A STUDY OF THE MORPHOLOGY AND PEDOGENESIS OF A MEDIAL CHERNOZEM DEVELOPED IN LOESS

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This is to certify that the

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THESIS





### A STUDY OF THE MORPHOLOGY AND PEDOGENESIS

#### OF A MEDIAL CHERNOZEM DEVELOPED IN LOESS

By

### W. CLINTON BOURNE

### AN ABSTRACT

Submitted to the School of Graduate Studies of Michigan State University of Agriculture and Applied Science in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

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#### ABSTRACT

The soil investigated, Bozeman silty clay loam, is a well drained, medial Northern Chernozem developed in loess. The surface horizons are black and granular. In contrast, the subsoil horizons are dark grayish brown and compound, prismatic-blocky. The limecarbonate horizon is characterized by accumulations of lime along a very coarse network of prismatic-blocky structure planes and tubular pores. The underlying C horizons are massive with only disseminated lime.

The principal objective of the investigation has been to describe and measure those features of the soil believed to reflect the processes of soil genesis. The loess parent material is a pebble-free, calcareous, pale brown, massive silt loam similar to Wisconsin loesses of the Central Plains. Its outstanding characteristic under the microscope is the uniformity and simplicity in size, shape and arrangement of the individual grains and pores making up the matrix. Sizes equal to coarse silt predominate for both grains and pores.

Particle-size distributions for sixteen samples collected at five sites along an east-southeast traverse extending twenty-five miles from the bluffs of the Madison River indicate that local flood plains contributed significantly to the silt and sand fractions. Texture of the locess is finer as the distance from the probable local sources increases. The percentages of heavy minerals in the very fine sand and coarse silt

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fractions change significantly immediately to the leeward of the West Gallatin flood plains. This change is still apparent in the very fine sand ten miles distant, but not so in the coarse silt. This is evidence that local sources contribute mostly to the coarser fractions. It appears that at least two cycles of loess deposition have occurred. For this reason the quantitative pedogenetic changes were calculated relative to the  $C_{ca}$  horizon.

Examination of the soil under the binocular microscope revealed striking changes in the microfabric from horizon to horizon indicating that pedogenetic changes have occurred since the deposition of the parent material. The structure of the principal subsoil horizons appears as somewhat irregular blocks which fit together with many angular edges and flat surfaces. Most surfaces appear glazed, indicating the presence of clay films. The surface horizons are more coarsely porous. They show marked granular aggregation so that the pore-space lattice is made up complexly of pores of varying size. The general appearance is that of a black, complex, sugary mass.

The A horizons have clay contents 9.3%, 12.3% and 13.5% higher; the B horizons, 14.1% and 9.3% higher; and the  $C_{ca}$ , 6.9% higher than that of the C horizons. The volume weight data show the A horizons to be relatively lighter and the B horizons relatively heavier than the C horizons. The percentage of acid-soluble material indicates that slightly more than three per cent of lime has accumulated in the  $C_{ca}$ . The soil is neutral to alkaline, each horizon being slightly more alkaline than the one above.

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The mineral suites of all horizons are qualitatively alike. Quartz, the feldspars, volcanic glass, the amphiboles, the micas and chlorite predominate. In general, the mineralogical data indicate a relative decrease in the weatherable fractions from the C horizons toward the surface of the soil. Inversely, there is an increase in quartz and clays. It is probable that the differences largely are due to differential weathering during the process of soil formation. Clearly, of course, marked translocation of acid-soluble material has occurred and the organic matter has tended to remain at the depth of synthesis.

The cation exchange capacities of the clay fractions range from 55.2 to 63.2 for the A horizons; whereas they range from 70.0 to 80.3 for the B and C horizons. X-ray diffraction intensities together with potassium contents, cation exchange capacities and specific surfaces of the fraction less than two microns in diameter indicate that the clay minerals in Bozeman silty clay loam are illite, montmorillonite, kaolinite, and possibly halloysite. Apparently, the principal changes which have occurred in the clay fraction of the solum during the course of soil formation have been (1) a relative gain of illite in the A horizons, (2) a relative gain of montmorillonite in the B horizons, (3) a reconstitution of much of the clay of the A horizons to randomly interstratified, mixedlayer forms, and (4) an increase of total clay by formation. The shifts in relative illite and montmorillonite contents of the A and B horizons are likely the result of differential leaching. The marked increase in randomly interstratified minerals in the surface horizons may be largely

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due to a potassium equilibrium reaction among the clays.

The calculations of the amount of clay formed in each horizon show the relative rate of clay formation to be the highest in the lower A horizons. The rate of formation drops sharply in the  $B_{22}$  horizon. According to the calculations about a fourth of the clay gain in the  $B_{21}$ and about two-thirds of the clay gain in the  $B_{22}$  are due to translocation from the A horizons.



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### CHAPTER I

## INTRODUCTION

This is the report of a study of the morphology and pedogenesis of a Northern Chernozem soil from Montana.

The soil investigated, Bozeman silty clay loam, is the more humid member of a climo-sequence of soils developed in loess under grassland vegetation in intermountain valleys and on outwash plains of western Montana. It was selected, after extensive field investigations, as a suitable beginning point for the study of pedogenesis and soil genetical relationships over a large area. The chernozemic zones of great soil groups (Chernozems, Chestnuts and Browns) occupy the major portion of Montana. As far as the author is aware, no similar study has been made of a Montana soil.

The principal objectives of the investigation have been (1) to describe and measure those features of the soil believed to reflect the processes of soil genesis, (2) to measure the progress of soil formation, (3) to evaluate the relative importance of the various processes involved, and (4) to assess the value of the data obtained with respect to its usefulness in characterizing natural soils and tracing the course of soil genesis.

The techniques used have involved the study of field relationships, the macro and micropedological study of the undisturbed soil profile, the mineralogical analysis of the soil separates, and the measurement of the more important physical and chemical properties of the soil.

During the course of the study, particular attention has been focused upon the development of the color, texture and structure profiles of the soil.

## CHAPTER II

# REVIEW OF THE LITERATURE

In this review of the literature the principal objective has been to outline briefly the development of the present day concepts of the Chernozem great soil group. Particular emphasis has been given to papers bearing on the characteristic environment, morphology and genesis of the modal soils. Additional references bearing upon the origin, characteristics and uniformity of the loess, which is the parent material of the soil studied, also are reviewed.

