

AN ABSTRACT OF THE THESIS OF

Matthew Ikechuku Sunday Ezenwa for the degree of Master of Science
in Soil Science presented on June 21, 1976

Title: COMPARATIVE STUDY OF SOME SOIL CHARACTERISTICS
OF GRASSLAND AND FOREST /GRASSLAND TRANSITION
IN WESTERN OREGON

Abstract approved: Redacted for Privacy
[Signature] Gerald F. Kling

Vegetation as a soil-forming factor was studied under forest transition and grass while attempting to keep other soil forming factors constant in the western part of the Willamette Valley, Oregon. The physiographic units consist chiefly of the interior foothills and slopes formed from an old basaltic flow with some interspersed sedimentary formations.

The soils and vegetation of these physiographic units are developing under a temperate humid climate characterized by dry summers and mild, rainy winters.

The foothills and slopes support stands of Douglas-fir (Pseudotsuga menziesii) and Oregon white oak (Quercus garryana). Scattered through the area are non-forested openings.

Many ecologists have looked into the successional relationships of the communities within the Willamette Valley. One school believes

that the open conditions became established during a time of warmer and perhaps dried hypsithermal climate that occurred near the middle of the postglacial time. Fire control activities motivated by the settlers are, however, thought to be responsible for the successional change--replacement of a prairie by Quercus savanna, then by Quercus/Pseudotsuga, then by Pseudotsuga, etc.

The soils under both vegetation types were similar (classified as series) and they ranged from Mollisols to Ultisols.

To gain insight into the effects of vegetation on soil development, physical and chemical analyses were carried out and values were obtained for the following: particle sizes, pH, organic matter and organic carbon, total N, exchangeable cations and free Fe oxides.

The establishment of two different vegetational assemblages has had a direct impact on the soils. The soil parameters that are affected by the plant cover are directly or indirectly associated with the organic matter with the exception of exchangeable sodium.

Generally, there are some differences in both color and structure of the soils under these two vegetation types, but they are not adequately quantified to show statistical differences.

The profiles range from very strongly acid in soils derived from sedimentary rock, to neutral in the soils derived from basaltic rock. The pH of soils under grass is significantly higher than that of the forest transition soils.

There is subsurface accumulation of silicate clays but there is not much sign of the accumulation of oxides.

The exchangeable potassium contents of forest transition soils are significantly higher than those of the grassland soils. This is attributed to greater recycling of potassium by tree species.

The percentages of total nitrogen in the A horizons of the grassland soils are significantly higher than those of the A horizons under forest transition. The carbon/nitrogen ratios of forest transition soils are significantly higher than those of grassland soils. All these are attributable to the qualities of the organic matter in the soils of the two vegetation types.

The overall exchangeable sodium in the forest transition soils is significantly higher than that of the grassland soils. The reason for this might be that the area under study is very close to the ocean and that the trees are able to trap wind-borne salt spray from the ocean, which is eventually washed down the profile. Research is needed to prove this hypothesis.

The soils of the study area do not possess features of podzolic soils even though they occur in a latitude and under a forest vegetation where podzolic soils are commonly formed. This can be attributed to the climate of this area and the nature of the parent material of the soils under study.

Comparative Study of Some Soil Characteristics
of Grassland and Forest/Grassland Transition
in Western Oregon

by

Matthew Ikechuku Sunday Ezenwa

A THESIS

submitted to

Oregon State University

in partial fulfillment of
the requirements for the
degree of

Master of Science

Completed June 1976

Commencement June 1977

Redacted for Privacy

Redacted for Privacy

Redacted for Privacy

Date thesis is presented

June 21, 1976

Typed by Mary Jo Stratton for Matthew Ikechuku Sunday Ezenwa

ACKNOWLEDGEMENT

The author thanks each of the following for assistance rendered toward the completion of this thesis and the degree for which it is a requirement.

To Dr. Gerald Kling for his concern and assistance in guiding the author to this goal and beyond; to Dr. G. H. Simonson for introducing me to the soils of the study area and for serving on my committee; to Dr. W. K. Ferrel of the School of Forestry, and Dr. Chester T. Youngberg for their suggestions and for serving on my committee.

Thanks are also extended to my wife, Victoria, who missed a lot of activities because of scholastic pressures; to L. D. Johnston of the Department of Botany and Marvin Rowley of the School of Forestry who assisted me in plant identification; to Marvin Kauffman and Toney Crudele who assisted me in chemical and physical analyses respectively, and to others who offered their assistance during this period of scholastic endeavor and contributed a substantial part, the effect of which is realized only by the author.

My final thanks go to the Food and Agriculture organization of the United Nations who funded this project.

TABLE OF CONTENTS

	<u>Page</u>
INTRODUCTION	1
Geography	3
Geology	3
Soils	4
Climate	6
Vegetation	7
LITERATURE REVIEW	13
Organic Matter	15
Clay Fraction	20
Soil Reaction	24
Total Nitrogen	27
Cation-Exchange Capacity and Base Saturation	31
Pedogenetic Oxides	35
MATERIALS AND METHODS	40
Field Investigation	40
Laboratory Investigations	43
Sample Handling and Preparation	43
Physical Methods	43
Chemical Methods	44
RESULTS AND DISCUSSION	45
Physical Properties	45
Soil Morphology	45
Particle Size Distribution	61
Chemical Properties	66
pH	66
Cation Exchange Capacity and Bases	67
Organic Matter	82
Total Nitrogen	84
Free Iron Oxides	86
SUMMARY AND CONCLUSIONS	90
LITERATURE CITED	93
APPENDIX	106

LIST OF TABLES

<u>Table</u>		<u>Page</u>
1	Locations, names of soil series, and elevations of the study sites.	40
2	Particle size distribution of soils of the study area.	62
3	Paired F-tests on some soil properties.	64
4	Chemical properties of soils under grass and forest/grass transition.	71

LIST OF FIGURES

<u>Figure</u>		<u>Page</u>
1	Stand of <u>Quercus garryana</u> intermixed with <u>Pseudotsuga menziesii</u> .	8
2	Large <u>Quercus garryana</u> trees shelter abundant conifer reproduction.	11
3	Map of Benton County area, Willamette Valley, western Oregon.	41
4	Site 1, profile of Witzel series under forest cover.	46
5	Site 1, profile of Witzel series under grass cover.	47
6	Site 2, profile of Dixonville series under forest cover.	48
7	Site 2, profile of Dixonville series under grass cover.	49
8	Site 3, profile of Ritner series under forest cover.	50
9	Site 3, profile of Ritner series under grass cover.	51
10	Site 4, profile of Dixonville series under forest cover.	52
11	Site 4, profile of Dixonville series under grass cover.	53
12	Site 5, profile of Bellpine series under forest cover.	54
13	Site 5, profile of Bellpine series under grass cover.	55
14	Site 6, profile of Bellpine series under forest cover.	56

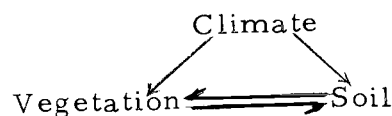
<u>Figure</u>		<u>Page</u>
15	Site 6, profile of Bellpine series under grass cover.	57
16	Site 7, profile of Bellpine series under forest cover.	58
17	Site 7, profile of Bellpine series under grass cover.	59
18	Depth functions of clay fraction.	65
19	Depth functions of pH for selected soils.	68
20	Depth functions of exchange capacity for selected soils.	70
21	Depth functions of exchangeable calcium for selected soils.	75
22	Depth functions of exchangeable magnesium for selected soils.	77
23	Depth functions of Ca:Mg ratio for selected soils.	78
24	Depth functions of K^+ for selected soils.	80
25	Depth functions of exchangeable sodium for selected soils.	81
26	Depth functions of percent organic carbon for selected soils.	83
27	Depth function of percent total nitrogen for selected soils.	85
28	Depth functions of C:N ratio for selected soils.	87
29	Depth functions of free iron oxide for selected soils.	89

COMPARATIVE STUDY OF SOME SOIL CHARACTERISTICS OF GRASSLAND AND FOREST /GRASSLAND TRANSITION IN WESTERN OREGON

INTRODUCTION

Investigators in soil genesis have attempted to isolate particular factors of soil formation for detailed study (Jenny, 1941, 1958) and have also tried to find groups of soils within which all soil forming factors but one were constant. Within each group, any difference from soil to soil could be studied as a function of the single soil forming factor which does vary from site to site. An example of such a group is the biosequence--a group of soils differing in formation only with respect to the biotic factor. Other such sequences are the toposequence, the climosequence, the chromosequence and the lithosequence, within which soil formation can be studied with respect to relief, climate, time, and parent materials respectively.

The biotic factor in pedogenesis is difficult to assess because of the dependence of both vegetation and soil on climate and the interaction of soil and vegetation. Jenny (1941) depicts the interrelationship of these three factors thus:



We concern ourselves here with the lower part of the above triangle, specifically the influence of vegetation on the soil. Field

sites can be found where vegetation is the most important variable producing difference in soil properties. In many places, these vegetational effects could be interpreted as microclimatological influences brought about by vegetation differences, but we shall assume a constant climate by considering only the regional climate so that in areas where this remains constant, we can determine the influence of vegetation on the soils.

Located at the Western part of the Willamette is a series of lower interior foothills some of which are up to 466 meters in elevation. These hills were formed from basalt with some interspersed sedimentary formations. They are generally unsuitable for agriculture due to steep topography. Some of them have stony and very clayey soils (Youngberg 1954). They support stands of Douglas-fir (Pseudotsuga menziesii) and Oregon white oak (Quercus garryana). Scattered through the area are also non-forested openings.

Similar forest stands and their adjacent grasslands (openings) have long been of interest to foresters, ecologists, range specialists, soil scientists, and others. Frequently, openings in forested areas have obvious reasons for their existence such as timber cutting activities or burning. It is believed, however, that natural openings which are mainly grasslands can often be related to differences in geology, soils, drainage, microclimate, and probably other basic factors, either singly or in combination.

The objective of this study is to determine the effects of the vegetation types on some physical and chemical characteristics of the soil.

Geography

The study sites are located on the foothills and slopes in the Western part of the Willamette Valley. The Willamette Valley is located in Western Oregon about 64 kilometers inland from the Pacific Ocean. It is bounded by the Oregon Coast Range on the west and the Cascade Mountains on the east and extends from the Columbia River southward to a point near the town of Cottage Grove, Oregon, where the two ranges converge. North to south the valley is about 200 kilometers in width. The northward flowing Willamette River with its many tributaries is the major drainage system.

Physiographically, the valley is a structural depression with hills of moderate relief separating broad alluvial flats. The valley floor has a very gentle, north-facing slope with elevation increasing from 50 meters at Salem to only 129 meters at Eugene, 130 kilometers to the south. As a result, the Willamette River is a sluggish stream with many meanders, especially from Oregon City southwards.

Geology

The geology of the Willamette Valley is complex and for a full

understanding of the various geological formations, Baldwin (1964) and Beanlieu (1971) are recommended.

The Willamette Valley is bordered on the west by a variety of sedimentary and volcanic rocks of Eocene age. They include submarine pillow basalts, conglomerates, and tuffaceous sandstones and siltstones which are actually eastward extensions of Coast Range formation. In the southern portion of the valley, these Eocene rock formations probably extend to the western margin of the Cascade Range. Marine sedimentary rocks of Oligocene and Miocene age outcrop along the eastern margin of the valley.

The floor of the northern Willamette Valley is underlain by thick, nonmarine sedimentary deposits of Plio-pleistocene age. These deposits are present but not as thick in the southern part of the valley. During the Wisconsin glaciation late in the Pleistocene epoch, the entire valley as far south as Eugene was inundated more than once and partially filled with silt and lacustrine deposits. Recent alluvial deposits occur along the Willamette River on floodplains cut into the Pleistocene lake beds.

Soils

Soils on the valley floor are derived from silty alluvial and lacustrine deposits. Their morphology largely reflects differences in landform position and soil drainage. Well-drained soils situated on

the Willamette River floodplain are deep, moderately darker-colored and range from sandy loam to silty clay loam in texture, with the fine textured soils occupying the older higher floodplain. These soils generally lack subsurface horizons of clay accumulation (Balster and Parsons 1968). Soils occupying terrace positions generally exhibit a greater degree of profile development, typically having silt loam surface horizons and underlain by silty clay loam soil horizons. Depressions on terraces contain very poorly drained Albaqualfs (Planosols).

Soils derived from igneous and sedimentary rocks situated along the edges of the valley and on low hills are mainly Mollisols, Altisols and Ultisols. Soils developing in the very extensive deposits of sandstone exhibit a wide range of characteristics, ranging from brown or yellowish-brown shallow, stony loam to deep reddish silty clay loam showing some clay accumulation and a dark surface horizon of high organic matter content.

Soils developed from siltstone or shale parent materials resemble those derived from sandstone in some aspects, but generally they are noticeably finer textured. Typically, they have a silt loam surface horizon and a silty clay or clay-textured B horizon. Most of the soils developed from basalt in this area are deep, relatively stone free, and well developed. Their surface textures are generally clay

loam or silty clay loam with clay subsoils. In some locations the basaltic soils tend to be fairly shallow and stony.

Generally, the principal soils on the valley floor are Argixerolls (Prairie soils) with lesser amounts of Haploxerolls (Prairie soils), Haplaquolls (Humic Clay Soils) and Albaqualfs (Planosols). On the adjacent uplands and rolling hills, soils are largely Haplohumults (Reddish Brown lateritic soils) with some Haplozerolls (Regosols), Argixerolls (Western Brown Forest Soils) and Xeraqualfs (Red-yellow Podzolic soils).

Climate

In Thornthwaite's climatic classification scheme (Thornthwaite 1931), the Willamette Valley is classed as a humid, microthermal climate with a deficiency of summer precipitation. Average yearly precipitation is about 100 cm, occurring mostly between the months of November and May. Snowfall is most uncommon during the winter and snow rarely remains on the ground for long periods. Very little precipitation occurs during the summer months--about 5% of the total (Wells 1941).

Temperature extremes are rare as high summer temperatures are tempered by an afternoon ocean breeze that occurs almost daily and winter temperatures are moderated by warm-moist ocean winds. The mean winter temperature is about 4°C ; for the spring, 10°C ; for

the summer, 20°C; and for the fall months, 11°C (U.S. Weather Bureau 1938-56). Subzero winter temperatures are rare.

Vegetation

The general vegetational picture of the Willamette Valley today is one of farmlands and pastures interspersed with wooded areas of varying extent (Thilenius 1964). The river bottoms are occupied by riparian facets of Populus tricarpa, Salix spp., Acer macrophyllum, Alnus rubra, and other deciduous trees. Pseudotsuga menziesii, Abies grandis, and Pinus ponderosa may also be found. The steeper and rougher uplands areas are either in unimproved pastures of indigenous and exotic annual and perennial grasses and forbs or support stands of Quercus garryana intermixed with Pseudotsuga menziesii (Fig. 1).

Many ecologists have looked into the successional relationships of the communities within the Willamette Valley. Before the Willamette Valley was settled, much of it was occupied by prairies and oak savannas (Smith 1949, Habeck 1961, Johannessen et al. 1971). Dense forests were confined primarily to the mountain foothills and floodplains. Indians were probably responsible for most of the fires which created and maintained open conditions (Kirkwood 1902, Morris 1934, Johannessen et al. 1971). Douglas (1914) also emphasized the burned nature of the country. He states he was told by the Indians that the



Figure 1. Stand of Quercus garryana intermixed with Pseudotsuga menziesii.

burning was for the purposes of ". . . urging the deer to frequent certain parts, to feed, which they (Indians) leave unburned, and of course they (deer) are easily killed" and ". . . that they (Indians) might better find wild honey and grasshoppers, which both serve as articles of winter food." Other evidence (Ugolini and Schlichte 1973) indicates that the open conditions (the prairies) became established during a time of warmer and perhaps drier climate than occurred near the middle of the postglacial time. This climatic period, which has been called the Hypsithermal Interval by Deevey and Flint (1957), had considerable effect on vegetation as shown in pollen sequences taken from both North America and Europe. According to Flint and Brandtner (1961), the period of maximum warmth occurred between approximately 7500 and 4500 years B.P.

The replacement of a prairie and Quercus savanna by a closed Quercus forest has been the most notable change since the settlement of the Willamette Valley. Habeck (1961, 1962) has documented this change, using land survey records from the 1850's. Thilenius' (1964, 1968) detailed analysis of Quercus garryana stands confirmed the fact that most closed canopy Quercus forests have originated since 1850. He found scattered large trees of open-grown form and averaging 237 years old, which were surrounded by smaller Quercus of forest-grown form, averaging 74 to 105 years of age.

Fire control activities motivated by the settlers are, however, believed to be responsible for this major successional change. The hypothesis that open Quercus savannas were maintained by fire is strengthened by the fire resistance displayed by large, isolated trees during fall field burnings.

Various opinions have been put forward regarding the species which will succeed Quercus garryana and constitute the forest climax of the Willamette Valley. Quercus garryana provides a favorable environment for establishment of Pseudotsuga menziesii seedlings (Sprague and Hansen 1946, Collins 1947, Owen 1953). Large Quercus garryana trees often shelter abundant conifer reproduction (Fig. 2). Some old Quercus garryana snags of open-grown forms can be found in the foothills. They are the victims of the conifer seedlings they sheltered. As a result, Sprague and Hansen (1946) felt that Quercus garryana stands would be replaced by Pseudotsuga menziesii and that these stands might, in turn, be replaced by climax forests of Abies grandis or Abies grandis and Pseudotsuga menziesii. Habeck (1962) supported this idea but added that Quercus garryana would be succeeded by Pseudotsuga menziesii or Acer macrophyllum or both. Johannessen et al. (1971) emphasized the fact that coniferous stands have increased in density since 1853 in the valley and have in some cases been grown, logged, and regenerated on sites which were grasslands or savannas prior to that time. Other authors such as Thilenius



Fig. 2. Large Quercus garryana trees shelter abundant conifer reproduction.

(1964) and Franklin and Dyrness (1973) indicated that several successional segments are likely, depending on local conditions.

Although many of the earlier prairies or grasslands have been lost to forest and woodland transition, some still occupy a few areas. New ones have also been created by settlers by clearing or burning or both. Habeck (1961) provided a list of grasses which may have been characteristic of dry and moist sites in the Willamette Valley. Species on well-drained sites included Agrostis hallii, Agropyron caninum, Bromus carinatus, B. vulgaris, Danthonia californica, Elymus glaucus, Festuca octoflora, F. californica, F. rubra, F. occidentalis, Melica subulata, Poa scabrella, Sitanion jubatum, and Stipa lemmonii. Habeck (1961) also suggested that a large number of forbs were probably present on the native prairies. There is a very high proportion of introduced species in the existing communities, including all the annual grasses, a few perennials and forbs.

LITERATURE REVIEW

In soil genesis, the development of a differentiated profile can be thought of as occurring at least in early stages of development, in two phases (Simonson 1959, Hallsworth 1965). These phases are a stationary phase of a mineral skeleton and a mobile phase consisting of additions, removals, transfers and/or transformations of clays, sesquioxides, exchangeable cations and anions, soluble salts and organic matter. These phases and the dynamics of their interactions are functions of the five soil forming factors one of which is biotic. Hole (1961) suggested a sixth factor of soil formation--space--which helps to determine biologic and hydrologic characteristics of a soil. He further recognized the possibility that horizon differentiation may be inhibited or disturbed by "propedistropic" factors, such as colluvial activity as a function of gravity, the occurrence of tree blowdown or the retardation of water movement by restricting layers in the soil.

For any one soil, the relative importance of pedogenic processes varies and the result is the variety of profiles seen in any landscape.

The role of the multilayered biotic community that surmounts and permeates the horizonated soil body is significant in soil genesis (Buol et al. 1973). Gleason and Cronquist (1964) liken the lithosphere to an easel, the pedosphere to a canvas, and the biosphere to a painting

on the canvas. Yet interactions occur among the three and each layer is changing in character, the surficial one the most rapidly. Wilde (1954) points out that the biologic factor in soil formation is so influential and versatile that even soil reaction is not always a dependable criterion in pedologic theory, as shown by the existence of calcareous Podzol soils and very acid non podzolic mull soils. Polynov (1951) noted that the relative mobility from land to sea of the four elements $\text{Na} > \text{K} > \text{Ca} > \text{Mg}$ has been changed by soil and biota to $\text{Ca} > \text{Na} > \text{Mg} > \text{K}$.

Jenny (1941) has set up a theoretical model $S = f(o, r, p, t, \dots)$ (organisms, climate, relief, parent material, and time), which means that the soil is a function of organisms acting together as an independent variable when the other factors, climate, relief, parent material, and time are kept constant. Here we can consider the biotic factor, organisms, as one of a number of flux potentials (Jenny 1961) introduced into the ecosystem from outside. It should be remembered however, as we are considering organisms as an independent variable, that other factors still play an active part in the holocoenotic environmental complex (Billings 1965).

Hallsworth (1965) indicates that each soil profile can be thought of as consisting of several profiles, one for each property considered. For example, a soil might possess a textural profile, an organic matter profile, etc. It will be more convenient to treat these profiles as created by different types of vegetation under different headings.

Organic Matter

The soils owe many of their properties to the kinds of vegetation they support (Thorp 1948). The main additions to most soils are organic matter from the surface vegetation and their contained elements (Birkland 1974).

Four major kinds of organic horizons in mineral soils have been widely recognized: mor, mull, sward, and Orterde (Muller 1887, Wilde 1946, 1958, Buol et al. 1973). A key for the classification of forest humus types has been introduced by Hoover and Lunt (1952). Most of these horizons correspond with the O1, O2 and A1 in the new Soil Survey terminology (Soil Survey Staff 1962). The mor and mull are mainly associated with forest soils whereas the term sward, coined by Wilde (1958) for the dominantly rhizogenous A1 horizon, is associated with grassland soils. The B horizon of a humus Podzol (Humod), called Orterde by German pedologists, is a soft, weakly granular layer that is dark reddish brown to black and with an organic matter content ranging as high as 15% dry weight.

Another form of dark organic matter in soils is the humic film ("organ" of Brewer 1964) on the surface of peds of B horizons as in some Argiudolls (Brunizems), Natrustolls (Solodized Solonetz soils) and Hapludalfs (Gray Brown Podzolic Soils).

An often-cited example of vegetation-soil relationships is the comparison of forested and grassland soils at the forest-prairie boundary of mid-continental United States and Canada (Jenny 1941, Smith 1950, White and Riecken 1955, Bailey et al. 1964, Ruhe 1969, Pettapiece 1969). According to these authors, Alfisols are present in the deciduous forests, and they possess an A/E/B/C soil horizon sequence. Mollisols in the grassland possess an A/B/C soil horizon sequence. Transitional soils (intergrades) commonly occur between these two kinds of soils in the field, and their properties are intermediate between the two end members.

Although soils may have an equally high content of organic matter at the surface, the distribution with depth varies with vegetation. Thorp (1948) indicates that while it is altogether possible that the total percentage of living and dead organic matter of soils of many forested areas is equal to or greater than that of the Chernozem and Prairie soils (Mollisols) the organic matter of most forest soils is not quite so dark-colored as that of the grasslands and is distributed through greater thickness of soil materials. White and Riecken (1955), Jenny (1941), Floate (1965) and others show that forested A horizons are relatively thin, and that the organic matter content decreases rapidly with depth, whereas grassland A horizons are thick, and the organic matter content remains high for a considerable depth. Buol et al. (1974) opine that about 90% of the total organic matter of

prairie tessera is below ground. Wilde (1950) in support of the light colored nature of forest soil indicates that in Indiana, the light colored A horizon of forest soils may contain considerable accumulations of light colored organic matter with as much as 6% in the first 15 cm.

According to Thorp (1948), the ratio of humified soil organic matter to roots in most forest soils is less than in the grassland and generally, the majority of forest soils in the United States are light colored except the Brown Forest soils and a few areas of forested Rendzinas which have dark colored surface soils not greatly different in superficial appearance from Chernozems. Another exception is the very thin layer found immediately under the decaying leaves of the forest. Shields et al. (1968) proved that the gray wooded soils (Boralfs) are significantly higher in color value (Munsell notation) per unit of organic matter than are the Chernozemic soils (Mollisols). According to Shields et al. there are differences in the ultraviolet reflectance spectra of these two soils and in the E4:E6 ratios of alkali extracts.

Thorp (1948) feels that the difference in the amount and distribution of organic matter between the dark colored grassland and light colored forested soils can be ascribed partly to the fact that forest soils are generally more acid in reaction than grassland soils. The grassland soils organic matter is developed largely through the decay

in place of a myriad of grass roots rather evenly distributed through the soil, and by decomposition of grass tops which die back annually. The "mild" or non-acid humus so formed in the grassland is more stable than the relatively soluble acid humus of the forest. Thorp also opines that part of the differences may be due to the fact that the molds (fungi) are more active in the organic matter of forest than bacteria, whereas the reverse is true in most grassland.

Alfanas'yeva (1966), in his study of the thick Chernozemes under grass and tree coenoses, has shown that throughout the year the CO_2 concentration of the soil air in a forest Chernozem is higher than in a Chernozem under grass. This is due to the structure of woody coenoses and more intense exchange of substances in the oak stands. These conditions, together with the different rhythm of the seasonal development of the oak forest, create better conditions for the eluvial process in soils under a forest with the same macroclimate than in the soils under grass vegetation. According to Alfanas'yeva, the top layers of forest Chernozems receive more plant residues every year than do virgin Chernozems of the weed grass steppes, but they have less humus, suggesting that the biological cycle is more vigorous in a forest than in a steppe.

Campbell et al. (1967), in their soil humus studies, show that organic components of soil are composed of at least three fractions when considered on a dynamic basis: (1) decomposing plant residues

and the associated biomass which turn over at least once every few years, (2) microbial metabolites and cell wall constituents that become stabilized in soil and possess a half-life of 5-25 years, and (3) the resistant fractions, which in grassland soils are composed of humic components ranging in age from 250 up to 2500 years. The results of Campbell et al. may help to explain why the grassland soils have higher content of organic matter.

