in parent materials and variations in climate found along an elevation gradient. These influences are expressed in secondary mineralogy, cation retention, elemental distribution in soils and runoff waters and rates of soil development. Parent materials include welded ashflow rhyodacitic tuffs, unwelded airfall ash breccias, andesite and basalt flows and flow breccias.

Total chemical analyses of soil/bedrock profiles augmented by evaluations of secondary mineralogy and analyses of runoff water chemistry indicate that podzolization is the primary pedogenic process above 1,150-1,333 M. Continuous leaching of mobile rock-forming cations (Si, Ca, Mg, K and Na) occurs below this elevation, resulting in the accumulation of Fe, Al, Ti, P and Mn towards the surface of soils which are not forming under the influence of a substantial snowpack.

The importance of lithic textures, exclusive of bedrock chemistry is illustrated by the significant differences in soil mineralogy and rates of formation between soils forming on Oligocene or Miocene rocks. At the Andrews Forest study site, Miocene pyroclastic rocks were deposited primarily as an airfall ash while Oligocene pyroclastic rocks

included several welded ashflow tuffs. The contrast in textures due to the different modes of emplacement appears to be responsible for differences in weathering rates and pathways since initial chemistry is very similar for the two rock types.

Chemical evidence of buried profiles demonstrates the longevity of biological influences on the chemical composition of soils. The results of this study generally emphasize the importance of biological processes in the control of ionic distributions of biologically essential cations in the soil profile and the regulation of soil activity which influences inorganic weathering reactions.

Chemical and mineralogical differences noted in two profiles, developed on the same parent material in different geomorphic positions illustrate the importance of drainage and soil mantle stability as they related to oxidation of Fe, Ti and Al the development of biologically cycled ion distributions in soil profiles.

INTRODUCTION

Background and Objectives

This work is part of the analysis of the coniferous biome ecosystem, initiated under the auspices of the National Science Foundation (NSF) through the International Biological Program. The research effort continues with grants from NSF to scientists in a broad range of fields at Oregon State University, Pacific Northwest Forest and Range Experiment Station, United States Forest Service, and University of Oregon. On-going projects focus on nutrient cycling, the aquatic-terestrial system interface, and erosion processes in natural and man-manipulated systems. A computer simulation is being developed to predict system level responses in nutrient fluexes, erosion rates and vegetation succession patterns to various management practices.

The thesis is separated into two phases; a literature review, followed by a report on the field study project carried out at the H. J. Andrews Experimental Forest. The literature survey reviews the role of five factors in the rock weathering process: parent material, climate, atmosphere, soil water and soil organisms. General principles of rock weathering learned and presented in the literature review apply to the discussion of the field project. The Interdisciplinary Master's

degree program at University of Oregon encourages a detailed literature review, to assure integration of the three diverse disciplines being studied in the program.

Information from the field study in this project is classified as process evaluation and will aid in conceptualizing weathering processes and to some extent in assessing the relative weatherability of the different rock types underlying the H.J. Andrews Experimental Forest in the western Cascade Mountains of Oregon. Traditionally, in ecosystem nutrient models weathering has been ascribed to a "black box" or unknown commodity category for which a rate function has been calculated by subtracting from the annual dissolved solid content of runoff waters, input from precipitation and the estimated net annual nutrient uptake by vegetation. The ionic storage capacity of the soil and litter can be ignored assuming that in an undisturbed system a steady-state quasi-equalibrium is reached between weathering flux and the various pathways of ion removal from the soil-bedrock environment (Johnson, 1971; Bricker and Garrels, 1966; and Vitousek, 1977). Unfortunately this approach ignores many of the actual processes which interact to weather rocks. This study is designed to better quantify weathering processes in a total ecosystem context.

In order to observe processes as slow and sensitive to disturbance as those involved in rock weathering a quantitative examination of soil chemistry and soil mineralogy compared to bedrock mineralogy and chemistry was selected to investigate the long-term effect of weathering.

In conjunction with a study of stream water chemistry, this information provides a basis for evaluating weathering processes. Comparative mineral weatherability, influence of climate on weathering and the many ways in which organisms interact with soil and rock has been investigated to varying degrees in many previous studies, but only in a few studies have they been integrated into a single study of rock weathering. This is, as often as not, the result of taking a less than complete interdisciplinary view of weathering. In this study an interdisciplinary approach was considered to be of primary importance.

The experiment involved selecting sites where soils appeared to be forming on stable surfaces of a single parent material. Samples were taken from each of the most extensive bedrock units represented in the H.J. Andrews Experimental Forest.

Detailed total chemical analysis of soil through bedrock profiles, soil clay mineral identification and thin section study of parent material textures and mineralogy provides a view of the long term effects of rock weathering. Unweathered and partially altered bedrock samples provide benchmarks for comparisons, with soil samples. In addition, analysis of one soil/bedrock sequence by textural class as well as depth reveals in detail concentrations of particular elements into various soil size fractions.

Chemistry of runoff waters and soil solution from several of the bedrock types were used to make an observation of current elemental mobilization trends. Synthesizing information about weathering on each time scale provides sufficient information to hypothesize about ion

mobilization processes and the roles played by the different factors in the weathering process in the central western Cascades.

LITERATURE REVIEW

Introduction

The interdisciplinary nature of the study of rock weathering is revealed by the variety of early approaches to the subject. Records indicate that thinkers as early as Strabo (25 A.D.) had noticed aspects of soil forming processes such as clay illuviation and rocks turning to soil. Processes of erosion and physical weathering were conceptualized earliest because they could be observed without extensive scientific background or instrumentation.

The first modern geochemists began publishing results of analyses early in the twentieth century which indicated that soils differed in many striking ways from the parent materials of which they were derived (Merril, 1906; Leath and Meade, 1915; Goldschmidt, 1922; and Goldich, 1938).

During the latter half of the 19th century basic concepts of soil science had come to light through the work of scientists on three continents. In Asia (V.V. Dokuchaev, 1883, K.D. Glinka), North America (V.R. Williams, C.F. Marbut, and E.W. Hilgard), and Europe (P.E. Muller and Gedroiz), scientists were gaining knowledge which allowed them to develop better agricultural methods and breed better plants. The fact that they worked in very different soil forming situations enabled soil science to be founded on a broad base of environments.

The synthesis of geochemical, geomorphic and soils concepts made possible rapid strides in carrying out rock weathering research during the 20th century. Understanding rock weathering processes requires a fundamental knowledge of mineralogy, petrology, soil science, hydrology and physiology of organisms commonly found in the soil.

The following literature review summarizes research in each of these disciplines as it relates to rock weathering. The organization follows the simple conceptualization put forth by Reich (1943):

> Rock weathering is the response of materials, which were in equilibrium within the lithosphere, to conditions at or near its contact with the atmosphere; the hydrosphere; and perhaps more importantly the biosphere.

At the heart of this statement on rock weathering is the notion that igneous rocks are a product of their environment of formation, the high temperature or pressure conditions which preceeded and prevailed during their formation. They must now reequilibrate with the very different surficial environment. The amount of time a rock requires to reach equilibrium with the surficial environment depends upon its inherent resistance to weathering which is derived from its mineralogical and chemical composition. The environment in which it is weathering at the surface is related to the climate, hydrology of the soils, geomorphic situation, and type and quantity of organisms which are present. Hydrologic character of a particular weathered rock and soil is determined by its initial mineral composition and texture, subsequent disturbance (faulting, intrusion by other rocks or rapid erosion rates) and the quantity of weathering agents active in its environment.

Role of Parent Material

The degree of organization of the atomic constituents of rocks and minerals dictates their resistance to weathering. Numerous studies have linked magma temperature and cooling history with strength and complexity of inter- and intra-crystal bonds (Goldich 1938; Loughnan 1969; and Keller 1954). Mineral resistance to weathering has been empirically determined by several workers (Pettijohn 1941; Graham 1941; and Jackson 1968; see Figure 1). Their conclusion has been that those minerals formed at the highest temperatures are most susceptible to weathering forces (Goldich 1928; Loughnan 1969; and Keller 1954).

Factors which would affect weatherability that might be controlled by the energy of formation include structural complexity and degree of substitution. In general minerals formed at higher temperatures have fewer interconnecting bonds and a greater percentage of substituted elements (Marshall, 1964), particularly Al for Si in silica tetrahedra.

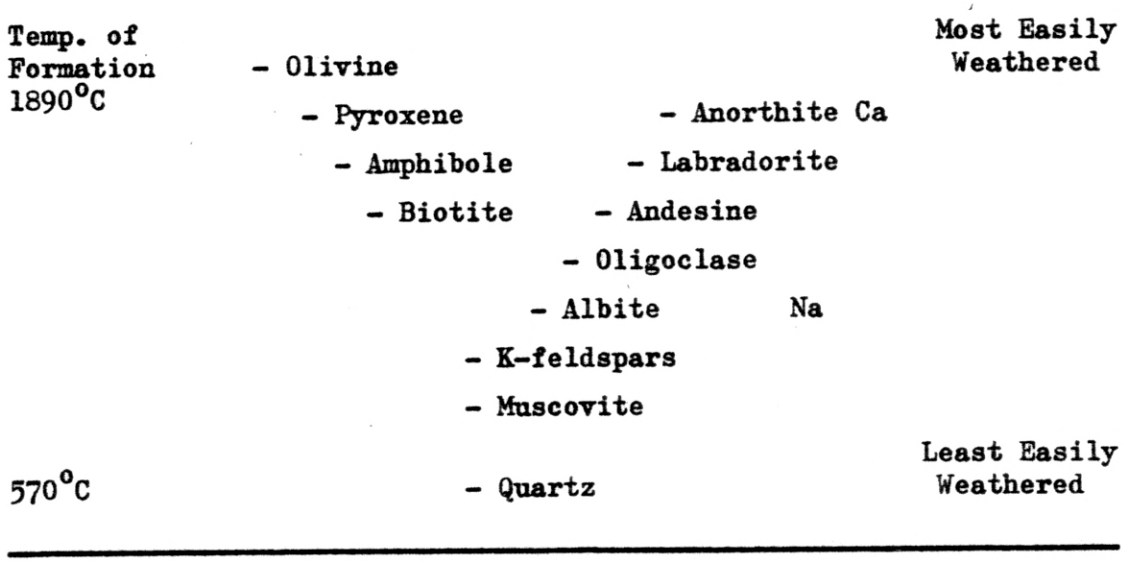
The plagioclase series demonstrates quite clearly how important substitution can be in determining the weatherability of a mineral. The number of bonds between tetrahedral building blocks are the same for both minerals, since they are members of the tectosilicate family. Yet Ca-rich plagioclase-end members (anorthite, labradorite) are much more weatherable than the Na-end members which form at a lower temperature. Marshall (1964) discovered that the rate of Al^{+3} substitution for Si^{+4} was significantly greater in the higher temperature end member of the series. Charge imbalance and the difference in structural irregularities caused by the difference in ionic radii would increase the energy of

attraction by the structure for H^+ and other weathering agents in the soil, and enhance the chances for exposure to weathering agents.

Goldich based his theory on observations of weathered rocks and soil in thinsection as well as on chemical changes. These results have been corroborated by other empirical evidence. Pettijohn (1941) determined which minerals were most common in sedimentary rocks (Table 1). Graham (1941) observed the rates at which minerals decayed when subjected

Figure 1

Bowen's Reaction Series Compared to Goldich's Weatherability Index



to leaching by acid clay slurries (Table 1) and Jackson (1968) studied a broad range of soils to see which minerals were present in very Ancient edaphie environments (Table 1). Modern thermodynamic approaches have illuminated the relationship between bonding energies and weatherability (Reesman and Keller 1968; Huang and Kiang 1973; Huang and Keller 1973; Rai and Lindsey 1975).

TABLE 1

Ordering of Minerals from Least to Most Weather Resistant

Pettijohn (1941)
(Observed in Sediments)

1	olivine
2	actinolite
3	diopside
4	hypersthene
5	sillimanite
6	augite
7	zoisite
8	shene
9	topaz
10	andalusite
11	hornblende
12	epidote
13	kyanite
14	staurolite
15	magnetite
16	illmanite
17	apatite
18	biotite
19	garnet
20	menzanite
21	tourmaline
22	zircon
23	rutile
24	muscovite
25	anatase

Graham (1941)
(Leached with Acid Clays)

1	olivine
2	apatite
3	anorthite
4	bytownite
5	biotite
6	augite
7	hornblende
8	garnet
9	epidote
10	labradorite
11	andesine
12	staurolite
13	orthoclase
14	microcline
15	albite
16	oligoclase
17	quartz
18	muscovite
19	zircon
20	tourmaline
21	rutile
22	ilmenite
23	anatase
24	kyanite
25	sphene
26	magnetite

Jackson (1968)
(Abundances in Well Weathered Soils)

1,2	Gypsum, halite
3,4	calcite, apatite
5,6	olivine, pyroxene
7,8	biotite, glauconite
9, 10,11	albite, anorthite and volcanic glass
12,13	quartz, cristobilite
14,15	muscovite, sericite
16	vermiculite
17	montmorillonite
18,19,20	kaolinite, halloysite and allophane
21,22	gibbsite, beohmite
23,24	hematite, goethite
25,26,27	anatase, rutile and zircon

The three different mineral weatherability lists differ mainly in the range of minerals they cover (Table 1).

There have been many attempts to explain the reasons for the particular ordering of minerals on these empirical lists by calling on modern concepts of bonding energy (Keller 1954), structural packing arrangements (Keller 1954; Fairburn 1943) and cation electronegativities (Gruner, 1950). It is clear that mineral structure must play an important role in weathering resistance, but chemical characteristics of the individual cations are also important. Reiche (1943) developed an index which stressed chemical leachability of rock-forming elements alone and found some correlations in addition to many clear contradictions to the observed orders.

The most modern approach has been to "weather" individual minerals in the laboratory. Using concentrations of the weathering products in distilled H_2O after a given amount of time the Gibbs free energy of formation (ΔG_f^o) can be estimated from the solubility constant K (Reesman and Keller, 1968; Huang and Kiang, 1973; Huang and Keller, 1973; and Rai and Lindsey, 1975). The advantage of this approach is that weathering resistance of each mineral is acknowledged to be a result of both structural and chemical factors.

The values listed in Table 2 are from a compilation of the efforts of many workers reported in a paper by Rai and Lindsey (1965). They theorize that the greater the energy of formation the greater the weathering stress that is required to weather an element, mineral or rock.

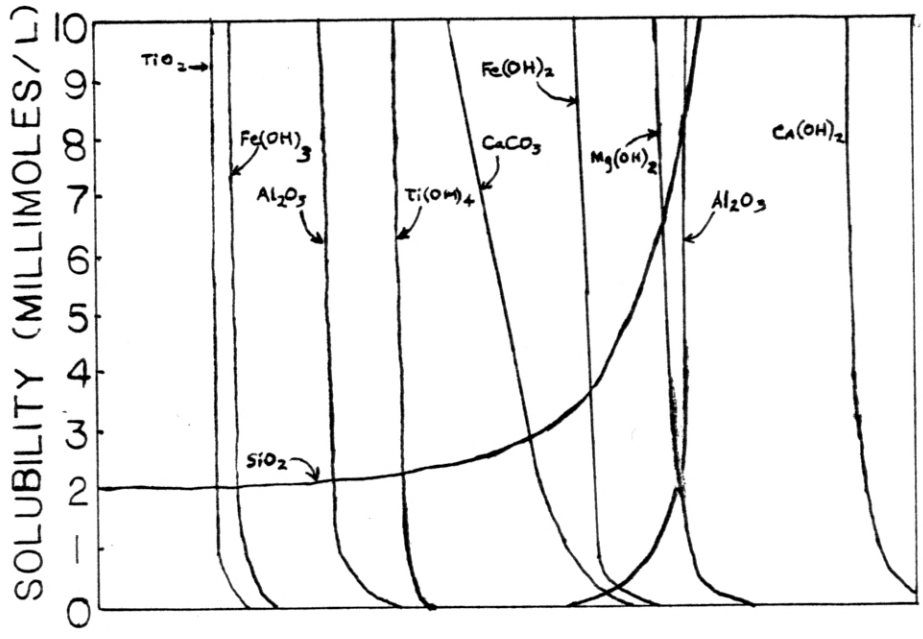
TABLE 2

(ΔG_f^o) Gibbs free energy of formation for rock forming elements and selected mineral

Chemical Species	State*	G_f^o(kcal/mole)
Fe^3	aq	- 2.52
OH	aq	- 37.94
H_2O	lq	- 56.688
Na	aq	- 62.539
K	aq	- 67.70
Mg^2	aq	- 108.90
Al^3	aq	- 116.00
Ca^2	aq	- 132.18
$Fe(OH)_3$	am	- 169.17
SiO_2 (glass)	am	- 202.927
SiO_2 (soil)	am	- 203.418
SiO_2 (quartz)	c	- 204.646
$Al(OH)_3$	am	- 273.974
$Al(OH)_3$ (Gibbsite)	c	- 275.20
H_4SiO_4	aq	- 312.566
$NaAlSi_2O_6.H_2O$ (analcime)	c	- 734.262
$CaAl_2SiO_6$ (pyroxene)	c	- 745.130
$NaAlSi_3O_8$ (Na-glass)	am	- 873.25
$KAlSi_3O_8$ (K-glass)	am	- 883.05
$NaAlSiO_3O_8$ (low albite)	c	- 883.988
$KAlSi_3O_8$ (microline)	c	- 892.817
$CaAl_2Si_2O_8$ (anorthite)	c	- 955.626
$KAl_2(AlSi_3O_{10})$ $(OH)_2$ (muscovite)	c	- 1330.103
$Mg_5Al_2Si_3O_{10}$ $(OH)_8$ (chlorite)	c	- 1954.800
$Al_2Si_2O_5$ $(OH)_4$ (halloysite)	c	- 898.419
$Al_2Si_2O_5$ $(OH)_4$ (dickite)	c	- 902.14
$Al_2Si_2O_5$ $(OH)_4$ (kaolinite)	c	- 904.72

The next step for proponents of the thermodynamic approach will be the quantification of the energy of leaching which acts on minerals in any given soil situation. A major handicap in taking this step is lack of sufficient understanding of the processes involved in rock weathering. The multiplicity of factors which interact to determine leaching energy include temperature, moisture, soil permeability, decomposer activity, litter production, metabolic activity of vegetation and texture and geochemistry of soil/parent material. It will be many years before these activities can be quantified but the thermodynamic data on minerals can be utilized for comparing weathering resistances between minerals.

Figure 2. Solubility in relation to pH for some componenents released by chemical weathering



Compilation of the activities of different ions in mineral structures soil solution chemistry or in organisms has been made possible by years of pure inorganic chemistry research. Summaries have been made by several workers (Loughnan, 1971). Figure 2 demonstrates the solubilities of several common rock forming elements at 25°C and 1 atm pressure over the entire pH range (Loughnan, 1971). Precipitation averages near pH 5.5 and soil solution lies between pH 6.5-7.5 at the study site as a standard of comparison (McCorison and Sollins, unpublished data).

The elements which would not be in solution in the soil pH range reported are Ti, Fe and Al. During the development of laterites, which are especially common in warm tropical climates on extremely old land surfaces of basic lithologies, the soil contains almost exclusively, Al_2O_3, Fe_2O_3 and TiO_2.

In temperate climates, similar trends are noted, but because climatic factors are not as intense, the rates of such laterizing reactions are much slower. Weathered bedrock samples collected from ancient landforms in temperate regions indicate that if the weathering process could be allowed to continue, a laterite would result. Hendricks and Whittig (1968) present results of analyses of bedrock through saprolite samples of Pliocene age hypersthene andesite from the southern Cascade Mountains near Mount Lassen, California. The rock is comparable in coding history, composition and weathering environment to some of the rocks found in the central western Cascades. (See Table 3.)

TABLE 3

Saprolite Samples Showing Increasing Weathering*

	Hypersthene Andesite	1	2	3	4
SiO_2	57.0	45.7	41.3	38.9	39.9
Al_2O_3	16.7	22.3	29.1	31.6	31.8
Fe_2O_3	2.0	5.2	9.8	11.1	11.3
FeO	4.7	4.2	1.1	.50	.40
TiO_2	1.0	1.0	1.2	1.3	1.3
MnO	.12	.24	.26	.27	.26
P_2O_5	.15	.07	.03	.03	.03
CaO	7.2	4.8	1.2	.62	.43
MgO	6.1	7.2	2.7	1.4	.27
Na_2O	3.1	1.2	.86	.79	.62
K_2O	1.2	.24	.28	.11	.15
H_2O	.35	7.4	11.8	12.6	12.9

*Hendricks and Whittig, 1968

Al, Fe, Mn, and H_2O increase at the expense of all other cations. K, Mg and Ca are leached more rapidly than Na. If the process was to continue for millions of years a residue closely resembling a laterite would develop. In most soil forming situations such long episodes of undisturbed leaching are rarely encountered.

Bedrock chemistry and mineralogy have been shown to directly influence soil factors such as clay mineral formation and texture (particle size distribution) (Jackson and Sherman, 1953; Loughnan, 1969). An interesting study involved a unique conglomerate in Virginia which contained boulders of granite, gneiss, gabbro, large feldspar crystals

from a pegmatite and quartzite, weathering under nearly identical conditions for the same period of time (Barnhisel and Rich, 1967). Based on the large quantity of Al absorbed on montmorillonite clays, the deposit had been weathering for a long period of time under warm humid conditions. Using X-ray diffraction, electron microscopy and cation extraction techniques, three representative boulders (gabbro, granite and gneiss) were studied.

Clay mineral formation was dependent on initial parent material mineralogy. The sodic and postassic feldspar crystals of granite and gneissic rocks weathered to kaolin minerals, whereas the plagioclase, augite and accessory minerals in the gabbro boulder weathered to montmorillonite. Mg and K occupied most of the exchange sites inhabited by Al on montmorillonite while Na was the most abundantly available cation in the kaolinite system.