# A. CHERNOZEMS

Chernozems were first recognized and described on the great steppes of Russia where they are co-extensive with grassland vegetation and temperate, subhumid climates. According to Glinka (1914) the first attempt to explain their origin was made in 1763, by Lomonosoff who said that their formation was not mineralogical or geological, but belonged to the other two natural kingdoms, vegetable and animal. Rupprecht, in 1866, first suggested that the Chernozems have developed from the steppe grass vegetation. In 1883, Dokutschajeff added climate as an important factor in their formation. Thus, they are still recognized today, --a great group of soils developed under prairie vegetation and cool to warm temperate, subhumid climate.

1. <u>Morphology and Environment</u>: Glinka (1914) recognized four principal varieties of Chernozems in European Russia--(1) Northern

Chernozem, (2) Ordinary Chernozem, (3) Fat or Thick Chernozem, and (4) Southern Chernozem. He described the morphology of the Ordinary Chernozem as follows:

In ordinary Chernozem the thickness of the humus horizons  $(A_1 + A_2)$  reaches an average thickness of about 28 inches, of which the thickness of the  $A_1$  horizon is less than half, and often only a third. The transition from horizons  $A_1$  to  $A_2$  is more abrupt than is that in Fat Chernozem. Horizon  $A_2$  is spotted and streaked; dark spots and tongues alternate with tongues and spots of material like the parent rock. In the loams and silt loams, horizon  $A_1$  is granular, but the structure is faintly developed. It is sometimes lumpy, the lumps break up however into granules. In horizon  $A_2$  the granular structure becomes nut structure and this passes into a lumpy prismatic structure below. In the upper 5 to 10 centimeters horizon  $A_1$  is marked by a faintly developed platy structure, though it disappears under cultivation.

His use of  $A_1$  and  $A_2$  as horizon designations is not quite clear, but apparently they are equivalent to the A and B horizons, or solum, as now used in American literature. Certainly the  $A_2$  does not refer to a light colored, eluviated layer as in most current American literature.

The other varieties differ from the ordinary Chernozem in being thicker and darker or in having grayer surface soils and slightly compacted, prismatic subsoils. According to Nikiforoff (1936) the rich brown subsoils of Southern Chernozems are characterized by considerable compactness and conspicuous lumpy or cubical structure, together with dark and often glassy coatings on the sharp, angular aggregates and in the pores. All of the varieties are described as having lime carbonate and gypsum accumulations in the lower subsoil horizons. These accumulations were considered to be due to the action of pedogenetic processes, as shown by the fact that they occur not only in Chernozems that have developed on calcareous material but also in those developed on granite, basalt or other kinds of rocks.

The Chernozem great soil group was introduced into the classification of soils in the United States by Marbut (1927, 1935). Apparently to him their most important characteristics were the dark, friable surface soils relatively high in organic matter, and the horizon of carbonate accumulation. The upper, lime-free subsoil horizons were usually described as transitional. However, soils with subsoils distinctly finer textured than the surface soils were recognized. Total chemical analyses presented by Marbut (1935) for soils without marked textural profiles show little differential movement of silica and sesquoxides, suggesting that the weathering of primary minerals to new forms has not been an important process in Chernozem genesis.

Baldwin, Kellogg and Thorp (1938) define Chernozems as a zonal great soil group of well draised, black of very dark grayish-brown, friable soils, extending to depths, ranging up to three or four feet and grading through lighter color to whitish lime accumulations, developed under tall- and mixed-grass prairies in temperate to cool, subhumid climates.

Thorp, Williams and Watkins (1948) recognized minimal, medial and maximal Chernozems. The minimal soils are described as having very dark grayish brown, granular surface soils and dark brown, granular subsoils of about the same texture. Calcium carbonate

horizons begin at three to five feet. The medial soils differ by having moderately finer textured subsoils than surface soils, with blocky or prismatic structure. The maximal soils have distinctly finer textured subsoils and strong compound prismatic-blocky structure.

McClelland et al. (1959) describe Northern Chernozems developed from glacial till in North Dakota as having A horizons five to fourteen inches thick that are black, granular loams of neutral reaction with sand grains that are unstained and partly free of adhering organic particles. The B horizons are brown to very dark grayish brown, usually moderate compound prismatic-blocky and five to fourteen inches thick. There are sometimes texture B horizons with clay films on ped faces that range from thin and patchy to thin and continuous. The  $C_{ca}$  horizons are strongly calcareous with weak structure. The soils occur on gentle to rolling, convex to concave topography and are well or moderately well drained. Particle-size distribution data for Barnes and Aastad profiles show moderate accumulation of fine clay in the lower A and the B horizons. Nygard et al. (1952) and Westin (1953) report for similar soils developed in loam till in Minnesota and South Dakota, A and B horizons with some three to five per cent more clay than the C horizons.

In accordance with the above definitions, the Bozeman silty clay loam profile selected for this study is a medial Northern Chernozem developed in loess. The Bozeman series was first recognized and described in the Gallatin Valley Area, Gallatin County, Montana by DeYoung and Smith (1931). The profile selected for study is typical of the series

in that area. It is described in detail in the chapter on Characteristics of the Soil.

The climate of the Chernozem soils zone in the plains region of North America is classed by Thornthwaite (1948) as dry-subhumid. It is characterized by cold winters and warm to hot summers. Mean monthly January temperatures range from  $0^{\circ}$  F. in North Dakota to  $30^{\circ}$  F. in central Kansas. Mean July temperatures range from  $65^{\circ}$  F. to  $80^{\circ}$  F. Mean annual precipitation ranges from about fifteen inches to about twenty inches in the northern part and from about twenty-two inches to thirty inches in the southern part of the zone (<u>Climate and Man</u>, 1941 Yearbook of Agriculture). About two-thirds falls during the growing season. Normally, at the beginning of the growing season, the soil is moist into the lime carbonate horizon, but as the season warms and vegetative growth becomes vigorous, it dries rapidly. Usually the subsoil is dry from midsummer to the end of the growing season. The surface soil, however, may be rewetted from time to time by summer rains.

Weaver (1954) has described the vegetation as being dominated by mid- and tall-grasses with an understory of short grasses, sedges and forbs. Tall forbs, including deep-rooted legumes are scattered through the association.