Studies of the composition of humic acids (Alexandrova et al. 1968) extracted from the humification products of vegetable residues, showed that variations in elemental composition and the total amount of functional groups in the first stages of humification are conditioned by the chemical composition of plant residues. The characteristic humification reactions are carboxylation of the high molecular products of hydrolytic breakdown of protein, tannin and lignin, and their intercondensation. The double-component system of humus substances, consisting of humic acid and fulvic acid, forms by breakdown of the high molecular weight products of humification when they interact with the mineral part of the soil. The less soluble and less dispersible substances remain in the soil phase at the place of formation, giving humic acids and their organo-mineral compounds. The more dispersible and more soluble compounds form fulvic acids and migrate down the soil profile, also giving a number of organo-mineral derivatives.

Shields (1965), in his study of a sequence of loamy till, cultivated grassland, transitional and forest soils in Saskatchewan, reveals regular differences in humic composition and in degree of aromatic condensation as estimated from a spectral parameter, the E4:E6 ratio. Generally, the black grassland soils have larger amounts of humic acid and the humic acid is more strongly condensed than in Gray Wooded and Brown soils. This also establishes the importance of the quantitative characteristics of humic constituents in determining the relative contribution of soil organic matter to soil color value as mentioned earlier. Lowe (1969) and Dormoar (1967) also share the same view.

Mekeague (1968) utilizes humus composition parameters such as humic:fulvic acid ratio and is able to differentiate Podsollic B horizons from A horizons of various soils. He finds general relationship between degree of degradation and humus composition and also finds that fulvic acids predominated in the more podzolic soils.

Clay Fraction

The distribution of clay-size particles in many moderately to strongly developed soils is marked by relatively low contents in the A and C horizons with the maximum amount in the B horizon, generally in the upper part of the B (Birkeland 1974). Several processes may account for this distribution. One such process is that the

constituents of clay are derived by weathering higher in the profile and that they move downward in solution with the percolating water and precipitate as clay minerals in the B horizon. The second process is that the clays have formed in place from mineral weathering in the B horizon. The third is that the clays have moved as particles in suspension in the downward percolating water to accumulate in the B horizon because of flocculation, constrictions in the pores through which the water moves, or because the base of the B horizon marks the lower limit of most water movement (McKeague and Arnaud 1969). It should be noted that there may be a combination of these processes in the clay formation in the B horizons of most soils, but there might be greater preponderance of one process than the other.

In a situation where the third process, translocation, plays the major role a striking difference will be found in the morphology of soils formed under different vegetation types. Jenny (1941) has shown in his study of soils under forest and under prairie in Illinois that the average content of colloidal clay particles ($< 1\mu$) was nearly the same in two profiles, 28% for prairie and 26% for timber, but significant differences were found in the distribution patterns. The concentration of the clay particles in the B horizon was most marked under the forest cover, a feature that is due to enhanced translocation of clay from the surface horizon to lower parts of the profile. Similar results have been reported by Smith et al. (1950) and McComb and

Riecken (1961). McComb and Riecken in addition mentioned that translocation of clay into B horizon resulted in the decrease in permeability of the B horizon of forest soils in Iowa.

Sharander (1950), in a study of loess-derived soil in Missouri, revealed that there was a significant difference between the means of the surface soil clay content of soils derived under prairie, under forest, and under prairie forest transition vegetation, that differences in surface clay content were most marked between forest and prairie soils with the transition soils occupying a position about midway between the other two and that the differences in surface clay content were most marked between the less developed soils and were slight for the highly developed soils. Bray of Illinois was also quoted as stating that in the later stages of development, the difference caused by vegetation became very slight (Jenny 1941). Some others have concerned themselves with the mode of translocation of clay particles. In Duchaufour's (1970) opinion, complexes in which the cation is at the center of an octahedron consisting of six organic ligands that are not chemically interlocked predominate in a Podzolic soil. In this case, the hydrophilic humus molecules form a shell around clay particles and promote their dispersion. Several researchers (Antipov-Karatayev and Tsyuvupa 1961, Coutts et al. 1968, and others) have established some chelate compounds, such as Fe-D TPA, for example, are irreversibly fixed on the surface of clay particles and determine

the mobility of the latter. There is an indication (Aleksandrova 1967, Dudas and Pawluk 1969) of the possibility of formation of clay-humus complexes in soil. The Ca, Fe and other cations can serve as bridges in the ionic bond of amino acid with a mineral. However, artificially obtained clay humus complexes (Mel'nilcova and Kovenza 1974) without a cation bridge are stronger (a chemical bond formed between clay and amino acids). In the first case a correlation is found between the clay content and mobile forms of iron, and in the second case, between the clay content and total iron. The possibility of formation of mobile organo-mineral complexes in soil can explain the stability of soil suspensions and the mobility of clay along the profile under favorable hydrophysical conditions.

The most striking result of Sharander's (1950) studies is that the Missouri soils developed under grassland contained more montmorillonitic clay minerals than did the soils developed under forest. Giesecking (1949) and Whiteside and Marshall (1944) got similar results. Peterson (1946) reported that loess-derived Tama silt loam, a Prairie soil and its forested equivalent, the Clinton silt loam in Iowa had essentially the same clay minerals--both high in montmorillonite--but he did present some evidence to indicate that there might be more kaolinite in surface of the forest soils than in the prairie; the base exchange capacity of prairie soil being higher (84 ME) than that of forest soil (61.9 ME).

Soil Reaction

Over the past years soil acidity and pH have been closely associated and one has most often been thought of in connection with the other. This idea resulted from purely chemical systems in which a lower pH corresponds to a stronger acidity. Moreover, based upon the older notion of acidity as being the ability to donate hydrogen ions, soil acidity has been visualized as arising due to existence of hydrogen ions on the exchange site, broken edges of the clay minerals, or on the organic matter present in the soil. This concept of soil acidity was widely accepted and went undisputed as late as the 1940's. Even then methods were being devised to determine the hydrogen ion concentration in the soil systems (Mehlich, 1942, 1943). Existence of proton donor species other than the hydrogen ion was either not known or ignored, hence the findings of some researchers such as Veitch (1902), and Schofield (1955) that a salt-exchangeable aluminum definitely occurred in soil systems, passed unnoticed.

Jenny (1961a), Coleman (1953), Low (1955) and other authors brought up again the idea of aluminum being responsible for soil acidity. Their idea became widely accepted and this has been further substantiated by much literature appearing on the subject since then (Turner, Nichol and Brydon 1963, Yuan 1963, Coleman and Thomas 1964, McLean et al. 1964, Ayres, Hagihara and Stanford 1965,

Clark 1966, Pionkee and Corey 1967, and others). In the recently accepted theory of soil reaction, the proton producing species in soils are: H_3O^+ or H_9O_4^+ , $\text{Fe}(\text{OH}_2)_6^{3+}$, and $\text{Al}(\text{H}_2\text{O})_6^{3+}$ present on the exchange sites, hydrogen ions present on the mineral edges, organic matter, and aluminum present as $\text{Al}_m(\text{OH})_{n-x}(\text{H}_2\text{O})^+$ in the crystalline and amorphous materials of clays (Jackson, 1956), hydroxy- Fe^{+3} present in intergrade minerals (Coleman and Thomas 1967) and probably manganese present in various forms and different locations (Yuan 1963).

Since organic matter is contributory to the pH of the soil and since different types of vegetation produce different types of organic matter, we must expect some differences in pH of soils under different vegetation covers. It has long been established that many species, particularly the conifers, tend to intensify soil acidity due to their acid producing litter (Ovington 1953, Brady 1974). Indeed, many investigators refer to the weathering of rocks that takes place under certain coniferous trees as "acid weathering." An extreme example is found in New Zealand, where local podzolization occurs under Podocarpus sp. (Jackson and Sherman 1953) and Kauri pine (Agathis australis) (Crocker 1952). The influence of the tree is such that the spodosols occur only under each tree and not beyond the influence of the tree, to form "basket spodosols." Zinke (1962) demonstrated the effect of a single shore pine tree, Pinus contorta, on surface soil

properties indicating that the soil was much more acid beneath the tree crown relative to non-vegetated sand of the same parent material. It should be noted that the tree leaf litter and bark litter contribute to the acidity and that bark litter is more acid and has a lower cation and nitrogen content.

Comparison of soils developed under prairie and under forest in Illinois reveals that in all horizons the pH is lower under forest than under prairie (Jenny 1941). The lower A horizon and the upper B horizons are highest in acidity. Contrary to the above findings, studies in Washington state reveal that there is no marked difference between soils formed under grass vegetation (Lotspech et al. 1961, Ugolini and Schlichte 1973). Differences in acidity between the Prairie and Grey-Brown Podzolic soils have been advanced as an explanation for the difference in organic matter. Smith et al. (1950) indicated that this explanation seemed untenable. They opined that the vegetation and organic matter differences of the Chernozem and Gray-wooded soils of Alberta, Saskatchewan and western Minnesota paralleled the relationships between the Prairie and Gray-brown Podzolic soil of the corn belt, yet the Gray-wooded soils had the same reaction profile as the Chernozem, as indicated by Newton, Ward and Bently (Smith et al. 1950). They, therefore, argued that if reaction was the controlling factor it would seem that the Gray-wooded soils should resemble the Chernozems rather than the Gray-Brown Podzolic soils.

Total Nitrogen

According to Jenny (1930), the order of importance of the soil-forming factors in determining the nitrogen contents of loamy soils within the United States as a whole is as follows: climate > vegetation > topography = parent material > age. The vegetation factor of soil formation is concerned with the pool of plant species available to a given location, not the quantity of vegetative growth produced. The latter is considered to be controlled by the other soil forming factors especially climate (Jenny 1941).

The nitrogen in soil, especially that in the epipedon, occurs largely in organic combination, thus the process of nitrogen accumulation bears a close relationship to the accumulation of organic matter (Schreiner and Brown 1938). As a rule, the carbon-nitrogen ratio of the undisturbed topsoil in equilibrium with its environment is about 10 or 12 to 1 in the temperate region (Tisdale and Nelson 1975). It narrows in the subsoil in many cases, partly because of the higher content of NH_4^+ nitrogen and the generally lower amounts of carbon.

Other factors being equal, soils developed under plants with extensive root systems generally have higher nitrogen and organic matter contents than those developed under plants with restricted root systems. Under grass vegetation, where considerable debris is added in forms of root excretions and as sloughed off roots, dark-colored

soils (Mollisols) are formed which contain relatively high amounts of nitrogen. Light colored soils such as Gray-brown podzolic soils (Udalfs) are formed with relatively low amounts of nitrogen under forest type vegetation where most of the plant debris is added to the soil in the form of fallen leaves. In commenting on the unusually high nitrogen values obtained by Smith et al. (1951) for some tropical soils of Puerto Rico, Joffe (1955) suggested that the human factor was partly responsible, namely through cultivation of plants having sod characteristics.

The reason for the high nitrogen content of grassland soils, as compared to forest soils, has been the subject of considerable conjecture. It is common knowledge that the nitrogen and organic matter level of most soils cannot be maintained, or increased under tillage without putting them into grass sods; consequently, the conditions in grassland soils are considered favorable for the retention of available nitrogen as humic substances (Stevenson 1965). Harmsen (1951) concluded that the accumulation of nitrogen and organic matter in grassland soil cannot be explained solely by the higher amounts of plant residues produced, because luxuriously growing crops, together with higher applications of organic matter in the form of manures, composts, or green manures cannot avert completely the depletion of nitrogen and organic matter during cropping. It is claimed that lack of sufficient aeration in densely populated grassland soils contributes

to the preservation of nitrogen and organic matter but Harmsen (1955) disagreed with this and indicated that well drained grassland soils were generally better aerated than the arable land.

Further explanation for high nitrogen contents of grassland soils has been given by Mattson et al. (1943). They claimed that high base saturation in the soil leads to the absorption of atmospheric oxygen and the fixation of ammonia by lignin, or lignin-nitrogenous constituents of the organic matter. In their opinion grassland soils which always contain more bases should have higher nitrogen contents and lower C/N ratios than the forest soils with lower base status. Their theory was used by Moodie (1951) to explain the variable results obtained by application of the hypiodite method for studying the nature of the organic matter.

According to Walker et al. (1954) the main source of nitrogen in the grassland seems to come from symbiotic nitrogen fixation by associated legumes. Stevenson (1965) indicated that the rhizosphere was a zone of intense microbial activity due to the presence of abundant energy from grass roots excretion and sloughed off roots and that fixation of nitrogen in this zone through the activities of non-symbiotic-fixing heterotrophs, such as *Azotobacter* and *Clostridium*, might facilitate the production of humus substances in grassland soils.

Some authors have proposed that substances secreted from living roots suppress nitrification and consequently nitrogen is preserved and can accumulate to a high level (Theron 1951, Theron and Haylett 1953, Soulides and Clark 1958). Other authors think that clay minerals help in restricting the decomposition of organic matter and can accumulate protein rich organic matter in their interlayers (Grim 1953, Pinck et al. 1955).

According to Stevenson (1965), in forest soils differences in the profile distribution of nitrogen occur by virtue of the manner in which the leaf litter becomes mixed with mineral matter. In soils formed under deciduous forests on well-drained sites adequately supplied with calcium such as Gray-brown podzolics (Udalf) the litter becomes well mixed with the mineral layer through the activities of earthworms and other organisms of microfauna. In this case, the nitrogen profile shows little if any accumulation in the B horizon. On the other hand, in soils formed under coniferous forests on soils low in available calcium (Spodosols) the leaf litter does not become mixed with the mineral matter, but forms a mat on the soil surface. The nitrogen profile of a Spodosol generally shows a bulge in the B horizon as a result of the eluviation of soluble organic constituents partly as metal complexes, into this horizon.

Cation-Exchange Capacity and Base Saturation

Colloidal material in soils carries an electrical charge. Although both negative- and positive-charge sites exist, the origin of the negative sites has been studied most extensively. The negative charge is approximated by the cation-exchange capacity (CEC) of the material expressed in milliequivalents per unit dry weight of the material and most of the cations attracted to these sites are exchangeable. The CEC varies in amount and origin with the soil material, hence it is important for both soil fertility nutrition studies and for genesis. It is also widely used in soil classification considerations.

In humus, the charge originates from the dissociation of H^+ from carboxyl and phenolic groups at the particle surface and within the particle. In clay particles, the charge originates at the clay edge or along the surfaces parallel to the ab crystallographic plane (Carroll 1959). Charges along the clay edge originate from unsatisfied (broken) bonds at the edge of the particle, say, between Si-O or Al-OH, or from the dissociation of H^+ from OH^- . Charges originating in this way are common in the 1:1 clays and is dependent on particle size, because the number of exposed edges increases as particle size decreases. Isomorphous substitution in the clay lattice commonly results in a net negative charge, mainly along the interlayer surfaces

This is probably the main origin of charge in the 2:1 type clays, although some charge originates at the clay edge.

The highest values of CEC are in organic matter (Carroll 1959, Grim 1968, Powers 1932). According to Powers, decomposition results in a higher concentration of lignin and a consequent increase in base exchange capacity. The variation of CEC with clay mineral is due to a combination of ionic substitution and its extent, the degree of hydration, and the number of exchangeable sites at the edges of particles. Because organic matter can have such a high CEC the presence of a small amount of organic matter can greatly affect the CEC of soil. In contrast, nonclay minerals and rock fragments have a negligible effect on the CEC, except zeolite (Gedroiz 1922, Carroll 1959, Grim 1968, Birkland 1974). One can, therefore, roughly estimate the CED of the clay fraction, and hence of the possible clay mineral present, by knowing the CEC of the soil, the amount of clay present and possibly the amount of organic carbon. Multiple regression equations relating CEC with clay and organic matter contents have been calculated by William (1932), Helling et al. (1964), Hallsworth and Wilkinson (1958), and Mohamed and Cohan (1961).

Joffe (1949) reports Russian work on Chernozems where the clay and organic matter contributions are additive. Naturally, the exchange capacity of the clay fractions will vary considerably

depending on the nature of the clay fractions, but the contribution attributed to the organic matter is remarkably similar for the various soils.

The formation of clay-organic matter complexes might be expected to modify the additive function established above, and Esminger and Giesecking (1941) noted a marked drop in the exchange capacity of montmorillonitic soil after the absorption of gelatin. Hendricks (1941) has also shown that certain large organic molecules when absorbed on montmorillonite cover up some of the exchange sites and reduce the cation exchange sites and the CEC. McLean (1952), on the other hand, considers the reduction in exchange capacity of a mixed system of beidellite clay and humic acid was due to a loss of exchange capacity by the humic acid rather than obstruction of exchange sites on the clay. For many grassland soils (not high in clay), Walker (1956) has found that there is a high correlation between carbon and CEC. According to Walker's view, it is possible in grassland soils rich in organic matter that the exchange sites of the clay minerals are completely blocked by organic molecules, particularly if the content of montmorillonitic clay is low. It is said that hydrous oxides of iron and aluminum, on the other hand, tend to behave as positively charged colloids under soil conditions and thus may neutralize negative charges on organic colloids (Walker 1956). It may, therefore, be possible to find grassland soils rich in organic

matter with high CEC, but low base saturation in relation to the soil pH. This appears to be so for soils containing allophane or its variants (Soils of North Island, N.Z. Dept. Sci. Ind. Res. 1955).

The variation in quantity and quality of organic matter in soils of different vegetation cover will obviously bring about variation in both CEC and base saturation of these soils. In forested soils the main input of organic matter, as mentioned earlier, is by litter fall to the soil surface, whereas in grassland soils the organic matter input is both by litterfall and by root decay at depth. The overall percentage base saturation tends to be higher in grassland than in forests. This difference could be a function of greater leaching in the forests along with greater annual biomass production and cycling of cations in grassland (White and Reicken 1955). The greater leaching in the forests relative to the grasslands could be due, in part, to lower evapotranspiration rates and the presence of more chelating agents and more acid leaching waters under a forest canopy. On the other hand, base status and pH of the A1 horizon are higher for the forested soils than for the associated grassland soils. Jackson and Sherman (1953) have proposed that tree roots penetrate more deeply than grass roots to the less weathered material and are able to return more bases to the soil surface. Tarrant (1949) has also proposed that there is a direct exchange of cations between roots and the soil so that nutrient elements need not necessarily go into solution to

be made available to the plant. Micorrhizal activity also provides a means of nutrient intake for the plant, thus adding to total nutrient availability. A compilation of data of several authors (Chandler 1941, Jenny 1941, Kik 1943, Fraps et al. 1945, Metz 1952) shows that probably more calcium is returned to the soil surface by the annual leaf fall of the indigenous deciduous trees than is returned by the die-back of the foliar parts of the prairie grass. Proportionately more calcium than magnesium is found in the leaf tissue of trees as compared to the aerial parts of the grass. The higher ratio of the exchangeable calcium to exchangeable magnesium in the A1 horizon of the forested soil in comparison to that of the associated grassland soil of the Mississippi Valley can be attributed, therefore, to the biotic factor of soil formation. Barshad (1960), on the other hand, feels that the ratio of exchangeable calcium to exchangeable magnesium can be used as an indication of soil development since the ratio narrows with increased weathering and leaching.

Pedogenetic Oxides

The oxides and hydroxides of Fe and Al are the products of weathering in many soils, the formation of which depends on the weathering conditions and on the parent material. Silicic acid is usually depleted more rapidly than Al and Fe in soils that are subject to strong weathering and leaching with the result that amorphous

hydrous oxides of Al and Fe are accumulated in the soil. Under more moderate conditions, the formation of silicate clay minerals together with hydrous oxides may be dominant and these products affect many soil properties by forming interlayers and coatings, and also by their interaction with organic matter (Schwertmann 1966). The extent of formation of hydrous oxides as well as further transformations, such as movement and reprecipitation or the formation of crystalline forms through aging, makes these materials, particularly the various forms of Fe, useful indicators in soil genesis and classification (Gorbunov et al. 1961, Oades 1963, Mitchell et al. 1964).

Available data indicate that Fe and Al probably move in the soil as soluble metallo-organic chelating complexes. Fulvic acids are thought to be the common chelating agents for a number of soils (Kononova 1961, Wright and Schnitzer 1963, Schnitzer 1969). The stability of the chelate that forms is a function of size and valency of the metal ion, with increasing stability associated with smaller size and higher valencies. The process of translocation envisaged begins with the production of fulvic acids in the O or A horizon. A stable fulvic acid chelate is formed with Al^{3+} and Fe^{3+} , or with aluminum and iron hydroxy ions, and because these chelates are water-soluble they move downward with the percolating soil water. The precipitation of these complexes can be brought about by small changes in ionic content (Kundblad 1933, Schwitzer 1969) or by microbial action

(Schuylenborgh 1965). Oades (1963) indicates that ferric ions in soils are hydrolytically precipitated as hydrous oxides and that further transformations depend mainly on pH, temperature, redox potentials and presence of other compounds. Schwertmann et al. (1964) indicate that the crystallization of artificial and natural ferric hydroxides is significantly retarded or even inhibited by organic compounds.

Goethite (α -Fe OOH) and hematite (α -Fe₂O₃) are the Fe oxides most commonly encountered in soils, but Lepidocrocite and Maghemite (γ -Fe₂O₃) are encountered to lesser extent. Goethite is the predominant form under the temperate, humid climate such as western Oregon. Its formation in soils is generally attributed to at least two different processes (Oades 1963): direct slow precipitation of ferric ions in solution at low pH, and aging of ferric hydroxide gel. It is said that low temperature, high pH, and the presence of various ions, particularly SO₄⁻ and HCO₃, favor the formation of goethite whereas the formation of hematite which is commonly found in tropical and subtropical soils is favored by high temperature and low pH (Schellmann 1959).

In soils of the temperate humid climate, the free Al oxide content is generally low. In such climates Al normally participates either in the formation of clay minerals, or it forms gibbsitic interlayers in expandable layer silicates (Schweitmann 1966).

As mentioned earlier, since there is a tendency for more fulvic acid to form in forest soils than in the prairie soils, it should be expected that the Fe and Al oxides will be higher in the B horizons of forest than in the prairie soils.

According to Birkland (1974), conditions for chelate formation and movement of Fe and Al are best for Spodosols, but movement can also take place in other soils, as shown by some with E horizons. Generally, the degree of podzolization declines in a transect going from boreal forest to grassland (Kononova 1961), that is, in a transect from Spodosols to Alfisols to Mollisols. In such a transect, the humic acid:fulvic acid ratio increases from about 0.5 to 1.5-2.0. Birkland (1974) reports that the humic acids in the spodosols are similar to the fulvic acids in that they are dispersed and mobile. The calcium-humic acid complexes that are both relatively stable and immobile form in Mollisols and thus are not capable of translocating Fe and Al.

Ugolini and Schlichte (1973) report that the pedogenic Fe is higher in amount and more mobile in the forest than in the grassland in western Washington. They indicate that amorphous Fe reaches a peak in the B horizons of the forest but not in the prairie soils. Birkland (1974) indicated that the ratio of free Fe_2O_3 in A and/or E horizons relative to that in the B horizons is lower in forested than in grassland soils. These authors interpret the differences as meaning

that the organic fractions of the forest have greater ability to complex Fe than the prairie.

Chester et al. (1957) indicated that in all soils they studied, iron oxide showed a marked effect on aggregation with a tendency to be more important in the smaller aggregate size range. They found that the effect of clay was very small with the exception of the high clay soil (often 30% clay) in which its effect was only exceeded by that of the free iron oxide. Lutz (1936) has also indicated that free iron oxide content has been shown to be positively correlated with stability of soil aggregates.

MATERIALS AND METHODS

Field Investigation

Thorough investigation was made to select landscape units in the western part of the Willamette Valley to represent the two vegetation types. Soil survey techniques including air photo interpretation and preliminary augering were used to ascertain that each unit was underlain by the same parent material. Eventually, seven landscape units were chosen, each unit consisting of the grassland and an adjacent forest. A site number was assigned to each of the units as shown in Figure 3.

The location, name of soil series and elevation of each study site are as shown in Table 1.

Table 1. Locations, names of soil series, and elevations of the study sites.