Of special significance in this study was the fact that the entire deposit was well drained. 2:1 clays are commonly formed in poorly drained situations where the cations are concentrated over time. In the well drained Virginia deposit, formation of the montmorillonitic clay was hypothesized to be the result of basic rock chemistry in combination with the effect on microdrainage of the fine texture to which a basic rock is more typically inclined to weather. Grain size of the gabbro and granite were nearly identical initially. The absence of quartz and presence of more easily weathered minerals such as pyroxene and plagioclase resulted in a more rapid breakdown to a much

finer textured matrix than did the granite or gniessic rocks which contain much coarse grained, weathering resistant quartz. The coarser texture encourages more rapid drainage and provides less opportunity for concentration of nutrients.

Texture alone has been shown to be a significant factor in rocks of similar chemical composition. A pyroclastic rock would be more prone to breakdown to clay sized particles because of the higher quantity of fine grained, easily weatherable volcanic ash (refer to table of thermodynamic values, Table 2), than a coarsely crystalline granite of the same composition (Loughnan, 1969).

Janda and Birkland (1970) found in a north-south transect of the Sierra Nevada Mountains, that presence of basic parent materials indicated soils would contain 2:1 clays. On this trend, the relationship between abundance of moisture and clay mineral formation was superimposed. In the xeric regions to the south, less 1:1 clay was found in all soils as compared to the more moist northern region, but in all cases, 2:1 clay minerals were most abundant in soils derived from basic rocks.

An example of how the two very different rock types can interact to create a geomorphic situation which enhances the weathering environment for both units and significantly affects soil formation, vegetation patterns and stream configuration, is found in the H.J. Andrews Experimental Forest (Swanson and Jame, 1975). Well indurated younger flow rocks (Sardine Fm.) overlie several hundred meters of massively bedded pyroclastic rocks (Lower Sardine and Little Butte Fm.) throughout

much of the eastern part of the Forest. The tremendous differences in the internal strengths of the two rock types leads to a very unstable "cap rock" situation which is a result of the slower weathering rates of less porous or permiable crystalline flow rock in comparison to the permeable, more easily weathered, glass-rich ash flow and airfall ash deposits. Large deep seated earthflows have resulted, some with relief exceeding 500 M.

The earthflow consists of a series of deep seated slip planes, which creates poorly drained benchy topography. Greater percolation by and retention of ground water in the grinding, churning slump areas perpetuates the contrast in weathering rates with adjacent landforms since exposure of more mineral surface area to greater water fluxes leads to greater weathering rates and more rapid rates of erosion.

On the other hand, rapid surface erosion and active creep processes on steep slopes can inhibit chemical weathering by mixing soil horizons, maintaining rapid drainage, removal of the litter layer or surface horizons and interrupting the plant/soil interface.

Role of Climate and Atmosphere

The role played by climate in physical or mechanical weathering processes are most apparent in regions where conditions are too extreme for most vegetation to survive. In artic and in alpine regions freeze-thaw processes and loading and unloading due to changes in glacier distribution, account for much of the rock decay by mechanical processes.

In extremely arid regions, wind erosion, thermal expansion and contraction and desiccation (partially chemical) are the dominant weathering processes. In regions where vegetation is dense, the surface of rocks are protected from most physical processes.

Reynolds (1971) has demonstrated that clay formation and chemical weathering do occur quite readily above the tree line in the North Cascade Mountains of Washington. Highly leached and oxidized crusts were found intact on south facing slopes where they were somewhat protected from mechanical erosion processes. Because mechanical processes were so active, little thought had been given to the role of chemical weathering but analyses of stream waters revealed that Ca, Mg, K and Na cations and bicarbonate ion were present in quantities greater than the world average.

When considering the influence of climate on chemical weathering, it is important to bear in mind that the microclimate of the actual weathering site can be significantly different from the overall climate due to micro site textures (Barnhisal and Rich, 1967). There are, however, several macroclimatic influences which are apparent throughout the world.

Weaver (1949) found that oxisols (highly leached, Fe and Al rich soils, Soil Survey, 1970) were found on a greater variety of parent materials and over a much larger area in tropical regions compared to temperate regions, where they were usually restricted to very old surfaces formed on basic rock types.

Continental-scale transects have yielded correlations between deeper, more weathered soils and regions receiving higher rainfall and subjected to higher temperatures in the United States (Jenny and Leonard, 1934), and in Australia (Simonett and Banleke, 1963). On the west coast of the United States a similar investigation of soils on a climatic transect along the east slope of the Sierra Nevada range revealed that kaolinitic clays predominated in the more humid northern areas with 2:1 clays increasing in the more xeric southern Sierras (Janda and Birkland, 1970).

In terms of physical chemistry of weathering reactions, such trends in clay mineralogy are understandable. If weathering reaction products become too concentrated in the immediate region of the source, chemical weathering energy will be absorbed by the formation of complex secondary minerals, thereby minimizing continued leaching of primary minerals. An apparent equilibrium condition is reached, until weathering products are removed. Therefore, in a situation where rainfall is so sparce that evapotransporation quickly removes moisture from the soil, weathering products are highly concentrated in residual moisture which favors decreased weathering rates of primary minerals, and formation of 2:1 clays. In a region where cooler temperatures and higher rates of precipitation combine to maintain weathering products at a much lower level in the soil solution, leaching of primary minerals will be more continuous and synthesis of kaolinitic clays favored. Weathering patterns enter a third phase when soils are frozen for extended periods in the winter, which will be described below.

Weathering studies over large scale temperature gradients reveal that weathering is more intense in warmer regions. This can be attributed to the effect of higher soil temperatures on chemical reactions as well as much higher rates of decomposition. Jenny and Leonard (1934) found that weathering rates increased from north to south along a thermal gradient and west to east along a moisture gradient in the United States.

Cortes and Franzmeier (1972) studied soil formation on a relatively recent ash deposit in the Central Cordillera of Colombia. The highest sites approached 3,500 M, the lowest were near 1000 M. Soil mineralogy indicated that parent material was consistant throughout the transect. They found that as temperature increased the degree of B horizon development increased, clay content increased, organic matter content decreased, degree of weathering of primary minerals increased and the content of amorphous material, kaolinite and halloysite increased.

Considerable evidence of the role climate plays in soil formation and rock weathering is provided by the differences observed in soils which form on the north-facing side of a hill as opposed to those forming on the south-facing slopes. Cooper (1960) found that north facing soils in Michigan were deeper, retained more moisture over the year, were more enriched in organic material and contained less subsoil clay than their south-facing counterparts. Similar findings have been reported from a variety of physiographic regions, S.E. Alaska (Bishop and Stevens, 1964), interior Alaska (Krause, Rieger and Wilde, 1959), Cumberland Plateau, Kentucky (Franzmeier et al., 1969) and North Carolina (Losche, McCracken

and Davey, 1970). Dyrness (1967) observed that soils forming on south-facing slopes in the H.J. Andrews Forest were shallower and more gravelly in comparison to those on north-facing slopes. Microclimate differs greatly between north and south aspects due to solar radiation input. The most important results of this difference in energy flux are differences in metabolic rates of plants and decomposers, soil temperature as it affects rates of chemical reactions, degree of desiccation of the soil, particularly as it relates to the fire frequency and intensity, and snowpack longevity.

Gas exchange between minerals and the atmosphere in soils is primarily carried out by organisms, but it can occur inorganically as well. In both cases soil pH is very important. Oxidation of Fe, Mn and Al to highly immobile compounds occurs most easily from pH 6 to 8. Poorly drained soils exclude both oxygen and aerobic organisms. Anerobic organisms aid in the reduction of the same metals to very mobile forms. This explains in part the tremendous differences between soils and vegetation in moist areas compared to communities and pedons forming in better drained sites.

The seasonal water table can be identified in a soil by the blue-gray mottled color of the soil below the zone of extended wetness which is due to the color of iron and manganese in their reduced states. Bright orange mottled spots and lines indicate where aerobic communities of microbes can survive. They typically follow root channels or other large soil pores (Buol, Hole and McCracken, 1973).

Rock texture influences soil texture by its influence on micro-drainage and the development of soil structure. For example, coarse grained soils do not retain water long enough to support a large population of organisms. Excessively fine textured soils limit root extension, plant growth and drainage. The aggregation of soils into distinct peds is the result of organism activity and repeated wetting and drying and the precipitation of amorphous gels (Taskey, unpublished data). Development of soil pores and root channels facilitates gas exchange, which is significant in the weathering process by virtue of its enhancement of growth of vegetation, aerobic microbial activity and movement of soil water. Fixation of atmospheric nitrogen by free living and rhizobial bacteria also augment development of plant and animal communities.

Role of Soil Water

Without water, weathering would proceed at an extremely slow rate. Water, as it enters the soil from the atmosphere, is not in itself the most active weathering agent, but it serves as the medium through which organic substances reach rocks, soil particles, and mineral surfaces; it sustains organisms which produce highly reactive substances; it carries weathering products away from the site of the reaction, allowing endothermic weathering reactions to continue and it is capable of doing a large amount of physical work if present in large quantities or if frozen in the soil.

Chemical weathering is facilitated by water through hydration or attraction of dipolar water molecules to charged mineral surfaces. Once enveloped by water, hydrolysis can occur, which is the exchange of hydrogen ions for cations bound by less energetic forces in mineral structures. The hydrogen ion is the smallest and most highly charged ion among common terrestrial elements.

The concentration of H^+ ion in water on its arrival to the soil is not exceptionally high, but through enhancement by organic and inorganic processes it can reach very high levels. The soil, on the other hand, possesses a considerable buffering capability, which rapidly absorbs much of the H^+ through mineral decay or complexation with negatively charged substances. Measurements have indicated that extremely low pH conditions exist for only short periods of time in a soil as elemental transfers and mineral degradation quickly respond to the presence of excess H^+ ion (McColl, 1974).

Resistance of minerals to the presence of H^+ varies considerably, depending on mineral structure, degree of substitution, and mineral composition.

Acidity of precipitation varies as well, depending upon the quantity of available acidifying chlorine, sulphur, nitrate or other substances introduced into the atmosphere which react with airborn moisture, freeing H^+. The pH of precipitation in the H.J. Andrews Forest averages near 5.5 (McCorison, unpublished data), and appears to be relatively free of industrial acidity.

Inorganic reactions between certain minerals and water enhance soil solution acidity. Both free Fe^{3+} and Al^{3+} can remove hydroxyl ions from solution to form $Al(OH)_2$, $Al(OH)_3$ gibbsite, AlO(OH) diaspore and boehmite or FeO(OH) geothite and lepidocrocite, forcing solution pH lower (Huang and Keller, 1972). Inorganic H^+ enhancement is much less important than organic processes near the surface but becomes the dominant buffer of soil acidity and in the subsoil. Organic processes will be discussed in detail below.

The role water plays in chemical weathering makes residence time in the soil a significant matter. Residence time is a function of magnitude and frequency of precipitation events and drainage conditions which are regulated by soil texture, geomorphic setting and vegetative cover. Soil texture is closely related to parent material, climate, and interval of soil development. With such a range of factors, soil moisture conditions can often vary over short distances.

Both quantity and frequency of precipitation is important in soil genesis. Soils formed on Tama Glacial outwash was studied from northern Illinois to St. Louis (Goddard, Runge and Ray, 1973). In spite of higher temperatures and slightly more precipitation at the southern end of the transect, the highest clay contents were found at both ends of the transect. The authors hypothesized that the single common denominator, rainfall event frequency, must have been the most important factor in determining the severity of the weathering environment. Other workers have shown that very thin gel-like amorphous films form on soil particles when desiccation occurs, which would be washed into the soil on subsequent rewetting (Grossman and Lynn, 1967).

Sherman (1949, 1952) observed that two distinctive trends could be seen in the weathering history of a basalt flow in Hawaii based on the presence or absence of alternating wet and dry seasons. If soils were moist year round Fe and Ti were apparently not oxidized, which allowed them to be leached from the soil, leaving Al as the predominant, residual ion. With the occurrence of a dry season, oxidation and hydroxalation of Fe and Ti into insoluable compounds prevented their leaching, leaving them as the dominant constituents.

Chemical weathering is facilitated by the presence of water. Its importance is observable in the fact that soil moisture conditions, particularly on a microscale, strongly influence rock weathering pathways.

Role of Organisms

The presence of a rich biota in a soil is usually correlated with the predominance of chemical weathering processes. However, mechanical weathering does occur as a result of plant root expansion in rock fractures, abrasion and soil mixing as a result of uprooting of trees and a range of disturbances due to animal activities.

Groundcover mosses and herbs play an important role in suppressing surficial mechanical erosion caused by raindrops and frost heaving (Mersereau and Dyrness, 1972) and larger plants stabilize soils with intertwining root penetration (Swanson, 1970). Soil aggregation is also closely associated with secretions from plants and animals (Buol,

Hole and McCracken, 1974). Both aggregation and cavity forming activities of organisms enhance soil porosity which encourages unsaturated flow of soil water, thereby preventing surficial erosion.

Logs and large pieces of organic debris retard surficial erosion on hillslopes and are very important in the reduction of erosion by stablizing stream beds and channel banks.

Chemical weathering as a result of rock and soil interaction with organisms occurs via processes such as nutrient uptake, accumulation of byproducts from decomposition and plant metabolism in soil solution, CO_2 evolution during respiration and incipient conversion into bicarbonate ion, secretion of acids and enzymes from microbes and plants or the result of digestion by soil macrofauna.

Large quantities of nutrients are stored in the various compartments of the biological realm. Through the intricate complex of interactions between organisms and their substrates, nutrients and energy are transfered with a high degree of efficiency. In soil, an important portion of the biomaze is composed of microfauna and flora. The ability of different soil organisms to concentrate particular elements through selective metabolic membrane processes makes possible biogeochemical nutrient cyclying (B.N. Richards, 1974; Salisbury and Ross, 1969; Gist and Crossley, 1975).

Sources for nutrient cations in soil solution include decay of primary minerals (Loughnan, 1971), removal from charged sites on clay minerals, displacement of cations from soil organic matter via decomposer cycling (McLaren and Peterson, 1967) and leaching or sloughing

of nutrient ions from vegetation (Cole, Gessel and Dice, 1967). A brief review of these exchange processes and the roles played by each of the ions in higher plants will aid in understanding the trends of different elements demonstate in the soil.

Plants take up essential elements to failitate life sustaining biochemical reactions and the building of cellular material for growth and repair functions. While nutrient requirements of different plants vary considerably it is presumed that uptake mechanisms of the majority of vascular plants are nearly identical (E. Epstien, 1965; F.L. Stewart, 1963).

A portion of the nutrients taken up by plants are returned to the soil by leaf wash, while others remain in the plant until they are cycled through decomposer organisms. The figures in Table 4 demonstrate how nutrient content of Douglas-fir needles, sampled in an oldgrowth forest stand in the H.J. Andrews Forest, 1) changes with age, 2) differs from that of branches and twigs and 3) demonstrates how the nutrient composition of living tissue compares to that of the litter it eventually becomes.

TABLE 4

Nutrient Status of Current and Older needles, Woody material and Needle Litter of Douglas Fir

Dry Wt.	Current[a] Needles	Older[a] Needles	Branches[a] & twigs	Needle Litter[b] from H.J.A.
Total (tons/ha/yr)	2.63	9.36	57.0	1.74
Na (%)	.04	.04	.002	.03
K (%)	.60	.43	.20	.17
Ca (%)	.27	.50	.48	1.22
Mg (%)	.11	.10	.03	.082
P (%) of (eq/ton)	.16 (398)	.27 (461)	.04 (325)	.12 (735)

a) from C. Grier (unpublished data) from H.J. Andrews Forest
b) from Abee (1973) for W.S. 10 H.J. Andrews Forest

A variety of changes in cation concentration occur as current foliage ages and eventually becomes part of the litter layer. Sodium is present in negligible quantities initially and shows a 25% decrease from living tissue and litter, Ca is 452% higher in the litter than initial foliar amounts and shows a steady increase through each stage of development, Mg changes little between young and old needles, but drops 27% in the litter, and P showed a substantial increase from current to older foliage (69%) but decreased to a value 25% below that of the current needles in the litter.

In a study of leaf wash and stem flow from a 35 year old second growth stand of Douglas fir in western Washington an idea of the leachability of several cations was obtained for Douglas fir (Cole, Gessel and Dice, 1967). Of the K taken up annually 42% was returned to the forest floor by leaching from the leaves, while only 19% of the calcium, 6% of the phosphorus and 4% of the annual nitrogen budgets were returned to the soil each year.

An analysis of the role played by each of the nutrient cations in plant aids in understanding the results of both studies. A complete understanding of the role K plays in plants is lacking. It is clearly not a structural element in plant tissue, but rather a catalyst of many important enzymatic and energy transfer reactions, such as carbohydrate and nitrogen metabolism, neutralization of important organic acids, development of meristem tissue and possibly involved in stomatal response (Tisdale and Nelson, 1966). High concentrations

of K in leaf wash is most likely a reflection of the rapid fluctuations of K concentration in stomatal guard cells (Salisbury and Ross, 1969).

Calcium is an essential nutrient for plant growth. In its absence terminal bud and root tip development cease. There is evidence that Ca favors synthesis of protein in mitochondria, whose activity is closely associated with ion uptake (Salisbury and Ross, 1969). In view of its slow leaching rate and regular buildup in the litter, calcium must be closely involved with cell structure, highly selected for by cell membranes, and rapidly taken up by fungi and other organisms living on the leaf and attacking the dropped needle in the litter (Cromack, Tedd and Monk, 1975).

Magnesium is primarily located in the central position of the chloryphyll molecule. It is also important in the transfer of energy from phosphorylated compounds. Most plants translocate chlorophyll from the older needles, thereby preserving some Mg, before leaf abscision occurs (Tisdale and Nelson, 1966).

Phosphorous is involved in the transfer of energy throughout the plant via the many phosphorylated compounds such as ADP, ATP, NACP and NADPH. Isotope tracer studies have followed repeated translocations of marked P ions from the leaves to the roots and back again over a very short time interval (Salisbury and Ross, 1969). In spite of its mobility, the recycling pathways for phophorylated compounds are well established, protecting them from excessive leaching.

Manganese and iron are both required in trace amounts for proper plant development. Manganese is used in the activation of numerous

enzymatic reactions, including carbohydrate metabolism, phosphorlation reactions, the citric acid cycle and in certain photochemical processes (Tisdale and Nelson, 1966). Iron is necessary for several vital enzyme reactions as a divalent cation and for chloroplastic protein syntheses as trivalent Fe.

Mineralization of organic matter occurs through a complex series of interactions between different trophic levels of atmospheric and soil microorganisms. Studies in England, using pine needle (Pinus sylvestrus) litter (Kendrick and Burges, 1962) and in Australia using litter from Eucalyptus regans forests (MacAuley and Thrower, 1966) have revealed the colonization succession patterns of fresh litter. The first decomposer organisms are airborn leaf parasites that replace the fauna and flora that depend upon carbohydrate and hormonal leaf exudates of living leaves, during late summer before the leaves are shed. The leaf parasites are followed by innoculation of saprophytic imperfect fungi during the fall before and as the senescent leaves are dropped and they remain the major decomposer organism for about 6 months while the litter occupies the L-layer. Mites and springtails consume most of the hyphae and spores of the F_1-layer following sporulation in the late fall of the next year. After 2 $\frac{1}{2}$ years the needles enter the F_2-layer where moisture levels are more consistant which makes possible colonization by basidiomcete fungi. Increased grazing by soil meso fauna eventually terminates this phase, leaving a thick faecal mat mixed into surface mineral soil. Needles required nearly

seven years for complete deterioration to occur. Eucalyptus litter was decomposed much more quickly. It is important to note that bacteria played a limited role in the breakdown of acid conifer litter while more basic litter of deciduous species and grasses is quickly colonized by bacteria.

Through the production of organic substances such as citric, acetic, formic, fumaric and oxalic acids, carbohydrates, strong chelating agents such as 2-ketogluconic acid and certain enzymes, organism have been shown to breakdown primary minerals. Fungi common to almost any forest soil were placed in contact with biotite, muscovite, vermiculite, phlogopite, geothite, leucite, nepheline, olivine, wollastinite, apophyllite, huelandint and stilbite (Muller and Forester, 1960 and 1964; Nintikka and Naykki, 1967; Henderson and Duff, 1963 and Eno and Reuszer, 1950) and were found to produce significant increases in the mineral constituents in solution. Some of these studies have been criticized because they have been carried out in laboratory situations without a complete soil microbial population (Richards, 1974). On the other hand electron microscopy has provided ample evidence that mineral grains extracted from the rhizosphere are more highly etched and pitted than those from deeper horizons (Berner and Holdren, 1977; Graustien, personal communication, 1977). The combined influence of plant root secretions, root lyses and byproducts of soil microbial and mesofaunal activity create a potent source of weathering energy.

Several workers have found that concentrations of organic substances in soil solution near the surface reach 45-55% of the anionic component (Johnson, 1976), yet they represent only 10-15% at depth, where inorganic anions such as bicarbonate and sulphate dominate.

Three striking illustrations of the chemical weathering potential of organisms are the role lichens play in the colonization of fresh lava flows and recently glaciated bedrock (Schatz, 1963), the influence of "cheluviation" in the formation of spodisols and oxisols (Strobbe and Wright, 1959, Jenny, Arkley and Schultz, 1969, Graustien, 1976), and the unique role played by organic substances in formation of clays and other secondary minerals (Linares and Huertes, 1971).

The role of organic substances in spodisolization has been carefully studied (Strobbe and Wright, 1959). Iron, Al, Ti and P compounds are highly insoluable in the pH range 4-7, within which most soils fall. However, they can be in spodosals mobilized and translocated to the subsoil, leaving a leached, almost pure silica "albic" A horizon. These soils are widespread in the northeastern United States, in alpine regions and thrcughout cooler subpolar regions of each hemisphere. The linkage with climate is related to the extended presence of a significant snowpack which inhibits percolation of soil waters for several months. Groundwater spodosols have been recognized in many mild coastal regions also (Jenny, Arkley, Schultz, 1969). Organic substances build up as a result of plant metabolism and decomposer activity which are not inhibited by the insulating blanket of snow.