Topography occupied by Chernozems ranges from level to rolling or hilly; drainage from moderately well to well. The soils occupy both concave and convex slopes and all aspects. Well developed, modal Chernozems do not occupy excessively steep or exposed topographic



positions, nor do they occur in excessively drained materials.

2. <u>Genesis</u>: Nikiforoff (1936) and Byers <u>et al.</u> (1938) consider the outstanding features of Chernozems, which dominate all their other characteristics, to be a high content of humus in the A horizon, an accumulation of carbonates below the  $B_2$  horizon, and a saturation of the colloidal complex by the metallic ions, among which calcium is by far the most significant. They relate the formation of Chernozems to (1) Humification, (2) Carbonate accumulation and (3) Calcification of the colloidal complex.

Humification or humus formation is a biochemical process. In Chernozems, much of the vegetation produced is returned annually to the soil. Considering the gradual decrease of organic matter with depth and the lack of evidence of much humus translocation, the underground parts are perhaps the more important. In young soils humus tends to accumulate; in more mature soils it tends to remain constant. The average content is a function of the amount and composition of organic residues and the rate of their mineralization.

The development of a lime carbonate horizon through accumulation of alkaline earths is a characteristic process of Chernozem formation. Generally, it is believed to occur by leaching of carbonates from overlying horizons. However, some have suggested that under certain conditions it may be accumulated by upward capillary movement of water. Although it is recognized that most of the carbonates are inherited from the parent material, it is believed that some are formed

from the weathering of calcium-bearing minerals.

Calcification is the process of the saturation of the soil colloidal complex with metallic cations, chiefly calcium and to a lesser extent magnesium, potassium and sodium. It is an exchange reaction and depends upon an excess of these ions in the soil solution. Thus, an alkaline or neutral reaction, a slow leaching rate or a rapid biological recirculation of cations is requisite to the formation of Chernozems.

Leaching is not generally considered to be an important factor in Chernozem formation. The amount of moisture which penetrates beyond the lime carbonate horizon is very limited.

3. <u>Clay Minerals</u>: In the classical reviews of Chernozemic soil development, the formation and translocation of clay minerals have not been considered to be of consequence. Many authorities have held the view summarized by Grim (1953, p. 344), that in the weathering of calcareous sediments, there is substantially no alteration of the silicates until the carbonate is completely broken down and the calcium removed from the environment. Yet many soils recognized as Chernozemic (including Chestnuts, Chernozems and Brunizems) have distinctly more clay in the B than in the **A** horizons and more clay in the solum than in the C horizons (Thorp, <u>et al</u>. 1948; Winters and Simonson, 1951; Unpublished standard soil series descriptions, National Coöperative Soil Survey, USDA).

In the Brunizem zone of Illinois and Iowa, where leaching is somewhat greater than in the Chernozem zone, a number of studies of

loess-derived soils with AB/C and B/A clay ratios markedly higher than one have been reported (Bray, 1934; Smith, 1942; Beavers <u>et al.</u>, 1955; Davidson and Handy, 1954). Generally the higher clay content of the solum relative to the underlying loess has been attributed to pedogenetic processes. Bray (1934) states that the clay starts forming early in the weathering process before the carbonates have been leached away and its period of greatest formation and downward movement occurs before the soil has developed any great acidity.

Larson, Alway and Rhoades (1947) reported a moderate increase in total clay and a marked increase in fine clay in the B and A horizons over the calcareous C horizons for a profile of Holdrege silt loam, a medial Chernozem. Differential thermal analyses indicated that montmorillonite and illite were the dominant clay minerals. The cation exchange capacities of the fine and coarse clays were respectively 55. 3 and 16. 1 for the A, 88. 8 and 60. 4 for the B and 69. 9 and 47. 5 for the C horizons respectively. They believed that particularly the fine clay was formed largely during the period of soil development.

Foth and Riecken (1954) reported similar texture differences for two Moody and four Galvin silt loam profiles, Chernozem-Brunizem intergrades, in northwest Iowa. The marked decrease in fine clay occurred at from twenty-four to thirty-four inches. In the case of the Moody soils this break occurred at the beginning of the carbonate horizon; but in the case of the Galvin profiles the break occurred at the beginning of non-calcareous  $B_3$  or  $C_1$  horizons. The authors made no attempt to



explain the texture changes. However, as mentioned later, Pleistocene stratigraphy nearby in South Dakota would strongly suggest that loess may be stratified, the upper strata being about thirty inches thick.

Riegers and Smith (1955) reported the occurrence of a deep  $A_2$  in certain soils of the Prairie zone of the Palouse area of Washington and Idaho to be associated with a lithologic break in the loess parent material. They believed the development of the  $A_2$  and the underlying, more clayey  $B_2$  to be at least in part pedogenetic, but attributed major importance to the lower loess strata being initially more compact and possibly finer textured. Associated soils without  $A_2$  horizons did not exhibit lithologic breaks within the profile.

In general, the kinds of clay minerals in soils in temperate climates seem to depend more on the kind of parent material than on pedogenetic factors (Van Houton, 1955). For loessial chernozemic soils the predominant minerals usually reported are montmorillonite and illite (Beavers et al., 1955; Russell and Haddock, 1940; Schmehl and Jackson, 1957; Larson et al., 1947; Davidson and Handy, 1954). Beavers et al. (1955) and Larson et al. (1947) indicated that the surface soils are relatively higher in illite than the subsoils. Schmehl and Jackson (1957) emphasized the randomly interstratified nature of the clay minerals in surface soils.

The literature generally indicates that in the formation of clays, the reactants must first be in solution. Henin (1956) and his associates have prepared substances similar to clay minerals by allowing
very dilute (tenths of milligrams per liter) solutions of stable alkaline silicates or aluminates and stable acids to run together at the rate of several milliliters per day in a large glass vessel containing two liters of distilled water. Using Mg,  $Fe^{+2}$ ,  $Fe^{+3}$ , and Ni as cations and pH values more alkaline than 7.0, minerals of the montmorillonite type were prepared. When K was in the solutions illite was sometimes formed. In acid solutions oxides or amorphous substances were usually formed. Most of the reactions were conducted at temperatures of  $100^{\circ}$  C., but clays were formed at temperatures as low as  $20^{\circ}$  C. By extrapolation the rate of clay formation at  $100^{\circ}$  C. relative to  $0^{\circ}$  C. was 560:1.