Site	Location			Elevation (meters)
	Section	T. & R.		
1	22	10S, 5 W	Witzel	330
2	5	11S, 5 W	Dixonville	170
3	3	11S, 5 W	Ritner	300
4	18	11S, 5 W	Dixonville	200
5	35	12S, 5 W	Bellpine	130
6	3	13S, 6 W	Bellpine	200
7	9	12S, 6 W	Bellpine	210

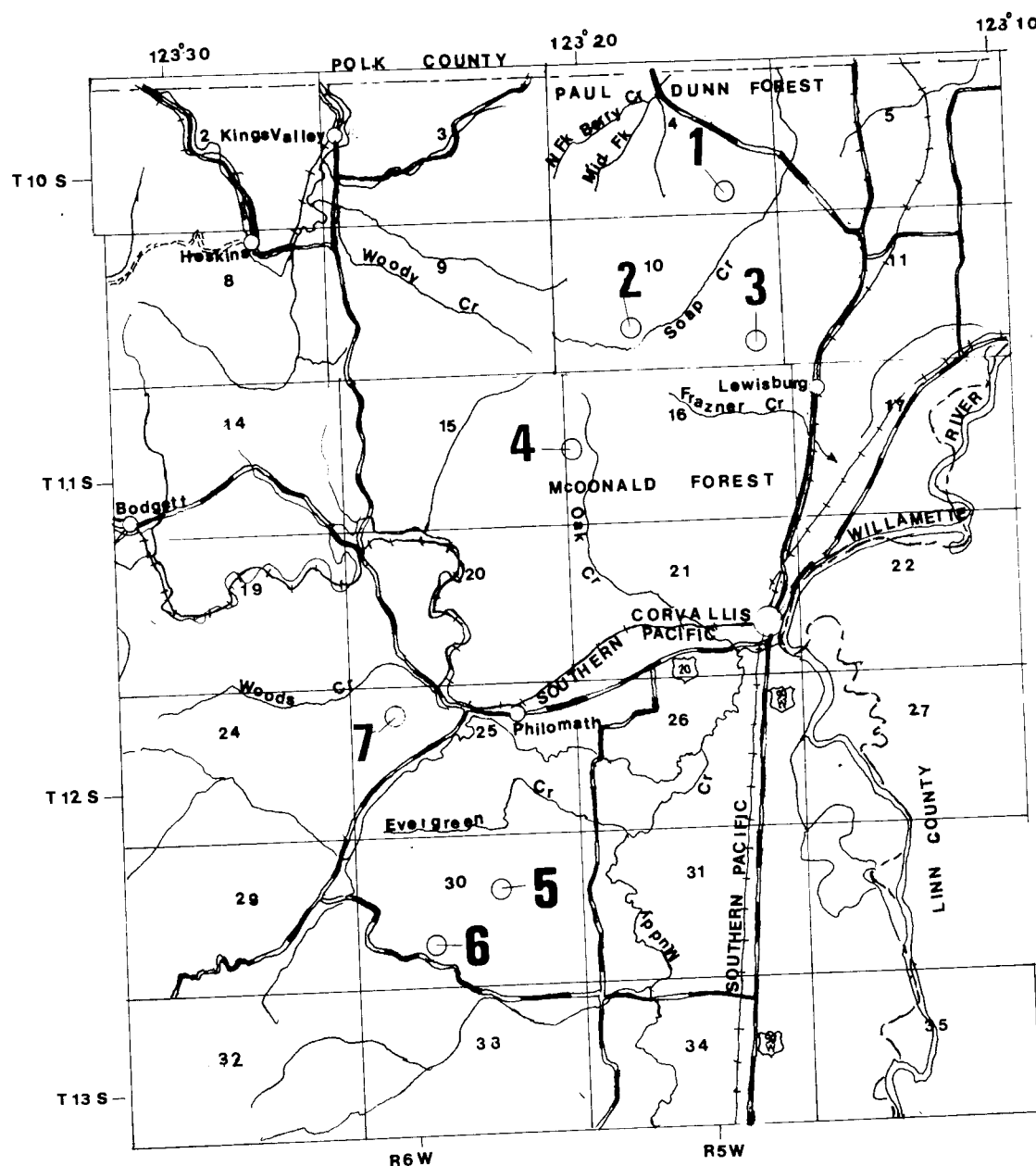


Fig. 3. BENTON COUNTY AREA, WILLAMETTE VALLEY, WESTERN OREGON

1 0 1 2 3 4 5 6 7 Kilometers

COMPILED FROM AERIAL PHOTOS DATED 1973. RECTANGLES

REPRESENT MAP SHEETS.

Surfaced Road	—
Rail Road	—+—+—+—+—
Dirt Road	==
Boundary Line	- - - -
Study Site	○

All the study sites were on south facing slopes. Some of the soil series are the inclusions of the dominant series which appear in the soil map (Knezevich 1975).

At each study site, two profile pits were opened, one in the grass and the other in the adjacent forest cover. A profile description was made for each exposed profile using the techniques of the Soil Survey Manual (Soil Survey Staff, 1951, 1962). Soil color was determined with the aid of Munsell color charts. The vegetation of each site was analyzed by estimation of species coverage. The age of the forest cover at each site was estimated making combined use of annual ring count of recently cut nearby tree stumps, internode counting of the existing conifer trees and the information obtained from Mr. Marvin Rowley of the School of Forestry, Oregon State University, Corvallis. Descriptions of soil profiles, vegetation, and other site characteristics are presented in Appendix I. The trees are listed in the order of dominance. Asterisks indicate that the tree species forms over 80% cover.

The content of coarse fragments, by volume, was estimated for each soil horizon by scrutinizing the pit face. Each profile was sampled by horizon for laboratory analysis.

Laboratory Investigations

Sample Handling and Preparation

The soil samples were collected during the summer of 1975 so that they were not wet but moist. These samples were air dried on clean tables in the laboratory, crushed and screened to remove all coarse fragments prior to laboratory analyses. The screening was done by two methods. One method involved the use of 2 mm sieve for physical analysis, the other involved the use of a Nasco-Asplin soil grinder for chemical analysis (Robert et al. 1971).

Physical Methods

Physical property analysis was performed by the author. Particle size distribution of the entire profile of each individual was determined by modification of the Engle and Yoder (1926), Kilmer and Alexander (1949) and Chu and Davidson (1953) pipette methods. Soluble salts were not removed since preliminary trials gave comparable amounts of sand, silt and clay whether salts were removed or not. Each 10 gram sample was treated first with 30% hydrogen peroxide to remove the organic matter, then dispersed with 10 ml of 5% Calgon solution using air under 30 pounds of pressure for 5 min. The material remaining on the 300-mesh sieve, after washing to remove the silt and clay, was fractionated by dry sieving.

Chemical Methods

Chemical analyses were performed in the Soil Testing Laboratory at Oregon State University in Corvallis.

Cation exchange capacity was determined by the ammonium acetate method of Schollenberger and Simon (1945). The soils were not initially leached for removal of soluble salts, nor were corrections made for dissolved carbonates.

Exchangeable calcium, magnesium, potassium and sodium were extracted by ammonium acetate at pH 7 (Peech et al. 1947) and determined on a Perkin Elmer Atomic Absorption Spectrophotometer.

Soil reaction, expressed as pH units, was determined with a Sargent pH meter model DR. with electrodes immersed in a 1:2 soil to water suspension (Jackson, 1958).

Organic carbon was determined by the Walkley-Black potassium dichromate procedure (Walkley et al. 1934). Ferroin or o-phenanthroline ferrous sulfate complex was used as the indicator instead of diphenylamine (Roberts et al. 1971). This method was assumed to recover 75 to 80% of the organic carbon.

The total nitrogen was determined by the Macro-Kjeldahl method (Bremner 1965).

Free iron oxides were extracted by shaking for 2 hr in sodium citrate-sodium dithionite solution and determined on a Perkin Elmer Atomic Absorption Spectrophotometer.

RESULTS AND DISCUSSION

Physical Properties

Soil Morphology

Soil profile descriptions for all the profiles investigated are presented in Appendix I. The photographs of these profiles are shown in Figures 4 through 17. The most obvious morphological differences between the forest transition soils and the grassland soils are the color and structure of A horizon. The A horizons of most forest soils show lighter colors than those of the grassland soils. A typical example is in site 2 (Appendix I) where the A1 horizon of forest soil is one unit higher in value than that of grassland soil. Similar differences are seen in sites 1, 4 and 7. This is consistent with the findings of Shields et al. (1968), Wilde (1950) and Thorp (1948) as mentioned in the literature review.

The difference in soil color might be attributed to differences in the color and quantity of organic matter in the soils. This idea has been presented by Wilde (1950) who indicated that the light-colored A horizon of forest soils in Indiana might contain considerable accumulations of light-colored organic matter with as much as 6% in the first 15 cm. Shield et al. (1968) also found differences in the ultraviolet reflectance spectra of forest soils and grassland soils in



Figure 4. Site 1, profile of Witzel series under forest cover.



Figure 5. Site 1, profile of Witzel series under grass cover.



Figure 6. Site 2, profile of Dixonville series under forest cover.



Figure 7. Site 2, profile of Dixonville series under grass cover.



Figure 8. Site 3, profile of Ritner series under forest cover.



Figure 9. Site 3, profile of Ritner series under grass cover.



Figure 10. Site 4, profile of Dixonville series under forest cover.



Figure 11. Site 4, profile of Dixonville series under grass cover.



Figure 12. Site 5, profile of Bellpine series under forest cover.



Figure 13. Site 5, profile of Bellpine series under grass cover.

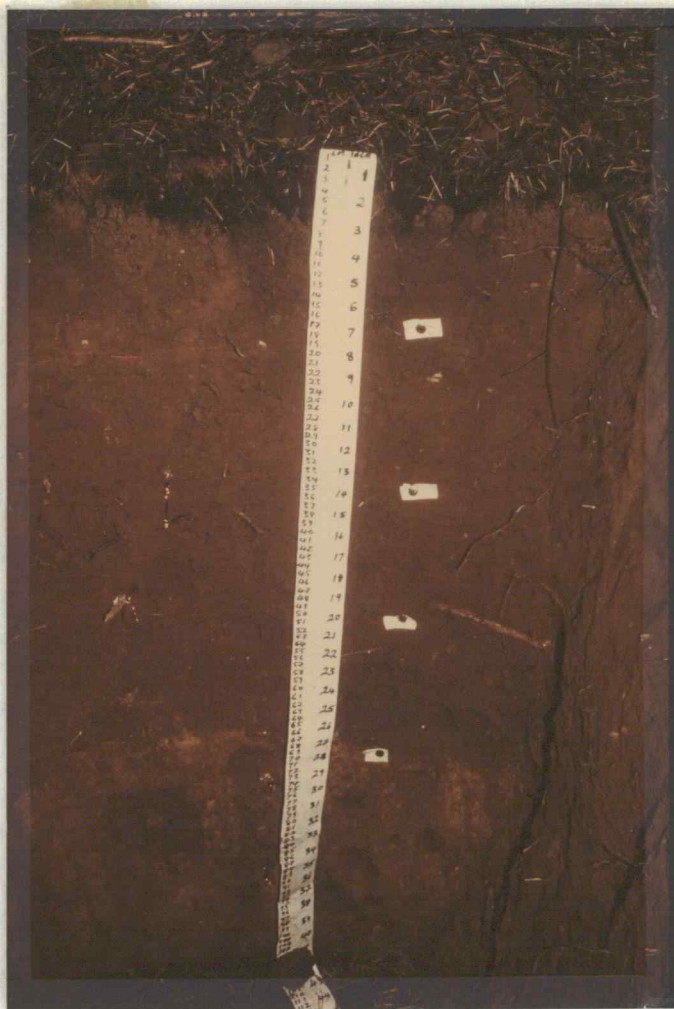


Figure 14. Site 6, profile of Bellpine series under forest cover.

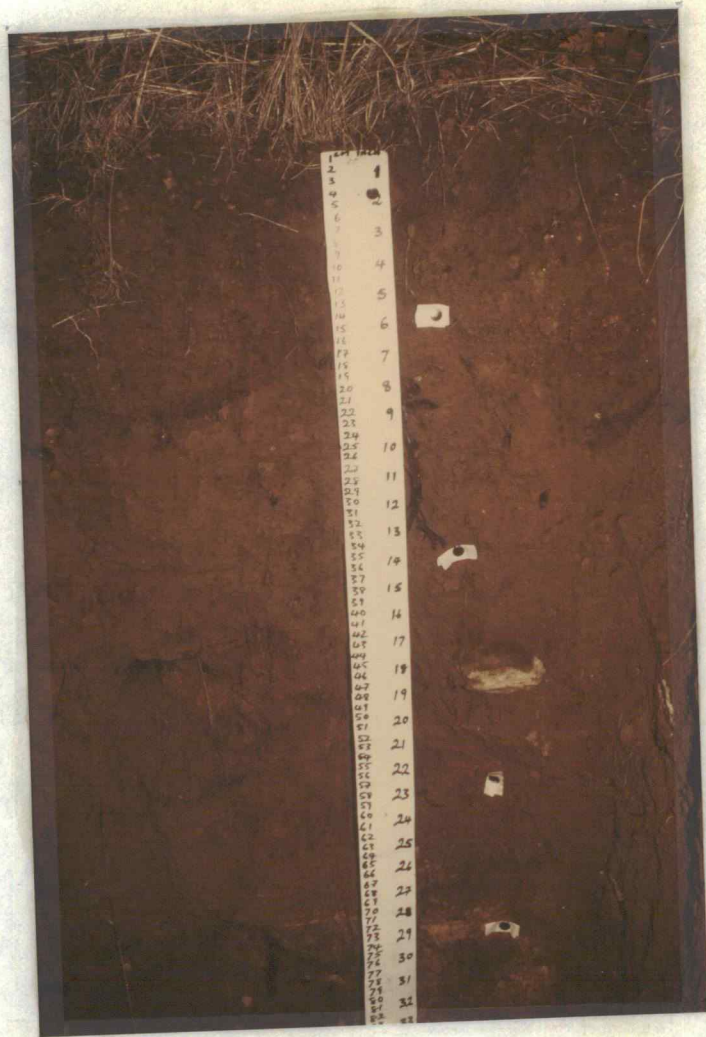


Figure 15. Site 6, profile of Bellpine series under grass cover.



Figure 16. Site 7, profile of Bellpine series under forest cover.

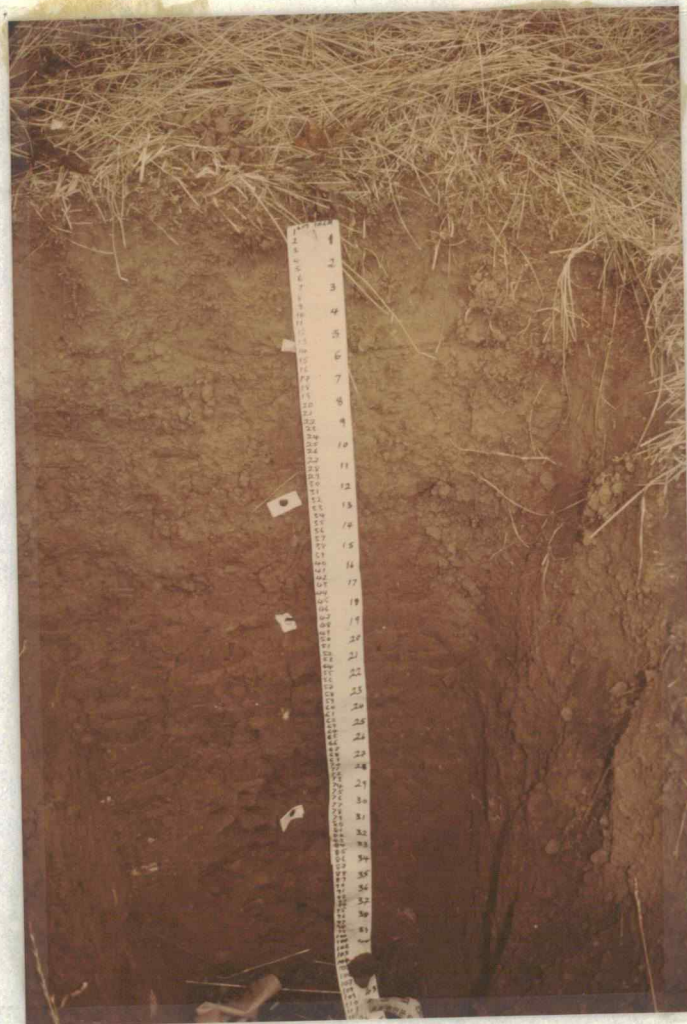


Figure 17. Site 7, profile of Bellpine series under grass cover.

the E4:E6 ratios of alkali extracts. Campbell et al. (1967), Lowe (1969) and McKeague (1968) share the view that the grassland soils contain more humic acid (darker colored) and less fulvic acid (lighter colored) than forest soils.

Difference in color of soils is also attributable to the loss of, or the form of, the free iron oxide in the soil (White and Riecken, 1955). Not much difference was found in the pedogenetic iron contents in the surface layers of the soil under forest and the soil under grass. However, sites 4 and 6 (Figure 29) showed some difference. The difference in pedogenetic oxide contents is further discussed under a separate heading. It should be noted that A2 (E) horizons are generally absent. Although this study area is in a more northerly latitude than some podzolic soil areas in the eastern United States, the soils of this area have formed in a humid, temperate climate where a prolonged snow cover seldom exists. The high pH and fine texture of parent material are also not favorable for podzolization. However, some characteristics, as will be seen later, indicate the formation of an incipient A2 horizon. Geis et al. (1970) have shown that the initial stages of soil formation under forest vegetation in Illinois produced appreciable chemical changes which occurred prior to the time at which physical and morphological differences become apparent.

Generally, the A1 horizons of the grassland soils show more granular structure than the A1 horizons of soils under forest transition. This can also be attributed to the differences in nature and quantity of organic matter in those horizons. The influence of clay on structure development is not clearly seen as there are high percentages of clay in the A1 horizons of all soils.

Particle Size Distribution

The particle size data for all the profiles investigated are presented in Table 2. The soils are dominated by silt, which usually ranges from 29% to 55%, and clay which ranges from 27% to 67%. Total sand never exceeds 29% and most of it occurs as the fine and very fine sand separate. Clay content is highest in the B horizon of all soils. Statistical data in Table 3 indicate that there is no significant difference in clay content of A1, A, B and A+B horizons between soils under forest transition and soils under grass. However, the distribution of clay-size particles in these soils indicates that they are moderately to strongly developed (Birkeland, 1974) as there are relatively low clay contents in the A and C horizons with maximum amounts in the B horizon. The higher content of clay in the B horizons under grass in sites 2 and 4 (Figure 18) may indicate that probably these sites have been occupied by forest over a long period before giving way to grass as a result of burning. The lower

Table 2. Particle Size Distribution.

Site	Horizon	Depth (cm)	Percent										Textural class
			Very coarse sand (2-1 mm)	Coarse sand (1-.05 mm)	Medium sand (.5-.25 mm)	Fine sand (.25-.1 mm)	Very fine sand (.1-.05 mm)	Total sand (2-.05 mm)	Coarse silt (.05-.02 mm)	Fine silt (.02-.002 mm)	Total silt (.05-.002 mm)	Total clay (<.002 mm)	
1F	A1	0-8	0.8	1.4	1.7	5.0	5.4	14.3	12.8	38.5	51.3	34.4	SiCL
	A12	8-15	0.7	1.2	1.7	4.8	6.7	15.1	11.5	37.0	48.5	36.4	SiCL
	B21	15-30	0.4	0.9	1.6	4.5	6.0	13.4	11.9	36.5	48.4	38.2	SiCL
	B22	30-46	0.5	1.2	1.2	2.2	6.7	11.8	17.6	34.0	51.6	36.6	SiCL
1G	A1	0-8	1.3	1.6	1.4	3.8	5.9	14.0	13.1	38.7	51.8	34.2	SiCL
	A12	8-15	1.2	1.6	1.6	4.0	5.2	13.5	13.1	40.0	53.1	33.4	SiCL
	B21	15-25	0.8	1.3	1.3	3.9	5.2	12.4	13.0	36.8	49.8	37.8	SiCL
	B22	25-40	0.8	1.6	1.9	4.9	6.5	15.7	11.3	33.5	44.8	39.5	SiCL
2F	A1	0-13	0.9	1.7	2.4	4.8	4.8	14.6	10.2	36.0	46.2	39.2	SiCL
	A3	13-28	0.9	0.2	2.0	4.3	4.4	11.8	8.4	36.8	45.2	43.0	SiC
	B2t	28-48	1.1	1.5	1.8	4.5	6.1	14.0	5.5	36.0	41.5	43.5	SiC
	B3t	48-61	2.4	1.6	1.1	3.8	7.6	16.5	11.9	31.8	43.7	39.8	SiCL
	C	61-127	2.5	1.5	0.9	3.8	9.6	18.3	5.2	45.0	50.2	31.5	SiCL
2G	A1	0-13	0.6	0.8	1.6	3.8	4.5	11.3	9.8	34.8	44.6	44.1	SiC
	A3	13-25	0.4	0.2	1.4	2.7	4.4	9.1	8.9	35.5	45.7	46.5	SiC
	B2t	25-43	0.1	4.7	1.1	3.0	3.2	12.1	6.2	33.0	39.2	48.7	SiC
	B3t	43-66	1.2	0.4	1.1	3.0	3.9	8.5	8.2	35.3	43.5	48.0	SiC
	C	66-127	0.1	0.4	1.0	3.8	6.7	12.0	12.3	38.4	50.7	37.3	SiCL
3F	A1	0-15	1.2	3.6	3.3	6.6	5.7	20.4	9.4	39.6	49.1	30.5	SiCL
	B21	15-43	0.7	2.4	2.6	5.4	4.9	16.0	13.5	33.8	47.3	36.7	SiCL
	B22	43-71	0.9	2.9	2.9	6.0	3.8	16.5	10.4	40.9	51.3	32.2	SiCL
	B3	71-102	1.1	3.1	3.8	9.5	8.3	25.8	11.3	35.8	47.1	27.1	SiL
3G	A1	0-15	1.8	3.1	2.8	4.6	6.0	18.3	11.8	40.0	51.8	29.9	SiCL
	B21	15-46	1.1	3.0	2.7	4.7	3.5	15.0	18.7	30.9	49.6	35.4	SiCL
	B22	46-66	1.1	3.2	3.0	5.3	4.3	16.9	11.2	35.9	47.1	36.0	SiCL
	B3	66-102	1.7	4.2	4.1	7.5	5.2	22.7	15.5	31.2	46.7	30.6	CL
4F	A1	0-15	0.4	0.7	1.7	3.3	4.7	10.8	11.0	36.9	47.9	41.3	SiC
	A3	15-33	0.3	0.9	1.6	3.2	3.8	9.8	11.2	37.6	48.8	41.4	SiC
	B21t	33-53	0.5	0.8	1.5	2.9	4.5	10.2	11.4	34.2	45.6	44.2	SiC
	B22t	53-76	0.1	0.2	0.4	2.1	6.4	9.2	16.3	29.5	45.8	45.0	SiC
	C	76-127	0.1	0.1	0.7	7.0	20.4	28.3	20.5	21.0	41.5	30.2	CL

(Continued on next page)

Table 2. (Continued)

Site	Horizon	Depth (cm)	Percent										Textural class
			Very coarse sand (2-1 mm)	Coarse sand (1-.5 mm)	Medium sand (.5-.25 mm)	Fine sand (.25-.1 mm)	Very fine sand (.1-.05 mm)	Total sand (2-.05 mm)	Coarse silt (.05-.02 mm)	Fine silt (.02-.002 mm)	Total silt (.05-.002 mm)	Total clay (<.002 mm)	
4G	A1	0-15	0.8	1.6	2.1	4.3	3.6	12.4	8.8	32.8	41.6	46.0	SiC
	A3	15-33	1.3	1.3	1.8	2.7	4.3	11.4	5.7	28.5	34.2	54.4	C
	B21t	33-61	0.9	0.9	1.0	2.2	2.6	7.6	3.0	22.5	25.5	66.9	C
	B22t	61-74	1.8	2.2	1.0	1.2	3.7	9.9	6.7	23.2	29.9	60.2	C
	C	74-127	1.0	2.6	2.2	4.7	4.9	15.4	17.9	28.1	46.0	38.6	SiCL
5F	A1	0-13	3.0	3.4	2.4	5.9	6.1	20.8	9.9	24.9	34.8	44.4	C
	B21	13-25	1.0	2.6	2.4	5.2	6.8	18.0	5.2	28.7	33.9	48.1	C
	B22t	25-38	1.3	2.5	2.2	5.4	5.5	16.9	9.9	29.6	39.5	43.6	SiC
	B23t	38-64	1.2	2.5	2.5	5.5	6.9	17.6	7.7	28.1	35.8	46.6	SiC
	II C	64-127	0.9	2.1	1.9	4.2	5.6	14.7	7.7	35.0	42.7	42.6	SiC
5G	A1	0-18	0.9	1.7	1.8	5.1	7.1	16.6	13.2	29.5	42.7	40.7	SiCL
	B21	18-33	0.2	1.2	1.2	4.0	6.4	13.0	13.2	29.7	42.9	44.1	SiC
	B22t	33-53	0.4	1.2	1.4	4.0	6.4	13.4	8.0	33.1	41.1	45.5	SiC
	B23t	53-74	0.4	1.2	1.6	4.4	6.2	13.8	7.8	22.8	30.6	55.6	C
	II C	74-127	0.4	0.9	0.9	2.2	3.4	7.8	3.8	32.2	36.0	56.2	C
6F	A1	0-15	1.1	1.4	1.3	3.1	4.3	11.2	9.6	32.3	41.9	46.9	SiC
	B21	15-31	0.4	1.3	1.3	3.3	4.7	11.0	8.3	27.0	35.3	53.7	C
	B22t	31-46	0.6	1.1	1.3	3.6	5.6	12.2	7.4	28.0	35.4	52.4	C
	B23t	46-66	0.5	1.1	1.3	3.9	5.8	12.6	9.2	26.0	35.2	52.2	C
	II C	66-127	0.7	1.4	2.4	10.8	1.3	29.2	16.8	21.1	37.9	32.9	CL
6G	A1	0-18	0.7	2.1	1.6	4.3	6.3	15.0	9.3	29.4	38.7	46.3	C
	B21	18-36	0.5	1.4	1.7	4.1	5.9	13.6	9.2	28.5	37.7	48.7	C
	B22t	36-58	0.6	1.2	1.5	3.9	5.9	13.1	8.6	29.3	37.9	49.0	C
	B23t	58-76	0.6	2.6	1.5	4.3	6.1	15.1	5.4	28.5	33.9	51.0	C
	II C	76-127	1.1	2.1	2.5	6.4	10.1	22.2	2.4	37.7	40.1	37.7	CL
7F	A1	0-15	0.8	1.3	1.1	2.1	8.4	13.7	18.2	31.0	49.3	37.0	SiCL
	A3	15-31	0.5	1.2	1.1	2.0	8.5	13.3	16.7	30.6	47.3	39.4	SiCL
	B21t	31-48	0.5	1.0	1.0	1.0	3.1	6.6	20.4	25.5	45.9	47.5	SiC
	B22t	48-69	0.8	1.6	1.3	2.2	8.5	14.4	5.6	33.1	38.7	46.9	SiC
	II C	69-127	1.1	2.9	2.1	3.9	3.3	13.3	26.3	23.8	50.1	36.6	SiCL
7G	A1	0-15	0.3	0.3	0.4	1.1	6.6	8.7	21.2	34.2	55.4	35.9	SiCL
	A3	15-36	0.3	0.7	0.7	0.6	6.4	8.7	20.1	32.7	52.8	38.5	SiCL
	B21t	36-56	0.3	0.4	0.5	0.8	4.2	6.2	13.0	29.0	42.0	51.8	SiC
	B22t	56-79	0.1	0.3	0.3	0.7	3.5	4.9	14.8	27.7	42.5	52.6	SiC
	II C	79-127	0.2	0.4	0.4	0.3	3.2	4.5	15.1	32.8	47.9	47.6	SiC

Table 3 . Paired t-Values of Some Soil Properties.