Spring thaw produces a very acid flush of melt water through the soil during which time Fe, Al, Ti and P are translocated with the other leached cations to the B horizon. The buffering capabilities of the soil and microbial degradation of organic substances ameliorates the acidity resulting in a reprecipitation of insoluable Fe and Al phosphate compounds and organic complexes in the striking "spodic" horizon. Since silica leaches at a steady rate from soils below pH 7 are due to its prominence in most soils it is preferentially concentrated in the surface soil after repeated acidic leachings.

In addition to leaching mineral constituents, organisms play a significant role in slowing removal of leachable ions by meeting their own nutrient requirements. Through microbial succession certain elements become concentrated in different organisms where they are foraged by other organisms. The chitinous remains and faecal wastes of mesofaunal exploitation serve as both a site of slow cation release and a source of readily usable nutrients for plants or other microorganisms.

The influence of organic substances on the process of secondary mineral formation has been demonstrated by several workers (Linnares and Huartes, 1972, Schnitzer, personal communication, 1977). In solutions of monomeric silica and aluminum, kaolinite will precipitate in the presence of a small quantity of fulvic and other organic acids. The reason proposed for this response has been an immediate formation of unstable complexes between the organic molecules and both Si and Al. The slow release of free Al and Si which follows promotes synthesis

of highly ordered kaolinite. In the absence of this intermediate complex very little and only poorly crystalized kaolinite was formed (Schnitzer, personal communication, 1977).

The importance of the interaction between plants and other soil organisms is clearly portrayed by the high concentrations of organisms found in the immediate vicinity of roots. The rhizosphere effect, as it is known by some (Richards, 1974), can be measured by comparing the size of the microbial population within the root zone, R, 1-15 mm from the root, to that of the soil away from the root, S (Table 5).

TABLE 5

The Rhizosphere Effect: Typical R/S Ratios for Various Microbes

Organism	Ratio
Bacteria	10-50
Fungi	5-10
Protozoa	1-3
Algae	1-2

From: B.N. Richards, 1974

Jenny and Grossenbacher (1963) have investigated the root soil boundary with an electron microscope. They find a mucilageneous gel supporting millions of rhizobial bacteria, fungi and viruses enfolding most roots. Figure 3 depicts large circumroot spaces observed in a soil in H.J. Andrews Forest with a light microscope.

Root exudates consisting of carbohydrates, hormones and enzymes and sloughed root material provide sustanence for the large variety of microbes while they in turn benefit the plant with secretions of mineral solubilizing organic acids and a pool of rapidly cycled nutients, a

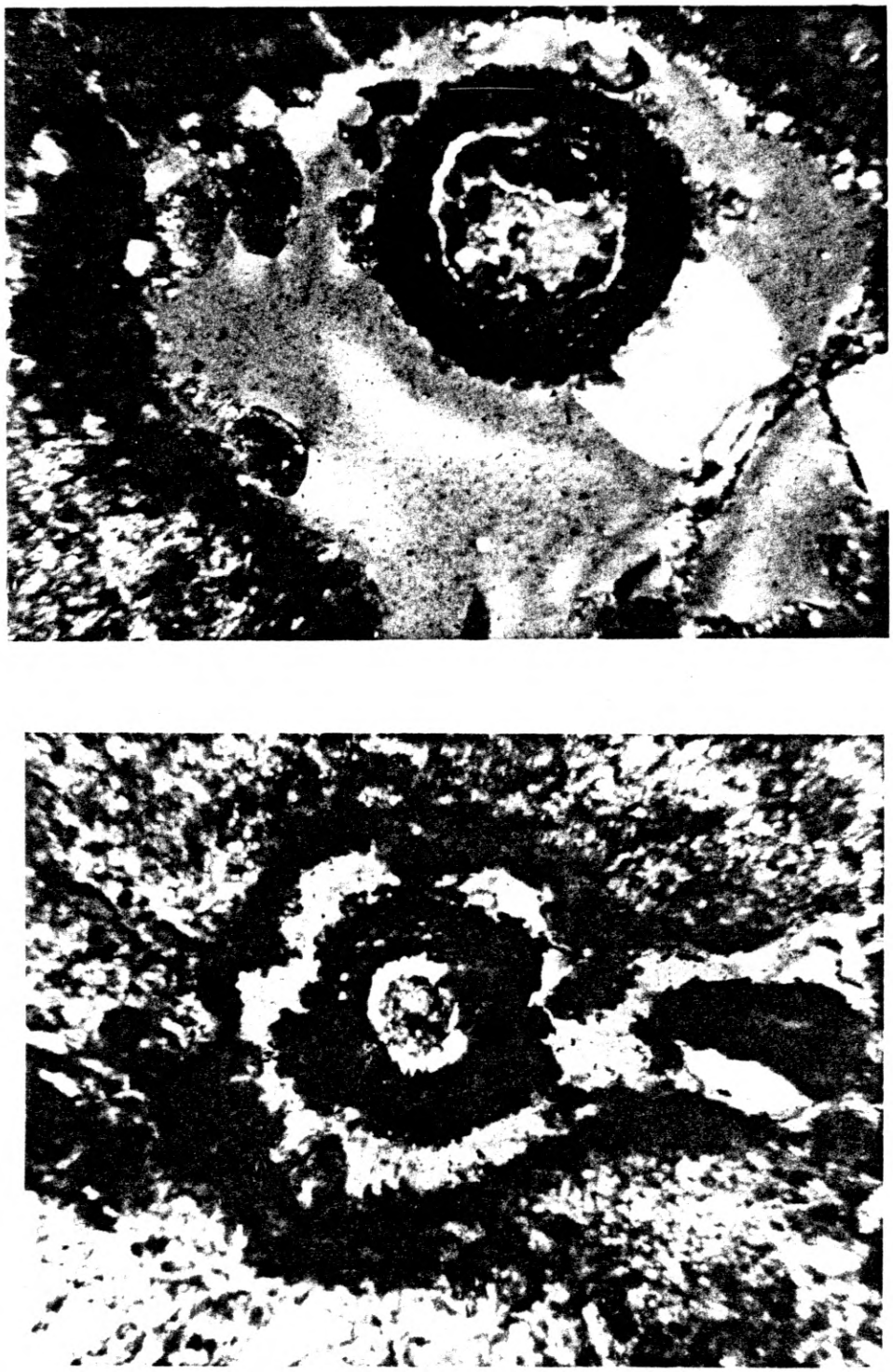


Figure 3. Picture of circumroot space, presumably occupied by fungi and microbe-rich mucilagenous gel, from soil in H.J. Andrews Forest (magnification 300X).

portion of which, become available to the plant. Extensions of fungal hyphae into the soil transport both water and nutrients to the immediate region of the root and by attracting burrowing organisms to the microbe-rich mucigel, the microorganisms enhance aeration and percolation in the region of the root which aids other symbiotic activities.

Another way that organisms enhance the weathering environment is through the evolution of CO_2 during respiration. Carbonic acid forms as CO_2 is assimilated by soil solution during carbonation. CO_2 content, of soil atmosphere reaches levels of .2 to 4.5% (Buchman and Brady, 1969) in comparison with the consistent average CO_2 concentration (.03%) of the atmosphere. Dissociated H^+, the most highly charged cation in most soils disrupts absorbed cations on secondary minerals, and readily removes components from primary mineral structures, while dissociated HCO_3^- bicarbonate ion, functions as an important anion in the leaching of cations from the weathering site (McColl, 1974).

The significance or organisms as an important source of chemical weathering away has heretofore been underestimated. The reasons for reluctant acceptance of the magnitude of the role of the biota have been soudly based on a lack of experimental justification. <u>In situ</u> studies of root exudates, determination of the organic fraction of soil solution and observation of plant/soil particle relations with electron microscopy are providing a clearer picture of the magnitude of the organic influence on chemical weathering.

THE STUDY AREA

Location

Sample sites were all in the H.J. Andrews Experimental Forest, located near Blue River, Oregon, about 75 km east of Eugene. The forest is situated on the western slope of the central western Cascade Range and encompasses the entire 6,100 ha (15,000 acres) of Lookout Creek drainage basin. Elevations range from 410 M (1,300 ft.) at the mouth of Lookout Creek to 1,630 M (4379 ft.) at the highest point, Carpenter Mountain, 14 km (8 mi.) to the northeast (fig. 4).

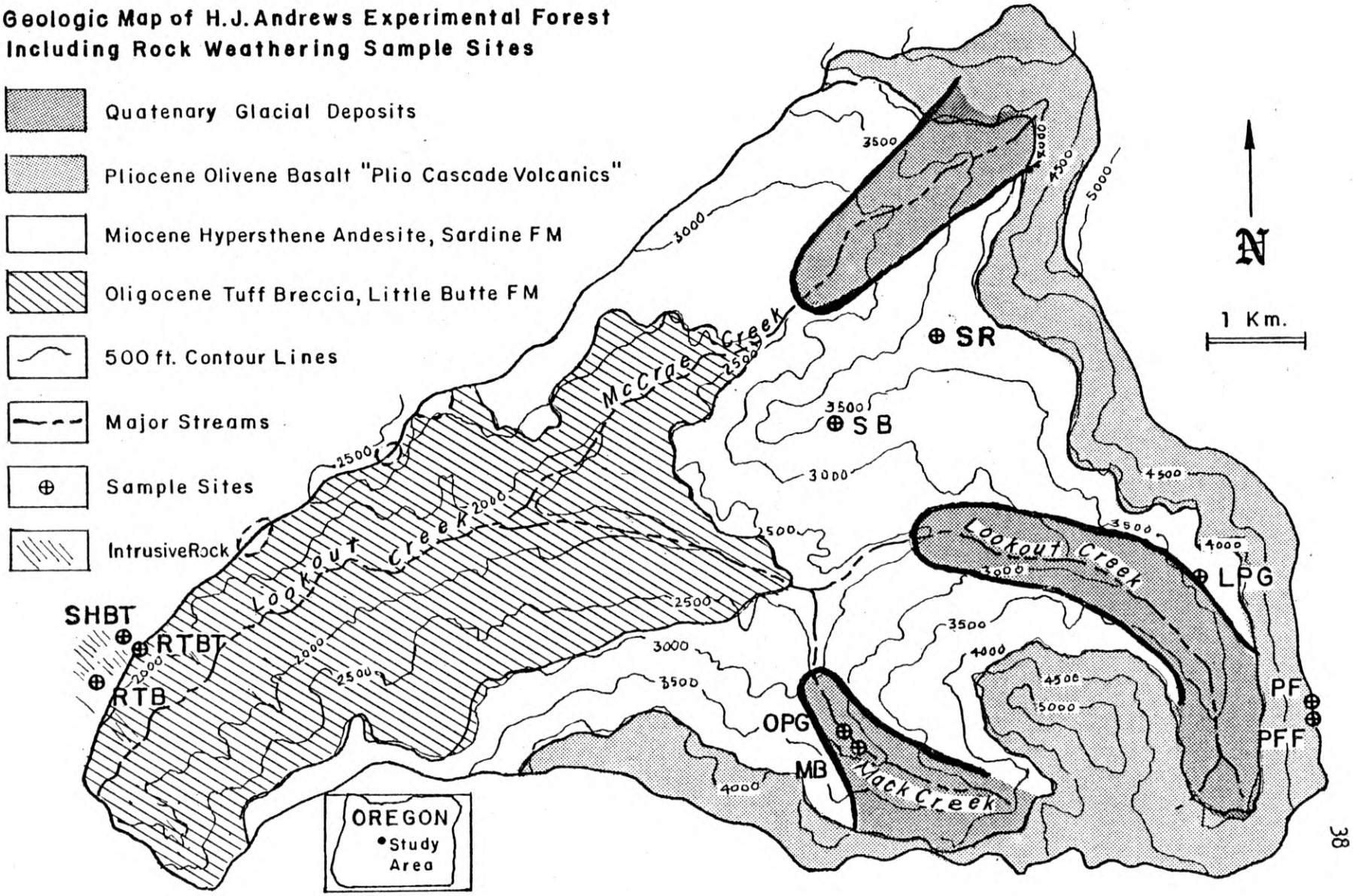
Climate

The climate of the Andrews Forest is characterized by heavy precipitation from late October to March. The growing season occurs mainly during the driest part of the year (fig. 5). Annual precipitation varies from 2,300 mm (89 in.) near the lower reaches of Lookout Creek to as much as 3,600 mm (140 in.) along the highest ridges (Bernsten and Rothacher, 1959). The snow pack above 945 M (2,700 ft.), generally persists from November to April. Precipitation is predominately in the form of rain below this elevation.

Mean temperature over a ten year period was 9.5°C (49.1°F), with a January mean of 1.7°C (35°F) and a July mean of 20.6°C (69.2°F).

Figure 4.

Geologic Map of H.J. Andrews Experimental Forest Including Rock Weathering Sample Sites



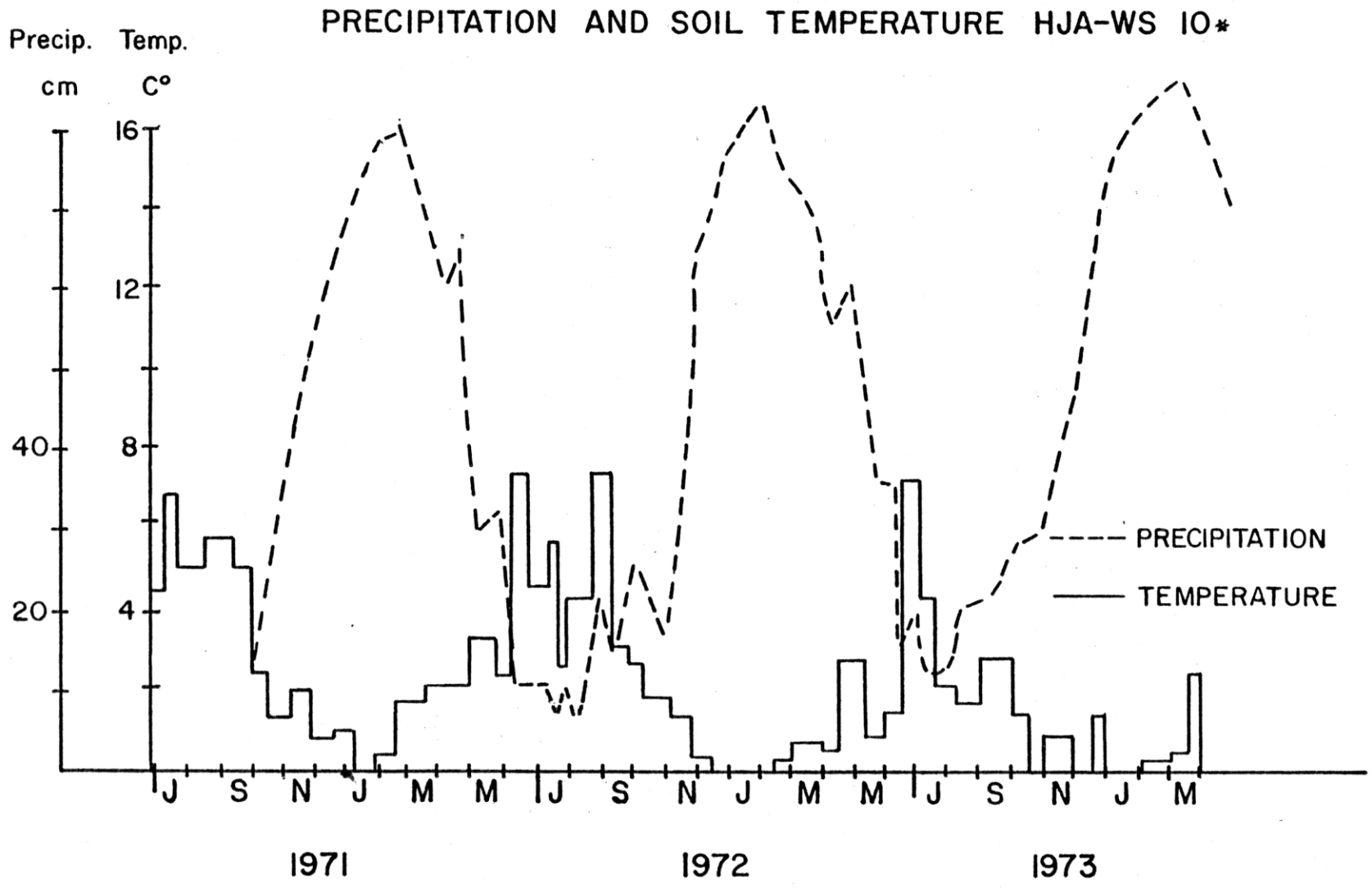


Figure 5. Precipitation and Soil Temperature HJA-WS 10
* Compiled by R. L. Fredriksen, R. A. Mersereau and A. B. Leveno, USFS, Corvallis, Or.

Temperatures rarely dip to -18°C (0°F). During the summer, particularly on exposed south and west facing ridges, temperatures may exceed 38°C (100°F) (Berntsen and Rothacher, 1959).

Vegetation

Vegetation distribution patterns in the H.J. Andrews illustrate clearly the impact of the two principal environmental stress gradiants on vegetation, temperature, and moisture. Fire and mass movement disturbances are important events in resetting the successional clock, which accounts for the large expanses of Douglas fir in the Cascades.

Dryness, Franklin and Moir (1974) described forest communities in the Central Cascades. They found that two major zones are present in the Andrews Experimental Forest. Tsuga heterophylla climax community dominates in areas which receive approximately 80 optimum growing days per year, while Abies amablis community is found at higher elevations where cooler temperatures limit the growing season. A transition zone occurs at intermediate elevations which includes elements of both communities.

Geology, Geomorphology and Soils

The H.J. Andrews Experimental Forest is underlain entirely by volcanic rocks which appear to have been erupted during three periods of volcanic activity (Pect et. al., 1964; Swanson and James, 1975) (fig. 4). Lookout Creek has incised 715-925 m (2,500-3,000 ft.) into

the volcanic units, exposing at its mouth the Little Butte Fm. (Oligocene-Early Miocene). Basal members of the exposed section are predominantly massive, green blocky breccias derived from pyroclastic flows and mudflow events. Upsection, welded ash flows include some lacustrian and airfall ash derived deposits. Included within the section of Little Butte Fm. pyroclastic rocks are a few basalt flows which have been hydrothermally altered by subsequent volcanic and plutonic activity and deeply percolating meteoric water. Locally the Little Butte Fm. is intruded by many feeder dikes from younger formations.

During middle and late Miocene thick sequences of tuffs, welded and unwelded pumice-bearing ash flows were deposited on the eroded Little Butte landscape, 150 M (500 ft.) relief. This unit has been named the Sardine Fm. by Peck et. al. (1964). Judging by the unwelded nature of the majority of deposits, large size of the lappilli and included lithic blocks, the vents appear to have been local. Predominantly hypersthene andesite flows were erupted infrequently during the sardine volcanic era.

The final period of activity initiated during the mid-Pliocene, 5-7 m.y. B.P.. The contact between Sardine Fm. and these "Pliocascade Volcanics" (informal name applied to Pliocene lava flows by E.M. Taylor, OSU, personal communication, 1973), is generally marked by fluvial and lacustrian deposits and pillow basalts. These occur at an elevation of about 1345 M (4,500 ft.) in the Andrews Forest area.

The Pliocene volcanic rocks are distinctive mineralogically and chemically from the Sardine rocks in that they contain olivine and their bulk chemical composition is characteristically basaltic. Pliocascade volcanism continued to about 3.5 m.y. b.p. High Cascade volcanism occurred from approximately 2 m.y. b.p. to the present, but lavas did not enter the present area of the Andrews Forest after 3.5 m.y. b.p.

Erosion of the present day Lookout Creek drainage commenced as Pliocene volcanism ceased. During the Pleistocene, repeated glaciation accelerated erosion. Cirques and U-shaped valleys were carved in the upper part of the drainage, the character of which has been modified by rapid stream erosion and repeated mass movement events. Fluctuations in the erosion-sedimentation regimes during and since the Pleistocene has resulted in the formation of stream terraces along many streams in the Cascades.

Other major processes involved in the formation of the present landscpae and soils are slow deep-seated mass movements, soil creep, shallow soil slides and debris torrent events. These erosion mechanisms and the landforms they create can be related to the manner in which a particular rock type weathers, or the juxtaposition of two types of rock which have different weathering and strength characteristics (Swanson and James, 1975).

Soils studies in the H.J. Andrews Forest have focused on continually refining the understanding of the complex soil-bedrock-

landscape relationship. Weathering rates and soil formation are intimately tied to landscape characteristics, such as slope, aspect and the nature of land sculpting processes as well as geologic substrate.

Rothacher et. al. (1967) described morphologic and hydrologic characteristics of soils on the three experimental watersheds located near the mouth of Lookout Creek. Using the previous soil classification shceme the soils were classified as Regosols, Reddish-Brown and Yellowish-Brown Laterites, and a lithosol (Soil Survey Staff, 1951). Over much of the steep watersheds pedogenic influences are minimal. The soils are quite shallow, except on toeslopes and in slump prone areas where soils are formed on deep deposits of unconsolidated soil and rock material such as massive earthflow and colluvial material which respons hydrologically as a deep soil. Infiltration rates are generally high; overland flow is restricted to stream channels and exposed bedrock surfaces. Dryness (1969) concluded that the stone content of the soils, which ranges from 5 to 76 percent by volume over the three watersheds, is the most significant factor causing variation in soil moisture storage capacity.

Dyrness (1967) made a survey of mass soil movements which occurred in the winter of 1964-1965. A disproportionately large number occurred where roads and landings were constructed on pyroclastic bedrock, particularly in areas of "greenish" tuff breccia. Slope angle was correlated with mass movement events. Eighty-three percent of the slides occurred on slopes between 45 and 100 percent. Soil mass movement was uncommon on slopes with a south or southwest aspect where

soils are generally more shallow, possess higher clay content and are redder than cooler north facing slopes. Morrison (1975) has observed similar relationships among aspect, soil depth, and slope stability in another Western Cascade drainage.