Brindley and Radoslovich (1956) examined weathered white clayey appearing materials on the surfaces of feldspar crystals in granite. X-ray analyses revealed no obvious clay minerals. The material appeared to be feldspars broken down to small particle size. When Na-feldspar flakes and powder were subjected to attack by 0.1 N HCl at hydrothermal conditions of  $280^{\circ}$  C. or  $430^{\circ}$  C. for periods ranging from a few hoars to fifteen days boehmite was formed from the flakes and a variety of products including kaolinite from the powders. According to Grim (1953), Niggli (1939) reports that under certain conditions in the alteration of feldspar, the breakdown reaches a colloidal state before the formation of secondary products. Grim reports that Demolon and Batisse (1946) have concluded that in the weathering of a granite there is a spontaneous change to the clay minerals.

Many authors report the possibilities of changing one type of



clay to another. De Mumbrum and Hoover (1958), Dyal and Hendricks (1952), and Kunze and Jefferies (1953) have changed expanding-lattice clays to illitic type clays by the fixation of K in the interlayer spaces. Wear and White (1951) reduced the cation exchange capacities of montmorillonitic clays by 16% to 27% by such treatment. Mortland <u>et al</u>. (1956) report the changing of biotite to vermiculite due to the extraction of K by growing plants.

#### B. LOESS

Although many theories as to the origin of loess deposits have been proposed (Schedig, 1934, in a review of the subject listed more than 20), it is now generally believed that they are of aeolian origin. The earlier proponents of the aeolian theory in America, such as Chamberlain (1897), postulated that the loess was blown from nearby major flood plains of Pleistocene rivers as they were exposed following the recession of floods. Many studies (Smith, 1942; Bollen, 1945; Swineford and Frye, 1951: Davidson and Handy, 1954) have shown that loess become progressively higher in clay content with distance away from major flood plains or dune-shaped, very sandy areas, such as the sandhills of Nebraska. Smith (1942) and Davidson and Handy (1954) show that thickness and carbonate content of the loess decrease with distance from the flood plains and suggest that some of the increased clay content may be due to weathering.

Beavers (1957) has recently suggested that only the silt and sand size fractions may be of local origin and that the clay may have

been blown in from distant arid areas. This, he believes, is the most logical explanation of the predominance of montmorillonitic clays in the loess of the Central States whereas the clays in the adjacent Wisconsin tills and flood plains are illites and kaolinites. There is a predominance of montmorillonitic clays in the arid Western States. He postulates that the clays were electrostatically attracted and adsorbed onto the silt and sand particles while still in the air, and that they were deposited together. This could explain the lack of particle size segregation and lack of stratification in loess.

Ruhe (1952) summarizes the views that thick loess deposits, even though without apparent unconformities, may be made up of strata of different ages. Schultz and Stout (1945) have described a succession of buried soils in the loess of southwestern Nebraska. Thorp <u>et al.</u> (1951) have discussed in detail the complexity of loess stratification in relation to buried soils and glacial till sheets. Along the margin of the early Wisconsin till in southeastern South Dakota the author of this thesis has observed geologic sections in which strata of loess of similar lithology are stratigraphically separated by the glacial till. Beyond the terminal moraine, these strata of loess blend into a single deposit in which the two strata are indistinguishable. The upper, younger loess is from twenty to forty inches in thickness and is finer textured than the older loess. It is about equal to the thickness of the sola. Thus the texture profiles of soils developed on loess in that area may be due partly to stratification.

#### CHAPTER III

### PROCEDURES

#### A. FIELD STUDIES

Using the "Soil Survey of the Gallatin Valley Area" by De Young and Smith (1931) as a guide, the distribution of loess and the soils developed in it were studied in relation to the geomorphology, climate and vegetation of the area. It was found that a climo-sequence of chernozemic soils with medial development exists. The Chernozem member of this sequence, Bozeman silty clay loam, was selected for detailed study. It was sampled in a native sod pasture on the Ft. Ellis Experimental Farm about four miles east of Bozeman, Montana, in Section 15, Township 2S., Range 6E. The samples were collected from the vertical wall of a freshly dug pit. Eight horizons extending to a depth of seventy-six inches were described and sampled. A half-gallon bulk sample and five, three inch by three inch, undisturbed, cylindrical cores were taken from each horizon. The bulk samples were air dried, crushed with a rolling pin and passed through a two-millimeter screen prior to being used for the analyses. Negligible amounts failed to pass through the screen. The core samples were oven dried at  $105^{\circ}$  C. for thirty hours but were otherwise undisturbed prior to use in analyses.

In connection with the studies of the stratigraphic and areal variability of the loess parent material, four additional profiles were sampled by soil horizons along a west northwest--east southeast traverse extending from the bluffs of the Madison River Valley to the location of the soil profile sampled for detailed analysis. The five sample locations, designated A, B, C, D and E are shown in Figure 1. Sites A and B are between the Madison and West Gallatin Rivers. Sites C, D and E are east of the West Gallatin River. Site E is the location of the profile of Bozeman silty clay loam studied in detail.

## B. MICROMORPHOLOGICAL STUDIES

The micro-descriptions were made by breaking the cores open along natural structural planes and examining them under the stereoscopic microscope at magnifications of 9x to 54x. Color, luster, pore space, aggregation of silt and sand particles, apparent degree of weathering of individual particles, species of minerals, and presence or absence of clay skins and other films were observed.

Thin sections were prepared by the method of Bourbeau and Berger (1948) by dropping oriented aggregates about one centimeter in diameter into vials two-thirds full of castolite to which four drops of hardener had been added. The vials were then placed in a vacuum desiccator and subjected to alternate vacuum and atmospheric pressure until all air bubbles were removed from the aggregates and plastic. After standing overnight at atmospheric pressure they were oven dried for four hours at  $100^{\circ}$  C., cooled, and the glass vials were broken off. Vertical and horizontal sections about two millimeters thick were sawed from the castolite impregnated aggregates, ground flat on one side, fastened to glass slides with Canada balsam, and reduced on a wet-grinding lap to thicknesses of 0.8 millimeters to 0.12 millimeters. The finished thin sections were examined and described under the petrographic microscope.

# C. PHYSICAL AND CHEMICAL PROPERTIES

1. <u>Volume Weight</u>: The oven-dried cores were weighed and their volume weights were calculated on the basis of their volumes at the time of sampling. The volume weights, which are reported in Table 14, are averages of five cores. The dried cores were saved to provide aggregates for micro-description under the stereomicroscope and for the preparation of thin sections.