Soil property	Al Horizon				A Horizon				B Horizon				A + B Horizons			
	Mean for F	Mean for G	Mean difference	Paired t-value	Mean for F	Mean for G	Mean difference	Paired t-value	Mean for F	Mean for G	Mean difference	Paired t-value	Mean for F	Mean for G	Mean difference	Paired t-value
pH	5.64	5.70	0.06	0.56	5.57	5.82	0.25	1.82	5.56	5.92	0.36	3.93	5.57	5.87	0.30	3.78**
% Clay	38.67	41.91	3.24	1.33	39.26	40.67	1.41	0.73	42.93	47.81	4.89	1.69	41.09	44.24	3.15	1.81
CEC	37.97	37.77	0.20	0.06	36.17	34.19	1.98	0.68	33.31	36.73	3.41	1.16	34.74	35.46	0.72	0.34
Ca ²⁺	16.37	15.09	1.29	0.55	14.81	14.56	0.26	0.11	14.17	15.64	1.47	0.93	14.49	15.10	0.61	0.44
Mg ²⁺	7.51	9.70	2.19	1.48	7.56	9.27	1.71	1.34	8.64	11.61	2.97	1.24	8.10	10.44	2.34	1.78
Ca ²⁺ :Mg ²⁺	2.31	1.80	0.50	2.36	2.10	1.82	0.28	0.99	1.40	1.57	0.17	0.60	1.75	1.70	0.05	0.26
K ⁺	0.95	0.65	0.30	2.87*	0.77	0.51	0.26	2.53*	0.26	0.23	0.02	0.62	0.51	0.37	0.14	2.16*
Na ⁺	0.20	0.14	0.05	2.18	0.20	0.15	0.05	1.87	0.22	0.16	0.05	2.15	0.21	0.15	0.05	2.94*
BS	60.12	63.25	3.14	0.77	59.73	62.64	2.90	0.59	59.85	64.31	4.45	1.00	59.79	63.47	3.68	1.15
Fe ₂ O ₃	8.32	8.66	0.34	0.79	8.46	8.81	0.35	0.93	8.71	8.39	0.33	1.30	8.59	8.60	0.01	0.04
Org. C	4.17	3.90	0.27	0.82	3.67	3.39	0.28	1.34	1.46	1.41	0.05	0.47	2.56	2.40	0.16	1.41
Total N	0.21	0.23	0.02	3.23	0.19	0.22	0.03	7.94**	0.10	0.09	0.01	0.83	0.14	0.15	0.01	1.67
C:N	20.02	16.79	3.23	2.91*	19.50	15.86	3.65	3.86**	16.33	14.13	2.20	3.59*	17.92	15.00	2.92	5.07**

F = Forest transition soils; G = Grassland soils

* Significant at 5 percent level.

** Significant at 1 percent level.

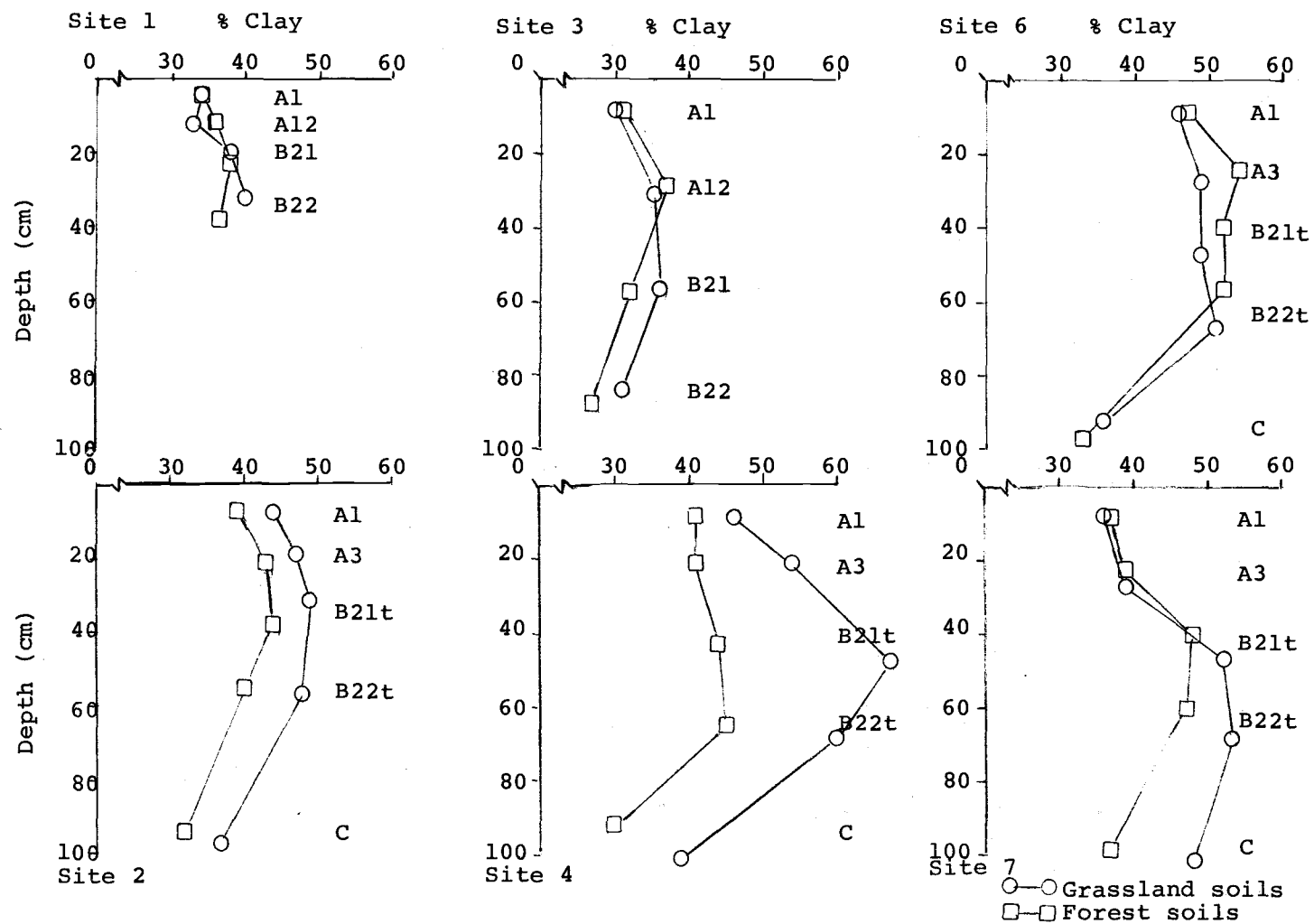


Figure 18

Depth Functions of Clay Fraction(%)

percentage of clay in the soils under forest in these two sites might be due to a shorter duration of the present forest cover or other landscape variables in weathering between these sites. The distribution of clay-size fraction has also influenced the structure of B horizons of those two profiles under grass which have coarse columnar to blocky structures (Figures 7 and 11). It is also possible that the higher clay content may have influenced occurrence of grass (favored grass over forest).

Chemical Properties

pH

The soils in the study sites are neutral (6.7) to very strongly acid (4.8) (Table 4). In the soils formed over the basaltic rock (sites 1 to 4), the lowest pH values were encountered in the surface horizons and the highest pH's were associated with the parent material whereas in the soils formed over sedimentary rock (sites 5 to 7) the lowest pH values were in the subsoil and the highest on surface horizons.

This suggests that the sedimentary rock is acidic but may reflect deeper leaching associated with an older soil landscape. There are significant differences between the pH values obtained from comparable horizons of forest transition and grassland soils (Table 3). This indicates that under these conditions, the difference in vegetation

types has affected the soil reaction. This is in consonance with the findings of Jenny (1941) and Smith et al. (1950) who found lower pH under forest than under prairie. On the contrary, Lotspech et al. (1961) and Ugolini and Schlichte (1973) found no marked difference between soils formed under forest and under grass vegetation. The marked difference in soil pH could be attributed to the differences in types of organic matter in these soils. The forest soils under study, except those in sites 1 and 7, were formed under forest covers where Pseudotsuga menziesii is dominant. It has been mentioned earlier in the literature review that the conifers tend to intensify soil acidity due to their acid producing litter (Ovington 1953, Brady 1974). Despite the fact that sites 1 and 7 under forest are dominated by oak, the presence of Pseudotsuga menziesii litter may have still increased the acidity of the soil. Site 6 did not show much difference in pH (Figure 19). This might be due to the young age (30 years) of the stand (Appendix I, site 6).

Most A horizons of the forest soils (Figure 19) show some decrease in pH (sites 1, 2, 4 and 7). This might be an incipient sign of podzolization or reflect recycling of bases.

Cation Exchange Capacity and Bases

The overall exchange capacity of grassland soil is not significantly higher than that under forest transition (Table 3). The

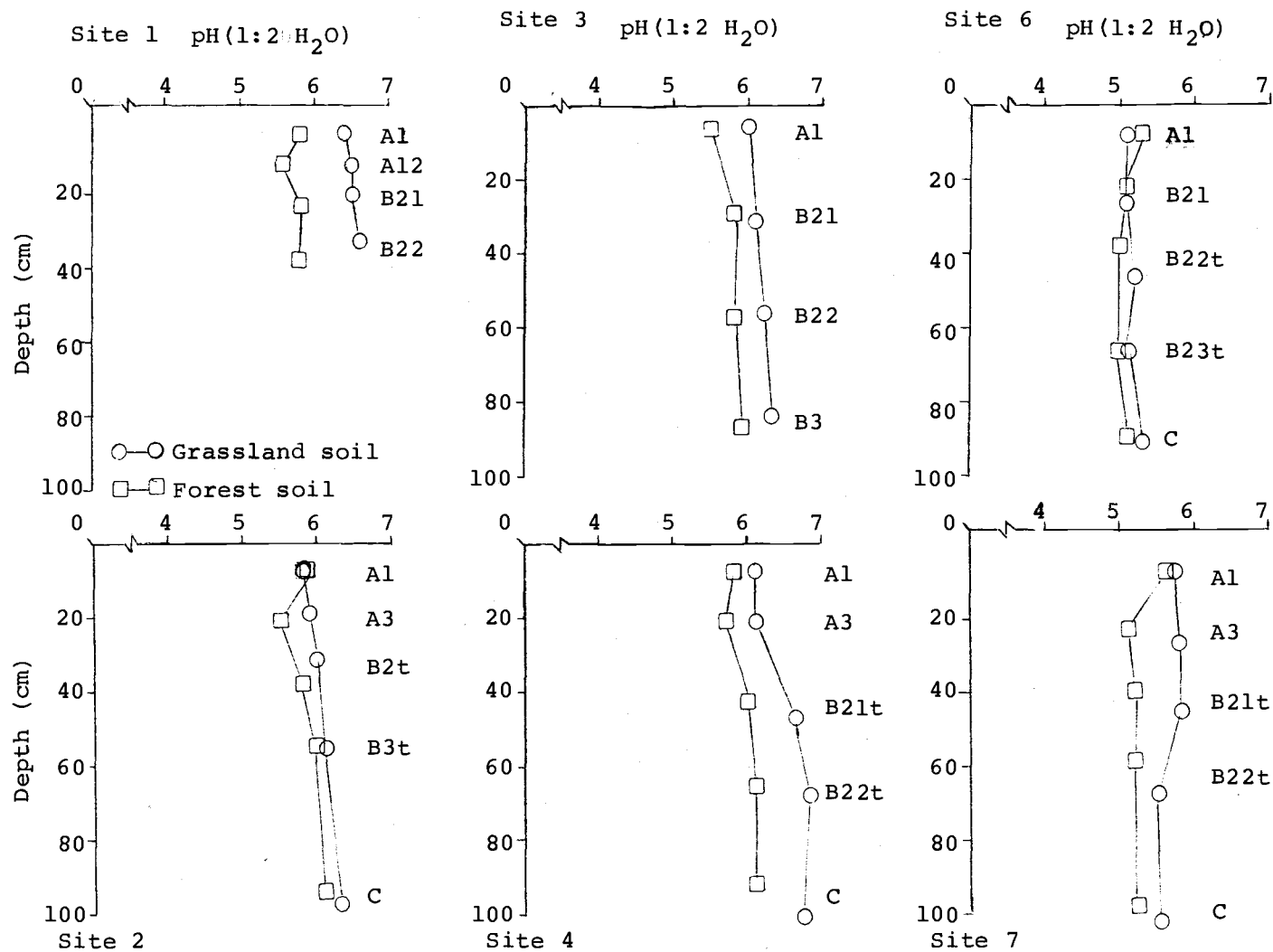


Figure 19

Depth Functions of pH(1:2 H₂O) for Selected Soils

mean for grassland is 35.4 meq/100 g whereas that of forest transition is 34.7 meq/100 g. Figure 20 shows contrast between individual sites. Sites 1 and 2 have higher CEC in the grassland soils while sites 3 and 4 have higher CEC in the forest soils. These variations are difficult to explain because they follow neither the trends in the organic carbon nor clay distribution as reported by William (1932), Helling et al. (1956), Hallsworth and Wilkinson (1958), and Mohamed and Coham (1961).

The overall percentage base saturation tends to be higher in grassland than in forests, but statistically, the difference is not marked (Table 3). The difference could be a function of greater leaching in the forests along with greater annual biomass production and cycling of cations in grassland as suggested by White and Riecken (1955). The greater leaching in the forest relative to the grasslands could be due, in part, to lower evapotranspiration rates and the presence of more chelating agents and more acid leaching waters under a forest canopy. The base saturation in the subsoils of sites 2 and 4 (Table 4) measure above 100%. This cannot be attributed to the pH dependent charges because, first, the soils in these sites are high in montmorillonitic type of clay; and second, the percent organic carbon is low in the subsoils of these sites. One possible reason might be that the subsoil in these sites contains appreciable amounts of zeolite which has high exchangeable bases. Another possibility

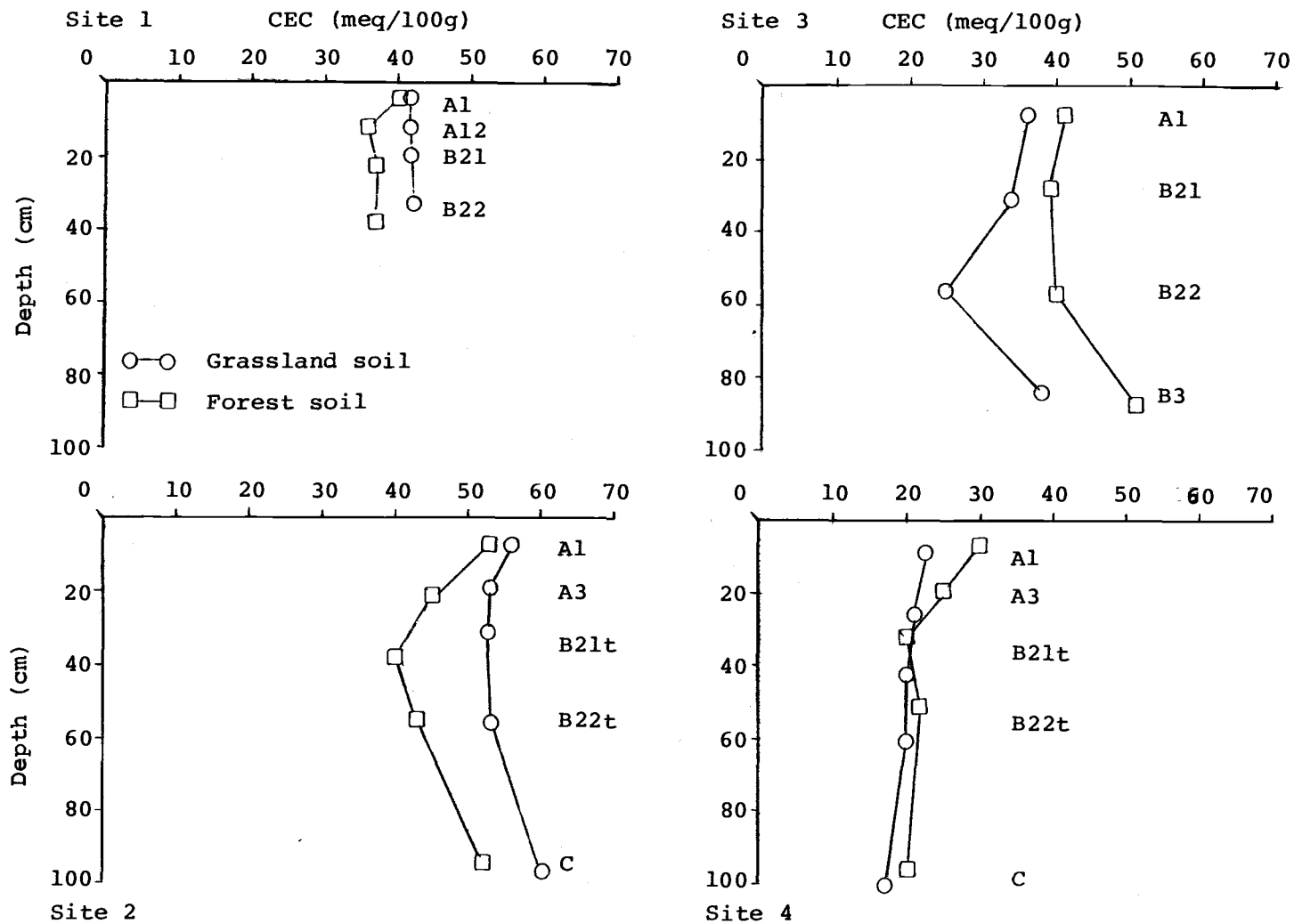


Figure 20 Depth Functions of Exchange Capacity (meq/100g) for Selected Soils

Table 4. Chemical properties of soils under grass and forest/grass transition.

Soil	Horizon	Depth (cm)	H ₂ O (pH)	Total OM (%)	Total OC (%)	Total N (%)	Ca ²⁺ -----	Mg ²⁺ (meq/100 g)	K ⁺ -----	Na ⁺ -----	CEC	BS (%)	Fe ₂ O ₃	C:N	Ca:Mg
1F	A1	0-8	5.8	5.7	3.3	0.19	17.5	7.4	0.81	0.19	39.5	65.6	9.2	17.4	2.4
	A12	8-15	5.6	4.5	2.6	0.15	15.7	7.6	0.32	0.19	35.9	66.4	10.2	17.3	2.1
	B21	15-30	5.8	3.7	2.2	0.14	15.9	9.2	0.23	0.19	36.8	69.3	9.9	15.4	1.7
	B22	30-46	5.9	2.7	1.6	0.10	16.7	8.9	0.18	0.21	37.2	69.9	9.5	15.5	1.9
1G	A1	0-8	6.4	5.6	3.2	0.22	25.0	11.0	0.36	0.17	41.9	87.3	8.8	14.6	2.3
	A12	8-15	6.6	4.2	2.4	0.19	25.0	11.0	0.25	0.15	42.3	86.0	8.8	12.8	2.3
	B21	15-25	6.5	3.2	1.9	0.15	25.0	11.0	0.18	0.17	42.5	85.5	9.1	12.4	2.3
	B22	25-40	6.6	2.7	1.6	0.12	25.0	11.0	0.16	0.19	41.2	88.3	8.4	12.9	2.3
2F	A1	0-13	5.9	8.1	4.7	0.21	30.0	13.0	1.03	0.28	53.4	83.1	7.1	22.4	2.3
	A3	13-28	5.5	5.0	2.9	0.15	29.0	13.0	0.45	0.37	44.9	95.5	7.1	19.2	2.2
	B2t	28-48	5.8	2.9	1.7	0.09	30.0	14.0	0.26	0.33	39.5	112.9	6.4	18.6	2.1
	B3t	48-61	6.0	1.6	0.9	0.05	34.0	15.0	0.19	0.39	42.9	115.6	4.8	18.4	2.3
	C	61-127	6.1	1.3	0.7	0.04	30.0	15.0	0.19	0.42	51.7	88.3	3.9	18.3	2.0
2G	A1	0-13	5.8	7.4	4.3	0.22	29.0	15.0	0.75	0.17	56.3	79.8	7.7	19.5	1.9
	A3	13-25	5.8	5.4	3.1	0.21	29.0	15.0	0.43	0.17	52.8	84.5	7.7	14.9	1.9
	B2t	25-43	6.0	3.3	1.9	0.11	32.0	17.0	0.22	0.21	53.0	93.4	7.2	17.2	1.9
	B3t	43-66	6.1	2.2	1.3	0.07	35.0	19.0	0.18	0.26	53.3	102.1	5.7	18.0	1.8
	C	66-127	6.3	1.5	0.9	0.05	40.0	19.0	0.16	0.39	60.4	98.5	4.3	17.0	2.1
3F	A1	0-15	5.5	5.1	3.0	0.15	19.0	13.0	0.40	0.30	41.2	79.4	11.3	19.7	1.5
	B21	15-43	5.8	3.3	1.9	0.12	13.7	13.0	0.14	0.26	38.5	70.3	11.9	15.8	1.1
	B22	43-71	5.8	2.2	1.3	0.08	14.1	15.0	0.11	0.33	40.4	73.1	11.4	15.8	0.9
	B3	71-102	5.9	1.3	0.7	0.04	24.0	21.0	0.11	0.50	51.4	88.8	9.3	18.3	1.1
3G	A1	0-15	6.0	4.0	2.3	0.18	14.1	13.0	0.19	0.13	36.1	75.9	11.9	12.9	1.1
	B21	15-46	6.1	2.2	1.3	0.10	12.4	13.0	0.11	0.13	33.7	76.2	12.1	12.6	1.0
	B22	46-66	6.2	1.6	1.0	0.08	11.2	13.0	0.09	0.13	25.4	96.3	10.8	11.9	0.9
	B3	66-102	6.3	1.1	0.7	0.06	13.7	16.0	0.09	0.26	37.7	79.7	10.8	11.0	0.9

(Continued on next page)

Table 4. (Continued)

Soil	Horizon	Depth (cm)	H ₂ O (pH)	Total OM (%)	Total OC (%)	Total N (%)	Ca ²⁺ -----	Mg ²⁺ (meq/100 g)	K ⁺ -----	Na ⁺ -----	CEC	BS (%)	Fe ₂ O ₃	C:N	Ca:Mg
4F	A1	0-15	5.8	6.2	3.6	0.18	24.0	9.2	1.64	0.15	40.4	86.6	5.4	20.0	2.1
	A3	15-33	5.7	3.4	2.0	0.11	18.1	9.9	1.00	0.17	37.1	78.7	5.3	17.9	1.6
	B21t	33-53	6.0	1.5	0.9	0.07	29.0	13.0	0.37	0.26	43.0	99.2	5.0	12.4	0.8
	B22t	53-76	6.1	1.1	0.6	0.04	32.0	19.0	0.26	0.28	48.0	107.3	3.2	16.0	0.3
	C	76-127	6.1	0.7	0.4	0.02	43.0	20.0	0.19	0.43	62.2	102.3	2.2	20.5	0.4
4G	A1	0-15	6.1	7.3	4.3	0.24	15.9	17.0	1.22	0.15	41.3	83.0	7.7	17.8	1.8
	A3	15-33	6.1	4.2	2.4	0.11	15.7	19.0	0.51	0.15	45.8	77.3	7.2	21.9	1.7
	B21t	33-61	6.6	1.8	1.0	0.08	25.0	32.0	0.26	0.23	55.2	104.1	5.0	13.0	1.4
	B22t	61-74	6.8	1.1	0.7	0.04	29.0	34.0	0.23	0.28	58.9	107.8	3.4	16.5	1.4
	C	74-127	6.7	0.7	0.4	0.03	32.0	36.0	0.15	0.23	60.7	112.8	3.0	13.7	1.3
5F	A1	0-13	5.7	8.8	5.1	0.23	10.1	4.9	0.94	0.17	30.1	53.6	9.3	22.1	2.1
	B21	13-25	5.5	6.1	3.5	0.19	6.8	4.3	0.42	0.15	24.8	47.0	10.5	18.5	1.6
	B22t	25-38	4.9	3.3	1.9	0.12	1.5	2.0	0.12	0.15	20.0	18.9	10.5	16.1	0.8
	B23t	38-64	4.8	2.1	1.2	0.09	0.3	1.1	0.08	0.13	22.1	7.3	10.9	13.4	0.3
	II C	64-127	4.8	1.7	1.0	0.07	0.3	0.8	0.06	0.19	20.9	6.3	9.5	14.4	0.4
5G	A1	0-18	5.6	6.7	3.9	0.24	6.4	3.5	0.18	0.15	23.0	44.6	9.5	16.3	1.8
	B21	18-33	5.3	3.1	1.8	0.13	3.7	2.2	0.06	0.10	20.6	29.4	9.2	13.9	1.7
	B22t	33-53	5.3	1.8	1.0	0.09	2.7	1.9	0.05	0.10	20.2	23.5	10.6	11.6	1.4
	B23t	53-74	5.2	1.5	0.9	0.06	2.5	1.8	0.04	0.09	20.0	22.1	9.8	14.2	1.4
	II C	74-127	5.0	1.0	0.6	0.05	2.4	1.8	0.04	0.09	16.5	26.3	9.8	11.4	1.3
6F	A1	0-15	5.3	6.8	3.9	0.22	4.5	1.9	0.21	0.13	25.0	26.9	10.3	17.8	2.4
	B21	15-31	5.1	2.9	1.7	0.23	1.5	1.0	0.09	0.15	22.2	12.3	11.3	7.4	1.5
	B22t	31-46	5.0	2.3	1.4	0.12	0.6	0.7	0.08	0.13	20.9	7.2	12.0	11.3	0.9
	B23t	46-66	5.0	2.0	1.1	0.07	0.4	0.6	0.08	0.15	29.3	5.5	11.7	16.3	0.6
	II C	66-127	5.1	0.4	0.2	0.07	0.6	0.6	0.07	0.15	35.7	4.0	12.6	2.9	0.6

(Continued on next page)

Table 4 . (Continued)

Soil	Horizon	Depth (cm)	H ₂ O (pH)	Total OM (%)	Total OC (%)	Total N (%)	Ca ²⁺	Mg ²⁺	K ⁺	Na ⁺	CEC	BS	Fe ₂ O ₃	C:N	Ca:Mg
							----- (meq/100 g) -----								
6G	A1	0-18	5.1	6.7	3.9	0.25	2.1	1.3	0.29	0.13	24.6	15.5	10.8	15.5	1.6
	B21	18-36	5.1	4.3	2.5	0.15	1.3	0.8	0.11	0.10	23.4	9.7	10.7	16.5	1.7
	B22t	36-58	5.2	3.1	1.8	0.12	1.0	0.6	0.08	0.10	24.4	7.3	11.2	15.0	1.7
	B23t	58-76	5.1	1.3	0.7	0.06	0.7	0.5	0.05	0.10	27.9	5.0	10.4	12.3	1.3
	II C	76-127	5.2	0.8	0.5	0.02	0.7	0.5	0.05	0.10	23.6	5.8	11.9	24.0	1.3
7F	A1	0-15	5.6	9.6	5.6	0.27	9.5	3.2	1.61	0.15	36.3	37.1	5.5	20.7	3.0
	A3	15-31	5.1	4.9	2.8	0.15	3.2	2.8	0.90	0.10	26.5	26.4	6.7	18.9	1.1
	B21t	31-48	5.2	2.7	1.6	0.09	2.4	3.6	0.64	0.10	24.0	28.1	7.6	17.2	0.7
	B22t	48-69	5.2	1.2	0.7	0.05	3.0	2.5	0.62	0.10	23.2	26.9	7.6	14.0	1.2
	II C	69-127	5.2	0.7	0.4	0.03	0.9	1.9	0.57	0.09	27.3	12.7	8.4	13.0	0.5
7G	A1	0-15	5.7	7.5	4.4	0.27	9.1	3.1	1.15	0.10	26.9	50.0	5.2	16.2	2.9
	A3	15-36	5.8	5.5	3.2	0.20	9.9	3.0	1.10	0.17	24.2	58.6	5.7	16.0	3.3
	B21t	36-56	5.8	2.0	1.2	0.07	9.3	4.0	0.89	0.13	30.2	47.5	7.4	16.7	2.3
	B22t	56-79	5.5	0.8	0.4	0.04	6.1	3.4	0.72	0.10	29.9	34.6	7.1	11.0	1.8
	II C	79-127	5.5	0.6	0.4	0.03	2.5	2.3	0.46	0.09	27.6	19.4	4.8	11.7	1.1

might be experimental error but the CEC and bases determinations on these soils were determined up to four times, each time yielding nearly identical results. The ammonium acetate method (at pH 7) was used to determine the CEC, and this method commonly gives high values ($> 100\%$) of base saturation when the exchangeable bases are high.