Paeth (1970) and Paeth et. al. (1971) studied soils developed from pyroclastic rocks in the H.J. Andrews Forest to determine genetic and stability relationships. They found that the unstable soils were in the fine, montmorillonitic, mesic families of Ultic Hapludalfs and Typic Haplumbrepts (Soil Survey, 1970). The more stable soils which developed from the "reddish" tuff breccia are members of the clayey, mixed, mesic family of Humic Hapludults.

The unstable soils, derived from the greenish parent material, had much higher concentrations of montmorillontic 2:1 type clays throughout the profile than soils formed on the reddish breccia. Probably more significantly, the reddish breccias are parent materials of soils relatively rich in free iron-oxides and hydroxides, suggesting that iron compounds act as a cementing agent, which may be significant factor in the resistance demonstrated by those soils to shear failure.

Paeth (1970) concluded that undispersed psuedomorphs of clay after plagioclase, sesquioxide secondary minerals and glass increased the internal friction of the red tuff soils, reducing their potential to shear, while the soils containing 2:1 clays were susceptable to shearing when saturated.

Brown (1975) studied soils formed on three characteristic landform

sequences in the Andrews Forest, an alluvial terrace, glacial-colluvial terrace-side slope complex and a ridge-side slope sequence. He concluded that slope aspect and steepness could be consistently correlated with soil depth and chemistry. Warmer, shallower soils formed on south and southwest facing slopes were generally less intensively leached (higher base saturation) and had a lower percentage of clay, contrary to similar research in other regions. He found slope steepness affected soil development. Amorphous material content appeared to be greater while exchangable bases and base saturation lower in soils formed on steeper slopes in contrast with lower angle slopes found on ridgetops and gently sloping terraces at the foot of the hill slopes. Higher contents of organic matter in the soils formed on steeper slopes were interpreted as reflecting the lower microbial activity in such soils due to earlier desiccation than in soils formed on lower angle terrace surfaces which remained moist much longer in the summer.

Exposure to higher winds could create more woody litter in stands on upland slopes which would decay more slowly than predominately needle and twig litter in more protected areas, as well as promoting rapid evaporation. Phillips (1976) found that soil moisture was the major regulating factor of decomposer activity during the growing season, while temperature was the principal control in the winter.

Brown (1975) also found that the saprolite horizons had higher base saturations than overlying horizons, due in part to slow movement

of soil water through the dense, weathering bedrock material. This would promote adsorbtion onto clay mineral exchange sites. Others have proposed that less hydroxylation is found at depth in soils due to less frequent wetting and drying cycles (Buol, Hole, and McCracken, 1973; Jackson, 1968). This affords the deeper soil clay particles more negatively charged adsorbtion sites for other cations. It has also been suggested that translocation processes select for electrostatically stable clay micells which would be nearly fully saturated before they moved to the subsoil (Jackson, 1964). Disruption of soil forming processes by mass movement processes on steeper slopes would disrupt biological processes as well. In general in the western Cascades, the high rainfall and moderately acid coniferous litter gives rise to soils which have fairly low base saturation, in comparison with less steep areas and deciduous forests or grasslands. Soil variability is extremely high in such a mountainous area, which requires that each aspect of the multi-variable soil forming process be considered at each site.

METHODOLOGY

Field Methods

Sample Site Selection

Sample sites were selected on the basis of the following criteria: (1) geomorphic situations in which soils showed a minimum of mixing or movement by slope processes, (2) representation of the three major lithologies in duplicate, (3) opportunity for utilization of related research in the Andrews Forest, i.e., soils analysis, stream and soil water chemistry data, etc., and (4) accessibility with a drilling rig.

Soil Sampling Methods

Soil pits were excavated to bedrock or as close to the soil-bedrock contact as possible. Pit face samples were collected from each horizon or every 20-30 cm (8-12 in.) where horizonation was indistinct. On deep profiles, a 8.9 cm (3.5 in.) bucket auger was used to extract samples. Samples weighed approximately 800 gm moist weight. Vegetation and geomorphic features were noted with the profile description.

Samples were homogenized and split into three equal portions for each of three phases of analysis: x-ray flourescence, x-ray diffraction and thin section analysis.

Water Sample Data

Nutrient analyses of stream and soil water samples were available from R. Fredriksen and D. Moore, U.S.F.S., M. McCorison and J. Kristaponis O.S.U. and R. Naiman, O.S.U. All samples were analyzed by identical procedures in the central water analysis laboratory at the Forest Sciences Lab, Pacific Northwest Forest and Range Experiment Station, Corvallis, Oregon. Available information included nutrient content of precipitation (McCorison) litter leachate (Moore, McCorison and Kristaponis), soil solution at three different depths (McCorison and Kristaponis), and stream runoff data for Watershed 10 (R. Fredriksen), which is underlain by red tuff breccia (RTB) and buff colored tuff (SHBT and RTBT). Stream chemistry was also made available for Lookout Creek (R. Naimon), Watershed 8 (SR) (R. Fredriksen), and Devil's Club Creek (MB) (R. Naimon).

Laboratory Methods

Major Element Analysis

Samples were all analyzed as total bulk samples including organic matter, gravel, sand, silt and clay fractions, except for samples of a stoney glacial till, a soil derived from a Pliocene flow and part of the red tuff breccia. They were separated with various methods into lithic and soil components of size classes to be analyzed independently.

In one approach the lithic fragments were separated from the rest of the bulk sample because the parent material did not seem to be gradationally represented in the soil, but rather, present as nearly

unweathered rock fragments in a matrix of highly organic soil material. A glacial till and an alpine soil formed on Pliocene flow rocks showed such clear differences between minimally weathered parent material and organic soil matrix that soil materials were analyzed separately from lithic fragments larger than 1 cm.

In an attempt to see whether elements are concentrated in different size fractions of the soil the soil of RTB was separated into gravel, pebble, sand and clay size fractions using sieve sizes 5.6 mm, 1.0 mm, and .074 mm. The nontraditional sizes are due to a sieve shortage in the laboratory at the time of preparation.

Samples were analyzed for nine major elements using x-ray spectroscopy. Expressed as oxides, they were SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MgO, MnO, CaO, K_2O and P_2O_5. Sodium was determined using Instrumental Neutron Activation Analysis (INNA).

Sample Preparation

Sample preparation and XRF analytical methods used were developed by Norrish and Hutton (1964) for analyzing soils in Australia. The staff of the Geology Department at the University of Oregon has modified the methods for use with a Ge-7 x-ray spectrometer.

Samples were oven-dried at 105°C for 12 hours. Using the quadrant microsplit method, a subsample of approximately fifty grams was separated for analysis. It was ground and pulverized to less than 2 microns. Crystalline H_2O and organic residues were removed by heating to 450°C for eight hours. Fusion of a mixture containing .57 gm dried

sam ple, 3.00 gm. Spectroflux 45, and .004 gm sodium nitrate is carried out in a platinum crucible at approximately 1100°C. The homogeneous melt is poured into inert carbon molds and pressed flat. The glassy disc is annealed for 48 hours in a very dry environment. If properly prepared, a smooth surface will fill the mask slot in the sample piston on the spectrophotometer.

USGS standard rocks, AGV-1, BCR-1, and SiO_2-A were used as mutual standards throughout all analytical runs. Data reduction programs compare unknown samples with counts obtained from the standards, make adjustments for mass absorbtion coefficients for each element and delete the effect of interference among elements.

Sodium concentration in each sample was analyzed by INNA at the University of Oregon under the direction of Dr. Goles. The samples were irradiated at the Triga reactor, Oregon State University.

Clay Minerology by X-ray Diffraction

A sample of sufficient size for clay mineral analysis could only be separated from OGT, SR, SB, RTB, SHBT, and RTBT. Certain horizons were so deficient in clays that no sample was available, and the soils forming on Pliocascade Volcanics were so rich in organic matter and amorphous materials the clays size fraction could not be dispersed for separation.

A sedimentation procedure developed by Mackenzie (1956) was used to separate the clay fraction from each soil sample. Material which remained in suspension after 12 hours was considered to be the less than two micron size fraction.

In accordance with laboratory procedures established by Dr. M. Harward, OSU, (unpublished guide, 1971), the dispersed clay fractions were flocculated with 10 ml 1 N $MgCl_2$ and further concentrated by centrifugation at 6000 rpm for 15 minutes. Two-fifths of the sample were removed for K saturation. Excess $MgCl_2$ was removed from the sample by repeating a wash cycle (shaking in DH_2O for 30 minutes, followed by recentrifugation) until no precipitate of AgCl formed in the presence of $AgNO_3$. The two-fifths separated earlier are saturated with KCl and treated in the same manner as the Mg saturated slides.

The saturated clay samples are spread on clean petrographic slides with a microspatula (Theisen and Harward, 1962). Mg saturated clays are spread onto three slides while K saturated clays are spread onto only one slide.

Analytical runs include Mg in 54 percent relative humidity, Mg saturated with ethylene glycol and the other slide saturated again with glycerol. The slide saturated with potassium is analyzed in dry air, at 300°C and at 550°C.

Analyses were carried out in the clay mineralogy laboratory of Dr. M. Harward at Oregon State University, Soils Department, using a Norelco, model 21, x-ray diffractometer which had been modified with the addition of a sealed chamber which allows controlled atmosphere analyses to be carried out. Air, passing through either a column of drierite ($CaSO_4$) or bubbled through a supersaturated solution of $MgCl_2$, can be maintained at 10 percent or 54 percent relative humidity, respectively.

Thinsection Preparation and Microphotography

Thinsections were made in the standard manner with rocks which exhibited sufficient solid internal strength but for soils and weathered materials which were too friable to slice to .003 mm, impregnation with casting resin after the procedures developed by Norgren (1962) were used successfully.

Thinsections were used for making parent material descriptions and pointing out clay/parent material relations, root/mineral relations and other notable soil features. Microphotographs were taken at 30X, 50X, 200X, 312X, 500X and 1250X in both plain and polarized light.

RESULTS

Results are partitioned into, (1) sample site descriptions with a map showing their location in relationship to the different bedrock units, (2) detailed descriptions of parent material petrology and petrography, (3) bulk sample x-ray flourescence (XRF) chemical data, (4) XRF data for textural separates of red tuff breccia (RTB) and (5) traces of x-ray diffraction (XRD) analysis of clay mineral fractions.

Sample Site Descriptions

Descriptions of each sample site include texture and composition of parent material, geomorphic situation, vegetation zone (Franklin and Dyrness, 1973) and elevation of the site. Sample site names are based on bedrock classification and geologic period of origin.

Late Pleistocene Till (LPG)

LPG is a mixture of clay, silt, sand and platy lithic fragments, predominantly olivine basalt, in poorly sorted lenses. Soil development is slightly below 1.6 M.

The site is located near the backslope boundary of Pleistocene glacial terrace (fig. 4), in a shallow depression which is moist much of the year. A distinctive dark organic matter-rich epipedon and

moisture-loving vegetation are consistent with unusually high soil moisture. In this regard, it is observed that the area is buried by snow during 5-6 months and lenses of clay inhibit drainage. The site is located in a small meadow in the transition zone between Tsuga heterophylla and Abies amablis communities (Franklin and Dyrness, 1974) at 1250 M.

Pliocene Olivine Basalt Flow (Pliocascade Volcanics) (PFF)

Dark organic matter laden epipedon mixed with coarse fragments of slightly weathered olivine basalt parent material creates a loose, permiable solum. This site is located on the crest of Frissel Ridge at 1600 M. in the Abies amablis zone.

Pliocene Olivine Flow Breccia (Pliocascade Volcanics) (PF)

PF is a soil formed on flow breccia of PFF olivine basalt unit. Friable, porous nature of parent material has allowed weathering to proceed to much greater depths (1.4 M.) than in PFF. This site is at approximately the same locale with similar vegetation before logging in 1967. Xerophyllum tenax composes 80% of the cover today.

Pleistocene Fluvioglacial Deposit (OPG)

OPG Site is located on a wedge of colluvium from adjacent slope mixed with alluvium from a small stream which flows onto a Pleistocene glacial terrace. The deposit is characterized by poor sorting, mixture

of Sardine age tuff breccias and flow material. The stone content reaches 30-40% and clay skins are prevalent in the B-horizon. Sample was extracted from a site on a north facing slope above Mack Creek at 800 M. Vegetation is chiefly of the Tsuga heterophylla zone.

Miocene Hypersthene Andesite Flow (Sardine Fm.) (SR)

SR profile is located on a prominant valley filling Sardine andesite flow, which now stands out as a ridge crest. Phenocrysts of plagioclase and hypersthene are apparent in hand specimens. Despite bedrock integrity at 116 cm, preventing deeper penetration, chemical data indicate considerable weathering which is an indication of the length of exposure time to agents of weathering. A buried A-horizon is visible at 12 cm. The site is located at 1130 M in the Tsuga heterophylla to Abies amablis transition zone.

Miocene Tuff Breccia (Sardine FM.) (SB)

The SB parent material is massively bedded non-welded pyroclastic deposits bearing large pumice shards and lithic fragments is weathered to 5 or 6 M. In the deepest samples zeolites were identified in hand specimens. The site is located on the crest of the interfluve between McCrae and Lookout Creek at 810 M. Vegetation is within the Tsuga heterophylla zone.

Mack Creek Miocene Tuff Breccia (Sardine Fm.) (MB)

This rock unit consists of massively bedded, pyroclastic flows

or deposits which were subsequently altered hydrothermally. Rock texture and pumice content is similar to the nonwelded breccias of SB but the rock is much harder and weathering has proceeded at a much slower rate. The site is located on a steep north-facing slope at about 660 M beneath a 450 year old stand in the Tsuga heterophylla zone.

Oligocene Buff Colored Tuff/Ridgetop (Little Butte Fm.) (RTBT)

These weakly bedded deposits of airfall ash have variable texture with grains ranging in size from lapilli to silt to pebbles. Random inclusions of lithic fragments can be larger. Presence of leaf fossils in a shallow near shore environment indicate that deposition occurred in a lacustrian basin. RT and SH buff tuff are more deeply weathered than ashflow tuffs deposited above and below in the sequence. The site is located on a ridgecrest at approximately 660 M. The ridge top soil profile is more completely developed than most soils studied, which indicates uninterupted pedogenesis since freeze-thaw disturbance during the Pleistocene. Vegetation is typical of Tsuga heterophylla zone.

Oligocene Buff Colored Tuff/Sidehill (Little Butte Fm.) (SHBT)

The parent material of this unit is thought to be identical to RTBT. The major difference is evidence of colluvial slope processes and better drainage in SHBT. The site is located 4 M vertically below

the ridgetop site. The soil pit revealed little horizonation and the bowed tree boles suggest regular soil creep. Vegetation is well within the Tsuga heterophylla zone.

Oligocene Red Tuff Breccia (Little Butte Fm.) (RTB)

The parent material is a reddish welded tuff breccia which has weathered to a moderately shallow, skeletal soil. A significant portion of the soil gravel to pebble size fractions are composed of bedrock derived lithic fragments. In addition thinsection analysis reveals that some are concretionary clasts. The site is located on a broad ridgetop at 660 M in the Tsuga heterophylla zone.

Parent Material Petrology and Petrography

Petrology and petrography of each parent material was determined by thinsection analyses. Samples which lacked sufficiently unaltered bedrock specimens were studied by using least altered lithic fragments in gravelly soil samples.

Late Pleistocene Till (LPG)

Lithic fragments in LPG are primarily olivine basalt, for which a complete mineralogical description is found below (PFF). Glacial erosion and transportation to the terrace site occurred during the latter part of the Pleistocene based on the thickness of weathering rinds and degree of ferric iron buildup in the soil. Lenses of crudely

sorted silts and sands are interbedded with more poorly sorted deposits of boulders mixed with fine sediments. Nearly pure lenses of clay are randomly interspersed throughout the deposit.

Pleistocene Fluvioglacial Deposit (OGT)

Lithic fragments at the surface in OGT are predominantly hypersthene andesite flows or tuff breccias (Sardine Fm.) deposited from adjacent ridges by colluvial and alluvial processes. Descriptions of both rocks are found below (SR and SB). Road cuts provide a glimpse of a deeply weathered glacial till, presumably deposited earlier than LPG, underlying the fluvial and colluvial deposit sampled in OGT.

Pliocene Olivine Basalt (PFF)

Sixty-five percent of PFF is composed of bytownite-labradorite plagioclase. Plagioclase of this composition has been shown to contain CaO and Na_2O in the approximate quantities, 12-15% and 2-4% respectively. SiO_2 makes up 50% and Al_2O_3 30% of this type of plagioclase (Deer, Howie and Zussman, 1966).

Pyroxene is present as a fine grained titanoaugite making up approximately 25% of the groundmass. CaO constitutes approximately 10% and MgO, 15% of this mineral by weight.

Olivine makes up approximately 6% of the basalt. It is composed of 3-5% MgO and 50-60% Fe_2O_3. Magnetite was the most common accessory mineral at 3-4% and it contains 95% Fe_2O_3 by weight.

Based on petrographic observations of PFF the weatherability sequence of the primary minerals follows the sequence from most to least weatherable: olivine, aphanitic groundmas, plagioclase An_{60-70}, titanoaugite and magnetite.

Pliocene Olivine Basalt Flow Breccia (PF)

The parent material composition is nearly identical to PFF but due to its more rapid cooling history, crystalization was not allowed to proceed so far nor were all the volitile elements allowed to escape. The resultant texture, lower density and higher initial glass content, promotes more rapid weathering. Large accumulations of an amorphous material in the porous bedrock is also indicative of a rapid weathering rate and weathering conditions considerably different from those of adjacent PFF.

Miocene Hypersthene Andesite (SR)

The bedrock samples taken were not acceptable as unweathered parent material, so thinsection data were obtained from less weathered gravel fragments. Prominant plagioclase phenocrysts are bytownite An_{70-90} which make up 20-30% of the total rock volume. CaO can reach 14-17% in bytownite (Deere, Howie and Zussman).

The pyroxene mineral which makes up approximately 10% of the rock is pigeonite. It is the Mg-rich member of the hypersthene-enstatite solid solution series of minerals in which MgO can reach 12% while CaO makes up only 4% by weight (Deere, Howie and Zussman).

Groundmass minerals include microlitic plagioclase which appears to contain about half as much CaO as the larger crystals (Deere, Howie and Zussman).

From limited thinsection information it appears that Ca-rich plagioclase deteriorates faster than pigeonite in samples of SR.

Miocene Tuff Breccia (SB)

SB is a non-welded pyroclastic breccia which contains moderately undeformed shards of pumice, 20-25% up to 25 cm in length, fine grained ash matrix, 30-35%, some lithic fragments 10-15%, plagioclase phenocrysts (An_{40}), 10-15%, and microlitic plagioclase (An_{60-70}), pigeonite, magnetite and other accessories, 10%.

The plagioclase phenocrysts resist weathering more effectively than the groundmass of microlitic plagioclase An_{60-70}, pigeonite and magnetite. It appears that pumice vessicles serve as ideal microsites for secondary mineral formation. Lithic inclusions exhibit the greatest resistance to weathering.

Mack Creek Tuff Breccia (MB)

MB parent material seemed to be identical to SB in hand specimens so no thinsections were made.

Oligocene Buff Colored Tuff/Ridgetop and Sidehill (RTBT and SHBT)

The buff colored member of the Little Butte Fm. is similar to SB in that, it to, is an airfall ash deposit. The rock is fine grained

with sparse inclusions of lithic fragments. Almost no pyroxene and few phenocrysts of plagioclase were noticed. Only 10% of the water lain airfall ash could clearly be identified as plagioclase (An_{50}).

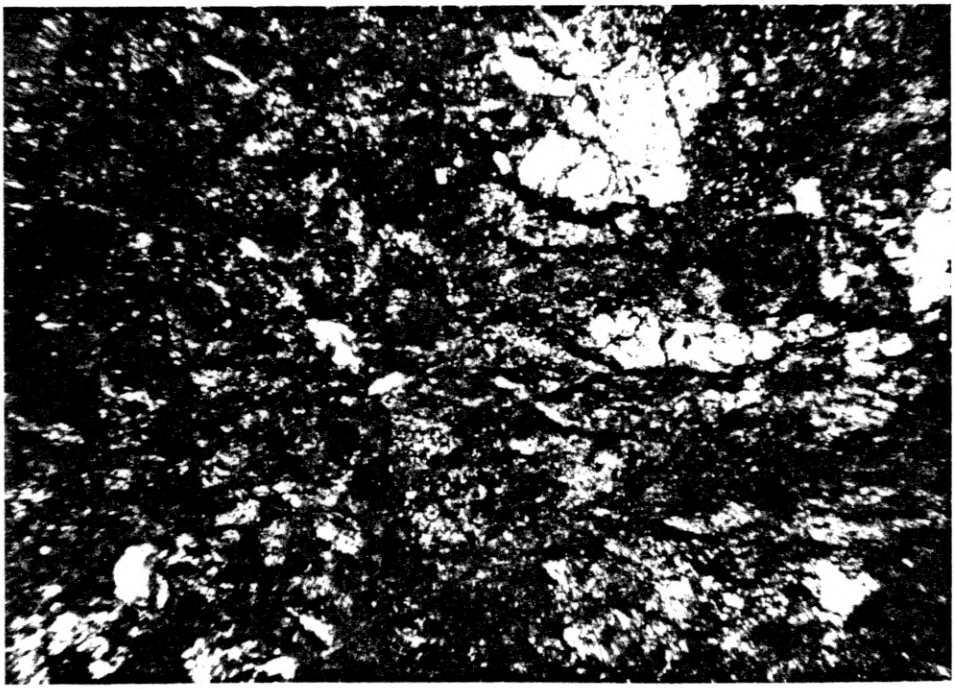
The glassy matrix material appeared to be the most easily weatherable material present. Photomicrographs of typical examples of RTBT and RTB (described below) illustrate the textural differences between welded and non-welded parent materials.