2. <u>Specific Gravity</u>: To determine the specific gravity or density of the whole soil about a half gram of soil was immersed in a solution of KHgI<sub>3</sub> adjusted to a density of 2.70. Entrapped air was removed by intermittent aspiration. Then the density of the solution was reduced by steps of 0.05 until the soil sank on being centrifuged.

3. Organic Matter: Organic matter content was determined by oxidizing duplicate ten-gram samples of each horizon with hydrogen peroxide. To the weighed oven-dry samples twenty-five milliliter portions of five per cent hydrogen peroxide were added. After frothing ceased the samples were heated to 90° C. and five milliliter portions of thirty per cent hydrogen peroxide were added at forty-five minute intervals until all of the organic matter was removed. Heating was continued for thirty minutes. After cooling, the samples were decanted into tared filter crucibles and washed five times prior to oven drying at 105° C.

The loss of weight on an oven-dry basis was calculated as organic matter.

4. <u>Acid Solution Loss</u>: Acid soluble material was determined by treating ten-gram soil samples, from which organic matter had been previously removed by hydrogen peroxide oxidation, with fifty milliliters of 0. 2N HCl for twenty-four hours. After treatment the samples were washed in tared filter crucibles five times with 0. 1N HCl and ten times with distilled water prior to oven drying at 105<sup>°</sup> C. and weighing. The loss of weight on an oven-dry basis was calculated as solution loss.

5. <u>Particle-Size Distribution</u>: Particle-size distribution was determined by the pipette method using ten-gram samples from which organic matter and acid soluble material had been previously removed as described above but omitting the oven-drying steps. Dispersion was effected by neutralization with NaOH to pH 8.0 and shaking overnight in a reciprocal shaker. Sand was removed prior to pipetting with a 300mesh screen and separated into size fractions by dry screening. Particlesize distribution of the mineral fraction has been reported in Table 11, as percentages of the oven-dry whole soil, and in Table 1, as percentages of the oven-dry soil after removal of organic matter and acid-soluble material.

The clay and silt remaining in the suspension after pipetting were separated by settling and decanting repeatedly (8 to 10 times) until the supernatant liquid was clear at the end of the settling period. These separates along with the sands were saved for mineralogical analyses.

The geometric means of the weight-diameter distributions

 $(M_{\not p})$  and their standard deviations  $( \sigma )$  were calculated in terms of the Phi  $( \not p )$  scale as described by Krumbein and Pettijohn (1938). The Phi scale is a transformation of the Wentworth grade scale, which is logarithmic, to an arithmetic scale. Phi is equal to the  $\log_2$  of the particlesize diameter. The weight-diameter distributions were obtained by multiplying the frequencies of the particle-size classes by their median diameters.

Analyses of variance were calculated according to the single factor and two factor models of replication summarized by Krumbein and Miller (1953). The step by step procedure for making the calculations is given by Dixon and Massey (1951).

6. <u>pH Measurements</u>: pH was measured with a Beckman glass electrode pH meter, Model 2 in 1:5 soil water suspensions stirred at intervals for an hour. The final stirring was made thirty seconds before the readings were taken.

7. <u>Cation Exchange Capacity</u>: Cation exchange capacity of clays was determined by ammonium acetate replacement following the procedure outlined by the United States Salinity Laboratory (1954).

8. <u>Specific Surface</u>: Total, external, and internal specific surfaces of the whole soil were measured by the ethylene glycol retention method of Bower and Gschwend (1952) as outlined by the United States Salinity Laboratory (1954).

9. <u>Total Potassium</u>: Total potassium was determined with a Beckman flame photometer following hydrofluoric acid digestion of the

clay by a modification of the method described by Treadwell and Hall (1942). Two milliliters of concentrated  $H_2SO_4$  and four milliliters of concentrated HFl were added to 0.5 grams of oven-dry clay in a platinum crucible and the mixture was digested in a heated sandbath until dense fumes appeared. Another four milliliters of HFl were added and the solution then was evaporated to dryness. To the residue, fifteen milli-liters of 1N HCl was added. This solution was boiled for five minutes, cooled and filtered. The filtrate was diluted to about seventy-five milli-liters, neutralized to the phenophthalein end point with  $NH_4OH$  and diluted to exactly 100 ml. After settling, the potassium content of the supernatant liquid was determined with a Beckman flame photometer.

10. <u>Heavy Mineral Separation</u>: The very fine sand (50-100 microns) and coarse silt (20-50 microns) separates were divided into density fractions of more than 2.95, 2.75 - 2.95, and less than 2.75. The heavy liquid used was 1, 1, 2, 2-tetrabromoethane. One- to two-gram aliquots of the separates were placed in fifty milliliter long-tapered centrifuge tubes, covered with about twenty milliliters of the tetrabromoethane at a density of 2.95, evacuated for fifteen minutes, mixed by swirling, and then centrifuged at 1,000 r. p. m. and a radius of 5 inches for 5 minutes. The swirling and centrifuging were repeated twice. Then, the bottom part of the tube containing the heavy minerals was frozen by placing it in a scooped out cavity in a one-pound block of drv ice for three or four minutes. The top part of the tetrabromoethane containing the light minerals was then poured off, and the sides of the tube were washed

clean with acetone. After the bottom part had thawed the heavy minerals were transferred to filter paper, washed with acetone, dried and weighed. The lighter fraction was again divided into two fractions by repeating the procedure using tetrabromoethane adjusted to a density of 2.75 by dilution with acetone. The procedure proved to be simple, rapid and accurate. In the case of the very fine sand the fractionation was particularly sharp, there being virtually no contamination of the heavier fraction by lighter minerals.

11. <u>Mineral Identification</u>: All of the density fractions were examined under the petrographic microscope to identify the mineral suites. In the heavy fraction with density of more than 2.95, the percentage composition of green hornblende, brown hornblende, garnet, zircon and opaque or coated grains was determined quantitatively by making duplicate counts of 300 or more grains in randomly selected fields.