Considering the individual bases, there is no marked difference in exchangeable calcium between the soils under forest transition and soils under grass (Table 3). It can be seen that soils formed on the basaltic parent material (sites 1-4) have higher calcium than those formed on sedimentary rock (sites 5 and 6, Figure 21). Generally there are higher percentages of calcium in the subsoils of sites 1 to 4 than on the topsoil, whereas in sites 5 and 6 the reverse is the case. This is attributable to the nature of the parent materials and possibly the age of the landscape. The higher content of calcium in the surface horizons of the soils of sites 5, 6 and 7 is probably due to the recycling of nutrients by the plants. Sites 3, 4, 5 and 6 show higher contents of exchangeable calcium in the A horizons of soils formed under forest transition than in those formed under grass. Jackson and Sherman (1953) have proposed that tree roots penetrate more deeply than grass roots to the less weathered material and are able to return more bases to the soil surface. Tarrant (1949) has proposed that there is a direct exchange of cations between roots and

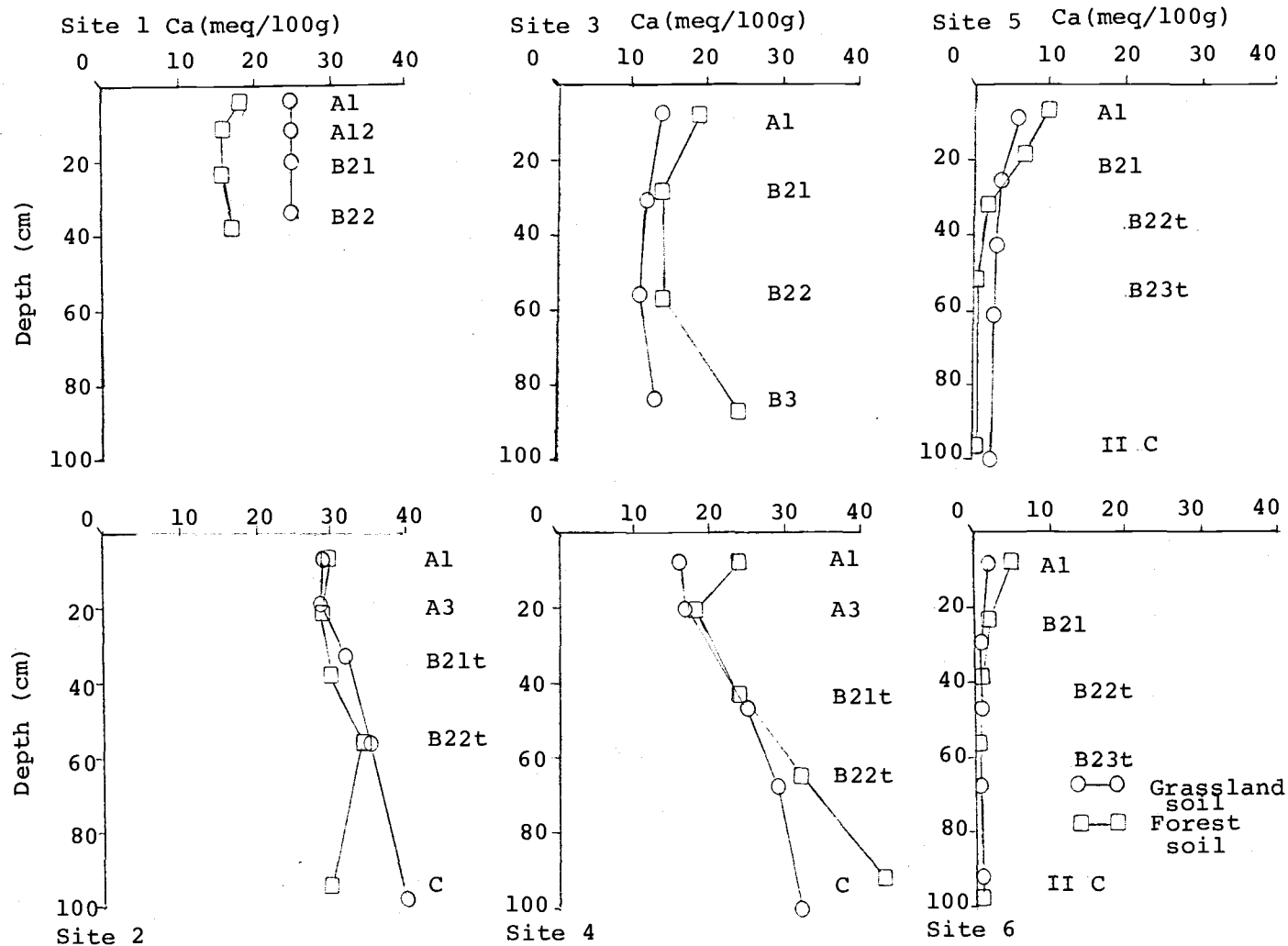


Figure 21 Depth Functions of Exchangeable Calcium (meq/100g) for Selected Soils

the soil so that nutrient elements need not necessarily go into solution to be made available to plants. As mentioned earlier in the literature, mycorrhizal activity also provides a means of nutrient intake for the plant. As far as accumulation of exchangeable calcium is concerned, there is no indication of different tree species effects.

There is also no significantly marked difference in exchangeable magnesium between the grassland soils and forest transition soils (Table 3). Magnesium shows an almost similar pattern as calcium (Figure 22). White and Riecken (1955) found proportionately more calcium than magnesium in the leaf tissue of tree as compared to the aerial parts of the grass. They also showed that the exchangeable calcium-magnesium ratio in the lower layers of the profiles was smallest in the Gray Brown Podzolic soils, intermediate in the transitional soil and greatest in the Brunizem of the biosequence under study. They confirmed that weathering was more rapid under forest vegetation and the calcium-magnesium ratio has been used as a measure of soil weathering (Bray and DeTurk 1930, Jenny 1941). It can be seen in Figure 23 that some sites such as 1, 5, 6 and 7 showed narrower calcium-magnesium ratios in the lower layers of the profiles as indicated by White and Riecken. Generally, there is no marked difference statistically between soils under forest transition and soils under grass in calcium-magnesium ratios. The magnesium content of forest transition soils is about 1/2 (14.49 meq/100 g for Ca

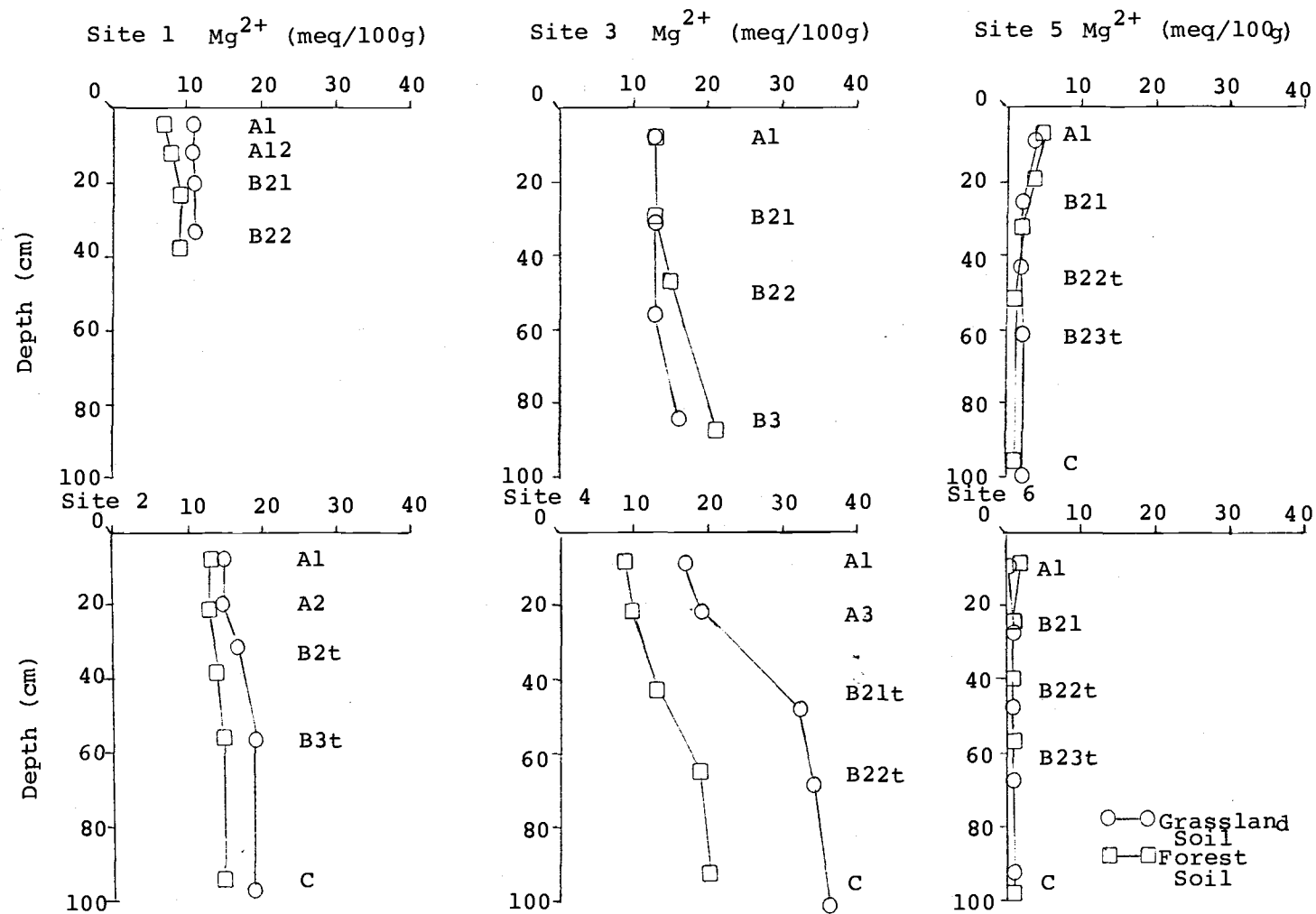


Figure 22 Depth Functions of Exchangeable Magnesium (meq/100g) for Selected Soils

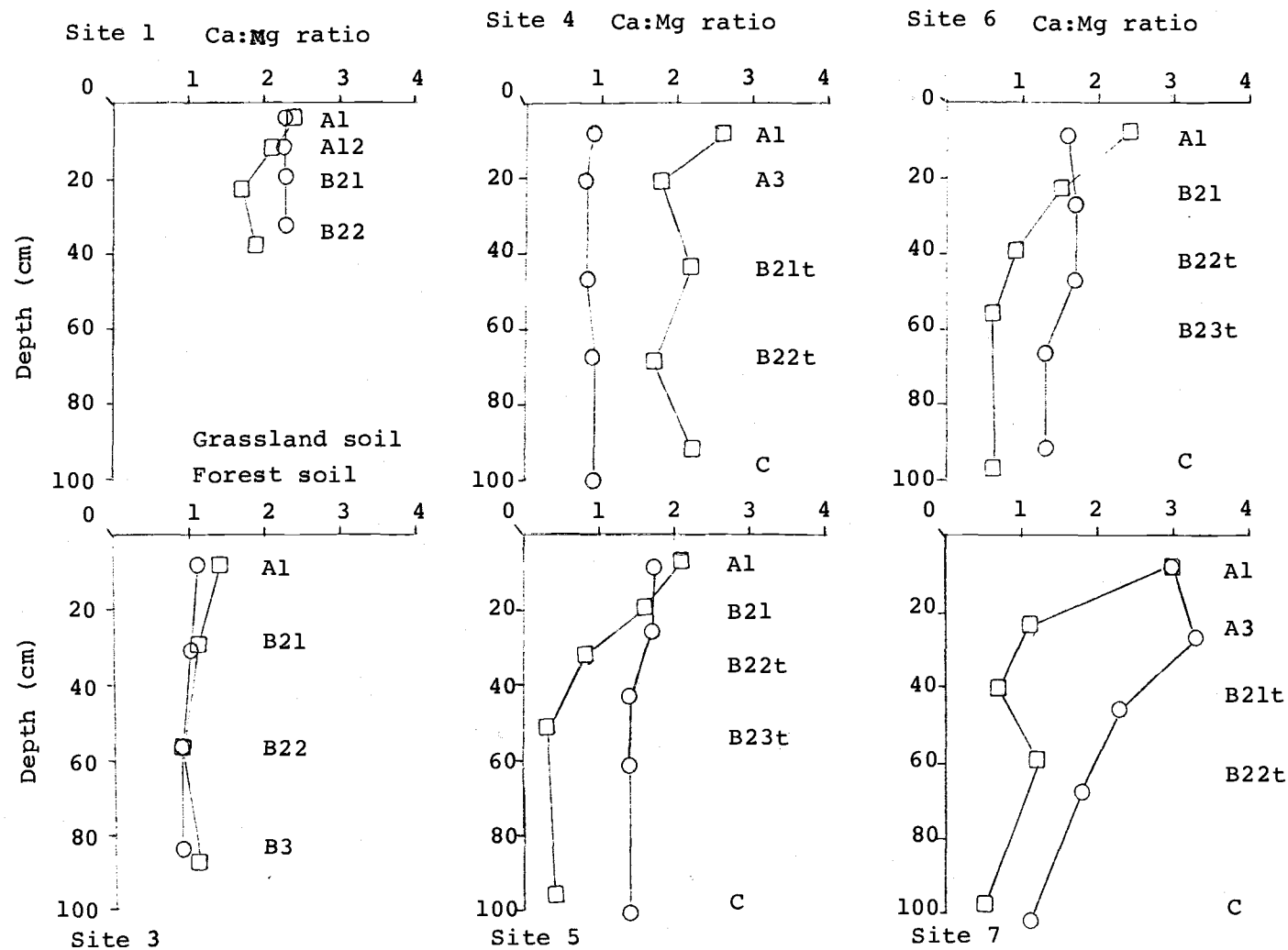


Figure 23 Depth Functions of Ca:Mg ratio for Selected Soils

and 8.10 meq/100 g for Mg) of that of calcium. This is not in consonance with the findings of Wilde (1946) who found magnesium content to be $1/5$ to $1/3$ that of calcium.

The exchangeable potassium contents of the forest transition soils are significantly higher than those under grass (Table 3). This is clearly shown in Figure 24. It can also be seen that the difference is greater in the A1 horizons. This can be attributed to mineral recycling as forest trees are known to require more potassium than the grasses.

Surprisingly, the sodium content of soils under forest cover is significantly higher than that of the soil under grass (Figure 25). Although it never exceeded 0.5 meq/100 g, even this amount is somewhat unusual in a soil under such a high rainfall. In both forest and grass, the sodium content is higher in the subsoil than in the surface soil. The trend with depth is in line with the usual pattern for well-drained soils. The higher quantity of sodium in soils under forest transition is probably a consequence of the proximity of the Pacific Ocean. The study area is only 64 km east of the Pacific Ocean, and the prevailing winds are from the west. During periods of strong winds salt spray may be carried several kilometers inland, to be trapped by vegetation and hence washed down by rain into the soil. The age effects of these forest stands do not reflect much except in

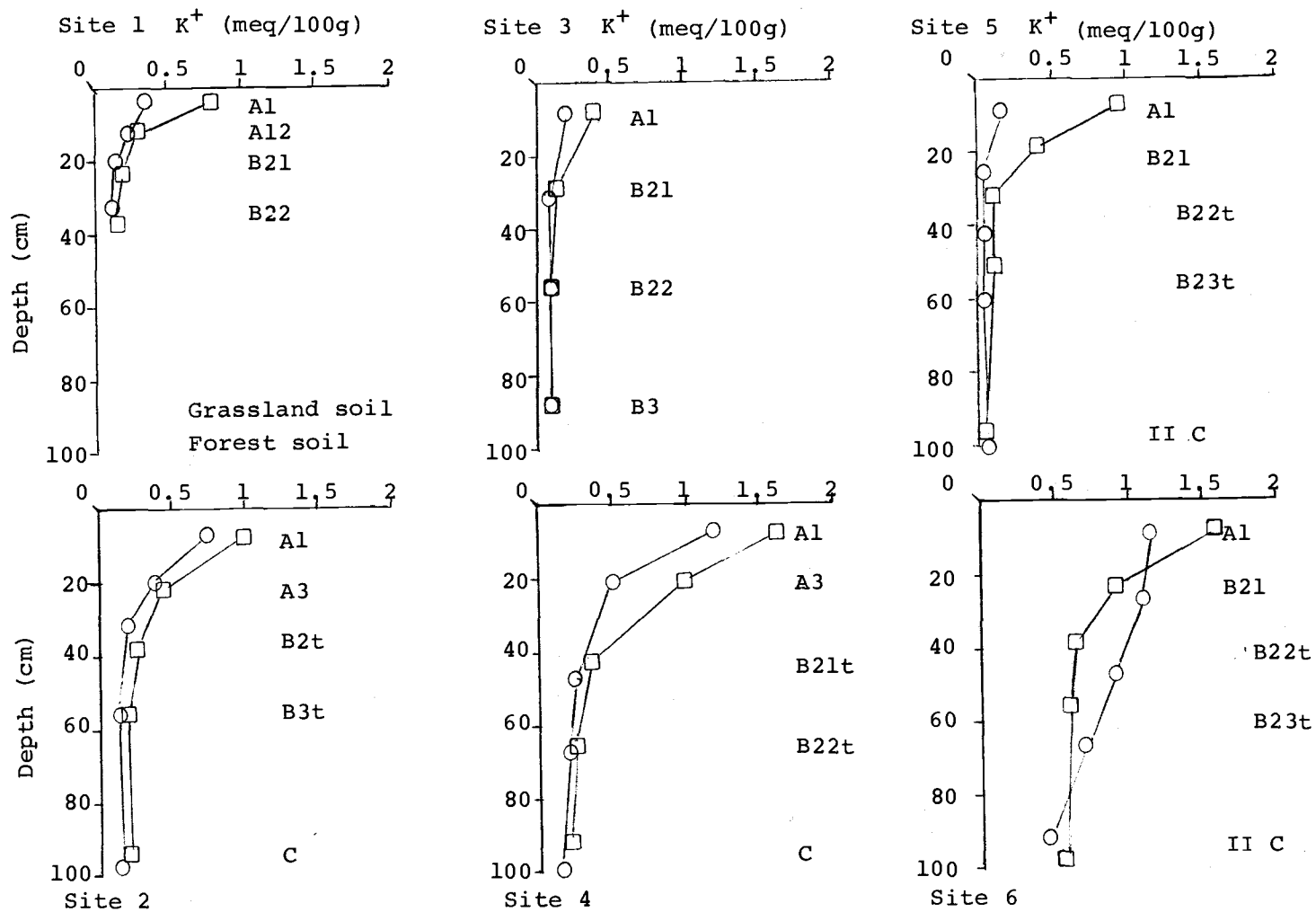


Figure 24

Depth Functions of K^+ for Selected Soils

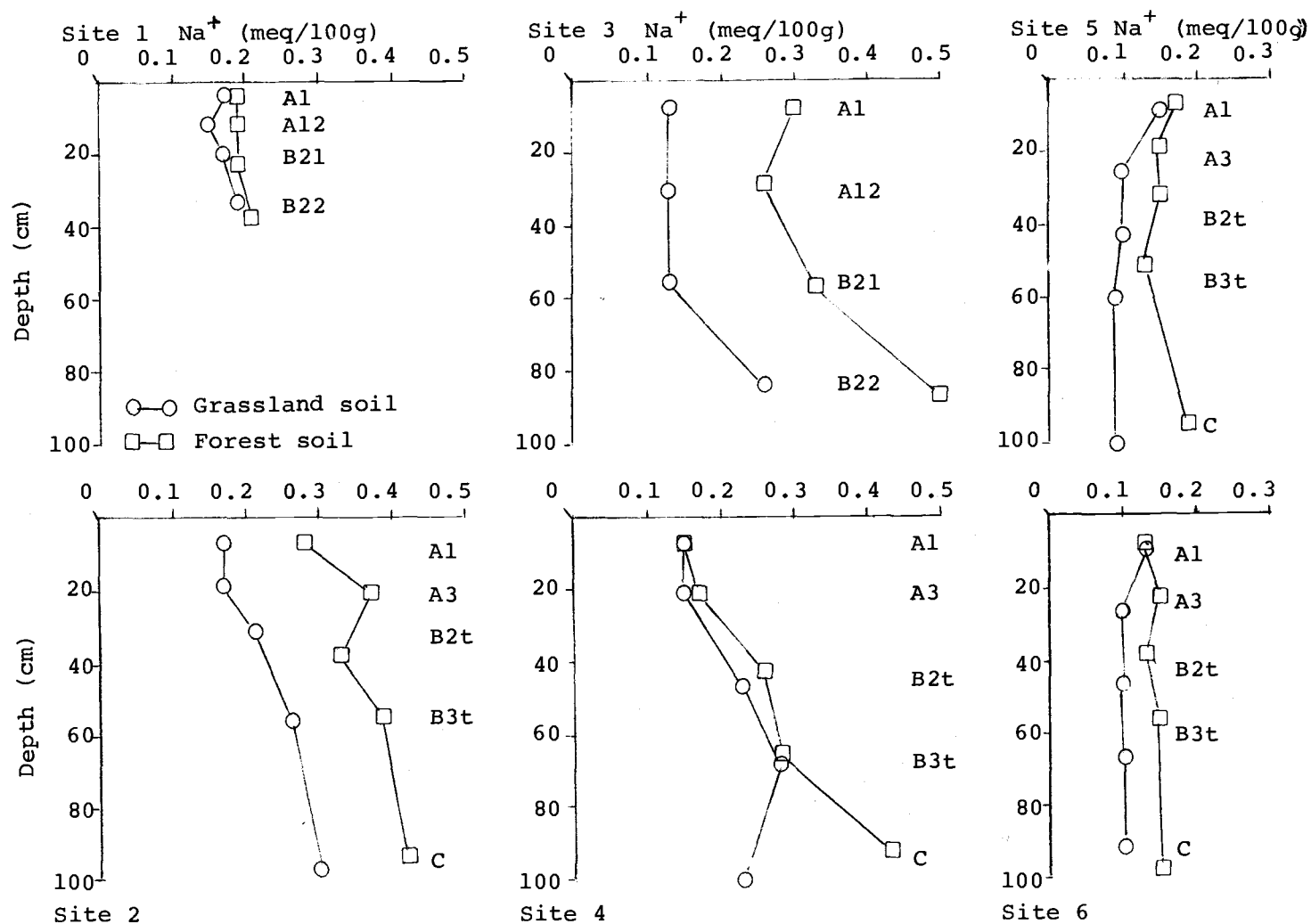


Figure 25 Depth Functions of Exchangeable Sodium (meq/100g) of Selected Soils

site 6 where the difference in sodium content is minimized because the forest stand is only 30 years old.