Oligocene Red Tuff Breccia (RTB)

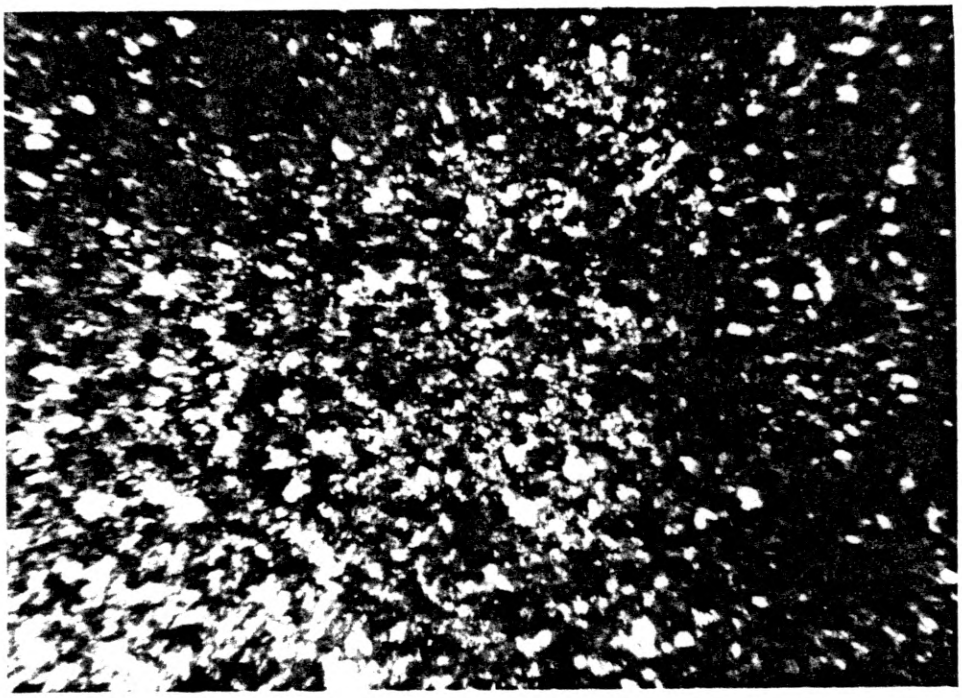
RTB is a welded ashflow tuff breccia made up in large part of foreign lithic fragments (85%). Included in the intersticial matrix material with glassy material are large crystals of plagioclase. Stresses induced by settling and movement in the ashflow during cooling makes difficult the chemical identity of these plagioclase. The slower cooling history of a hot ashflow in comparison to the rapid cooling which airfall ash grains experience is apparent by the presence of more weathering resistant minerals and less volcanic glass in RTB. This relationship is augmented by a more compact, impermeable texture which would slow chemical weathering.

Secondary Mineralogy

Diffractograms of samples which yielded sufficient clay material for analysis are presented in figures 7-12. The rationale used to interpret results of each treatment are presented below.



Little Butte RTB



Little Butte RTBT

Figure 6. Visual Comparison of ashflow tuff breccia (RTB) and airfall tuff breccia (RTBT) at 300X.

Interpretation Procedure

Montmorillonite and other 2:1 clays exhibit a distinctive 14 Å peak (6.5° 2° θ) saturated with Mg in 54 percent relative humidity. The peak will reflect the expansion of the clay mineral structure to 16.9 to 18 Å (5.5° 2 θ) upon solvation with organic molecules ethylene glycol or glycerol if the mineral is a smectite or montmorillonite. Neither vermiculite nor chloritic minerals can expand to accomodate the large organic ions. A 2:1 mineral which has a large quantity of hydroxal ions attached to its surface would not show regular or complete expansion on saturation by organic molecules nor complete collapse upon heating to 550°C (Jackson, 1964). Most of the 2:1 clay mineral in this study is characterized very incomplete expansion and collapse. OGT, SB and SR (Figs. 7, 8 and 9) best illustrate expansion.

Vermiculite clays collapse to 10Å (10° 2 θ) when saturated with K ion. Smectites had montmorillonites can only collapse to 12 Å with K saturation at room temperature if they are not heavily hydroxylated (Jackson, 1964). 2:1 clays in SR and RTBT exhibit much less collapse than those in SB, SHBT, OGT and RTB, with K saturation and heating to 550°C. All profiles containing 2:1 clays indicated less hydroxylation deeper in the profile, based on the larger portion of the peak which responds to solvation and heat treatments.

Well ordered kaolinite is indicated by a first order 7.15 Å (12.5° 2 θ) peak while less ordered, halloysite, the hydrated 1:1 clay mineral is represented by a 7.33 to 10.0 Å peak depending upon its degree of hydration. The most distinctive change in 1:1

minerals occurs at 550°C when they collapse completely. SB and SHBT best display 7Å peaks due to kaolinite clays. The deepest sample from SB (fig. 9) contains the most hydrated halloysite mineral.

2:1 clay can exhibit a secondary peak at 7 Å if it possesses sufficient crystalline regularity. In this study 2:1 clays generally display a broad range of d-spacing and degrees of hydroxylations with very broad 2:1 peaks which would indicate that a scattering of the secondary reflection occurs. In soil profiles where the deterioration of the 7 Å peak is minimal with K saturation, that peak is considered to be almost wholly due to the presence of 1:1 clays. RTBT exemplifies a soil containing both types of clays.

The interpretation of each profile is integrated into the discussion of all the results in Chapter VI.