12. <u>X-Ray Analysis of the Whole Soi</u>l: The percentage content of quartz, calcite, dolomite, and an approximation of the feldspar content of the whole soil was determined by x-ray diffraction. The apparatus used was a Norelco, Type 42202, geiger counter x-ray spectrometer with tungsten filament, copper target and nickel filter. Power settings were at twenty kilovolts and thirty-five milliamperes. Slits of  $1^{\circ}$  and 0.006 were used, and the goniometer was run at a speed of  $1^{\circ}$  per minute. The data were plotted by an automatic strip chart recorder with a time constant of two seconds, a multiplier of one and scale factors of sixteen or sixty-four as required to keep the peaks on the chart.

The samples of whole soil were scanned in triplicate from  $25^{\circ}$  to  $32^{\circ}$  while being rotated at the rate of 100 revolutions per minute in a Norelco flat speciman spinner. Previously they were ground to pass a 300-mesh screen and mounted in open ring holders. A firm, smooth scanning surface was obtained by pushing upward on the rubber stopper which served as the bottom of the holder, while a glass slide was held on the surface. A quartz standard was scanned before and after each sample run and used to adjust the peak heights to a standard peak height of 3840 for the 3.35 Å diffraction plane of pure quartz.

In addition to the samples from each horizon of the soil, samples of pure quartz, calcite, dolomite, the feldspars and known mixtures of these were also scanned. Data from these knowns were used to plot peak-height or peak-area vs. percentage-content curves from which the percentage content of each mineral in the soil samples was determined. The peaks used were those for d-spacings of 3, 35 Åfor guartz, 3.04. Å for calcite, and 2.88. Å for dolomite. In the case of feldspars the area of the combined peaks from 2 theta =  $27.0^{\circ}$  to  $28.6^{\circ}$ (d = 3, 31, A - 3, 14 A) was used, and the curve was taken to be a straight line from the average of the areas of orthoclase, microcline. albite, and esine, labradorite and anorthite at 100% to an area of 0.0 at a feldspar content of 7.5%. Thus, the true values for the feldspars may vary considerably from those determined. However, the indicated relationships between horizons should be valid. The data for the soil are given in Tables 8, 9 and 10. Those for the pure minerals are in Tables 21 and 22 and Figures 6, 7 and 8 in the Appendix.

13. <u>X-Ray Analysis of the Clay</u>: To aid in the identification and quantitative estimation of the clay minerals, x-ray diffraction data were also obtained for the fraction less than two microns in diameter with the same equipment, using a copper target and nickel filter with power settings of twenty kilovolts and thirty-five milliamperes. One degree scatter slits and a 0.003 counter slit were used. Scanning was at the rate of one degree per minute. Interpretation of the curves was in accordance with the method of Johns and Grim (1954).

Oriented films were prepared for x-raying by plating the clays on porous ceramic tile by a modification of the method described by Kinter and Diamond (1956), suction rather than centrifuging being used to remove excess water. A l'' x 2'' x 3/8'' porous ceramic tile was cut to a length of about 1.25 inches and ground flat. A suspension containing approximately thirty milligrams of Na-clay which had been treated overnight with five drops of glycerol was poured into the well of the holder. As the moisture was drawn through the tile, the clay was deposited uniformly on the surface as an oriented clay film. After plating was completed, five drops of a twenty per cent solution of glycerol in water was applied to the surface and drawn through to insure complete glyceration. The tile with the deposited, glycerated Na-clay film was air-dried and then stored in a desiccator for at least twenty-four hours prior to x-raying. After x-raying, it was returned to the special holder and leached by suction three times with five to six drops of IN KCl and then six to seven times with distilled water. The films were then air-dried and stored in a desiccator. The films were not disturbed by the process of leaching.

However, some films were ruined by curling during the drying process, apparently due to failure to completely remove the excess KCl. Data from the ruined films were discarded and replaced by data from new mounts successfully carried through the complete cycle. Qualitative identification of the clay minerals was made by means of strip-chart rate curves over the arc  $2^{\circ}$  to  $30^{\circ}$  two theta with the scaler set at a time constant of four seconds, a multiplier of one, and a scale factor of four.

The samples were first x-rayed as glycerated Na-clays. Afterward they were saturated with potassium, leached of glycerol, and air-dried as previously described. They were then x-rayed repeatedly after ovendrying at  $110^{\circ}$  C. for two hours, after oven drying at  $300^{\circ}$  C. for two hours, and after oven drying at  $575^{\circ}$  C. for two hours. Shifts in the basal-spacing peaks with treatment were used to identify the clay minerals present.

Quantitative measurements of peak intensities were made with the scaler over a specified arc of two theta on a different set of clay films prepared as described above. Settings of the x-ray were as before. A scale factor appropriate to the intensity of the peak was used. All counts were converted to a common basis by multiplying by the scale factor. The arc scanned for each basal spacing is shown in Table 1. Each arc was scanned once.

The background was measured at points on either side of the arcs as shown in Table 1. Measurements were made in triplicate as seconds per 100 counts, the average being converted to counts per

# TABLE 1

d / n ( A)	Arc 2 <b>0</b> (degrees)	Seconds counted	Background points (degrees)
18.0	3.0 - 7.0	240	3, 7 <sup>a</sup>
10.0	7.5 - 10.5	180	7, <sup>a</sup> 11
7.2	11.5 - 13.0	90	11, 13.5
3.6	23.2 - 25.7	150	23, 25.7 <sup>b</sup> , 30 <sup>c</sup>
3.3	25.7 - 28.2	150	23, 25.7 <sup>b</sup> , 30 <sup>c</sup>

# ARC SCANNED AND BACKGROUND POINTS MEASURED FOR EACH BASAL SPACING

<sup>a</sup>7.8<sup>°</sup> measured on glycerated Na-clays of soil.

<sup>b</sup>Not used in calculations.

<sup>c</sup>29<sup>°</sup> measured on glycerated pure Na-clay mixtures.

second. The background for the 3.6 Å and 3.3 Å peaks was calculated by multiplying the simple averages of the counts per second at 23° and  $30^{\circ}$  by 150 That for the 7 Å peak was calculated by drawing a smooth curve through the plot of intensities against degrees 2 **G** for the measurements at 7°, 11°, 13.5° and 23° and taking the intersection of the curve with 12.3° as the average background. This intensity was multiplied by ninety to obtain the total background count. A separate curve was prepared for each treatment of each sample. In addition to the clays from the profile of Bozeman silty clay loam, samples of pure kaolinite, montmorillonite, and illite, and mixtures of these were x-rayed as shown in Table 15. Counts were made for glycerated Na-clays and for K-clay heated to  $300^{\circ}$ C. for two hours. For the soil clays, duplicate counts were made from different spots on the same clay film. For the pure clays, duplicate counts were made from different clay films.