Organic Matter

The data for the percentage of organic matter and organic carbon are presented in Tables 3 and 4 and also in Figure 26. Statistical analysis reveals that there is no significant difference between the organic matter and hence the organic carbon contents of soils formed under forest transition and soils formed under grass. However, Appendix I shows that the A horizons of the forest transition soils have higher values than those of grassland soils. This is not consistent with the findings of White and Reicken (1955), Jenny (1941), Floate (1965) and others. The lighter color of the A horizons of the forest transition soils suggests that the organic carbon of the soil is more of fulvic acid than humic acid (Thorp 1948, Shields 1965). Research in Iowa reveals that the decrease in the organic matter content of the A horizon of the transitional soil as compared to the associated Brunizem is accompanied by a decrease in the amount of clay in that horizon and an increase in the amount of clay in the B horizon (White and Riecken, 1955). In the study area in western Oregon, this covariance does not occur. This may be attributable to the fact that there is a warmer climate (drier summers and milder winters) in western Oregon than in that part of the

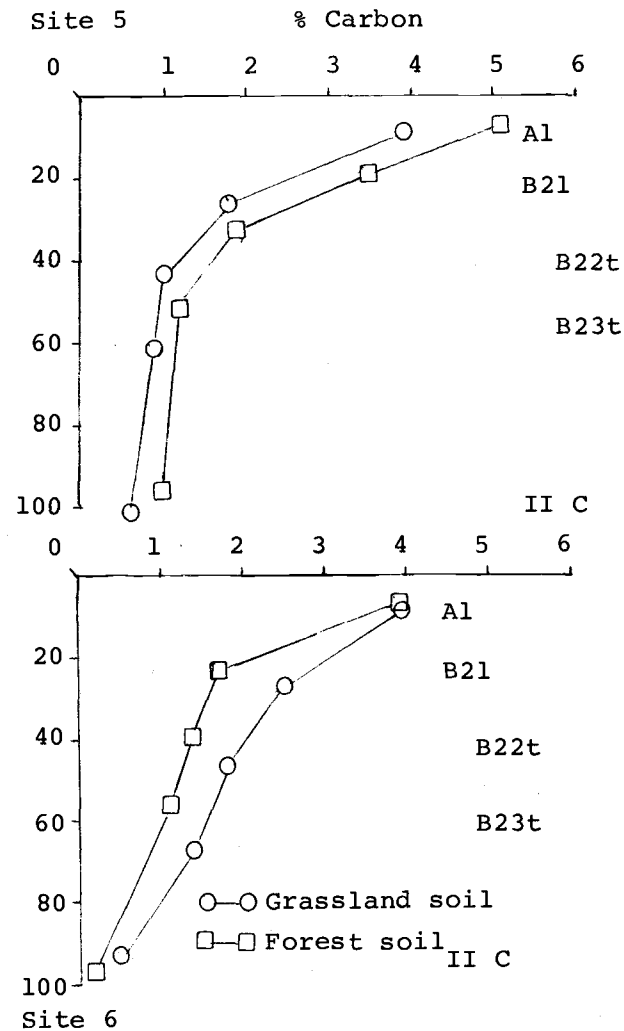
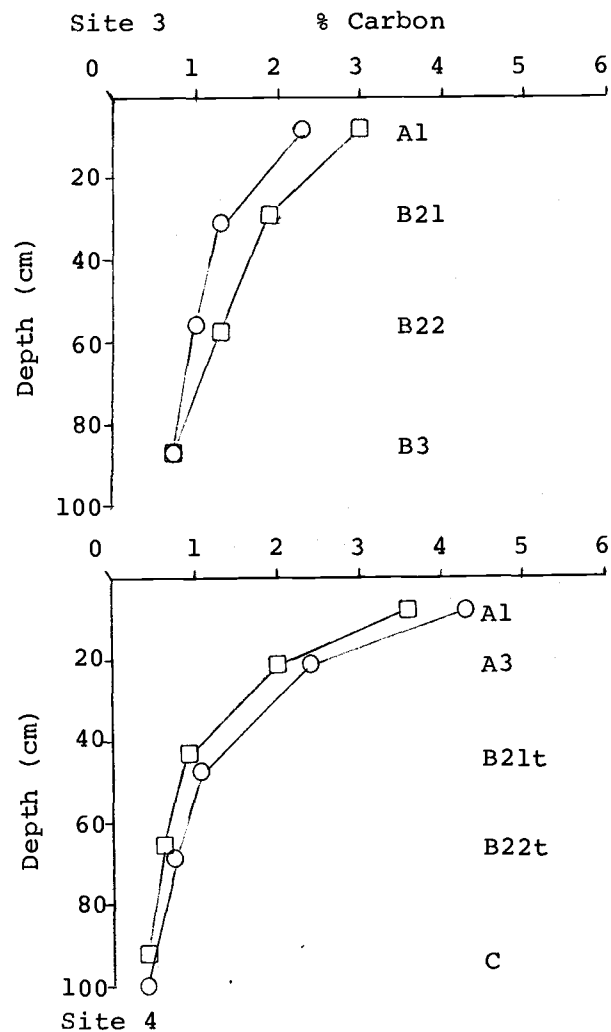


Figure 26 Depth Functions of % Organic Carbon for Selected Soils

mid-continent of America with relatively high organic matter accumulation under forest. However, there are indications of illuviation of clay in the B horizons as shown in Figure 18. Table 4 shows high contents of organic matter in these study sites. There is up to 9.6% organic matter in the A1 horizon of the forest transition soil in site 7. It is said that organic matter is consistently higher in fine-textured soils than in coarse, sandy soils (Tarrant, 1949). The reason for this has been discussed in the literature (Grim 1953, Pinck et al. 1955).

Total Nitrogen

The data for the total nitrogen are presented on Tables 3 and 4. The nitrogen contents of A horizons of grassland soils are significantly higher than those of the forest transition soils. The values for the A horizons average 0.19 for forest transition soils and 0.22 for grassland soils. It can be seen in Figure 27 that the total nitrogen values decreased with depth as did those of organic carbon. It is said that nitrogen in soil, especially that in the epipedon, occurs largely in organic combination, thus, the process of nitrogen accumulation bears a close relationship to the accumulation of organic matter (Schreiner and Brown, 1938). Ugolini and Schlichte have found the average nitrogen percentages of A1 and B2 horizons of grassland soils to be twice those of the comparable forest soils. They attribute

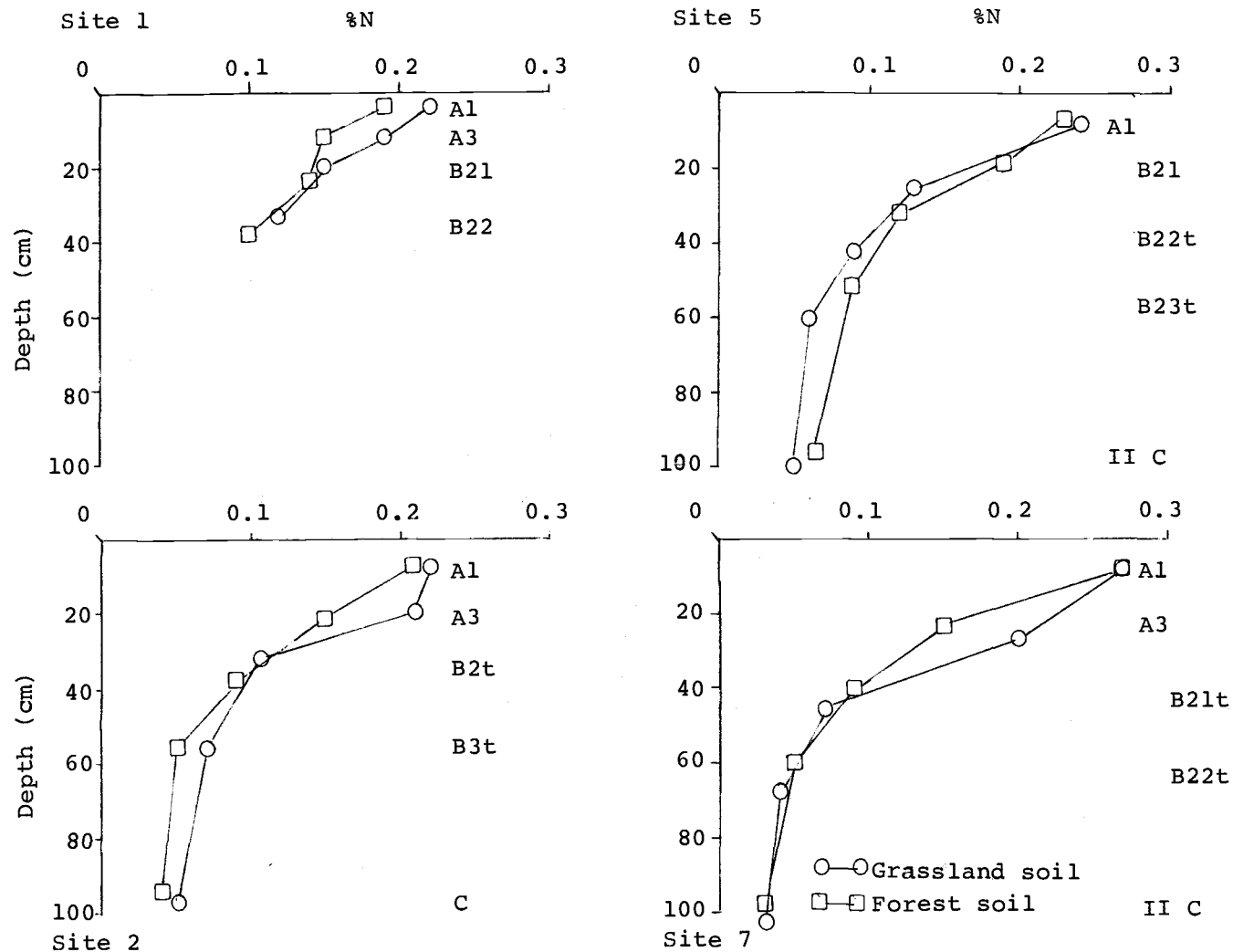


Figure 27 Depth Functions of % Total Nitrogen for Selected Soils

this to the higher content of nitrogen in the organic litter produced by the grassland vegetation. The nitrogen content of grasses, forbs and mosses is generally higher than the more woody litter which accumulates on the forest floor. Other reasons why the grassland soils should have higher percentage of total nitrogen have been given in the literature (Mattson et al. 1943, Harmsen 1951, Stevenson 1965).

The carbon/nitrogen ratios also reflect the difference in the character of the plant litter which is incorporated into the soils. The carbon/nitrogen ratios of forest transition soils are significantly higher than those of the grassland soils (Table 3). Figure 28 shows that the difference is somewhat greater in the epipedon but still extends down into the subsoil. The reason for the difference in the carbon/nitrogen ratios has been given by Mattson et al. (1943). They claimed that the high base status in the soil leads to the absorption of atmospheric oxygen and the fixation of ammonia by lignin, or lignin-nitrogenous constituents of the organic matter. In their opinion, grassland soils, which contain high bases, should have higher nitrogen contents and lower carbon/nitrogen ratios than the forest soils with lower base status.

Free Iron Oxides

Considerable efforts have been made to establish quantitative

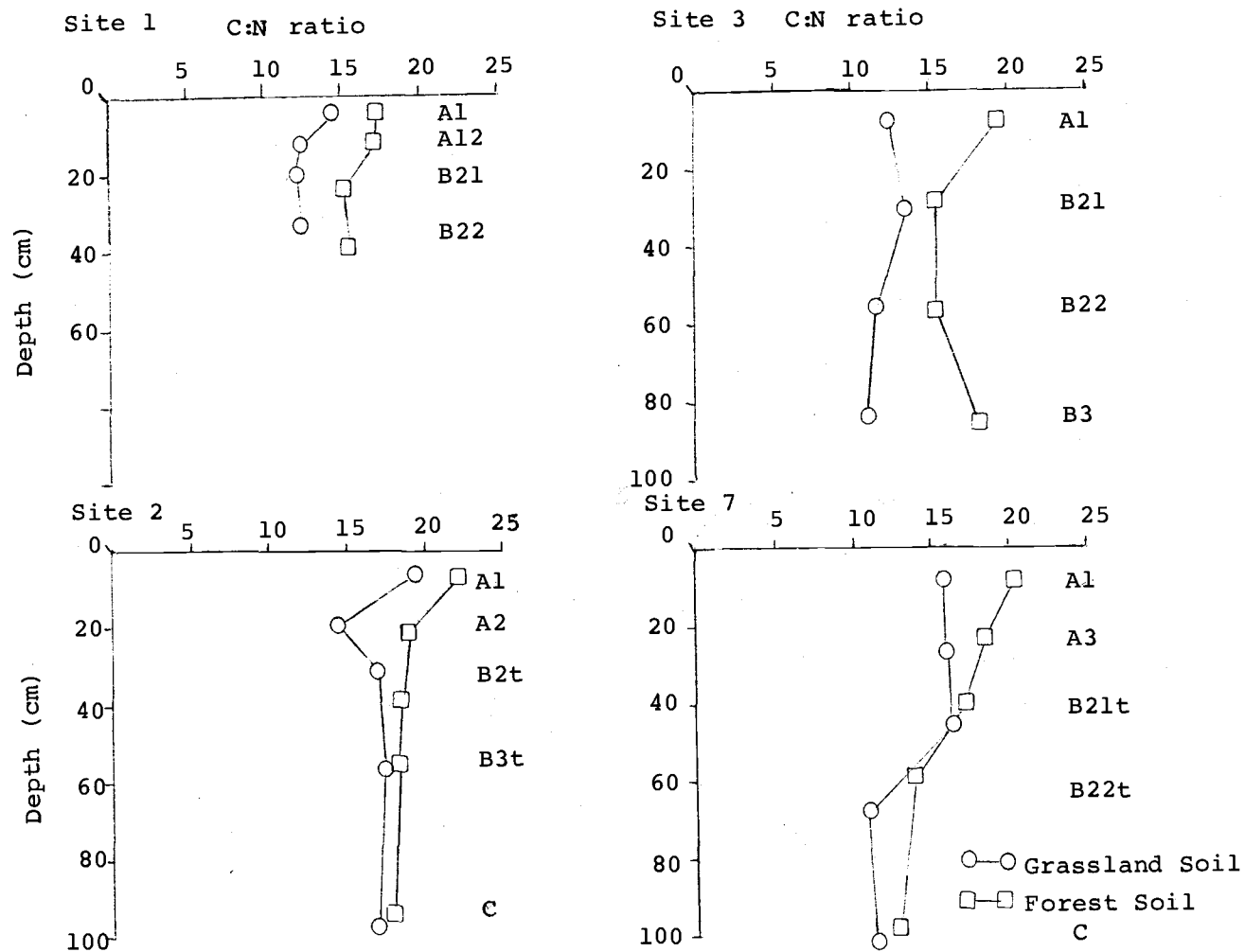


Figure 28 Depth Functions of C:N Ratio for Selected Soils

criteria for separating diagnostic soil horizons (McKeague 1968) and for interpreting pedological processes. A separation of the A1 horizon from humic rich B horizons, the identification of cambic and spodic horizons and the distinctions among Podsol and Brunisolic soils and Humic Podsol have been successfully accomplished by a number of ratios involving pyrophosphate extractable carbon, humic and fulvic acids, pyrophosphate extractable iron and aluminum.

Because the morphology of the soils under study did not indicate presence of iron-humus illuvial horizons only the dithionate extractable iron was measured and these data are presented in Table 4. There is no significant difference between the free iron oxide in soils formed under forest transition and those formed under grass (Table 3). In the A1 horizon, the pooled average percentage free iron oxide of the grassland soil is 8.3 whereas that of forest transition soil is 8.7. In the B horizon they are 8.7 and 8.4 respectively. It can be seen that there is an incipient movement of free iron oxide from A to B in the forest transition. In sites 1 and 3, which have relatively old stands, signs of larger buldges appear at the lower part of A and upper B horizons. It is predicted that in 500 to 1000 years time the differences between these will be marked. The percentage free iron oxide in the C horizons of most soils is very low. This is an indication that vegetation and profile weathering have great influence on the formation of the pedogenic oxides.

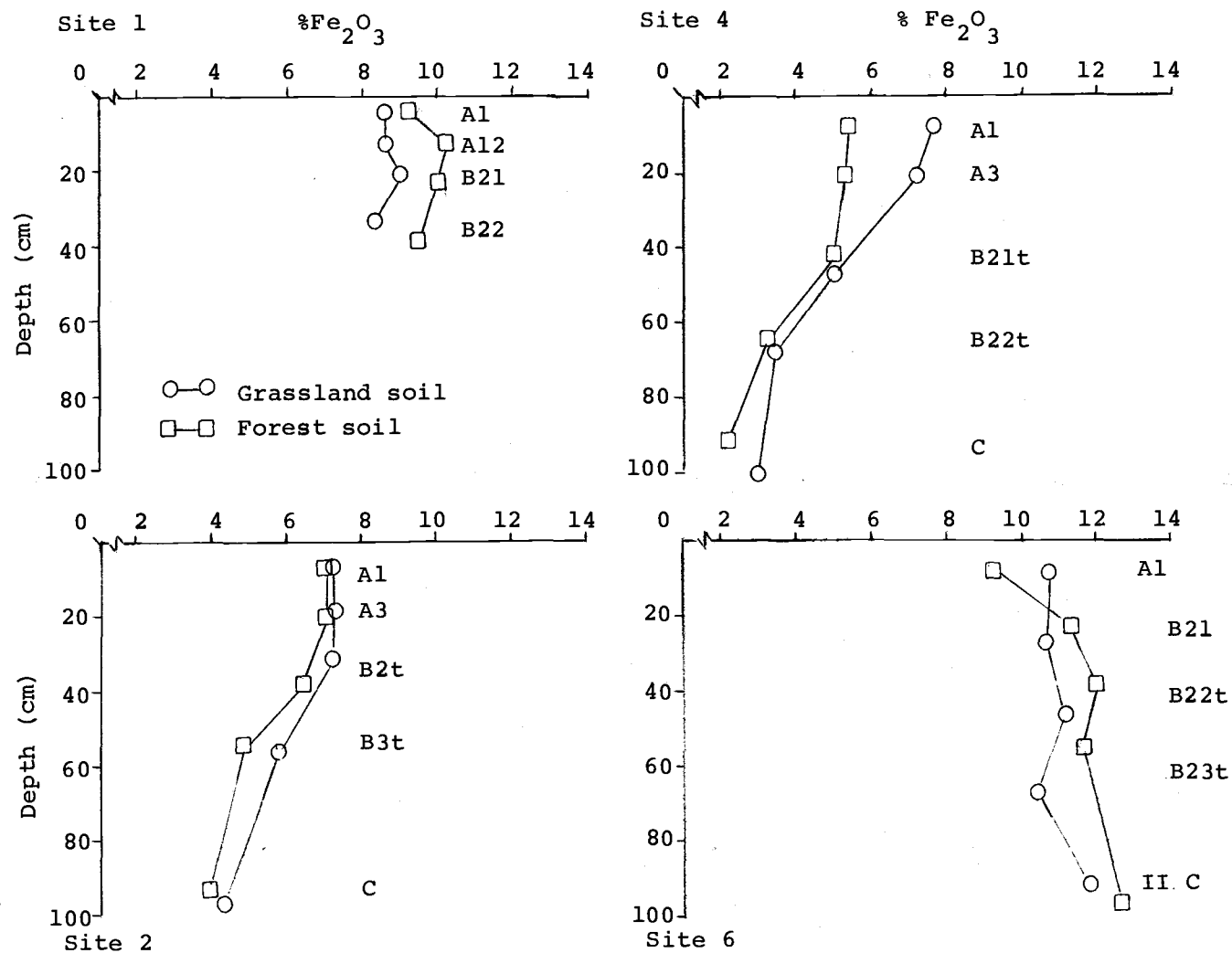


Figure 29 Depth Functions of Free Iron Oxide (%) for Selected Soils

SUMMARY AND CONCLUSIONS

Vegetation as a soil-forming factor was studied under forest transition and grass, while attempting to keep other soil-forming factors constant, in the western part of the Willamette Valley, Oregon. The physiographic units consist chiefly of the interior foothills and slopes formed from an old basaltic flow with some interspersed sedimentary formations.

The soils and vegetation of these physiographic units are developing under a temperate humid climate characterized by dry summers and mild, rainy winters.

The foothills and slopes support stands of Douglas-fir (Pseudotsuga menziesii) and Oregon white oak (Quercus garryana). Scattered through the area are non-forested openings or grasslands.

Many ecologists have looked into the successional relationships of the communities within the Willamette Valley. Before the valley was settled, much of it was occupied by prairies and oak savannas. Dense forests were confined primarily to the foothills and flood-plains. Indians were responsible for most of the fires which created and maintained open conditions. Other evidence available indicates that the open conditions may have become established during a time of warmer and perhaps dried climate (Hypsithermal) that occurred near the middle of the postglacial time.

Fire control activities motivated by the settlers are, however, believed to be responsible for the successional change--replacement of a prairie by Quercus/Pseudotsuga forest.

The establishment of two different vegetational assemblages has had a direct impact on the soils. The soil parameters that are affected by the plant cover are directly or indirectly associated with the organic matter with the exception of exchangeable sodium.

Generally, there are some differences in both color and structure of the soils under these two vegetation types, but they are not adequately quantified to show statistical differences.

The profiles are very strongly acid in the sedimentary rock derived soils, to neutral in the soils derived from basaltic rock. The pH of soils under grass is significantly higher than that of the forest transition soils. This may be related to the types of organic matter in those soils.

There is subsurface accumulation of silicate clays but without co-accumulation of Fe oxides.

The exchangeable potassium contents of the forest transition soils are significantly higher than those of the grassland soils. This is attributed to the higher circulation of potassium by tree species.

The percentage of total nitrogen in the A horizons is lower under forest transition. The difference is also attributed to the

difference in types of organic matter in the soils of the two vegetation types.

The carbon/nitrogen ratios of forest transition soils are significantly higher than those of grassland soils. This is also attributed to the qualities of organic matter in the soils under the vegetation types.

The exchangeable sodium in the forest transition soils is significantly higher than that of the grassland soils. The reason for this might be that the area under study is very close to the ocean and that the trees are able to trap wind-borne salt spray which is eventually washed down into the profile. Research is needed to explore this hypothesis.

The soils of the study area do not possess features of podzolic soils even though they occur in a latitude and under a forest vegetation where podzolic soils are commonly formed. Although there are a few signs of initial podzolization in some of the soils the overall absence is attributed to the climate of this area and the nature of the parent material of the soils under study.

LITERATURE CITED

- Aleksandrova, L.N. 1967. Organo-mineral derivatives of humic acids and methods for studying them. *Pochvovedeniye*, No. 7.
- Alexandrova, L.N., V. Th. Arshavskay, E.M. Dorfman, M.F. Lyuzin and O.V. Yurlova. 1968. Humus acids and their organo-mineral derivatives in soil. *Trans. Intern. Cong. Soil Sci.* p. 143.
- Alfnas'yeva Ye A. 1966. Thick Chernozem under grass and tree coenoses. *Sov. Soil Sci.* 6:615-625.
- Antipov-Karatayev, I.N. and I.G. Tsyurypa. 1961. Forms and conditions of migration of substances in the soil profile. *Pochvovedeniye*, No. 8.
- Ayres, A.S., H.H. Hagihara and G. Stanford. 1965. Significance of extractable aluminum in Hawaiian sugar cane soils. *Soil Sci. Soc. Amer. Proc.* 29:387-392.
- Bailey, L.W., R.T. Odell and W.R. Boggess. 1964. Properties of selected soils developed near the forest-prairies border in east-central Illinois. *Soil Sci. Soc. Amer. Proc.* 28:257-263.
- Baldwin, E.M. 1964. *Geology of Oregon*. 2nd ed. Cooperative Bookstore, University of Oregon, Eugene. 165 p.
- Balster, C.A. and R.B. Parsons. 1968. *Geomorphology and soils, Willamette Valley, Oregon*. Agric. Exp. Sta. Oregon State Univ. in coop. with the Soil Conserv. Serv. U.S.D.A.
- Barshad, I. 1960. The effect of the total chemical composition and crystal structure of soil minerals on the nature of the exchangeable cation in acidified clays and in naturally occurring acid soils. *Proc. and Papers. Seventh Int. Cong. Soil Sci.* 2:435-455.
- Beaulieu, J.D. 1971. Geologic formations of western Oregon (west of longitude 121°30"). *Oreg. State Dept. Geol. & Miner. Ind. Bull.* 70. 72 p.
- Billings, W.D. 1965. *Plants and the ecosystem. Fundamentals of Botany Series*. Macmillan, London. 154 p.

- Birkeland, P.W. 1974. *Pedology, Weathering and Geomorphological Research*. Oxford Univ. Press, New York.
- Brady, N. 1974. *The Nature and Properties of Soils*. Macmillan, New York.
- Bray, R.H. and DeTurk, E.E. 1930. Chemical aspects of some Illinois soils as related to their genesis and morphology. *Second Int. Cong. Soil Sci.* 5:118.
- Bremner, J.M. 1965. Total nitrogen. *In*: C.A. Black (ed.), *Methods of Soil Analysis*. Agron. Monograph 9. Amer. Soc. Agron., Madison, Wisc. Part 2. p. 1149-1178.
- Brewer, R. 1964. *Fabric and Mineral Analysis of Soils*. John Wiley and Sons, New York.
- Buol, S.W., F.D. Hole and R.J. McCracken. 1973. *Soil Genesis and Classification*. The Iowa State Univ Press, Ames.
- Campbell, C.A., E.A. Paul, E.N. Rennie and K.J. McCallum. 1967. Factors affecting the accuracy of the carbon-dating method in soil humus studies. *Soil Sci.* 104:81-85.
- Carroll, D. 1959. Ion exchange in clays and other minerals. *Geol. Soc. Amer. Bull.* 70:749-780.
- Chandler, R.F., Jr. 1941. The amount and mineral nutrient content of freshly fallen leaf litter in the hardwood forest of central New York. *J. Amer. Soc. Agron.* 33:859.
- Chester, G., O.J. Atloe and O.N. Allen. 1957. Soil aggregation in relation to various soil constituents. *Soil Sci. Soc. Amer. Proc.* 21:272-277.
- Chu, T.Y. and D.T. Davidson. 1953. Simplified air-jet dispersion apparatus for mechanical analysis of soils. *Proc. of Highway Research Board* 32:541-547.
- Clark, J.S. 1966. The pH values of soils suspended in dilute salt solutions. *Soil Sci. Soc. Amer. Proc.* 30:11-14.
- Coleman, N.T. and M.E. Harward. 1953. The heats of neutralization of acid clays and cation exchange resins. *J. Amer. Chem. Soc.* 75:6045-6046.