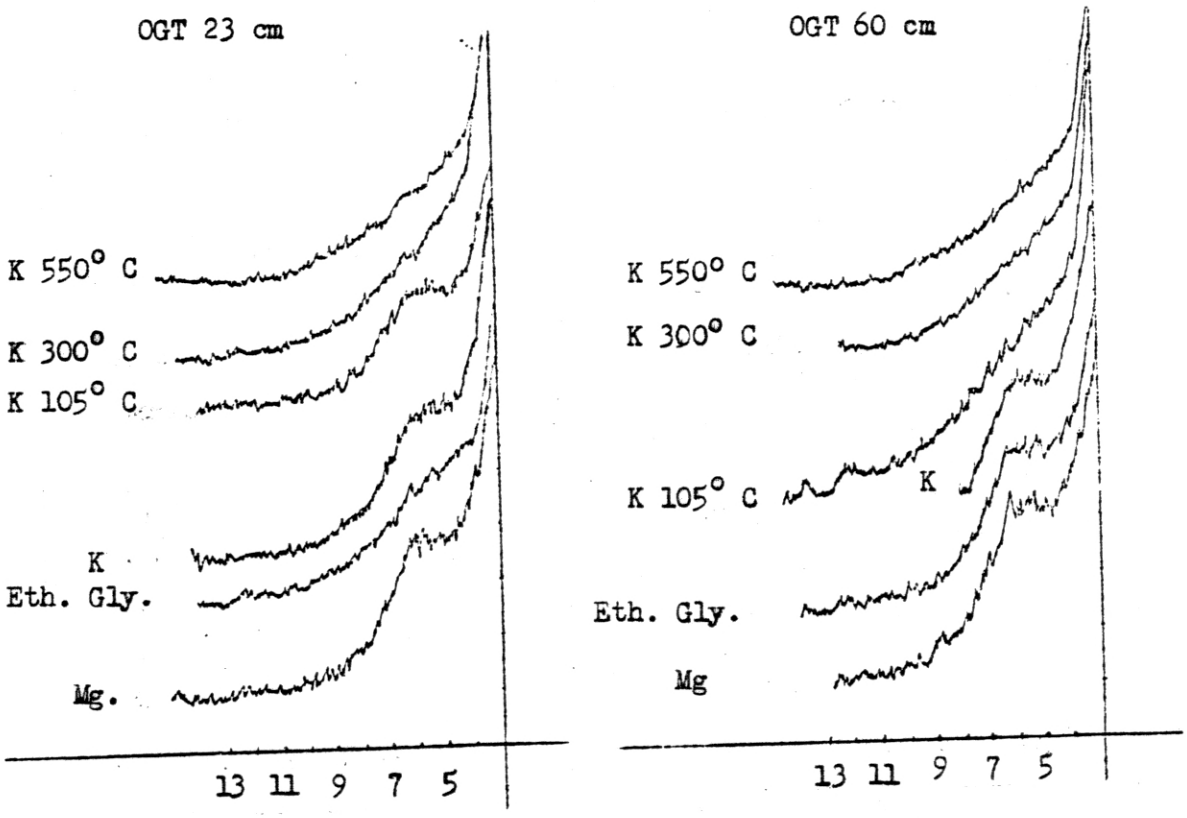


Figure 7. X-ray Diffraction Results for OGT

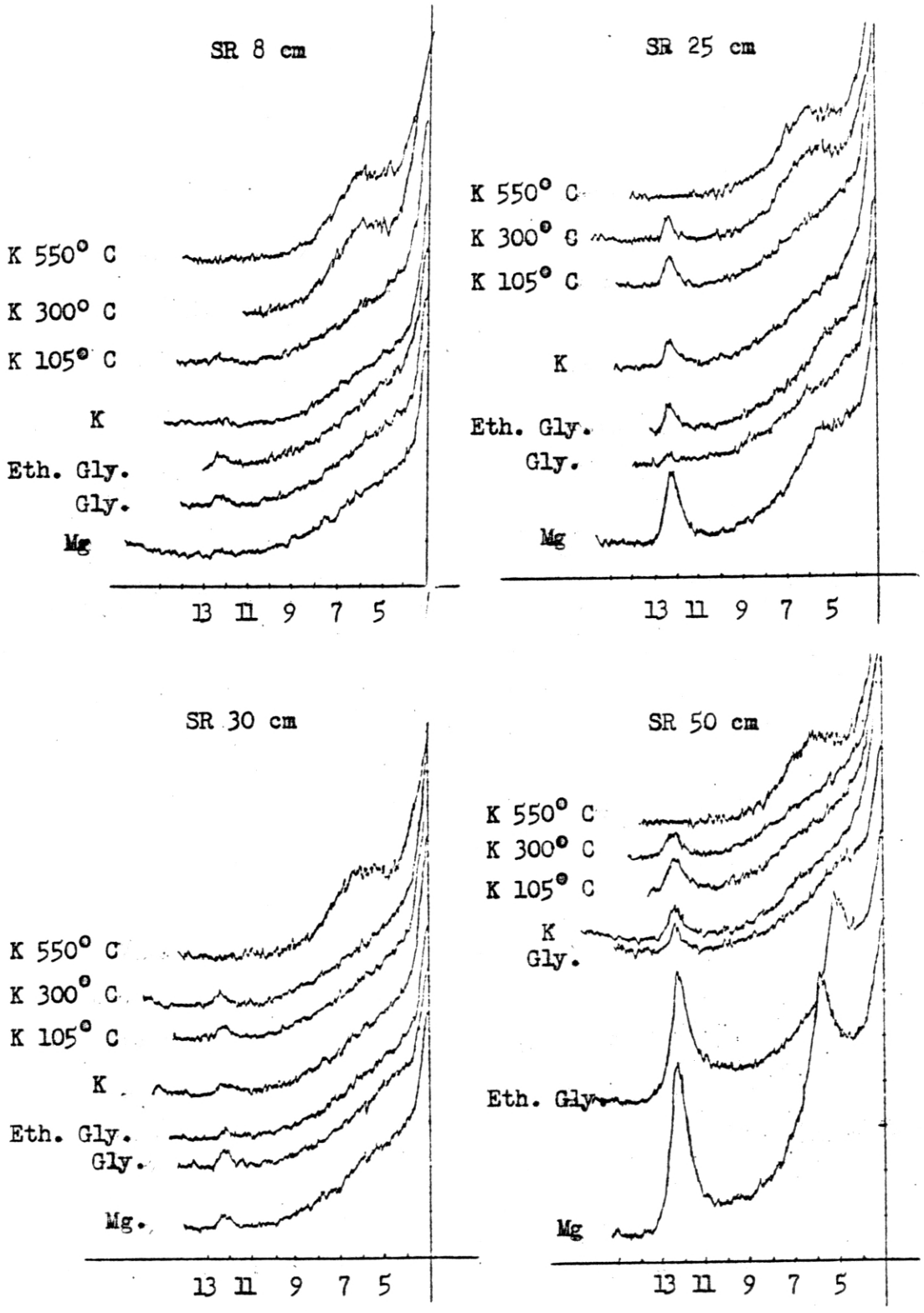


Figure 8. X-ray Diffraction Results for SR

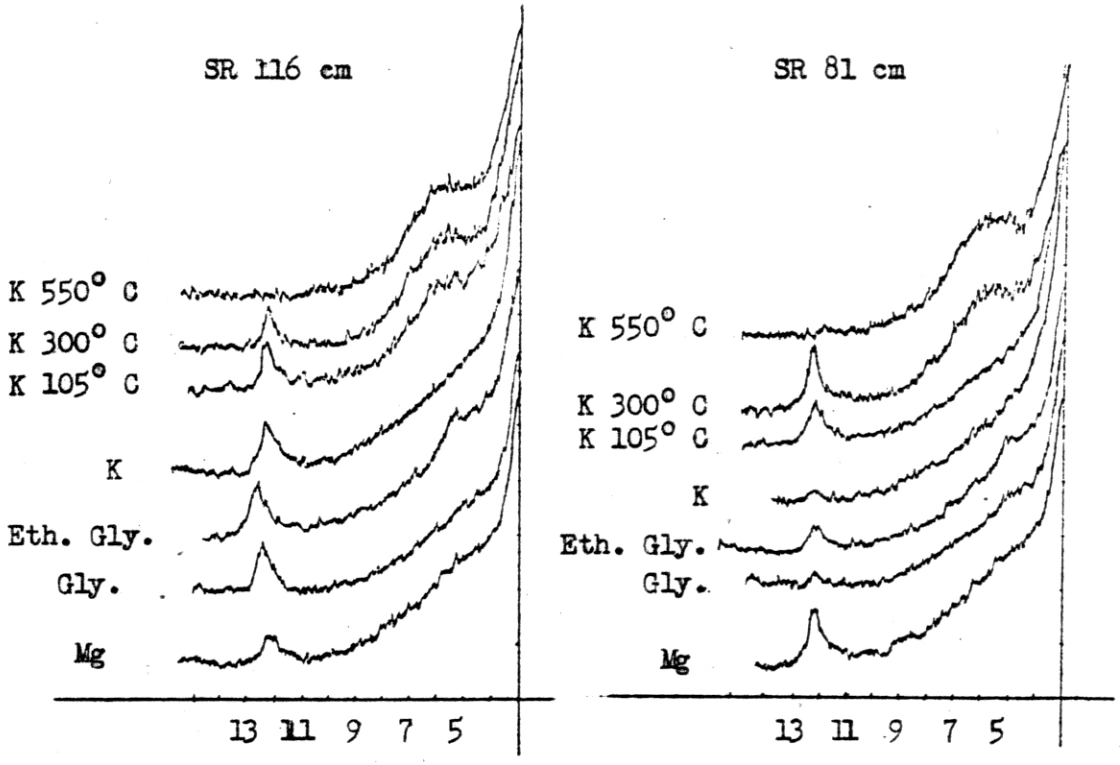


Figure 8. Continued

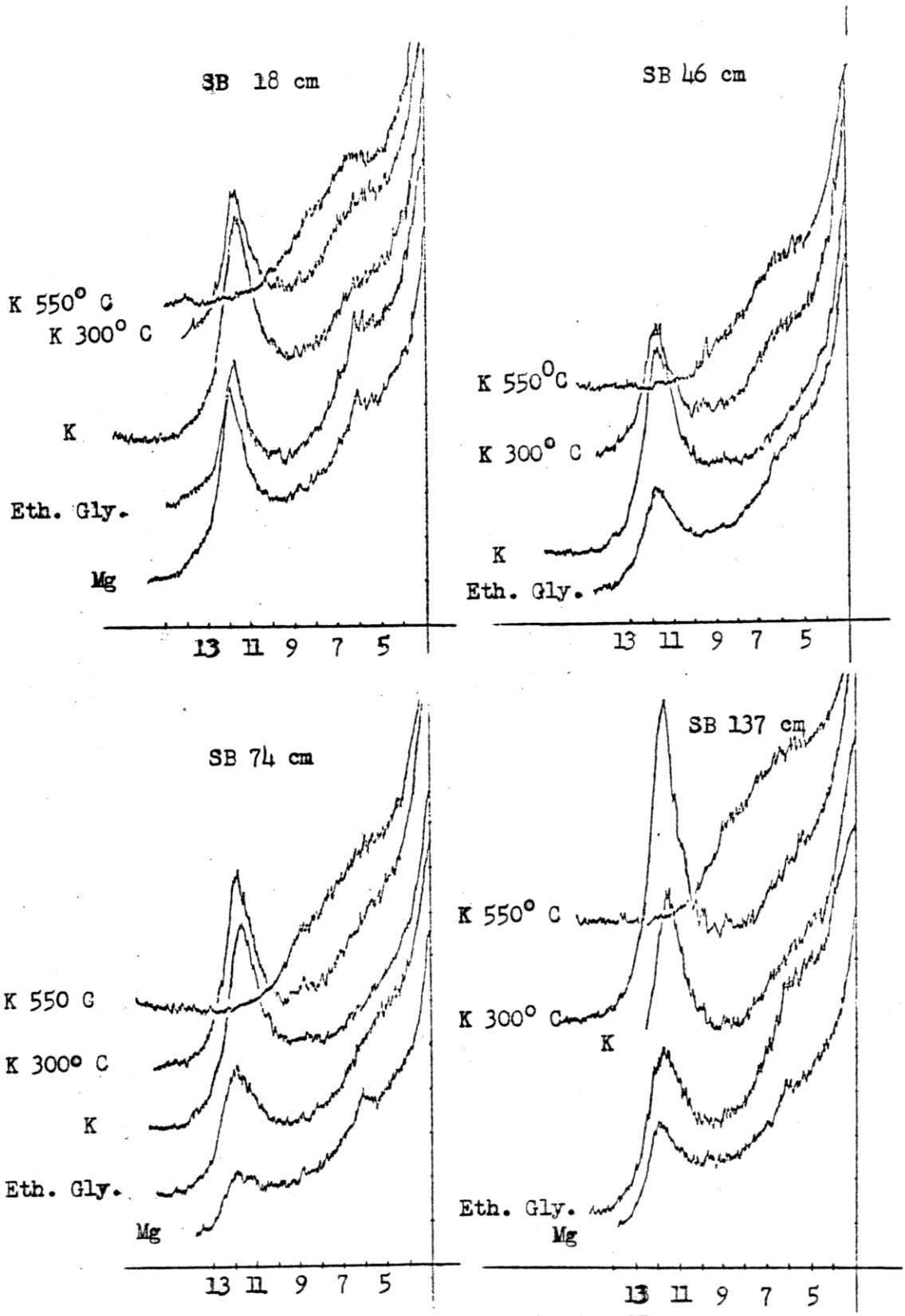


Figure 9. X-ray Diffraction Results for SB

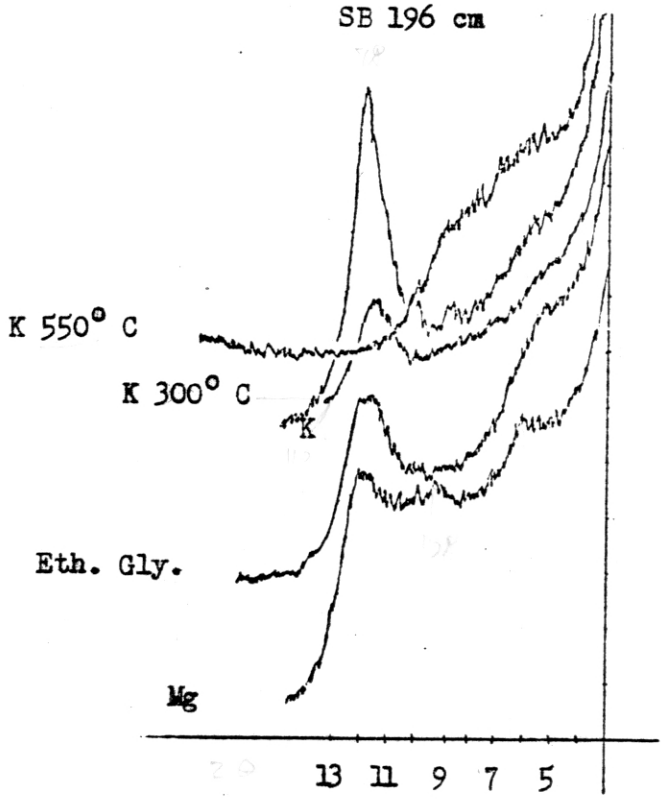


Figure 9. Continued

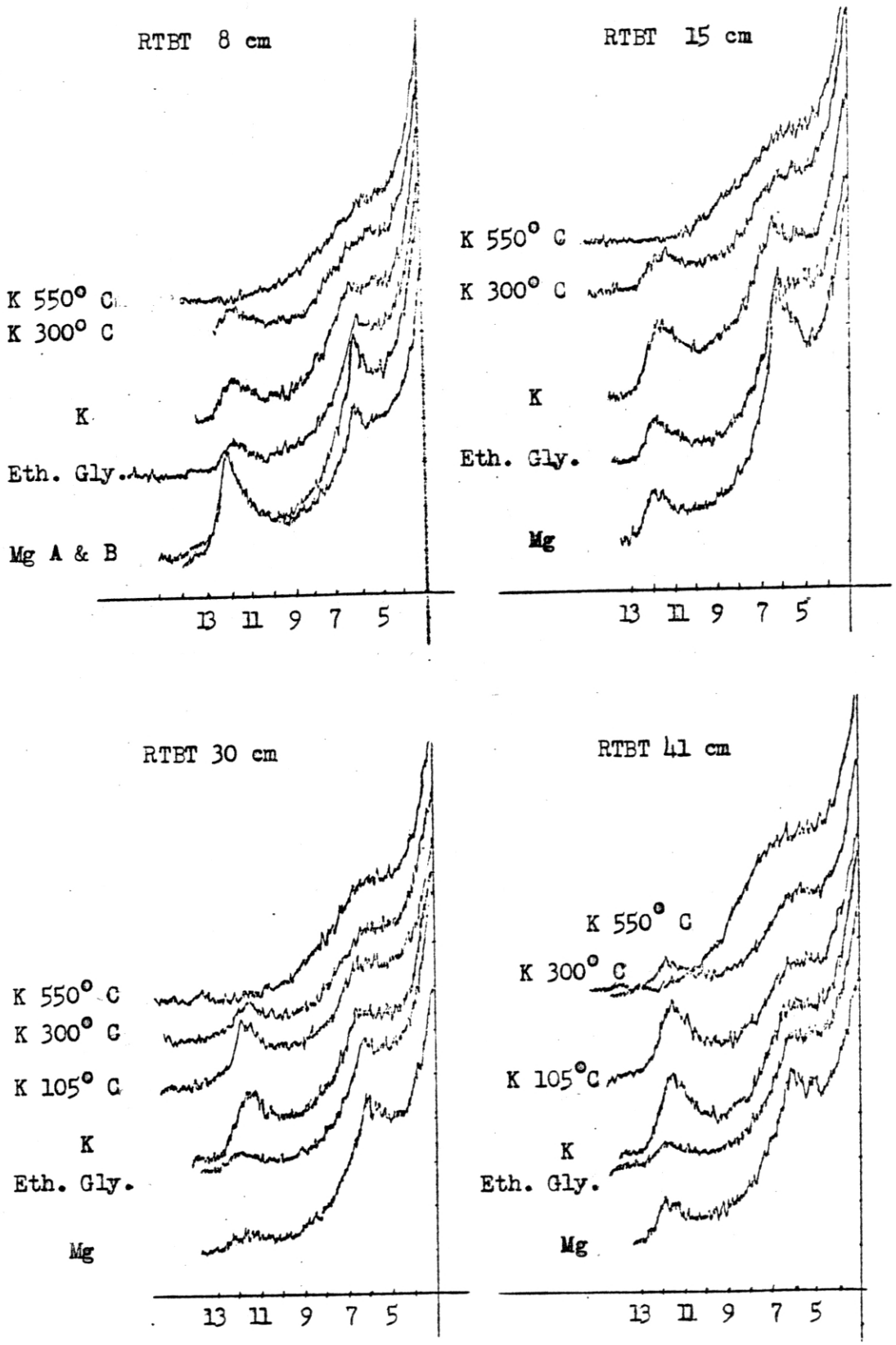


Figure 10. X-ray Diffraction Results for RTBT

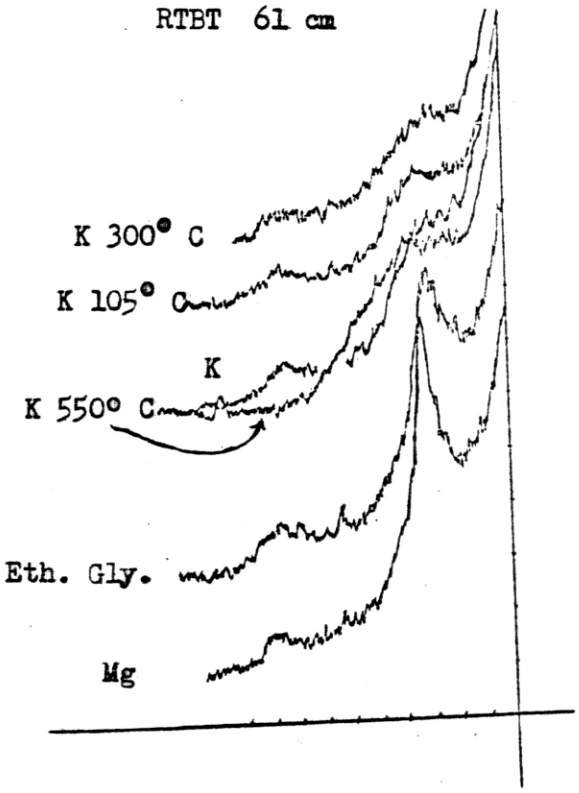


Figure 10. Continued

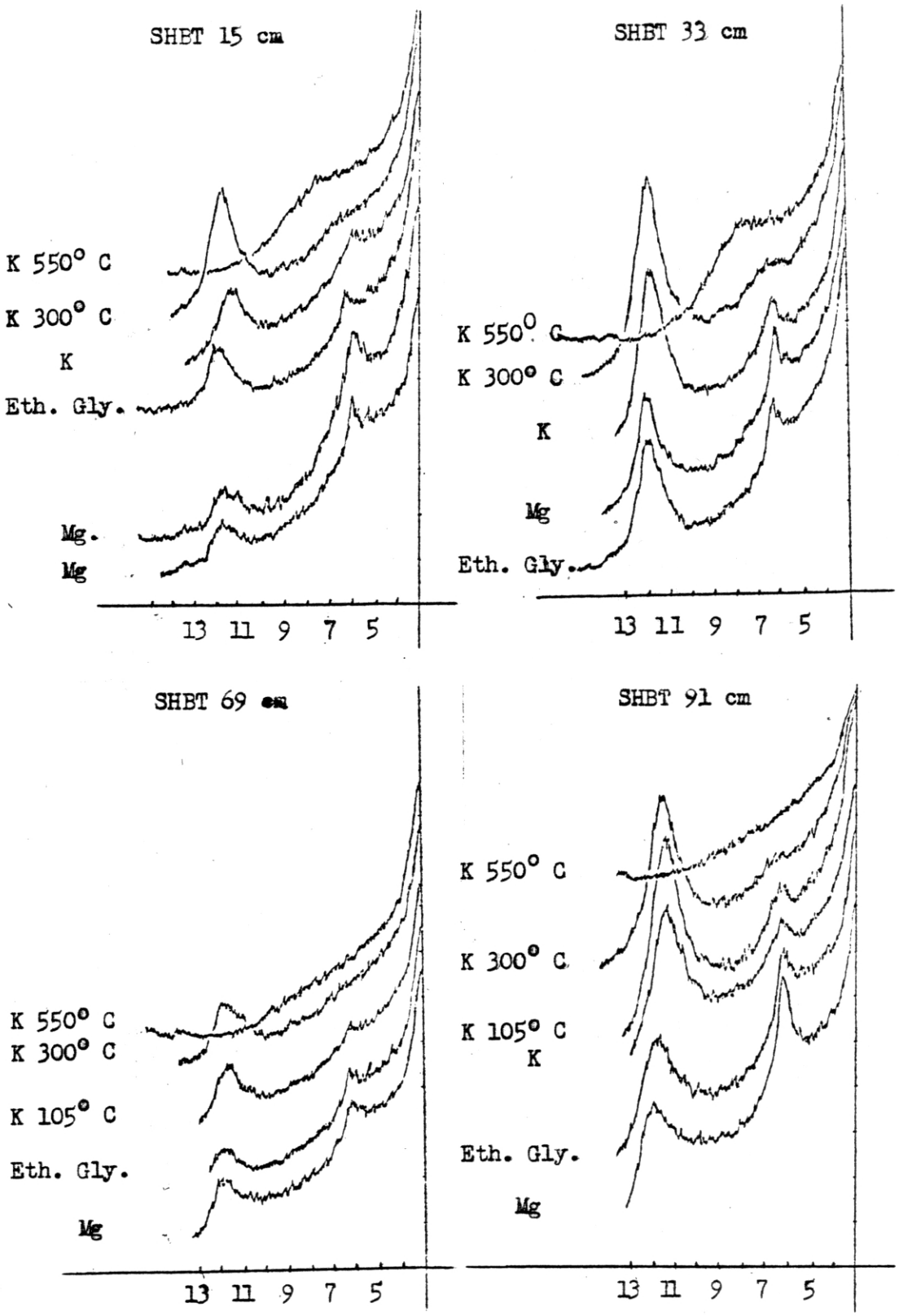


Figure 11. X-ray Diffraction Results for SHBT

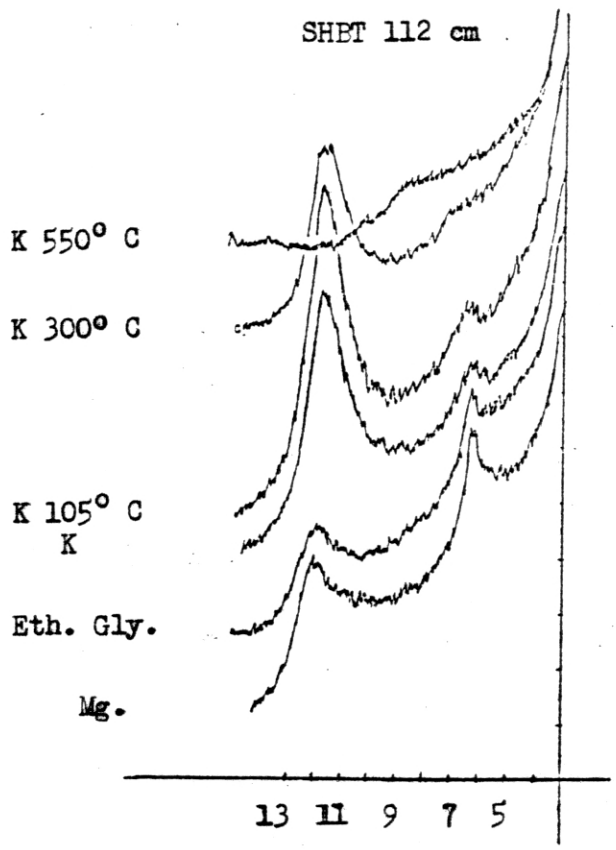


Figure 11. Continued

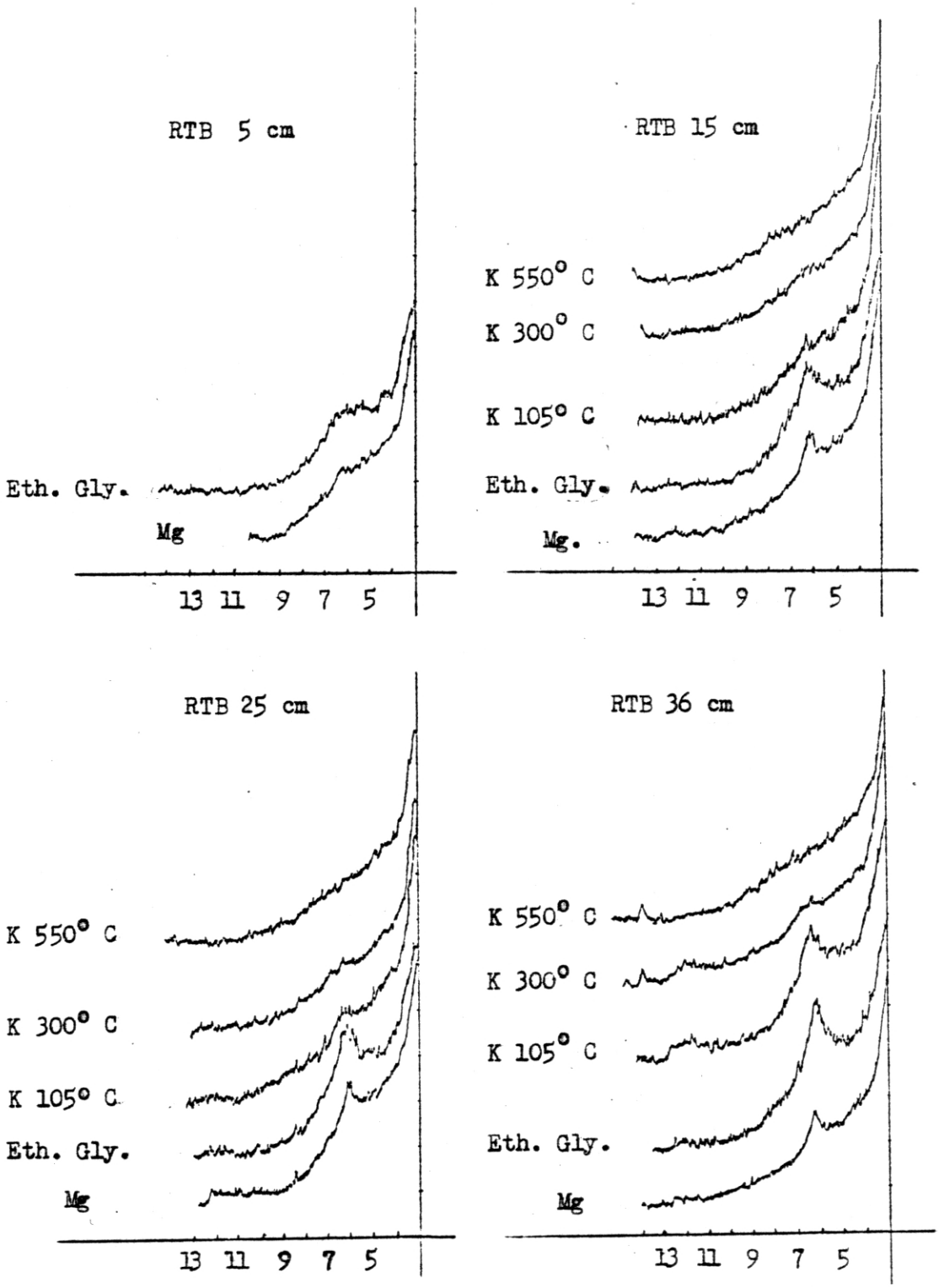


Figure 12. X-ray Diffraction Results for RTB

Total Chemistry of Bulk Samples

Results from x-ray fluorescence analysis (XRF) of the total chemical composition of bulk soil and bedrock samples are presented in Table 6. Results of XRF analysis of RTB textural separates follow in Table 7. Some samples were subdivided into lithic fragments and soil matrix since the two materials were so differentially weathered. PFF and LPG were composed of nearly unweathered lithic fragments in a highly organic soil matrix. In an attempt to determine whether cations are segregated into size fractions of a soil, RTB was subdivided and analyzed as 4 textural subsamples for each level of the soil profile.

Symbols used in the tables include R, which indicates that lithic fragments were separated from the soil matrix and are analyzed separately, and S, is the remaining soil material analyzed separately. Undesignated samples are bulk samples. In PF, R* indicates that the clasts analysed are solid and F indicates that the material is friable and incompetent. PM was used when the deepest sample extracted appeared to represent parent material. In most cases where PF is indicated this was probably not the case.

Integration of the implications of ionic distributions in the soil profiles in comparison to parent materials with secondary mineralogy and optical observations is presented in Chapter VI.

TABLE 6

TOTAL CHEMISTRY OF BULK SAMPLES (%)

Sample	Depth	SiO_2	TiO_2	Al_2O_3	MgO	Fe_2O_3	MnO	CaO	Na_2O	K_2O	P_2O_5	TOTAL
Pleistocene Glacial Till (LPG)												
7-17 cm	S	47.29	1.60	25.36	5.26	11.88	0.25	3.77	2.26	0.87	0.46	99.00
36-46 cm	S	46.61	1.68	24.72	5.85	12.40	0.22	3.69	2.17	0.85	0.45	98.63
56-66 cm	S	45.80	1.72	24.79	6.93	13.03	0.20	3.69	1.84	0.80	0.43	98.59
87-97 cm	S	45.73	1.78	23.85	7.39	13.54	0.21	3.64	1.71	0.75	0.40	99.00
109-119cm	S	46.09	1.74	23.00	7.63	13.45	0.20	3.89	1.86	0.78	0.37	99.01
17-25 cm	R	49.71	1.50	18.35	8.30	11.49	0.20	3.89	2.78	0.75	0.31	100.36
87-97 cm	R	49.22	1.45	18.05	8.59	11.28	0.19	7.01	2.90	0.76	0.27	99.72
Pleistocene Fluvialglacial Deposit (OGT)												
13-23 cm		53.80	1.10	24.58	2.89	8.72	0.15	3.29	2.30	0.77	0.26	97.86
50-60 cm		55.46	1.02	22.48	3.15	7.97	0.13	4.00	2.74	0.88	0.17	98.00
Pliocene Olivine Basalt Flow (PFF)												
0-10 cm	S	49.91	1.70	22.46	5.17	12.20	0.14	5.24	2.85	0.80	0.13	100.59
15-25 cm	S	49.37	1.70	23.70	5.12	12.38	0.14	4.83	2.76	0.76	0.16	100.92
40-50 cm	S	46.51	1.63	26.26	4.96	12.33	0.14	4.14	2.48	0.76	0.23	99.44
15-25 cm	R	48.52	1.50	17.97	9.31	12.18	0.20	8.46	2.90	0.34	0.13	101.51
40-50 cm	R	47.74	1.49	18.33	9.54	12.12	0.21	8.02	2.80	0.34	0.14	100.73
PM		49.25	1.44	17.95	8.89	11.46	0.20	8.74	2.98	0.33	0.12	101.35
Pliocene Olivine Basalt Flow Breccia (PF)												
15-25 cm		47.59	1.74	22.37	6.66	12.65	0.21	5.72	2.53	0.58	0.24	100.29
40-50 cm		55.88	1.78	23.05	6.67	12.78	0.19	5.47	2.39	0.55	0.24	109.00
90-100 cm		44.83	1.93	34.37	8.38	14.16	0.23	6.10	1.95	0.31	0.24	112.51
230-240 cm	F	44.79	1.60	19.70	8.55	12.30	0.20	7.07	2.98	0.33	0.18	97.70
230-240 cm	R*	48.18	1.67	30.72	9.20	12.15	0.21	8.03	2.30	0.41	0.18	113.04

S, soil free of lithic clasts; R, lithic clasts greater than 1 cc.; F, friable, loose breccia; R*, solid clast

TABLE 6 (Cont.)

Sample	Depth	SiO_2	TiO_2	Al_2O_3	MgO	Fe_2O_3	MnO	CaO	Na_2O	K_2O	P_2O_5	TOTAL
Miocene Hypersthene Andesite (SR)												
	0-18 cm	55.52	1.23	28.07	1.63	10.16	0.27	0.76	0.68	0.37	0.32	98.99
	15-25 cm	57.32	1.17	27.75	1.40	9.88	0.09	0.21	0.28	0.30	0.13	98.53
	25-30 cm	55.95	1.22	28.10	1.46	10.03	0.23	0.62	0.53	0.36	0.22	98.72
	40-50 cm	58.15	1.16	28.11	1.27	9.89	0.06	0.09	0.20	0.25	0.13	99.31
	71-81 cm	58.10	1.19	27.44	1.30	9.98	0.06	0.06	0.25	0.22	0.14	98.74
Miocene Tuff Breccia (SB)												
	8-18 cm	55.06	1.23	25.03	2.37	7.83	0.14	2.12	2.02	0.64	0.09	96.52
	36-46 cm	54.49	1.21	28.27	2.18	8.82	0.09	1.15	1.00	0.48	0.02	97.71
	64-74 cm	51.80	1.12	26.07	2.32	8.65	0.10	1.07	0.96	0.48	0.00	92.56
	97-107 cm	58.28	1.39	18.04	4.60	8.92	0.18	3.51	1.13	2.26	0.10	98.41
	127-137 cm	59.21	1.06	25.71	2.22	7.64	0.09	1.20	1.29	0.49	0.00	98.89
	186-196 cm	54.67	1.43	25.39	2.98	9.24	0.10	1.13	1.12	0.49	0.00	96.53
	295-305 cm	54.87	1.12	26.34	2.25	8.18	0.10	1.19	1.41	0.42	0.00	95.88
	325-335 cm PM	58.78	1.25	15.34	4.02	7.31	0.14	5.10	2.17	2.12	0.19	96.43
	325-343 cm PM	60.66	1.25	15.29	4.35	7.56	0.16	5.52	2.35	2.20	0.17	99.51
Miocene Mack Creek Breccia (MB)												
	5-15 cm	52.78	1.26	25.79	4.32	9.97	0.16	2.28	1.47	0.60	0.15	98.77
	26-36 cm	54.65	1.12	24.34	3.81	8.30	0.12	2.93	1.99	0.75	0.05	98.06
	31-41 cm	54.10	1.15	22.93	3.79	8.67	0.13	3.46	2.15	0.91	0.08	97.37
Oligocene Buff Colored Tuff/Ridgetop (RTBT)												
	0-8 cm	59.38	1.52	31.73	3.84	10.08	0.34	1.51	0.87	1.84	0.16	111.27
	5-15 cm	60.30	1.48	31.54	3.29	9.44	0.16	1.05	0.96	2.25	0.11	110.58
	20-30 cm	59.05	1.48	20.83	3.09	9.73	0.15	0.78	1.04	2.42	0.22	98.77
	31-41 cm	63.71	1.33	17.19	2.73	8.74	0.12	0.61	1.24	3.14	0.20	99.02
	51-61 cm PM	67.80	1.10	29.33	3.02	6.76	0.15	1.81	3.29	3.11	0.07	116.44

TABLE 6 (Cont.)