#### CHAPTER IV

## CHARACTERISTICS OF THE LOESS MANTLE

Although barely mentioned in the geologic literature of the area (Berry, 1943; Peale, 1896), one of the more important proto soils or mantle formations of the Gallatin Valley Intermountain Basin in southwestern Montana is a pale brown, uniformly silty, calcareous loess which is as much as fifteen feet thick. Since it occurs across an area with marked variation in precipitation (13" - 18") and is the parent material for zonal soils ranging from Browns to Chernozems, within a distance of twenty-five miles, its characteristics are of considerable interest to soil scientists. The study reported here was made to assess the relative uniformity of this material in both its areal and vertical distributions.

## A. FIELD OBSERVATIONS

The Gallatin Valley is in southwestern Montana within the Northern Rocky Mountains physiographic province. It lies about fifty miles north of Yellowstone National Park. Bozeman is the principal town.

The valley is one of a system of intermountain basins underlain by late Tertiary sediments. The elevation of the upper rim is approximately 5, 500 feet. Elevation at the confluence of the Missouri River with the Gallatin, Madison, and Jefferson rivers is about 4,040 feet. The boundary on the east and the south is marked by maturely dissected, complex mountains which rise to peaks of 9,000 to 11,000 feet. To the north are maturely dissected, complex uplands, known as the Horseshoe Hills, which rise to a general elevation of about 6,000 feet. On the west is an escarpment which drops some 300 to 700 feet to the floodplain of the Madison River, beyond which is an extension of the Tertiary basin and maturely dissected complex uplands.

The Tertiary alluvial plains in general slope toward a common point in the north-central part of the basin. Subsequent to their formation, they have been weakly faulted and maturely dissected. In the central part, they have been buried under Pleistocene alluvium.

The present broad landforms in the Gallatin Valley and its environs include (1) floodplains and recent terraces; (2) low gravel terraces and fans; (3) low, loess mantled, immaturely dissected terraces and fans; (4) loess mantled maturely dissected Tertiary alluvial plains; (5) loess mantled, maturely dissected, complex uplands; and (6) maturely dissected, complex mountains. The generalized distribution of these forms in shown in Figure 1.

Thickness of the loess ranges up to at least fifteen feet. It should be noted that it is not in all cases continuous over the landforms described as being mantled. However, on all forms, the loess occurs on both ridges and valley slopes, indicating that the period of loess deposition was largely subsequent to the dissection. Its absence in some areas may be due to more recent dissection locally or to local conditions which prevented deposition. Presumably its source would have been the



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SAMPLE SITES
FLOODPLAINS AND RECENT TERRACES
LOW GRAVEL TERRACES AND FANS
LOESS MANTLED IMMATURELY DISSECTED TERRACES AND FANS
LOESS MANTLED MATURELY DISSECTED COMPLEX UPLANDS
MATURELY DISSECTED COMPLEX MOUNTAINS

FIGURE 1: GENERALIZED GEOMORPHOLOGY OF THE GALLATIN VALLEY

AREA, MONTANA

recent floodplains and gravel terraces shown in Figure 1 or similar landforms farther west and north. The alluvium in these areas was derived from a large variety of sedimentary, metamorphic, and igneous extrusive and intrusive rocks (Ross <u>et al.</u>, 1955). Since the loess is only eighteen inches to thirty-six inches thick on the loess-mantled immaturely dissected terraces and fans of the central part of the basin, it would seem that these landforms may also have been active floodplains and contributors to the loess during the earlier part of the deposition. This would suggest that the loess may have been deposited in distinct cycles separated by intervals during which there was no deposition.

Its relationships to the landforms of Recent and Pleistocene ages would indicate that the loess was deposited rather late in the Pleistocene period.

# B. PETROGRAPHY OF THE LOESS

The loess in the Gallatin Valley is a pebble-free, calcareous, pale brown, massive silt loam with characteristics observable in the field similar to those of the extensive Wisconsin loesses of the Central Plains of the United States.

1. <u>Microfabric</u>: The outstanding characteristic of the undisturbed loess viewed under the stereoscopic microscope is the uniformity and simplicity in size, shape, and arrangement of the individual grains and

<sup>&</sup>lt;sup>\*</sup>John Montagne, by personal communication, reports that near Amsterdam, Montana, the loess overlies a post-Tertiary fault with over fifty feet of throw.

pores making up the matrix. Sizes equal to coarse silt predominate for both grains and pores. Clay is not evident. The free calcium carbonate is mostly uniformly distributed. There is no apparent orientation of particles nor stratification. The individual grains are angular to subangular. They vary from equidimensional to flattened or elongated. There are a few volcanic shards.

2. <u>Mechanical Analysis</u>: The particle-size frequencies and the weight-diameter distributions in Phi units are given in Tables 2 and 3. Representative cumulative frequency curves are shown in Figure 2. These data show the Gallatin Valley loess to be similar in texture characteristics to Wisconsin (Peorian) loesses of the Great Plains and Corn Belt States as reported by Swineford and Frye (1951), Watkins (1945), Bollen (1945), and Kay and Graham (1943). It would thus appear that it was deposited under similar conditions.

Analysis of variance of the means of the weight-diameter distributions  $(M_p)$  of the samples from the twenty-eight to forty-eightinch depth at each site shows variations among sites to be significantly greater than the differences between duplicate analyses at the one per cent level. The least significant difference at the five per cent level is 0.11.