- Coleman, N.T. and G.W. Thomas. 1964. Buffer curves of acid clays as affected by the presence of ferric ion and aluminum. *Soil Sci. Soc. Amer. Proc.* 28:187-190.
- _____. 1967. The basic chemistry of soil acidity. *In: Soil Acidity and Liming*, Robert W. Pearson and Fred Adams (eds.). Madison, Wisc. p. 1-41. (Amer. Soc. Agron., Series No. 12)
- Collins, Duane Francis. 1947. Potential timber values as compared to grazing values of the Oregon State College foothill pasture lands. M.S. thesis. Oregon State Univ., Corvallis. 82 p.
- Couts, J.R.H., M.F. Kandil and J. Tinsley. 1968. Use of radioactive ^{59}Fe for tracing soil particle movement. II. Laboratory studies of labeling and splash displacement. *J. Soil Sci.* 19:325.
- Crocker, R.L. 1952. Soil genesis and the pedogenic factors. *Quart. Rev. Biol.* 27:139-168.
- Deevey, E.S. and R.F. Flint. 1957. The post glacial hypsithermal interval. *Science* 125:182-184.
- Dormaer, J.F. 1967. Infrared spectra of humic acids from soils formed under grass or trees. *Geoderma* 1:37-45.
- Douglas, David. 1914. Journal kept by David Douglas during his travels in North America 1823-1827. London, Wesley and Son. 364 p.
- Duchaufouy, Ph. 1970. Principles of Soil Science. Moscow.
- Dudas, M.J. and S. Pawluk. 1969. Naturally occurring organic clay complexes of orthic black Chernosems. *Geoderma* 3:5.
- Eagle, E.B. and D.R. Yoder. 1926. Application of the Robinson method to the determination of clay. *J. Amer. Soc. Agron.* 18(11):1016-1025.
- Ensminger, L.E. and J.E. Gieseking. 1941. The absorption of proteins by montmorillonitic clays and its effect on base-exchange capacity. *Soil Sci.* 51:125-132.

- Flint, R.F. and F. Brandtner. 1961. Climatic changes since the last interglacial. *Am. J. Sci.* 259:321-328.
- Floate, M.J.S. 1965. Distribution of organic matter and phosphorus fractions in a topographic sequence of soil in southern British Columbia. *Can. J. Soil Sci.* 45:32-36.
- Franklin, J.F. and C.T. Dyrness. 1973. Natural vegetation of Oregon and Washington. U.S.D.A. Forest Serv. Res. Paper PNW-8. 417 p. (Pacific N.W. Forest and Range Exp. Stn., Portland, Oreg.)
- Fraps, G.W. and J.F. Fudge. 1945. Chemical composition of sixty-four species of range position grasses grown on a Victorian clay loam. *J. Amer. Soc. Agron.* 37:251.
- Giesecking, J.E. 1949. The clay minerals in soils. *Adv. Agron.* 1:159-200.
- Gleason, H.A. and A. Cronquist. 1964. The Natural Geography of Plants. Columbia Univ. Press, New York.
- Gorbunov, N.I., G.S. Czyaderich and B.M. Tuniki. 1961. Methods of determining nonsilicate amorphous and crystalline sesquioxides in soil and clays. *Soviet Soil Sci.* 11:1252-1259.
- Grim, R.E. 1953. Clay Mineralogy. McGraw-Hill, New York.
- _____. 1968. Clay Mineralogy. McGraw-Hill, New York. 596 p.
- Habeck, James R. 1961. The original vegetation of the mid-Willamette Valley, Oregon. *Northwest Sci.* 35(2):65-77.
- _____. 1962. Forest succession in Monmouth township, Polk County, Oregon, since 1850. *Proc. Montana Acad. Sci.* 21:7-17.
- Hallsworth, E.C. and G.K. Wilkinson. 1958. The contribution of clay and organic matter to the cation exchange capacity of the soil. *J. Agr. Sci.* 51:1-3.
- Hallsworth, E.G. 1965. The relationship between experimental pedology and soil classification. In: E.G. Hallsworth and

- D.V. Crawford (eds.), *Experimental Pedology. Proc. of the Eleventh Easter School in Agric. Sci., Univ. Nottingham.* Butterworth, London. p. 35-372.
- Harmsen, G.D. 1955. Microbes in the young soils (in Dutch).
In: Directie Wieringermeer (ed.), *Origin and Development of the Wieringermeer Polder.* H. Veenman Enzonem, Wageningen. p. 116-128. (English summary)
- Harmsen, G.W. and D.A. VanSchreren. 1955. Mineralization of organic nitrogen in soil. *Adv. Agron.* 7:299-398.
- Helling, C.S., G. Chester and R.B. Corey. 1964. Contribution of organic matter and clay to soil cation-exchange capacity as affected by the pH of the saturating solution. *Soil Sci. Soc. Amer. Proc.* 28:517-520.
- Hendrick, S.B.J. 1941. Base exchange of the clay mineral montmorillonite for organic cations and its dependence upon adsorption due to Vanderwaal's forces. *J. Phys. Chem.* 45:65-81.
- Hole, F.D. 1961. A classification of pedoturbation and some other processes and factors of soil formation in relation to isotropism and anisotropism. *Soil Sci.* 91:385-477
- Hoover, M.D. and H.A. Lunt. 1952. A key for the classification of forest humus types. *Soil Sci. Soc. Amer. Proc.* 16:368.
- Jackson, M.L. 1956. *Soil chemical analysis (advanced course).* Madison, Univ. of Wisconsin, Dept. of Soils. p. 481.
- _____. 1958. *Soil Chemical Analysis.* Englewood Cliffs, N.J., Prentice-Hall. p. 151-154.
- Jackson, M.S. and G.D. Sherman. 1953. *Chemical weathering of minerals in soils.* Advances in Agron. Academic Press, New York.
- Jenny, Hans. 1930. A study on the influence of climate upon the nitrogen and organic matter content of the soil. *Missouri Agr. Exp. Sta. Res. Bull.* 152:1-6.
- _____. 1941. *Factors of soil formation.* McGraw-Hill, New York. 281 p.

Jenny, Hans. 1958. Role of the plant factor in the pedogenic functions. *Ecology* 39:5-16.

_____. 1961. Reflections on the soil acidity merry-go-round. *Soil Sci. Soc. Amer. Proc.* 25:428-432.

Joffe, J.S. 1949. *Pedology*. Somerset Press, New Jersey. 662 p.

_____. 1955. Green menuring viewed by a pedologist. *Adv. Agron.* 7:141-187.

Johannessen, Carl L., William A. Davenport, Artimus Millet and Steven McWilliams. 1971. The vegetation of the Willamette Valley. *Ann. Assoc. Am. Geogr.* 61(2):286-302.

Kik, M.C. 1943. Nutritive studies of forage plants. *Arkansas Agr. Exp. Sta. Bull.* 434.

Kilmer, V.J. and L.T. Alexander. 1949. Methods of making mechanical analysis of soils. *Soil Sci.* 68:15-24.

Kirkwood, J.E. 1902. The vegetation of northwestern Oregon. *Torrey* 2(9):129-134.

Knezevich, C.A. 1975. Soil survey of Benton County area, Oregon. U.S.D.A. Soil Conservation Serv. Oregon Agr. Exp. Stn. 119 p.

Kononova, M.M. 1961. *Soil Organic Matter*. Pergamon Press, New York. 450 p.

Lotspeich, F.B., J.B. Secor, R. Okazaki and H.W. Smith. 1961. Vegetation as a soil forming factor on the quillayute physiographic unit in western Clallam County, Washington. *Ecology* 42:53-68.

Low, P.F. 1955. The role of aluminum in the titration of bentonite. *Soil Sci. Soc. Amer. Proc.* 19:135-139.

Lowe, L.E. 1969. Distribution and properties of organic fractions in selected Alberta soils. *Can. J. Soil Sci.* 49:129-141.

Lundblad, K. 1933. Studies on podzols and Brown Forest Soil. Part I. *Soil Sci.* 37:137-155.

- Lutz, J.F. 1936. The relation of free iron in the soil to aggregation. *Soil Sci. Soc. Amer. Proc.* 1:43-45.
- Mattson, S. and E. Koutler-Andersson. 1943. The acid-base condition in vegetation, litter and humus. Vol. 1. *Lantbruks-Hogskolan Ann.* II:107-134.
- McComb, A.L. and F.F. Riecken. 1961. Effect of vegetation on soils in the forest-prairie region. *In: Recent Advances in Botany.* Univ. of Toronto Press. p. 1627-1631.
- McKeague, J.A. 1968. Humic-fulvic acid ratio. Al, Fe and C in pyrophosphate extracts as criteria of A and B horizons. *Can. J. Soil Sci.* 48:27-35.
- McKeague, J.A. and R.J. St. Arnaud. 1969. Pedotranslocation eluviation-illuviation in soils during the Quaternary. *Soil Sci.* 107:428-435.
- McLean, E.O. 1952. The effect of humus on cationic interactions in a beidellite clay. *Soil Sci. Soc. Amer. Proc.* 16:134-137.
- McLean, E.O., W.R. Hourigan, H.E. Shoemaker and E.R. Bhumbla. 1964. Aluminum in soils. V. Form of aluminum as a cause of soil acidity and a complication in its measurement. *Soil Sci.* 24:163-175.
- Mehlich, A. 1942. Base unsaturation and pH in relation to soil type. *Soil Sci. Soc. Amer. Proc.* 6:150-156.
- _____. 1943. The significance of percentage base saturation and pH in relation to soil differences. *Soil Sci. Soc. Amer. Proc.* 7:167-174.
- Mel'nikova, M.K. and S.V. Koverys. 1974. Effect of the physiochemical properties of soil on the movement of clay suspensions along the profile. *Soviet Soil Sci.* p. 682-688.
- Metz, L.J. 1952. Weight and nitrogen and calcium content of annual litter fall of forests in the South Carolina piedmont. *Soil Sci. Soc. Amer. Proc.* 16:335-338.
- Mitchell, B.P., V.C. Farmer and W.J. McHardy. 1964. Amorphous inorganic materials in soils. *Advances in Agronomy* 16:327-383.

- Mohammed, M.K. and A.A.I. Gohar. 1960. Contributions of clay, silt and organic matter to the potential fertility of Egyptian soils as measured by their cation exchange capacities. U.A.R. Minist. Agr. Publ. 42 p. (Abstr. in Soils and Fert. 24:7. 1961)
- Moodie, C.D. 1951. The hypoiodite method for studying the nature of soil organic matter fractions of different soils. Soil Sci. 71:51-65.
- Morris, William G. 1934. Forest fires in western Oregon and western Washington. Oreg. Hist. Quart. 35:313-339.
- Muller, P.E. 1887. Studien uber die naturalichien. Berlin, Humnsformen.
- Oades, J.M. 1963. The nature and distribution of iron compounds in soils. Soil and Fertilizers 26:69-80.
- Ovington, J.D. 1953. Studies of the development of woodland conditions under different trees. I. Soils pH. J. Ecol. 41:13-34.
- Owen, Herbert Elmer, Jr. 1953. Certain factors affecting establishment of the Douglas-fir (Pseudotsuga taxifolia[Lamb][Britt]) seedlings. M.S. thesis. Oregon State Univ., Corvallis. p. 71.
- Peech, M.L., T. Alexander, L.A. Dean and J.F. Reed. 1947. Methods of soil analysis for soil fertility investigations. U.S. Dept. Agr. Circ. 757. p. 25.
- Peterson, J.B. 1946. Relation of parent material and environment to the clay minerals in Iowa soils. Soil Sci. 61:465-475.
- Pettapiece, W.W. 1969. The forest grassland transition. In: Pedology and Quarternary Research, S. Pawluk (ed.). Univ. Alberta Printing Dept., Edmonton. p. 103-113.
- Pinck, L.A., R.S. Dyal and F.E. Allison. 1954. Protein-montmorillonite complexes, their preparation and the effects of soil microorganisms on their decomposition. Soil Sci. 78:109-118.

- Pionkee, H.B. and R.B. Corey. 1967. Relations between acidic aluminum and soil pH; clay and organic matter. *Soil Sci. Soc. Amer. Proc.* 31:749-752.
- Polynov, B.B. 1951. Modern ideas of soil formation and development. (Transl. from Russian) *Soil and Fertilizers* 14:95-101.
- Powers, W.L. 1932a. Characteristics of forest soils of the north-west United States. *Soil Sci.* 34:1-10.
- _____. 1932b. Characteristics of organic colloids in peaty soils. *J. Agr. Res.* 44:97.
- Roberts, S., R.V. Vodraska, M.D. Kauffman and E.H. Gardner. 1971. Methods of soil analysis used in the soil testing laboratory at Oregon State University Agricultural Experiment Station. Oregon State Univ., Corvallis.
- Ruhe, R.V. 1969. Soils, paleosols and environment. In: *Pleistocene and Recent Environments of the Central Great Plains*, W. Dort, Jr. and J.K. Jones, Jr. (eds.). Univ. Press of Kansas, Lawrence. 433 p.
- Schellmann, W. 1959. Experimentaelle Untesuchungen ueber die sedimentaere Bildung Goethd und Hematit *Chemie der Erde* 20:104-116.
- Schnitzer, M. 1969. Reactions between fulvic acid, a soil humic compound and inorganic soil constituents. *Soil Sci. Soc. Amer. Proc.* 33:75-81.
- Schofield, R.K. 1955. The measurement of soil pH. *Soil Sci. Soc. Amer. Proc.* 19:164-167.
- Schollenberger, C.J. and R.H. Simon. 1945. Determination of exchange capacity and exchangeable bases in soil--ammonium acetate method. *Soil Sci.* 59:13-24.
- Schreiner, O. and B.E. Brown. 1938. Soil nitrogen. *Soil and men*. U.S. Dept. Agric. textbook. p. 361-376.
- Schuylenborgh, J. van. 1965. The formation of sesquioxides in soils. In: *Experimental Pedology*, E.G. Hallsworth and D.V. Crawford (eds). Butterworth and Co., London. 414 p.

- Schweitmann, U. 1966. Die festen organischen bestandteile des bodens. In: Handbuch der Pflangenern ae-hrung und Duengung, K. Scharrer and H. Linser (eds.). Springer Verlag, Vienna and New York. Vol. II, p. 318-353.
- Schwertmann, U. and M.L. Jackson. 1963. Hydrogen-aluminum clays, a third buffer range appearing in potentiometric titration. *Science* 139:1052-1053.
- Schwertmann, U. 1964. Differenzierung der Eisenoxide des Bodens durch Extraktion mit Ammonium oxalate Loesung. *Zeitschrift fuer Pflanzenernaehrung, Duengung and Bodenkunde* 105:194-202.
- Schwertmann, U., W.R. Fischer and H. Papendorf. 1968. The influence of organic compounds on the formation of iron oxides. *Trans. 9th Cong. Int. Soc. Soil Sci., Adelaide*, 1:645-653.
- Sharander, W.D. 1950. Differences in clay contents of surface soils developed under prairie as compared to forest vegetation in the central United States. *Soil Sci. Soc. Amer. Proc.* 15: 333-337.
- Shields, J.A. 1964. Application of trichromatic analysis and organic matter fractionation to the study of soil color. M.S. thesis. Univ. of Saskatchewan, Canada.
- Shields, J.A., E.A. Paul, R.J. St. Arnoud and W.K. Head. 1968. Spectrophotometric measurement of soil color and its relationship to soil organic matter. *Can. J. Soil Sci.* 48:271-280.
- Simonson, R.W. 1959. Modern concepts of soil genesis--a symposium. Outline of a generalized theory of soil genesis. *Soil Sci. Soc. Amer. Proc.* 23:152-156.
- Smith, G.D., W.H. Alloway and F.F. Riecken. 1950. Prairie soils of the upper Mississippi Valley. *Advances in Agron.* 2:157-205.
- Smith, J.E. 1949. Natural vegetation in the Willamette Valley, Oregon. *Science* 109:41-42.
- Smith, R.M., G. Samuels and C.F. Cernuda. 1951. Organic matter and nitrogen build-ups in some Puerto Rican soil profiles. *Soil Sci.* 72:409-427.

Soils of Noth Island, New Zealand. Soil Bureau Bulletin (n. s.)
5, 1955. Dept. Sci. Ind. Res. New Zealand.

Soil Survey Staff. 1951. Soil Survey Manual. U.S.D.A. Handbook
18, U.S. Govt. Print. Office, Washington, D.C. p. 503.

_____. 1962. Supplement to U.S.D.A. Handbook 18,
Soil Survey Manual (replacing p. 173-188). U.S. Dept. Agr.,
U.S. Govt. Print. Office, Washington, D.C.

_____. 1970. Soil taxonomy of the national cooperative
soil survey. Selected chapters from the unedited text. Soil
Conservation Serv., U.S. Dept. of Agr., Washington, D.C.

Soulides, D.A. and F.E. Clark. 1958. Nitrification in grassland
soils. Soil Sci. Soc. Amer. Proc. 22:308-311.

Sprague, F.L. and H.P. Hansen. 1946. Forest succession in the
McDonald Forest, Willamette Valley, Oregon. Northwest
Science 20:89-96.

Stevenson, F.J. 1965. Origin and distribution of nitrogen in soil.
In: Soil Nitrogen, W.U. Bartholomew and F.E. Clark (eds.).
Amer. Soc. Agr. No. 10. Madison, Wisconsin. p. 17-41.

Tarrant, R.F. 1949. Douglas-fir site quality and soil fertility.
J. Forestry 47:716-720.

Theron, J.J. 1951. The influence of plants on the mineralization of
nitrogen and the maintenance of organic matter in soil. J. Agr.
Sci. 41:289-296.

Theron, J.J. and D.G. Haylett. 1953. The regeneration of soil
humus under a grass ley. Empire. J. Exp. Agr. 21:86-98.

Thilenius, J.F. 1964. Synecology of the white-oak (Quercus garry-
ana Douglas_ woodlands of the Willamette Valley, Oregon.
Ph.D. thesis. Oregon State Univ., Corvallis. p. 151.

Thilenius, John F. 1968. The Quercus garryana forests of the
Willamette Valley, Oregon. Ecology 49:1124-1133.

Thornthwaite, D.W. 1931. The climate of North America according
to a new classification. Geographic Review 21:633-655.

- Thorp, J. 1948. U.S. Dept. Agr. Yearbook "Grass." p. 55-66.
- Tisdale, S.L. and W.L. Nelson. 1975. Soil Fertility and Fertilizers. 3rd ed. Macmillan, New York.
- Tsyplerkov, V.P., I.A. Ferssherlcova and O.G. Rastvorova. 1974. Transformation of organic matter in forest soils. Soviet Soil Sci., Vol. 1, Jan-Feb.
- Turner, R.C., W.E. Nichol and J. Brydon. 1963. A study of lime potential. III. Concerning reactions responsible for the magnitude of the lime potential. Soil Sci. 95:186-191.
- Turner, R.C. and J.S. Clark. 1965. Lime potential and degree of base saturation of soils. Soil Sci. 99:194.
- Ugolini, F.C. and Ak Schlichte. 1973. The effect of Holocene environmental changes on selected western Washington soils. Soil Sci. 116:218-227.
- Veitch, F.P. 1902. The estimation of soil acidity and lime requirements of soils. J. Amer. Chem. Soc. 24:1120-1128.
- Walker, T.W., H.D. Orchiston and A.F.R. Adam. 1954. The nitrogen economy of grass-legume association. J. Brit. Grassland Soc. 9:249-274.
- Walker, T.W. 1956. The accumulation of organic matter in grassland soils. Trans. Intern. Cong. Soil Sci. 6th Congr. p. 409-419.
- Walkley, A. and I.A. Black. 1934. An examination of the Degtareff method for determining soil organic matter, and a proposed modification of the chromic acid titration method. Soil Sci. 37:29-38.
- Wells, E.L. 1941. Climate of Oregon. In: U.S. Department of Agriculture, Climate and Man; the Yearbook of Agric. p. 1075-1086.
- White, E.M. and F.F. Riecken. 1955. Brunizem, Gray-brown podsollic soil biosequences. Soil Sci. Soc. Amer. Proc. 19:504-505.

- Whiteside, E.P. and C.E. Marshall. 1944. Mineralogical and chemical studies of the Putnan silt loam soil. Univ. Mo. Col. Ag. Agr. Exp. Sta. Res. Bull. 386.
- Wilde, S.A. 1946. Forest soils and forest growth. Chronica Botanica Co., Waltham, Mass.
- _____. 1950. Crypto-mull humus: Its properties and growth effects (a contribution to the classification of forest humus). Soil Sci. Soc. Amer. Proc. 15:360-362.
- _____. 1954. Reaction of soils; facts and fallacies. Ecology 35:89-92.
- _____. 1958. Forest Soils. Their Properties and Relation to Silviculture. Ronald Press, New York.
- William, R. 1932. The contribution of clay and organic matter to the base exchange capacity of soil. J. Agr. Sci. 22:845-851.
- Wright, J.R. and M. Schnitzer. 1963. Metallo-organic interactions associated with podzolization. Soil Sci. Soc. Amer. Proc. 27: 171-176.
- Youngberg, C.T. 1954. Some site factors affecting the success of reforestation and aforestation activities in the Willamette Valley foothills. Soil Sci. Soc. Amer. Proc. 19:368-372.
- Yuan, T.L. 1963. Some relationships among hydrogen, aluminum and pH in solution and soil systems. Soil Sci. 95:155-163.
- Zinke, P.J. 1962. The pattern of individual forest trees on soil properties. Ecology 43:130-133.

APPENDIX

APPENDIX I

SITE CHARACTERISTICS AND SOIL
PROFILE DESCRIPTIONSSite 1

Description by M.I. Ezenwa, August, 1975.

Soil Series: Witzel very cobbly silty clay loam.

Soil classification (Soil Survey Staff 1970): Lithic Ultic Haploxeroll,
loamy-skeletal, mixed, mesic.

Parent material: Colluvium weathered from basaltic rock.

Drainage: Well drained.

Slope and aspect: 25%, south facing, convex.

Pit locations and elevations: NW 1/4, NW 1/4, sec.22, T10S, R5W,
elevation 330 meters.

Vegetation: 1F.

Quercus garryana, Pseudotsuga menziesii, Abies grandis,
Polystichum munitum, Rosa sp., Rhus diversiloba,
Brachypodium sp., Bromus vulgaris, miscellaneous mosses
and lichens. The age of the stand is between 140 (Douglas-fir)
and 200 (oak) years.
(Inventory, School of Forestry, O.S.U.)

1G.

Elymus caput-medusae, Cynosurus echinatus, Epilobium
sp., Avena fatua, Eriophyllum lanatum, Achillea millefolium,
and Chrysanthemum leucanthemum.

Soils 1F. (see Fig. 3)

Horizon	Depth (cm)	Description
01	4-1	Undecomposed needles, barks, oak leaves, twigs, cones and live mosses.

Horizon	Depth (cm)	Description
02	1-0	Dark brown decomposed litter.
A1	0-8	Dark-brown (7.5 YR 3/2)*, very cobbly silty clay loam; brown (7.5 YR 4.5/4) dry; fine-medium subangular blocky; firm, hard, slightly plastic; common fine roots; many fine tubular and interstitial pores; medium acid; clear smooth boundary.
A12	8-15	Dark brown (7.5 YR 3/2), very cobble silty clay loam; brown (7.5 YR 4/4) dry; medium subangular blocky; firm, slightly hard, slightly plastic; common fine roots, many fine tubular and interstitial pores; medium acid; clear smooth boundary.
B21	15-30	Dark, reddish-brown (5 YR 3/2), very cobbly silty clay loam; dark reddish-brown (5 YR 3/3) dry; medium subangular blocky; slightly hard, firm, plastic and sticky; common fine roots; many tubular fine pores; 40 percent cobbles, 15 percent stones; medium acid; irregular boundary.
B22	30-46	Dark reddish-brown (5 YR 3/2) very cobbly silty clay loam; dark reddish-brown (5 YR 3/3) dry; medium subangular blocky; slightly hard, friable, sticky and plastic; common medium roots; common very fine interstitial pores; 35 percent cobble stones, 20 percent stones; medium acid; clear irregular boundary.
11R	46+	Fractured basalt bedrock that has thin tongues of material from the B22 horizon in fractures.

Soils 1G. (see Fig. 4)

A1	0-8	Dark brown (7.5 YR 3/2) very cobbly silty clay loam; brown (7.5 YR 4/4) dry, very fine and fine granular; soft, very friable, slightly
----	-----	--

* The first Munsell color notation indicates moist color in all cases.

Horizon	Depth (cm)	Description
		sticky and slightly plastic; many fine roots; many fine interstitial pores; 30 percent pebbles and 20 percent cobble stones; slightly acid; gradual smooth boundary.
A12	8-15	Dark brown (7.5 YR 3/2) very cobbly silty clay loam; brown (7.5 YR 4/4) dry; fine granular structure; soft, very friable, slightly sticky and slightly plastic; many fine roots; many very fine interstitial pores; 30 percent cobble stones and 20 percent pebbles; neutral; gradual smooth boundary.
B21	15-25	Dark brown (7.5 YR 3/2) very cobbly silty clay loam; dark brown (7.5 YR 4/2) dry; weak fine subangular blocky structure parting to very fine and fine, granular structure; slightly hard, friable, sticky and plastic; many very fine roots; many very fine interstitial pores; 35 percent cobblestones, 15 percent stones and 10 percent pebbles; slightly acid; gradual irregular boundary.
B22	25-40	Dark brown (7.5 YR 3/2) very cobbly silty clay loam; dark brown (7.5 YR 4.2) dry; weak fine subangular blocky structure; slightly hard, friable, sticky and plastic; many very fine roots; many very fine interstitial pores; 40 percent cobblestones, 20 percent stones and 10 percent pebbles; slightly acid; clear irregular boundary.
IIR	40+	Fractured basalt bedrock that has thin tongues of material from the B2 horizon in fractures.