Sample	Depth	SiO_2	TiO_2	Al_2O_3	MgO	Fe_2O_3	MnO	CaO	Na_2O	K_2O	P_2O_5	TOTAL
Oligocene Buff Colored Tuff Breccia/Sidehill (SHBT)												
5-15	cm	59.26	1.73	21.80	2.66	10.07	0.31	0.85	0.79	1.50	0.19	99.15
23-33	cm	59.24	1.73	21.37	2.77	9.81	0.11	0.51	0.77	1.33	0.12	97.76
59-69	cm	59.14	1.75	21.91	2.70	10.22	0.15	0.55	0.67	1.49	0.09	98.66
81-91	cm	67.27	1.41	28.15	2.71	8.52	0.11	0.30	0.42	0.03	0.09	109.00
102-112	cm	60.78	1.52	20.38	3.35	9.86	0.15	0.33	0.31	0.12	0.06	96.87
120-130	cm	54.38	1.83	24.37	3.20	11.86	0.51	0.24	0.29	1.32	0.10	97.74
Oligocene Red Tuff Breccia (RTB)												
0-5	cm	55.63	1.51	20.81	2.10	9.78	0.32	2.43	3.21	1.31	0.73	97.83
5-15	cm	56.52	1.47	20.33	1.96	9.37	0.38	2.20	3.46	1.22	0.84	97.74
15-25	cm	57.23	1.54	19.62	2.03	9.63	0.19	2.16	3.53	1.41	0.40	97.73
26-36	cm	56.83	1.56	20.17	2.47	9.82	0.16	1.97	3.39	1.46	0.36	98.19
51-61	cm PM	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.40	0.14	99.36

TABLE 7

TOTAL CHEMISTRY OF RTB TEXTURAL SEPARATES (%)

Sample Depth	SiO_2	TiO_2	Al_2O_3	MgO	Fe_2O_3	MnO	CaO	Na_2O	K_2O	P_2O_5	TOTAL
Clay (0.0-0.074 mm)											
0-5 cm	60.38	1.94	18.91	1.76	7.93	0.32	2.34	2.56	1.40	0.38	97.42
5-15 cm	61.31	1.90	18.53	1.81	8.11	0.44	1.96	2.71	1.36	0.42	98.36
15-25 cm	59.70	2.11	19.47	1.93	8.93	0.17	1.54	2.78	1.33	0.31	98.23
26-36 cm	57.90	1.52	20.77	2.05	9.78	0.11	0.97	2.74	1.22	0.24	97.32
51-61 cm PM	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.19	0.14	99.15
Sand (0.074-1.0 mm)											
0-5 cm	58.74	1.60	19.50	2.45	8.60	0.46	4.07	2.79	1.16	0.27	99.65
5-15 cm	58.94	1.63	18.46	2.39	8.74	0.45	3.22	3.07	1.20	0.29	98.40
15-25 cm	58.22	1.72	18.53	2.41	9.42	0.20	3.01	3.36	1.16	0.17	98.49
26-36 cm	57.40	2.18	19.95	2.65	10.25	0.13	2.38	3.28	1.18	0.13	99.53
51-61 cm PM	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.19	0.14	99.15
Pebble (10.-5.6 mm)											
0-5 cm	54.70	1.60	21.13	1.96	10.78	0.31	2.02	3.07	1.31	0.98	97.86
5-15 cm	55.59	1.60	19.82	1.92	10.51	0.38	2.50	3.38	1.37	0.95	98.02
15-25 cm	56.94	1.64	19.89	2.04	10.32	0.20	1.90	3.72	1.34	0.49	98.49
26-36 cm	58.36	1.81	20.65	2.44	9.15	0.16	2.16	3.47	1.52	0.25	99.97
51-61 cm PM	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.19	0.14	99.15
Gravel (5.6 mm)											
0-5 cm	57.06	1.36	21.43	1.94	9.33	0.24	1.96	3.39	1.36	0.74	98.83
5-15 cm	55.45	1.35	20.56	1.84	9.63	0.34	2.15	3.69	1.41	0.87	97.28
15-25 cm	57.57	1.32	19.81	2.15	9.34	0.19	2.51	4.01	1.38	0.38	98.70
26-36 cm	57.95	1.41	20.70	2.10	10.71	0.13	1.44	3.56	1.44	0.43	98.19
51-61 cm PM	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.19	0.14	99.15

TABLE 7 (Cont.)

Sample	Depth	SiO_2	TiO_2	Al_2O_3	MgO	Fe_2O_3	MnO	CaO	Na_2O	K_2O	P_2O_5	TOTAL
Whole Sample												
	0-5 cm	55.63	1.51	20.81	2.10	9.78	0.32	2.93	3.21	1.31	0.73	97.83
	5-15 cm	56.52	1.47	20.33	1.96	9.37	0.38	2.20	3.46	1.22	0.84	97.74
	15-25 cm	57.23	1.54	19.62	2.03	9.63	0.14	2.16	3.53	1.41	0.40	97.73
	26-36 cm	56.83	1.56	20.17	2.47	9.82	0.16	1.97	3.39	1.46	0.36	98.19
	51-61 cm	58.03	1.44	19.66	2.24	9.16	0.17	3.18	3.93	1.19	0.14	99.15

DISCUSSION OF RESULTS

Introduction

It has been proposed at the outset in this study that an assessment of weathering history requires knowledge of (1) suites and degree of crystallinity of secondary minerals, (2) chemical changes during pedogenesis and (3) chemistry of the soil solution and runoff water. Various environmental parameters such as precipitation, temperature, slope angle, aspect, geomorphic position and geologic substrate must be considered in conjunction with this data. Due to the interdependent nature of these variables, hypotheses regarding weathering history of certain samples can be made with partial information.

Parent material can influence weathering by regulating rates of elemental release. In the case of volcanic rocks mineral composition is the result of initial chemistry and cooling history, e.g. more rapidly cooled magma results in the formation of less weathering-resistant minerals (Goldich, 1928) and a greater amount of substitution in minerals which has a weakening effect on crystaline structures (Marshall, 1964).

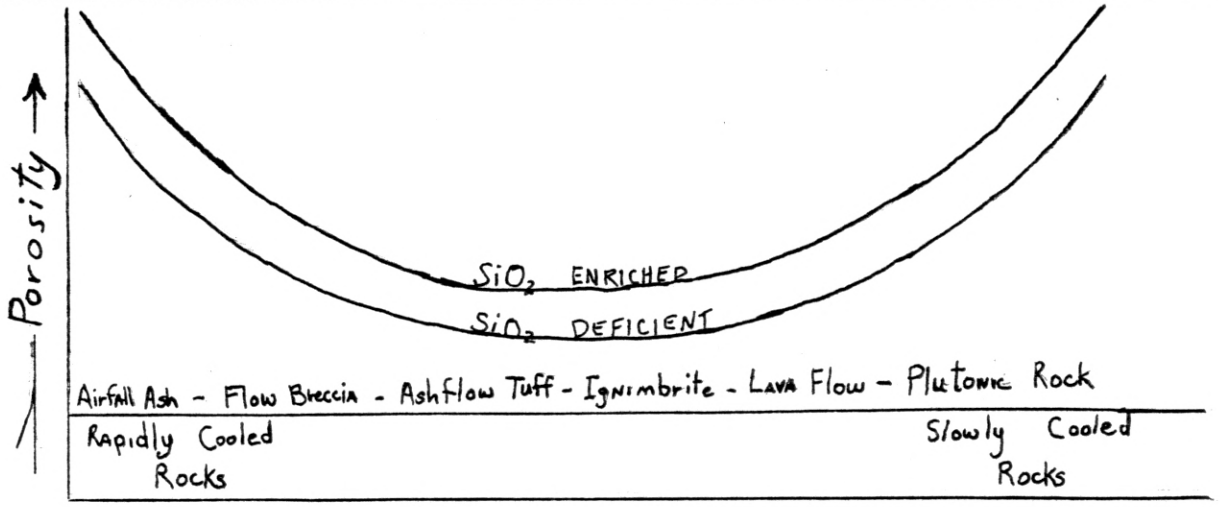
Elemental release is regulated by hydrologic characteristics of a soil as well as inherent mineral resistance to weathering. Rocks containing easily weathered minerals generally decay rapidly thereby

providing for rapid drainage and higher leaching rates. Highly impermeable rocks defy chemical weathering by limiting mineral exposure to agents of weathering. A rock which contains some of both kinds of minerals such as a coarse grained acidic rock type, granite, deteriorates rapidly in a physical sense, but the dominant minerals, quartz, plagioclase and mica slowly react to the large quantity of solution which passes around them. A coarse grained basic rock would deteriorate in a more uniform manner, encouraging formation of 2:1 secondary minerals by virtue of the rich supply of basic cations, which would limit porosity and slow leaching. In both cases the resulting weathering pathway is the result of the interaction between initial rock chemistry and texture and the physical forces which bring agents of weathering in contact with the rock. Rock density and texture is closely related to the cooling history.

Figure 13 demonstrates how porosity can be related in a general way to the cooling history of a rock.

Figure 13

Relationship between soil porosity and rock composition and cooling history



Porosity is expressed as a bimodal curve between two compositional limits as a response to initial rock density and mineral response to hydration. In the case of rapidly cooled lithologies, porous initial rock structure allows for rapid penetration of weathering agents. Course textured slowly cooled acidic rocks contain micaceous minerals which rapidly expand upon hydration, resulting in rapid physical deterioration of the rock. Basic rocks are deficient in quartz and are enriched with elements which readily form moisture and cation retaining 2:1 clays. As a result of coarse texture and a limited source of basic ions non-absorbative 1:1 clays tend to form in soils derived from slowly cooled siliceous rocks. Rocks in this study include airfall ash type breccia (SB, SHBT and RTBT), welded ash flow tuff breccias (RTB and MB) and lava flows (SR, PF and PFF) which range in composition from rhyodacite to basalt.

By integrating what is known about soil chemistry, secondary mineralogy, parent material chemistry and solution chemistry of runoff waters, much can be learned about the weathering pathways for rocks in the H.J. Andrews experimental forest.

PFF Pliocene Olivine Basalt Flow

As its name indicates, PFF is a highly unsaturated olivine basalt (Table 6). It has been weathering in an alpine environment on the crest of a 1600 M ridge beneath Abies Amablis vegetation since the recession of the Lookout Creek Glacier in the Pleistocene. Due to

continued cool temperatures the soil is very rich in organic matter. Snow blankets the site from late November into May or June.

Elemental distribution in the soil profile indicated that silica concentration increases towards the surface while Al, Fe and P concentrations increase with depth. In this study, these trends contrast with those of most of the other soils which formed in more mild temperatures at lower elevations.

Relative to parent material, elemental concentrations of Si, Ti, Al, Fe, K and P increase in the soil while Mg, Mn, Ca and Na decrease. However, Ca, Mg, Na and K exhibit the influence of biological cycling as they increase twoards the surface. The increase in Na towards the surface is much greater in PFF than in lower elevation sites. This could be related to the species of fungi or quantity of soil fauna which are found in alpine soils. The importance of K in biological cycling or as an absorbed cation on 2:1 clay minerals, is demonstrated by soil concentrations well in excess of parent material values.

Phosphorus increases towards the subsoil in the presence of Fe and Al. This trend, found only in the higher elevation samples, is an indication that Fe and Al phosphate compounds are being dissolved and mobilized to the subsoil. The iron, phosphate and aluminum in solution in soils forming at lower elevations synthesize highly resistant compounds which tend to accumlate towards the soil surface as weathering proceeds. Podzolization is driven by the accumulation of organic substances derived from decomposers and respiring roots beneath the

snowpack. Little moisture moves through the soil until warming occurs; this causes a flush of acid water to pass rapidly through the soil dissolving many insoluable Fe and Al compounds. They are mobilized to a horizon in the soil where the buffering capacity of the soil and microbial degration neutralize the solution, causing precipitation of metal-phosphate and organo-metalic complexes. Podzolization is well documented in regions of the world which experience 4-6 months of near or sub-freezing weather. (Strobbe and Wright, 1959)

Potassium is highly concentrated in PFF soil. This is probably related to fixation by organisms and adsorbtion onto 2:1 clays. High concentrations of Ca and Mg in PFF bedrock suggests that 2:1 clays should readily form. Possibly due to the profusion of organic and inorganic complexes with the clay particles, a sample could not be separated for x-ray diffraction analysis without reliance on chemical treatments, which were avoided in this study. This trait has been noted in Northwest alpine soils by several workers (personal communication, 1977, Randy Brown, OSU), however, a detailed explanation is lacking.

Elemental concentrations and optical observations support theories of mineral breakdown presented in the literature review. The greater leachability of Ca ions from plagioclase structures is indicated by the significant loss of Ca from bedrock to soil samples, while sodium shows much less change. Rapid decay of olivine is responsible for the large drop in Mg. The remaining Mg is possibly tied up both in more weathering-resistant pyroxenes and by adsorbtion onto montmorillonite clays.

PF Pliocene Olivine Basalt Flow Breccia

Despite identical chemical origins, contrasts in cooling histories of PF and PFF resulted in distinct rock textures which necessitated two different sampling methodologies. The wide range of textures in a flow breccia made a separation of lithic fragments from soil material impractical. Based on profile depths PFF appears to be deteriorating at a more rapid rate. This would be consistant with the greater abundance of glassier vessicular components in a flow breccia.

Chemical trends for Si, Al, Fe and Ti in PF are quite inconsistant, which might relate to the inherent chemical and textural variability of a breccia. High Al levels at 100 cm was associated with a deposit of friable, yellowish material which had an x-ray pattern indicative of amorphous material. Perhaps other concentrations of secondary minerals can be associated with some of the variation.

Although located in the alpine zone, PF does not strongly display a spodosolic elemental distribution. Al tends to decrease towards the surface and Si irratically increases towards the surface, but Fe and P concentrations do not follow expected trends. This sample site may have been disturbed 10-15 years ago by road construction and logging. A strong similarity between the two soils is the increase in K towards the surface from low initial values.

LPG Pleistocene Glacial Till

Parent material composition indicates that the source area for LPG

was in the region of PFF and PF (Table 6). Also located in the alpine zone but unlike the other two high elevation soils LPG formed in a seepy area which is moist well into the summer. Chemical composition trends are indicative that podozolic and non-podozolic processes are at work in LPG. Presumably soil moisture in PF and PFF diminish soon after cessation of precipitation and snow melt, termiating leaching. The increase in Ti concentrations with depth could be caused by the extensive periods of aneorobic conditions in the soil. PFF does not show an increasing Ti trend with depth in spite of a strongly pozolized horizon. Ti like Fe is very mobile in moist reducing environments. The facts that Si increases toward the surface and P levels remain steady to considerable depths in LPG soils indicates that podzolization also occurs.

Biological cycling of Ca, K, Mn, P and Na is evident. Perhaps Mg does not follow this trend because it is primarily retained in weathering resistant pyroxenes. Most of the easily weatherable olivine would have deteriorated during the first erosion-weathering cycle in the Pleistocene. Higher levels of K in LPG lithic fragments may be due to adhesion by K bearing 2:1 clay particles during the earlier transport/deposition episode. The K buildup in LPG is hypothesized to indicate presence of 2:1 clays but it was not possible to obtain verification by x-ray analysis.

SB Miocene Tuff Breccia

Documentation that Miocene age rocks in the western Cascades are more acidic than younger Piocascade Volcanics has been provided by

several workers (Peck et. al., 1964, White and McBirney, 1977). Miocene Sardine Formation rocks studied in the H.J. Andrews are consistant with this trend (Table 6). Sardine Fm. breccias are generally unwelded ashflows or airfall deposits whereas the Oligocene Little Butte Fm. breccias are welded ashflow or ignimbrite deposits (compare textures in fig. 6). The low density and high volcanic glass content of SB parent material has encouraged rapid rates of rock decay, formation of amorphous material and 1:1 type clays and a buildup of Al, Fe and Ti in the weathered residue. Si and the basic cations are highly leached.

The elemental distributions in SB are characterized by considerable variability with the exception of increases in the biologically cycled ions towards the surface from very low values in the subsoil. The variabiliy is both a function of the variability in texture and composition of a breccia and the inclusion of less weathered lithic fragments in the deep profiles. Based on a low Al concentration, and high K, Ca and Mg values the sample at 107 cm is an example of the effect of inclusion of less weathered material.

X-ray diffraction patterns indicate that 1:1 clays predominate over 2:1 clays (fig. 9) 2:1 clay of the highest crystallinity occurs near the surface presumably because rapid drainage leaches basic ions through the soil too quickly to allow formation of 2:1 clays. Broadening of the 7Å kaolin mineral peak with depth is caused by hydration of the mineral and is an indication of increasingly stable moisture conditions.

The degree of expansion upon treatment with organic solvents and collapse with heating to 550°C can be an indication of degree of

encrustation by Fe and Al hydroxal compounds on 2:1 clays (Jackson, 1963). SB appears to be the least hydroxylated profile studied (fig. 9). It also has lowest iron content of any profile studied (Table 6), which might be expected in view of its hydroxlyation.

A soft, white crystalline material was observed in hand specimens of the deepest SB samples. A diffraction reflection at 9.9° 2 θ (fig. 9) indicates that a mineral is present with an 8.9 Å d-spacing. The peak is most clearly displayed in samples which have been heated to 300° C which dehydrates halloysite, collapsing the peak towards 7.15 Å and 2:1 clays remain near 6.15 Å. The zeolites huelandite and clinoptilite each have a major peak near 8.9 Å. Presence of several types of zeolites including these two, have been acknowledged by others (Youngberg, et. al., 1971; Paeth, 1974 and Taskey, personal communication, 1977). It has been shown that zeolites form in breccia parent materials during diagenetic alterations incurred by heat and fluids from successive volcanic events in the region. The passage of molten magma through such pyroclastic bedrock provides the temperatures and movements of fluids necessary for zeolite formation (Peck, et. al., 1964; R. L. Hay, 1963). Formation of authnigenic soil zeolites has been hypothesized, but little data has been presented to support such an idea.

An important contrast exists between the type of clay minerals present and cation exchange capacity of SB SHBT with PFF and RTB. 1:1 clays dominate the former and 2:1 clays dominate the latter.

SR Miocene Hypersthene Andesite

It is apparent from elemental composition trends that the extractor did not penetrate unweathered parent material, in spite of encountering dense rock. High Al and Fe and low concentrations of Mg, Ca, Na, K and P signify that the bedrock is deeply weathered (Table 6). In spite of the highly leached soil, elemental concentrations toward the surface and in the buried A-horizon (30 cm) provide insight into the strong role played by organisms in cycling nutrients, and the durability of organo-metalic complexes in the soil.

The buried profile is obvious in the soil pit at 30 cm as a darker, more aggregated horizon. Notice how Si decreases and Al, Fe and Ti increase in both the buried and current A-horizon. Nutrient cations Mn, Mg, Ca, Na, K and P also increase dramatically. Although absolute dating of the burial has not been carried out, presence of charcoal in the soil above 30 cm is evidence that it might have occurred after a fire that burned through the area 130^{+} years ago or it may have occurred long before.

X-ray diffraction analysis reveals that clay mineral composition and degree of crystalization is somewhat similar for both the current and buried A-horizon (fig. 8). Peaks are broad and subdued, intimating that conditions of the clay forming environment are far from ideal. Activities in the surface soil such as, repeated wetting and drying cycles, flushes of organic substances from decaying litter and metabolizing organisms, disturbances by burrowing organisms, stresses from frost

heaving and mixing caused by sporatic tree blowdowns, would be significant interruptions in the clay formation process.

2:1 smectite clay mineral and 1:1 kaolin minerals are found in equalibrium with both the SR soil chemistry and moisture regimes. (fig. 8) Low cation retention in SR could be a function of the degree of hydroxalation of 2:1 clays as well as the low cation exchange capacity (CEC) associated with 1:1 type clays.

Because the deepest samples were highly weathered, it is not possible to make comparisons between soil and bedrock mineralogy. Ca is much more depleted than Na in the subsoil but due to greater selection for Ca by organisms, Ca is more concentrated at the surface than Na. Slower depletion of Mg than Ca or Na is indicative of the resistance to decay by pyroxenes compared to plagioclase.

MB Miocene Mack Creek tuff breccia

In spite of the presence of a very shallow soil mantle, the least weathered MB sample appears to be relatively more weathered in comparison to SB, parent material (Table 6); they appear very similar in hand specimen.

Silica, Ca, Na and K have been leached from the surface horizons while Al, Ti, Fe and nutrient cations Mg, Mn and P are concentrated towards the surface. Nutrient cycling is less pronounced in MB due to its youthfulness which is insured by soil creep.

Regarding elemental leachability, it is assumed that what has been said about SB could be said about MB. Ca and Na are being leached from MB at nearly equal rates.

OGT Pleistocene Fluvioglacial Deposit

Chemical concentration values are indicative of the fact that the parent materials for OGT are primarily derived from Miocene formations (Table 6). Although the terrace surface from which OGT was obtained correlates with an early Pleistocene glacial surface exposed 200 M downstream, there is strong evidence that colluvial and and post-glacial alluvial processes have been responsible for adding a substantial amount of material from adjacent slopes, underlain by Sardine Fm. rocks.

In view of the shallow profile, OGT data is most useful for comparisons with streamwater chemistry.

RTBT and SHBT Oligocene Buff Colored Tuff/Ridgetop and Sidehill Samples

RTBT and SHBT are presented together because they are derived from the same parent material. The sidehill sample, SHBT is located only 4 M vertically, below the ridgetop site, RTBT. Chemically, the two profiles contrast in the degree of variability in SHBT and relatively smooth trends present in RTBT (Table 6). Evidence of mixing and soil perturbation are corroborated by a lack of horizonization in the sample pit and presence of creep-induced bowed trees growing nearby.

The differences in clay minerals between the SH and RT sites is consistent with the anticipated differences in their permeability and porosity. As a result of mixing, soil creep and sidehill position, the SH site is more porous which encourages the formation of 1:1 type clays. 2:1 type clays predominate in RTBT due to concentration of nutrients as a result of slower drainage. Clay mineral crystallinity improves with depth as moisture conditions stablize.

Further documentation of the differences in Buff tuff site drainage conditions is provided by Fe+Ti: Al ratios. In RT periodic saturation and slower drainage have resulted in the mobilization of Fe and Ti on occassion. A lower Si:Al ratio indicates that the RT profile has been subjected to weathering forces for a long period of time. SH soil have a much higher Fe+Ti:Al ratio which indicates that Fe and Ti are rapidly oxidized after leaching occurs. Clays in RT were more encrusted by Al-hydroxal coatings. The BT samples indicate that very old soils, rich in Al have a much greater quantity of hydroxal coatings than a well drained, young soil such as SH where Fe is enriched in the surface soil by oxidation.

The comparison between the two BT samples illustrates the effect of drainage and geomorphic stability in elemental distribution, secondary mineral formation and mineral decay rates.

RTB Oligocene Red Tuff Breccia

A welded ashflow tuff (RTB) would be expected to weather more slowly than an airfall ash type deposit (BT). The higher Si:Al ratio

of RTB indicates that it is deteriorated at a slower rate than RTBT, since their parent material chemistry and geomorphic situation are nearly identical. The RTB site is located within the influence of headward erosion of the stream in WS 10 which would contribute to its youthfulness.