Site 2

Description by M.I. Ezenwa, August, 1975.

Soil series: Dixonville variant.

Soil classification (Soil Survey Staff 1970): Pachic Argixerolls, fine, mixed, mesic.

Parent material: Colluvium weathered from basic igneous rocks (basaltic).

Drainage: Well drained.

Slope and aspect: 25%, south facing, convex.

Pit locations and elevation: SE 1/4, NE 1/4, Sec. 5, T11S, R5W; elevation, 170 meters.

Vegetation: 2F.

Pseudotsuga menziesii*, Quercus garryana, Corylus cornuta, Polystichum munitum, Rosa sp., Pteridium aquilinum, Rhus diversiloba, miscellaneous mosses and lichens, Bromus vulgaris. The age of the stand is estimated to be about 80 years.

2G.

Bromus mollis, Daucus carota, Poa pratensis, Dactylis glomerata, Fragaria virginiana, Prunella vulgaris, Chrysanthemum leucanthemum, Epilobium sp., Eryophyllum lanatum and Pteridium equilinum.

Soils 2F. (see Fig. 5)

Horizon	Depth (cm)	Description
01	4-1	Undecomposed needles, barks, oak leaves, cones, twigs and live mosses.
02	1-0	Black, well decomposed litter.
A1	0-13	Very dark grayish-brown (10 YR 3/2) silty clay loam; dark brown (10 YR 4/3) dry; strong, fine, subangular, blocky; hard, friable, sticky and

Horizon	Depth (cm)	Description
		plastic; common very fine to fine roots; many very fine interstitial pores; fungal hyphae; medium acid; clear, smooth boundary.
A3	13-28	Very dark grayish-brown (10 YR 3/2) silty clay; dark grayish-brown (10 YR 4/2) dry; moderate, medium, subangular blocky; hard, friable, very sticky and very plastic; common very fine roots, many very fine tubular pores; strongly acid; clear smooth boundary.
B2t	28-48	Dark brown (7.5 YR 3/2) silty clay; dark brown (7.5 YR 4/2) dry; moderate, medium, subangular blocky; hard, firm, very sticky and very plastic; common fine roots; few fungal hyphae; common very fine tubular pores; few moderately thick clay films on peds; medium acid; clear, wavy boundary.
B3t	48-61	Dark reddish-brown (5 YR 3/2) silty clay loam; dark reddish-brown (5 YR 3/3) dry; moderate, medium subangular blocky; hard, firm, sticky and plastic; few medium and fine roots; common very fine tubular pores; common thick clay films; medium acid; clear, wavy boundary.
C	61-127	Variegated, olive brown (2.5 Y 4/4); pale yellow (2.5 Y 7/4) dry and light brownish-grey (10 YR 6/2) moist, light grey (10 YR 7/2) dry; saprolite; massive; hard, very firm, sticky and plastic; few clay films on fractured surfaces.

Soils 2G. (see Fig. 6)

A1	0-13	Very dark brown (10 YR 2/2) silty clay; dark greyish-brown (10 YR 4/2) dry; strong, fine granular; hard, friable, very sticky and very plastic; many very fine roots; many very fine interstitial pores; medium acid; clear, smooth boundary.
----	------	---

Horizon	Depth (cm)	Description
A3	13-25	Very dark greyish-brown (10 YR 3/2) silty clay; dark greyish-brown (10 YR 4/2) dry; moderate, fine, subangular blocky; hard, friable, very sticky and very plastic; many very fine roots; many very fine tubular pores; medium acid; clear, smooth boundary.
B21t	25-43	Dark brown (7.5 YR 2.5/2) silty clay; dark brown (7.5 YR 3/2) dry; weak, medium blocky to prismatic that breaks on handling to subangular blocky; hard, firm, very sticky and very plastic; common very fine roots; many very fine tubular pores; few moderate thick clay films on peds; medium acid; clear wavy boundary.
B22t	43-66	Dark brown (7.5 YR 3/2) silty clay; dark brown (7.5 YR 4/2) dry, weak, medium, blocky to prismatic that breaks on handling to subangular blocky; hard, firm, very sticky and very plastic; few patches of saprolite, common thick clay films on peds; medium acid; clear, wavy boundary.
C	66-127	Variegated, olive brown (2.5 Y 4/4) moist, pale yellow (2.5 Y 7/4) dry and light brownish-grey (10 YR 6/2) moist, light grey (10 YR 7/2) dry; saprolite; massive; hard, very firm, sticky and plastic, few clay films on fractured surfaces.

Site 3

Description by M.I. Ezenwa, August, 1975.

Soil series: Ritner variant.

Soil classification (Soil Survey Staff 1970): Xerochrept, clayey-skeletal, mixed, mesic.

Parent material: Colluvium weathered from basic igneous rocks (basaltic).

Drainage: Well drained.

Slope and aspect: 54%, south facing, convex.

Pit locations and elevation: SE 1/4, SE 1/4, Sec. 3, T11S, R5 W, elevation 300 meters.

Vegetation: 3F.

Pseudotsuga menziesii*, Quercus garryana, Pinus ponderosa, Acer macrophyllum, Polystichum munitum, Corylus cornuta, Elymus sp., mosses. The age of the stand is estimated to be about 150 years.

3G.

Cynosurus echinatus, Elymus caput-medusae, Bromus mollis, Bromus rigida, Epilobium sp., Hypericum perforatum, Chrysanthemum leucanthemum, Elymus glaucus, Rosa eglanteria and Berberis nervosa.

Soils 3F. (see Fig. 7)

Horizon	Depth (cm)	Description
01	5-1	Undecomposed needles, barks, cones and few oak leaves.
02	1-0	Black, well decomposed litter.
A1	0-15	Dark brown (7.5 YR 3/4) gravelly silty clay loam; dark brown (7.5 YR 4/4) dry; strong, medium, subangular blocky; hard, friable,

Horizon	Depth (cm)	Description
		sticky and plastic; many fine roots, many very fine interstitial pores; 20 percent fine and medium pebbles; strongly acid; clear, smooth boundary.
B21	15-43	Dark brown (7.5 YR 3/4) gravelly, silty clay loam; dark brown (7.5 YR 4.4) dry; moderate, medium, subangular blocky; hard, friable, sticky and plastic; many roots; many very fine tubular pores; 25 percent pebbles; medium acid; clear, smooth boundary.
B22	43-71	Dark reddish-brown (5 YR 3/2) cobbly silty clay loam; dark reddish-brown (5 YR 3/3) dry; moderate, medium, subangular blocky; hard, firm, sticky and very plastic; few very fine and fine roots; few very fine tubular pores; 35 percent cobbles and 20 percent gravel; medium acid; clear, smooth boundary.
B3	71-102	Dark reddish-brown (5 YR 3/3) cobbly silt loam; dark reddish-brown (5 YR 3/4) dry; moderate, medium, subangular blocky; hard, firm, slightly sticky and plastic; few fine and coarse roots; few very fine tubular pores; 40 percent cobbles and 20 percent gravel; medium acid; abrupt, irregular boundary.
R	102+	Fractured basalt bedrock with few tongues of B3 horizon in fractures, some red clay coatings on the rock surfaces.

Soil 3G. (see Fig. 8)

A1	0-15	Dark reddish-brown (5 YR 3/3) gravelly silty clay loam; dark brown (7.5 YR 4/4) dry; strong, fine, granular; hard, friable, sticky and plastic; many roots, many very fine interstitial pores; 25 percent fine and medium pebbles; medium acid; clear, smooth boundary.
B21	15-46	Dark reddish-brown (5 YR 3/3) gravelly silty clay loam; dark reddish-brown (5 YR 3/4) dry;

Horizon	Depth (cm)	Description
		strong, fine, subangular blocky; hard, friable, sticky and plastic; many fine roots; many very fine tubular pores; 35 percent gravel, slightly acid, clear, smooth boundary.
B22	46-66	Dark reddish-brown (5 YR 3/3) cobbly silty clay loam; dark-reddish brown (5 YR 3/4) dry; medium and fine, subangular blocky; hard, firm, sticky and plastic; 35 percent cobblestone and 20 percent gravel; slightly acid; clear, smooth boundary.
B3	66-102	Dark reddish-brown (5 YR 3/3) cobbly clay loam; dark reddish-brown (5 YR 3/4) dry; moderate, medium, subangular blocky; hard, firm, slightly sticky, slightly plastic; few very fine roots, few very fine tubular pores; 45 percent cobblestone and 10 percent gravel; slightly acid; abrupt, irregular boundary.
R	102+	Fractured basalt bedrock with few tongues of B3 horizon in fractures; some red clay coatings on the rock surfaces.

Site 4

Description by M.I. Ezenwa, August, 1975.

Soil series: Dixonville variant.

Soil classification (Soil Survey Staff 1970): Pachic Argixeroll, fine, mixed, mesic.

Parent material: Colluvium weathered from basic igneous rocks (basaltic).

Drainage: Well drained.

Slope and aspect: 23%, south facing, convex.

Pit locations and elevation: SE 1/4, SE 1/4, Sec. 18, T11S, R5W; elevation, 200 meters.

Vegetation: 4F.

Pseudotsuga menziesii, Quercus garryana, Acer macrophyllum, Corylus cornuta, Bromus vulgaris, Berberis sp. The age of the stand is between 40 (Douglas-fir) and 60 (oak) years.

4G.

Daucus carota, Bromus sp., Catula glamerata, Prunella vulgaris, Frageria vesca, Rosa sp., few scattered Q. garryana, Pseudotsuga menziesii and Pinus ponderosa.

Soils 4F. (see Fig. 9)

Horizon	Depth (cm)	Description
01	4-1	Undecomposed needles, oak leaves, cones, twigs, barks and some live mosses and lichens.
02	1-0	Dark brown, well-decomposed litter.
A1	0-15	Very dark greyish-brown (10 YR 2.5/2) silty clay; dark breyish-brown (10 YR 4/2) dry; strong, fine granular to subangular blocky; hard, friable, very sticky and very plastic; many very fine interstitial pores; medium acid; clear, smooth boundary.

Horizon	Depth (cm)	Description
A3	15-33	Dark brown (10 YR 3/3) silty clay; brown (10 YR 5/3) dry; moderate, medium, subangular blocky; white silt coatings, hard, friable, very sticky and very plastic; many very fine roots; many very fine tubular pores; medium acid; clear and wavy boundary.
B21t	33-53	Dark brown (7.5 YR 3/2) silty clay; dark brown (7.4 YR 4/2) dry; moderate, medium, subangular blocky; very hard, firm, very sticky, very plastic; common fine roots and few medium roots, clay and few silt coatings on ped surfaces; common very fine tubular pores; slightly acid; clear and wavy boundary.
B22t	53-76	Dark-reddish brown (5 YR 3/3) silty clay; reddish-brown (5 YR 4/3) dry; moderate, medium, blocky to subangular blocky; very hard, firm, very sticky and very plastic; few fine and medium roots; common very fine tubular pores; common clay coatings on the peds; slightly acid; clear and wavy boundary.
C	76-127	Variegated, light brownish-grey (10 YR 6/2), dark yellowish-brown (10 YR 4/4), and yellowish-brown (10 YR 5/6) saprolite; massive, hard, sticky and plastic; some reddish-brown clay coatings on the fractures; slightly acid.

Soils 4G. (see Fig. 10)

A1	0-15	Very dark brown (10 YR 2/2) silty clay; dark grayish-brown (10 YR 4/2) dry; strong, fine granular; hard, friable, sticky and plastic; many very fine interstitial pores; slightly acid; clear, smooth boundary.
A3	15-33	Very dark brown (10 YR 2/2) clay; dark reddish-brown (10 YR 4/2) dry; moderate, fine, subangular blocky; hard, friable, very sticky and very plastic; many very fine roots; many

Horizon	Depth (cm)	Description
		very fine tubular pores; slightly acid; clear, smooth boundary.
B21t	33-61	Dark brown (7.5 YR 3/2) clay; dark brown (7.5 YR 4/2) dry; moderate, coarse, angular blocky to blocky; very hard, firm, very sticky and very plastic; common very fine roots; common very fine tubular pores; common moderately thick clay coatings on peds; some pressure surfaces; neutral; gradual smooth boundary.
B22t	61-74	Dark brown (7.5 YR 3/2) clay; dark brown (7.5 YR 4/2) dry; moderate, medium, angular blocky to blocky; very hard, firm, very sticky and very plastic; common very fine tubular pores; common moderately thick clay coatings, some pressure surfaces, neutral; clear, smooth boundary.
C	74-127	Variegated, light brownish-grey (10 YR 6/2), dark yellowish-brown (10 YR 4/4), and yellowish-brown (10 YR 5/6) saprolite; massive; hard, sticky and plastic, some dark brown clay coatings on the fractures; neutral.

Site 5

Description by M.I. Ezenwa, August, 1975.

Soil series: Bellpine.

Soil classification (Soil Survey Staff 1970): Xeric Haplohumult,
clayey, mixed, mesic.

Parent material: Colluvium weathered from sedimentary rocks
(sandstone).

Drainage: Well drained.

Slope and aspect: 26%, south facing, convex.

Pit locations and elevation : NE 1/4, SE 1/4, Sec. 35, T12S, R6W;
elevation, 130 meters.

Vegetation: 5F.

Pseudotsuga menziesii^{*}, Quercus garryana, Corylus cornuta,
Rhus sp., Berberis nervosa and Dactylis glomerata. The age
of the stand is estimated to be about 130 years.

5G.

Cynosurus echinatus, Festuca sp., Bromus sp., Rubus
ursinus, Elymus caput-medusae, Daucus carota, Dactylis
glomerata, Pteridium aquilinum, Vicia sp., Rubus sp.,
Rosa sp. and scattered oak seedlings.

Soils 5F. (see Fig. 11)

Horizon	Depth (cm)	Description
01	5-1	Undecomposed needles, cones, barks, twigs and few oak leaves.
02	1-0	Dark brown, well decomposed litter.
A1	0-13	Dark brown (10 YR 3/3) clay; brown (10 YR 5/3) dry; moderate, fine to medium, granular; hard, friable, very sticky and very plastic; common very fine roots; many very fine inter- stitial pores; medium acid; clear and smooth boundary.

Horizon	Depth (cm)	Description
B21	13-25	Dark brown (10 YR 3/3) silty clay; dark brown (10 YR 4/3) dry; moderate, fine to medium, subangular blocky; hard, friable, sticky and plastic; common very fine roots, few white roots; many very fine tubular pores; worm casts present, few pebbles; strongly acid; clear and wavy boundary.
B22	25-38	Dark reddish-brown (5 YR 3/3) clay; reddish-brown (5 YR 4/3) dry; moderate, medium subangular blocky; hard, firm, very sticky and very plastic; few worm casts; common very fine roots, few coarse roots; many very fine and fine tubular pores; few clay coatings on peds; few pebbles; very strongly acid; clear and wavy boundary.
B23t	38-64	Dark reddish-brown (5 YR 3/3) silty clay; reddish-brown (5 YR 4/3) dry; moderate, medium, subangular blocky; common fine roots, common coarse roots, few worm casts, many very fine and fine tubular pores; few clay coatings on peds; few pebbles; strongly acid; abrupt and irregular boundary.
II C	64-172	Pink (5 YR 7/4) and yellow (2.5 Y 7/6) partly weathered sandstone, some dark reddish-brown (5 YR 3/3) clay coatings on fragments; tongues of the B23t horizons are present.

Soils 5G. (see Fig. 12)

A1	0-18	Dark yellowish-brown (10 YR 3/4) silty clay; yellowish-brown (10 YR 5/4) dry; moderate, fine to medium granular; slightly hard, friable, sticky and plastic; many very fine roots; many very fine and fine interstitial pores; medium acid; smooth, clear boundary.
B21	18-33	Dark brown (7.5 YR 4/2) silty clay; dark brown (7.5 YR 4/4) dry; moderate, fine, subangular blocky; slightly hard, friable, sticky and plastic; many very fine roots; many very fine

Horizon Depth (cm)

Description

		tubular pores; pebbles at lower boundary; strongly acid; abrupt and wavy boundary.
B22t	33-53	Dark reddish-brown (5 YR 3/3) silty clay; yellowish-red (5 YR 4/8) dry; moderate fine to medium subangular blocky; very hard, firm, sticky and plastic; common very fine roots; many very fine tubular pores; few fine weathered sandstone fragments; strongly acid, few clay coatings on the ped surfaces; gradual and wavy boundary.
B23t	53-74	Dark reddish-brown (5 YR 3/4) clay; yellowish-red (5 YR 5/6) dry; moderate medium, subangular blocky; very hard, very firm, very sticky and very plastic; few very fine roots, common tubular pores; few soft weathered pebbles; strongly acid; thick clay coatings on the peds; gradual and wavy boundary.
II C	74-127	Yellow (2.5 Y 7/6), pink (5 YR 7/4), partly weathered sandstone; some dark reddish-brown (5 YR 3/4) clay coatings on the fragments; tongues of the B23t horizon are present.

Site 6

Description by M.I. Ezenwa, August, 1975.

Soil series: Bellpine.

Soil classification (Soil Survey Staff 1970): Xeric Haplohumult, clayey, mixed, mesic.

Parent material: Colluvium weathered from sedimentary rocks (sandstone).

Drainage: Well drained.

Slope and aspect: 26%, close to the edge of a 35% south facing slope.

Pit locations and elevation: SE 1/4, NW 1/4, Sec. 3, T13S, R6W; elevation, 200 meters.

Vegetation: 6F.

Pseudotsuga menziesii* and Quercus garryana. Pseudotsuga menziesii formed about 95% of the cover and there was no understory. The age of the stand is about 30 years (ring count from stumps).

6G.

Bromus mollis, Pteridium aquilinum, Cynosurus echinertus, Elymus caput-medusae, Dactylis glomerata, Agoseris sp., Rosa englanteria, Rhus diversiloba, Rubus lacinatus and a few scattered Quercus garryana and Pseudotsuga menziesii seedlings.

Soils 6F. (see Fig. 13)

Horizon	Depth (cm)	Description
01	5-1	Undecomposed needles, twigs and cones.
02	1-0	Dark brown, well decomposed litter.
A1	0-15	Reddish-brown (5 YR 4/3) silty clay; yellowish-red (5 YR 5/6) dry; moderate, fine to medium subangular blocky; hard, friable, sticky and

Horizon	Depth (cm)	Description
		plastic; common fine roots; many very fine interstitial pores, strongly acid; clear, smooth boundary.
B21	15-31	Yellowish-red (5 YR 4/6) clay; yellowish-red (5 YR 5/6) dry; moderate fine and medium sub-angular blocky; very hard, very firm, very sticky and very plastic; common fine roots, few soft gravel, many very fine and fine tubular pores; strongly acid; gradual smooth boundary.
B22t	31-46	Dark reddish-brown (5 YR 3/4) clay; reddish-brown (5 YR 4/4) dry; moderate, medium, sub-angular blocky; very firm, very sticky, very plastic; few fine and medium roots; many fine and tubular pores; common thick clay coatings on the peds; soft yellowish-brown pebbles; very strongly acid; gradual smooth boundary.
B23t	46-66	Dark reddish-brown (2.5 YR 3/4) clay; reddish-brown (2.5 YR 4/4) dry; weak to moderate, medium, subangular blocky; very firm, very sticky, very plastic; few fine and medium roots; few fine and tubular pores; common thick clay coatings on the peds; soft yellowish-brown pebbles; very strongly acid; abrupt, irregular boundary.
II C	66-127	Pinkish-grey (7.5 YR 7/2) and reddish-yellow partly weathered sandstone; many thick, dark, red (2.5 YR 3/6) clay coatings on fragments; thin tongues of B23t extend into the C horizon.

Soils 6G. (see Fig. 14)

A1	0-18	Reddish-brown (5 YR 4/4) clay; reddish-brown (5 YR 5/4) dry; moderate, very fine to medium granular; slightly hard, friable, sticky and plastic; many very fine roots; many very fine interstitial pores; few gravel; strongly acid; clear and wavy boundary.
B21	18-36	Reddish-brown (5 YR 4/4) clay; yellowish-red (5 YR 4/6) dry; moderate, very fine and fine,

Horizon Depth (cm)

Description

		subangular blocky; hard, friable, very sticky, very plastic; common very fine roots; many very fine and tubular pores; few pebbles; strongly acid; clear and wavy boundary.
B22t	36-58	Reddish-brown (5 YR 4/4) clay; reddish-brown (5 YR 4/6) dry; moderate, fine, subangular blocky; very hard, friable, very sticky, very plastic; common very fine roots; common very fine tubular pores; few pebbles, clay coatings, strongly acid; gradual, smooth boundary.
B23t	58-76	Dark red (2.5 YR 3/6) clay; red (2.5 YR 5/6) dry; moderate, medium, subangular blocky; very hard, very firm, very sticky and very plastic; few very fine roots, common very fine tubular pores, common thick clay coatings on the ped surfaces; common yellowish weathered pebbles; strongly acid; abrupt and wavy boundary.
II C	76-127	Pinkish-grey (7.5 YR 7/2) and reddish-yellow partly weathered sandstone, many thick, dark red (2.5 YR 3/6) clay coatings on fragments; thin tongues of B23t extend into the C horizon.

Site 7

Description by M.I. Ezenwa, August, 1975.

Soil series: Bellpine variant.

Soil classification (Soil Survey Staff, 1970): Xeric Haplohumult, clayey, mixed, mesic.

Parent material: Colluvium weathered from sedimentary rocks (sandstone).

Drainage: Well drained.

Slope and aspect: 13%, south facing.

Pit locations and elevation: SW 1/4, NE 1/4, Sec. 9, T12S, R6W; 210 meters.

Vegetation: 7F.

Quercus garryana, Pseudotsuga menziesii, Acer macrophyllum, Pteridium aquilinum, Corylus cornuta and Festuca sp. The age of this stand is about 100 years for oak and 80 years for Douglas-fir (ring count from stumps).

7G.

Bromus rigidus, Bromus mollis, Festuca sp., Plantago lanceolata, Holcus lanatus, Poa pratensis, Rosa eglanteria, Rubus sp., mosses and scattered seedlings of oak and Douglas-fir.

Soils 7F. (see Fig. 15)

Horizon	Depth (cm)	Description
01	4-1	Undecomposed oak leaves, few Douglas-fir needles, twigs and barks.
02	1-0	Black, well decomposed litter.
A1	0-15	Dark reddish-brown (5 YR 3/3) silty clay loam; brown to dark brown (7.5 YR 4.5/4) dry; moderate fine to medium, subangular blocky;

Horizon	Depth (cm)	Description
		slightly hard, friable, slightly sticky and slightly plastic; many fine and medium roots; many very fine interstitial pores; few charcoal fragments; medium acid; clear, wavy boundary.
A3	15-31	Dark reddish-brown (5 YR 3/3) silty clay loam; reddish-brown (5 YR 4/4) dry; moderate, medium, subangular blocky; hard, friable, sticky and plastic; common fine to medium roots; many very fine and fine tubular pores; few unweathered pebbles, few pieces of charcoal; strongly acid; clear and wavy boundary.
B21t	31-48	Dark reddish-brown (2.5 YR 3/4) clay; reddish-brown (2.5 YR 4/4) dry; moderate, medium, subangular blocky; very hard, very firm, very sticky and very plastic; common fine to medium roots; few pieces of charcoal; many fine tubular pores; few weathered pebbles; strongly acid; thick clay coatings; gradual, smooth boundary.
B22t	48-69	Dark reddish-brown (2.5 YR 3/4) clay; reddish-brown (2.5 YR 4/4) dry; moderate, medium, subangular blocky; very hard, very firm, very sticky and very plastic; common fine to medium roots; many fine tubular pores; common soft yellow pebbles; strongly acid; thick clay coatings; abrupt wavy boundary.
II C	69-127	Reddish-yellow (5 YR 6/6) and (7.5 YR 7/8) partly weathered sandstone; many thick, dark reddish-brown (2.5 YR 3/4) clay coatings on the fragments.

Soil 7G. (see Fig. 16)

A1	0-15	Dark brown (7.5 YR 3/2) silty clay loam; dark brown to brown (7.5 YR 4/4) dry; moderate, fine to medium granular; slightly hard, friable, sticky and plastic; many very fine roots; many very fine interstitial pores; medium acid; gradual and smooth boundary.
----	------	--

Horizon	Depth (cm)	Description
A3	15-36	Dark brown (7.5 YR 3/2) silty clay loam; dark brown to brown (7.5 YR 4/4) dry; moderate, medium, subangular blocky; slightly hard, friable, sticky and plastic; many very fine roots; many very fine interstitial pores; medium acid; few pebbles; abrupt, wavy boundary.
B21t	36-56	Dark reddish-brown (2.5 YR 3/4) clay; reddish-brown (2.5 YR 4/4) dry; moderate, medium, subangular blocky; very hard, very firm, very sticky and very plastic; common thick clay coatings on peds; few fine roots, few pieces of charcoal; many fine tubular pores; medium acid; gradual, smooth boundary.
B22t	56-79	Dark reddish-brown (2.5 YR 3/4) clay; reddish-brown (2.5 YR 4/4) dry; moderate, medium, subangular blocky; very hard, very firm, very sticky and very plastic; few fine roots; common tubular pores; common soft, yellow pebbles; common thick clay coatings on the ped surfaces; strongly acid; abrupt and wavy boundary.
II C	79-127	Reddish-yellow (5 YR 6/6) and (7.5 YR 7/8) partly weathered sandstone; many thick, dark reddish-brown (2.5 YR 3/4) clay coatings on the fragments; thick tongues of the B22t horizon extend deep into the C horizon.