Higher initial Fe content in RTB and the presence of moist yet oxidizing soil conditions is manifest by the appearance of lepidocrocite (FeO·OH) in the soil (8.9° 2 θ) (fig. 12). Substantial quantities of 2:1 type clays in a soil derived from a relatively silicic bedrock is also an indicator of moist soil conditions. It also seems concievable that the large quantity of weathering-resistant lithic fragments might provide ample microsite situations which would aid in the formation of 2:1 clays and inhibit the development of 1:1 clays (Chichester, Youngberg and Harward, 1969). Despite relatively low Al concentrations 2:1 clay minerals respond to treatments as though they are hydroxylated. In an iron rich environment, it would be reasonable to postulate that RTB clays are coated with Ti and Fe-hydroxides which are mobilized infrequently under saturated soil conditions.

Chemical analysis of RTB textural separates reveals several very interesting elemental relationships in different size fractions (Table 7). In the bulk RTB sample, biologically cycled nutrients Ca, Mg, Mn, K, and P increase towards the surface from low levels in the subsoil. Al, Fe and Ti increase while Si decreases steadily from bedrock values. A notable feature is the strikingly large P_2O_5 values.

In comparison to airfall ash derived tuffs (RT and SHBT and SB), RTB parent material is more enriched in phosphorus and iron. It has been demonstrated that Fe and P are strongly attracted to one another resulting in the formation of Fe phosphate compounds (Bradley and Sieling, 1953), and the formation of lepidocrocite is an indication of high levels of free iron in RTB.

The clay-silt fraction (0.0-0.074 mm) has a higher Si:Al ratio (3.12:1) than the whole sample (2.8:1). Aluminum increases with depth. Fe concentration was lowest for all surface samples in the clay fraction and increases with depth.

Calcium and potassium demonstrate biological cycling most clearly in the clay fraction. Mg is not as important in the clay fraction as it is in other fractions, which reflects the order of cation preference by negatively charged sites on 2:1 clays (K, Ca, Mg and Na, respectively, Buckman and Brady, 1967).

The Si:Al ratio for the silt-sand fraction (0.074-1.00 mm) is 3.07:1 and both Al and Fe increase with depth from low levels at the surface. Ca and Mg levels in this fraction are much higher than in the other size fractions or bulk soil. Ca increases rapidly toward the surface soil. Without additional diffraction or electron microscopic analysis it is not possible to determine whether the Ca is in organic or inorganic forms. A large amount of Ca can be fixed by fungal mats and transformed to Ca-oxylate crystals. Burrowing and foraging soil fauna assimilate this readily available source of Ca into their exoskeletons

or internal faunas. In either form, the residues would be primarily concentrated in the silt-sand fraction. Phosphorus and K levels are very low in comparison to other size classes.

Based on thinsection observations, the pebble fraction (1.00-5.6 mm), is composed of resistant lithic fragments which are often coated by layered precipitates. The pebble class is highly enriched in Fe and P. Both increase towards the surface, which is unusual for Fe. It appears that Fe-phosphate compounds dominate the concretionary structures, based on the high concentrations of these two elements toward the surface in this size fraction. Evidence of Ca, Mg and K cycling is minimal, which would be expected since few exchange sites exist in sand, pebble or gravel size classes. Si is consistently depleted in comparison with the other fractions.

The gravel size class (5.6 mm) is most notable in its similarity to the bulk sample, indicating its proportional role in the bulk sample.

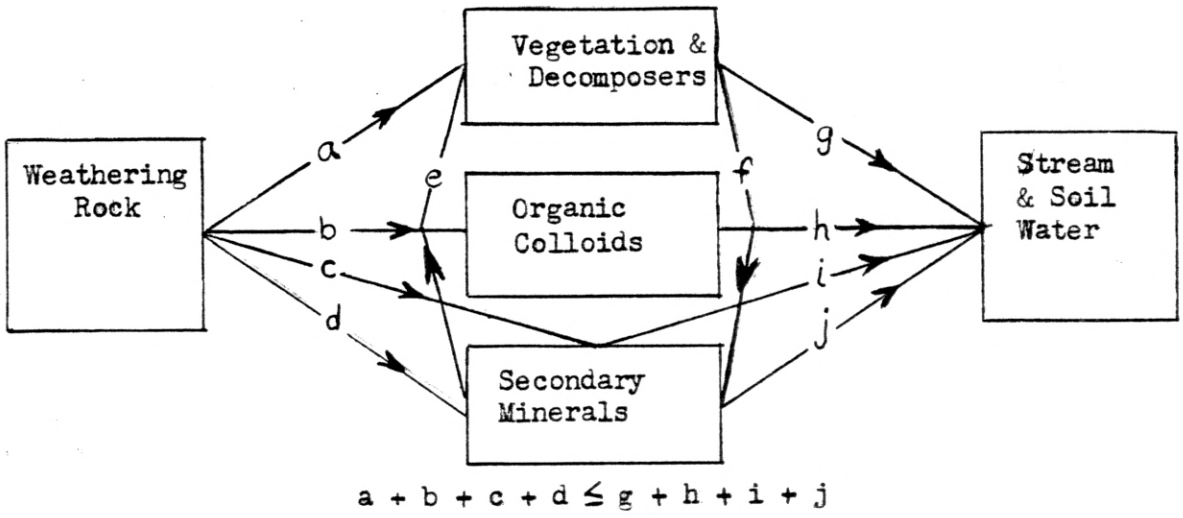
Stream Chemistry Compared To Bedrock Chemistry

Using total chemical profile data in comparison with elemental percentages of dissolved solids in streamwater where a stream is confined to a single bedrock unit weathering can be observed. Leachability of the major cations from various bedrock types can be observed if good bedrock samples have been taken. Speculation on differences between bedrocks can be made from differences in water chemistry where good bedrock samples are lacking. This model has been used in several nutrient cycling or weathering studies across the country where bedrock

remains constant (Marchand, 1974, Vitousek, 1977). In a region characterized by complex geology such as the Pacific Northwest, more care must be taken in drawing conclusions from differences between stream water chemistry and bedrock chemistry.

Carrying out this type of study in the H.J. Andrews Forest is advantageous because the forest age class structure has been mapped. Much of the remaining stand is mature old-growth. An assumption is made in this study that nutrients are returning to the mineral state nearly as rapidly as they are being removed from the soil. In an old-growth system where litter accumulation is rapid it is assumed that the vegetation-soil community has reached a steady-state equalibrium condition (Johnson, 1971). A younger forest would be preferentially selecting K, P, Ca, Mg, Mn and Fe ions from the soil at rates greater than they would be returned via litter production. Figure 14 diagramatically portrays the pathways of ions released by rock weathering in an old growth forest.

Figure 14. Elemental flux pathways in an old growth forest.



$$a + b + c + d \leq g + h + i + j$$

Ca, Na, Mg, K, P, and Si are the dominant cations in most streams in the western Cascades mountains. In the first two comparisons presented below (Tables 8, 9, 10) the term dissolved solids refers to the gross nutrient outflow minus net precipitation input. A comparison is made between the elemental percentages of bedrock and the percentages of dissolved load. Phosphorus concentration comparisons are based on soil values since P concentrations in the soil are up to five times greater than P concentrations in bedrock units. If the ratio is greater than one, the importance of an ion has increased from bedrock to stream, if it drops below one, it has decreased relative to the bedrock elemental distribution.

The most complete comparison is made between the stream water flowing from Watershed 10 (WS 10) and the two types of bedrock which underlie the stream. RTB is estimated to underlie 70% of the watershed surface while BT makes up the remaining 30%.

Table 8. Comparison of elemental percentages between bedrock and stream water RTB/BT and WS. 10

Dissolved Solids	Kg/Ha[a]	Elemental % of total D.S.	Elemental % of bedrock	Change (x)
Ca	47.84	12.4	1.96	6.3
Na	29.49	7.7	2.68	2.9
Mg	11.92	3.1	1.42	2.2
K	1.53	.4	1.36	.3
P	1.15	.3	.17	1.8
Si	100.66	26.1	27.46	1.0
Total Cationic Dissolved load	192.59	50.0	35.05	

[a]Dissolved load from R.L. Fredricsen for water years 1969-74, ave. precip. 242.16 cm.

A simple index of weatherability based on the degree to which an element increases in importance from bedrock to the stream predicts the following order of leachability from the WS. 10 ecosystem; Ca > Na > Mg > P > Si > K.

The ordering of cations in the solution chemistry of runoff water from WS. 10 bedrock is comparable to that derived from observing the leaching of bedrock as expressed by cation concentrations in the soil compared to initial concentrations in bedrock. In the WS. 10 comparison Ca has the highest concentration in solution and shows the greatest percentage decrease from bedrock to the lowest value in the profile (38% decrease). Presence of Na in solution follows Ca and correlates with an 18% decrease in the soil. Mg is next in solution and accompanies a 12% decrease from parent material to the lowest soil value. Phosphorus varies from the trend. It is next abundant in solution in spite of increasing to 6 times its parent material value in the soil. Of the nutrient cations, K is found at lowest levels in stream water and in fact shows a modest increase in the soils (23%).

The bedrock underlying WS. 10 has been intruded by manylarge dikes, which are rich in plagioclase (Swanson and James, 1975). Since cooling was rapid the plagioclase should be rich in calcium and moderately succeptible to weathering. The impact of the dikes is difficult to predict; however, data from one set of lysimeter tubes on Watershed 10 indicates that Ca in soil solution can reach very high values directly above and downslope from such a dike (M. McCorison, unpublished data, 1977).

Sodium is hypothesized to be more prevalent in stream water than Mg because Na is nearly biologically inert and not adsorbed on clay colloids. In spite of the fact that they are primarily located in minerals with nearly equal weatherability (Mg-pyroxenes, Na-albite), Mg is both taken up by organisms and attracted to colloidal exchange sites, while Na is quickly flushed through the system. Mg is thermodynamically less stable than Na according to Gibbs Free Energy of Formation (p. 25) but found less prevalent in solution which supports the selective uptake hypothesis also.

In a soil dominated by 2:1 clays K is both selectively adsorbed by secondary minerals and taken up by most organisms. Based on K buildup in soils and its rapid cycling through plants, K must be the most readily absorbed cation.

The large quantities of lithic fragments in RTB make accurate identification of sources of the elements difficult. Identifiable matrix minerals in thinsection were plagioclase, pyroxene and abundant volcanic glass.

The bedrock underlying guaged Watershed 8 (WS. 8) is a hypersthene andesite of the Sardine Fm. Since SR was sampled in another part of the experimental forest and because a good sample of SR bedrock is lacking it is not possible to make a WS. 10 type comparison for WS. 8. With an annual dissolved solid discharge budget for WS. 8 a direct comparison between the two watersheds will provide a basis to speculate on differences between the WS. 8 and WS. 10 parent materials (Table 9).

Table 9 Comparison of Dissolved Solid Load Carried by WS. 8 and WS. 10 Streams.

	Dissolved Solids Kg/ha		
	WS. 8	WS. 10	
	Hypersthene Andesite	Red Tuff Breccia	WS. 8 : WS. 10 (%)
Ca	46.22	47.84	94.8
Na	44.84	29.49	152.3
Mg	11.50	11.92	103.6
K	2.99	1.53	195.4
P	.64	1.15	55.6
Si	104.03	100.66	104.3

Parent material differences between WS. 8 and WS. 10 must be primarily responsible for the differences in runoff dissolved solid load. Weathering bedrock in the two watersheds are solubilizing similar quantities of Ca, Mg and Si, however, WS. 8 is loosing 50% more Na, 95% more K and half as much P as WS. 10. These figures imply that more Na source mineral (albitic plagioclase) is present in the bedrock, easily weatherable K source minerals or glass with fewer 2:1 clays are present in the system, and a environment exists in which P is less effectively fixed. Perhaps P is not as prevalent in the bedrock initially and the uptake requirements of the younger stand on WS. 8 limit P losses.

Weathering theories stated earlier would predict that WS. 10 should be weathering at a more rapid rate than WS. 8 since the flow rocks of WS. 8 were cooled more clowly creating a very impermiable substrate and a parent material with less volcanic glass than that underlying WS. 10. Factors that may be responsible for the anomalous weathering rates include the low weathering rates of lithic fragments which make up 65% of the RTB bedrock and the possiblity that an extensive complex

of joints and fractures exists in the Andesite underlying WS. 8, thereby providing a much greater surface area in contact with the runoff water than was used in the calculations of dissolved load per hectare. More information on parent material composition and structure is necessary to understand the solution chemistry of WS. 8.

The third stream with sufficient bedrock and water chemistry data for a comparison is called Devils Club Creek (DCC). It flows over Sardine Fm. MB (70%) and OGT (30%). Streamwater chemistry data collection for DCC has been on a periodic basis which necessitates using a different scheme for comparison, based on interion ratios in both bedrock and stream water. Na was chosen as the standard ion since it is nearly biologically inert. If a ratio increases between bedrock and stream water it indicates a decrease in the importance of the non-sodium ion in stream water. If it decreases, the ion has increased relative to Na in stream water.

Table 10. Comparison of bedrock and stream chemistry by interionic ratios with sodium, OGT/MB and Devil's Club Creek

Elemental % of bedrock		Ratios to Na		Average concentrations[a] (ppm) summer and fall, 1976	Ratios to Na	Change (x)
Ca	2.55	Na/Ca	.64	3.822	Na/Ca .64	1.00
Na	1.64	Na/Na	1.00	2.435	Na/Na 1.00	1.00
Mg	2.33	Na/Mg	.70	1.145	Na/Mg 2.13	.33
K	.77	Na/K	2.13	.350	Na/K 6.96	.31

[a]Concentration data from R.L. Naiman, Oregon State University, 1977.

In the DCC system, Ca and Na remain in the same proportion in the runoff while Mg and K diminish in the stream water in comparison to Na.

Without an annual dissolved load budget it is not possible to deterime weathering rates of the bedrock constituents or compare weathering rates of the parent material with the other systems. However, based on internal controls the weatherability index for MB/OGT is; Ca = Na > K = Mg. Since Na was present in lower quantities in soils in all systems studied it can be assumed that it is being removed from the DCC system in excess of its percentage in the bedrock, possibly at a rate similar to the other streams. The main difference between the two systems is the resistance to weathering demonstrated by Mg, or the fixation of Mg in the soil. Low stream water Mg values are supported by increasing Mg in soils over bedrock values. K does not increase in either MB or OGT soils.

Uses For Rock Weathering Research

In terms of forest management, this information can have some important implications. The lower nutrient status and less efficient nutrient retnetion mechanisms of weathered Miocene rocks would be important when considering fertilization programs or anticipated stocking levels. By the same token, highly efficient K fixation in the case of soils formed on Pliocene flows and P fixation in soils derived from Oligocene ashflow breccias would also be of economic interest when considering the benefits of a fertilization program. In the first case the fertilizer would be flushed through the system rapidly, in the latter two, a nutrient would be fixed quickly and perhaps made unavailable to plants. In cases where clay minerals interfered with K or P movement in

the soil an application of an organic acid which would release fixed K or P might be more advisable. Stocking levels should be considered in light of the carrying capacity of such soils. Similar types of analyses need to be made on more soils in the coniferous forest region.

Management questions raised as a result of this study include, what is the effect of physical disturbance of the soil surface in many clearcut operations? How sensitive are the large group of organic processes which survive on and process the litter layer on the soil? Would certain rock fertilizers be more appropriate for a long rotation crop such as Douglas-fir, than common agrictultural fertilizees which are expensive and not of the same duration as timber crops?

Information of the nature gathered in this study has several applications to ecosystem modeling. With basic mineralogical data, bedrock and soil chemistry and weatherability indices for major minerals and nutrients, a clearer picture of the magnitude and mechanisms of nutrient cycling. In addition organic and inorganic storage compartments in the litter and soil can be more clearly illustrated.

CONCLUSIONS

As a result of the braod based appraoch to weathering undertaken in this study, the conclusions address weathering processes in a genral way as well as specific relationships between the different bedrock units studied and/or the various environmental factors influencing weathering in the western Cascades area.

The most fundamental conclusion illustrated by this study is the relationship between geologic history of a rock and styles and rates of rock weathering. Two rocks of the same chemical composition will vary considerably in terms of texture and mineralogy given different modes of cooling and formation. Texture and mineralogy have a large influence on a rock's weathering environment.

Climatic influences on weathering processes were clearly defined in the study with the discovery that podzolization was occurring above 1100-1500 M, the zone of a continuous snowpack from November to June. Below this elevation leaching is a continuous process which concentrates residual Fe, Ti, Al and P towards the surface while most nutrient cations show increases near the surface from depleted values in the subsoil. The podzol soils are characterized by silica concentrated toward the surface and Fe, Al and P mobilized to the subsoil. This occurs as the snowpack melts carrying an intense acid wetting front through the soil which forms as melt waters incorporate concentrations of organic acids, enzymes and other organic substances that have accumulated beneath the snowpack.

Geomorphic position is a significant factor in weathering, particularly as it relates to soil drainage. The differences in clay mineralogy, degree of hydroxylation and quantity of oxidized Fe and Ti can be observed in two profiles which are forming on the same parent material in distinctive geomorphic positions.

The well drained sidehill site was characterized by 1:1 clay and large quantities of oxidized Fe and Ti while the poorly drained site is characterized by a greater percentage of 2:1 clays, the degree to which the clay is coated with hydroxal material and the large amount of Fe and Ti which has been mobilized in moist reducing conditions.

Influence of organisms is not directly observed in this study with the exception of a buried surface soil which appears very similar chemically and in its secondary mineralogy to modern surface horizons, indicating the long-term durability of biologically cycled and complexed cations.

Each of the different bedrock types studied provides unique highlights for the rock weathering picture in the western Cascades.

Pleistocene glacial and colluvial deposits can be traced to source areas by distinctive chemical signatures. Although, ample soil development has occurred since the retreat of alpine glaciation in the Pleistocene.

Pliocene olivine basalts are found only in the alpine soil zone where podozolization is taking place. Due to the low silica composition of parent materials and rapid erosion rates of the steep alpine zone

this process is detectable only by chemical analysis. Potassium increases well above bedrock values. The clay fraction could not be separated from the samples without using chemical pretreatments, indicating that the colloids were flocculated by interlayer K, and organic and amorphous materials.

Miocene aged rocks are represented by deeply weathered SR, a hypersthene andesite flow, and SB and MB, airfall ash breccias. The soil forming on these substrates are much richer in 1:1 type clays than 2:1 clays and demonstrated a lower K fixation capability than soils forming on either Pliocene or Oligocene parent materials. TiO_2 concentrations were consistently low for all Miocene samples.

The Oligocene airfall ash deposit, BT, revealed striking differences between the soils forming on a creeping sidehill slope position and one forming on the stable ridgetop. Basic differences related to the natures of the two Oligocene parent materials are apparent in a comparison between the welded ashflow tuff, RTB and the BT series.

Separation of the Oligocene welded ashflow tuff derived soil into clay, sand pebble and gravel fractions reveals that various elements are preferentially concentrated into different size classes. The clay fraction had the highest Si:Al ratio and best demonstrated increases of nutrient cations toward the surface. Calcium is concentrated in the sand fraction. Magnesium is less significant but also most concentrated in that fraction. The pebble size class is marked by concentrations of P, Al and Fe, increasing towards the surface. Thinsection analysis

indicates that this is related to the formation of Fe and Al phosphate concretions and aggregations on lithic fragments. The gravel fraction is most representative of parent material composition although increases in P towards the surface indicate that a coating of organic and phosphate compounds has occurred.

Comparisons between bedrock and stream chemistry reveal that runoff waters from WS. 10 which is underlain by RTB and BT, are carrying about six times as much Ca, three times as much Na, twice as much Mg and P, an equivalent amount of Si and one-third the quantity of K that it would be expected to carry based on a one to one relationship with parent material chemistry.

Runoff water chemistry of WS. 8 differed from that of WS. 10 by carrying twice as much K, 1.5 times as much Na and one-half as much P. Ca, Mg and Si were all very close in Kg/ha. Due to the absence of a soil-bedrock sequence from WS. 8 it is not possible to develope an element weatherability index by making a comparison of stream water chemistry to bedrock chemistry. Inferences can be made about the mineralogy of the andesite flow bedrock in comparison to the welded ash flow tuff of WS. 10 from variations in the stream chemistry of the two water shed.

Devil's Club Creek drained a MB/OGT terrain. Based on an elemental comparison of Ca, Mg and K with Na in both bedrock and the stream, Ca was found to be present in the same proportion with Na in each and both Mg and K were 2/3 less common in the stream compared to Na than they were in bedrock.

The broad based interdisciplinary approach used in this project created an organizational dilemna and prevented detailed mineralogical and weathering reaction studies. Also lacking for this approach was a uniform well crystallized bedrock substrate. By focusing on a series of volcanic and pyroclastic bedrock formations in the temperate humid climate of the Pacific Northwest with total chemical, mineralogical and optical methods, much about the weathering process in this region, which has not previously been widely discussed, has been initially observed. This includes the influence of the wide variety of volcanic rock textures found in this region, the effect of the Pacific Maritime climate along an elevational transect, and the biological effect of the coniferous biome on rock weathering processes. Some baseline rock and soil chemistry data will be useful in the nutrient cycling model for the coniferous biome.

The study demonstrated that an interdisciplinary approach is useful for understanding multi-faceted rock weathering processes.

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The importance of lithic textures, exclusive of bedrock chemistry is illustrated by the significant differences in soil mineralogy and rates of formation between soils forming on Oligocene or Miocene rocks. At the Andrews Forest study site, Miocene pyroclastic rocks were deposited primarily as an airfall ash while Oligocene pyroclastic rocks included several welded ashflow tuffs. The contrast in textures due to the different modes of emplacement appears to be responsible for differences in weathering rates and pathways since initial chemistry is very similar for the two rock types.

Chemical evidence of buried profiles demonstrates the longevity of biological influences on the chemical composition of soils. The results of this study generally emphasize the importance of biological processes in the control of ionic distributions of biologically essential cations in the soil profile and the regulation of soil activity which influences inorganic weathering reactions.

Chemical and mineralogical differences noted in two profiles, developed on the same parent material in different geomorphic positions illustrate the importance of drainage and soil mantle stability as they related to oxidation of Fe, Ti and Al the development of biologically cycled ion distributions in soil profiles.