The differences due to site can be best explained as a reflection of distance from the source of the material. Undoubtedly the Madison and Jefferson floodplains in the vicinity of Three Forks were major sources of dust. The West Gallatin floodplain immediately west of Site

	IA.
	MONTAN
	VALLEY,
	GALLATIN
TABLE 2	LOESS SAMPLES,
	FREQUENCY OF
	PARTICLE-SIZE

A	Horizon	Depth (inches)	Solution loss <sup>2</sup>	Medium sand >250µ <sup>3</sup>	Fine sand 100-250 µ <sup>3</sup>	Very fine sand $50-100 \ \mu^3$	Coarse silt 20-50μ <sup>3</sup>	Fine silt 2120µ <sup>3</sup>	Clay3 < 2 \$
	c2	32-40	17.51	. 38	2.06	18.60	49.75	22.24	6.97
В	000 642 642	36-48 64-76 96-114	19. 04 16. 33 16. 31	.13 .10 .21	. 36 . 32 . 45	10.12 10.78 11.66	54.99 54.62 49.08	26. 32 22. 32 26. 38	8.08 11.86 12.22
υ	C 25 C 25	28-48 96-114	22. 16 17. 09	. 10	. 79 . 39	14.07 13.13	51.82 58.28	23.40 17.21	9.72 10.89
D	C <sub>2</sub>	36-48	19, 30	.13	. 62	11.07	46.53	24.45	17.20
ш	A A A A A A A A A A A A A A A A A A A	0-4 4-8 8-13 13-21 21-28 28-48 48-48 64-76 64-114	8.35 8.95 6.44 5.03 4.87 19.08 15.57 16.26 14.15	. 48 . 26 . 25 . 25 . 25 . 17 . 11 . 11 . 48	1.76 1.25 1.23 1.33 1.91 1.91 1.91 1.17 1.17 1.17 1.7 2.46	12.52 12.24 12.04 14.80 17.44 16.50 9.41 16.41	32. 65 32. 17 32. 02 30. 69 31. 71 31. 71 31. 71 45. 79 43. 96 32. 96	24.02 22.52 21.61 19.50 20.06 20.33 24.61 26.82 26.05	28, 57 31, 55 32, 85 33, 46 28, 63 28, 62 19, 22 19, 22 19, 32 21, 64

l Averages of duplicate analyses.

<sup>2</sup>Percent by weight of oven-dry soil.

 $^{3}$  Percent by weight of mineral soil after solution loss.

TABLE 3	WEIGHT-DIAMETER DISTRIBUTIONS IN PHI UNITS OF LOESS SAMPLES, GALLATIN VALLEY, MONTANA
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			W	edium sand	Fine sand	Very fine sand	Coarse silt	Fine silt	Clay	Mg	(Ba
		1 H	Diameter Ø	units:							
		I	Range:	0-2	2-3.4	3. 4-4. 4	4.4-5.7	5.7-9.0	- 6		
			Median:	1.0	2.7	3.9	5.0	7.4	10.0		
Site	: Hor- izon	Depth (inches)									
A	c <sub>2</sub>	32-40	•.	0038	. 0556	. 7254	2.4880	1.6454	. 6970	5.61	1.74
щ	ပ်	36-48	•	0013	. 0097	. 3947	2.7495	1.9477	. 8080	5.91	1.70
	°℃	64-76	•	0010	.0085	. 4208	2.7319	1.6521	1.1860	6.00	1.84
	ۍ 4 %	96-114	•	. 0021	.0121	. 4546	2.4535	1.9520	1.2220	6.10	1.88
υ	ပ်	28-48	•	0020	.0214	. 5488	2.5910	1.7316	.9720	5.87	1.80
	$C_{2}^{\ell}$	96-114	•	.0010	.0106	.5119	2. 9135	1.2733	1: 0890	5.80	1.81
D	C <sub>2</sub>	36-48	•	0013	.0168	.4320	2.3260	1.8097	1.7200	6.31	2.06
ជ	Α,,	0-4	•	0048	.0475	. 4879	1.6340	1.7755	2.8570	6.81	2.38
		4-8	•	0026	.0340	. 4774	1.6079	1.6670	3.1550	6.94	2.39
	A,5	8-13	•	0025	.0334	.4697	1.6004	1.5991	3.2850	6.99	2.37
	B,15	13-21	•	0023	.0374	. 5773	1.5341	1.4430	3.3400	6.93	2.46
	B,1	21-28	•	0025	.0517	.6754	1.5912	1.4841	2.8630	6.67	2.45
	$C_{23}^{rr}$	28-48	•	0017	.0316	. 6438	1.7879	1.5039	2.6220	6.59	2.36
	ر م	48-64	•	0011	.0110	.3820	2.2899	1.8213	1.9280	6.43	2.09
	ີບົ	64-76	•	0010	.0148	. 3663	2.1949	1.9788	1.9320	6.49	2.08
	C 2 C	96-114	•	0048	. 0665	. 6397	1.6487	1.9276	2.1640	6.45	2.29





C must have been a secondary source. The floodplains of the Gallatin and East Gallatin rivers to the north of the traverse may have been another source, depending on the prevailing wind direction. The means of the weight-diameter distributions  $(M_{\phi})$  for the samples within the twentyeight to forty-eight inch-depth-range are 5.61 at Site A, 5.91 at B, 5.87 at C, 6.30 at D, and 6.58 at E. Thus, the texture is progressively finer as distance increases from the probably main source except that it is very slightly coarser at Site C where the adjacent West Gallatin floodplain is an undoubted secondary source. It is possible that some of the extra clay at Site E is the result of weathering. At that site, the twenty-eight to forty-eight inch depth is the calcium carbonate horizon  $(C_{ca})$  of the soil and shows evidence of change due to soil formation. However, the means of the weight-diameter distributions for the ninetysix to one hundred fourteen inch depth at Sites B, C, and E and for the sixty-four to seventy-six inch depths at B and E show the same trends. It is doubtful whether weathering at that depth has significantly altered particle-size distribution. The average means of all depths below twentyeight inches at each site are 5.61 at A, 6.00 at B, 5.83 at C, 6.30 at D, and 6.49 at E.

It is thus evident that in the Gallatin Valley, as elsewhere, the particle-size distribution of loess is directly related to the distance of transport from possible local sources. The heavy mineral data, discussed below indicate that this relationship may be largely a reflection of the absolute abundance of the coarser particles. To determine whether there might have been differences in texture of the loess with depth when originally deposited at Site E, the frequency distribution of the silt and sand size particles was recalculated to a clay-free basis and then multiplied by the mean diameters of the size classes to obtain the weight-diameter distributions (Table 4). This would be a valid reflection of the original distribution if clay formed by weathering <u>en situ</u> came proportionately from each of the size fractions. Although it is recognized that the finer fractions may tend to weather somewhat more than the coarser fractions, it is not believed that this would be sufficient to appreciably alter the calculations. Analysis of variance of the means of the weight-diameter distributions shows significant variations at the one per cent level. The least significant difference is 0. 122 at the five per cent level.

Therefore, it seems possible that there are real differences in the particle-size distributions at the various depths. The principal differences in the distributions are the lower sand contents of the fortyeight to sixty-four and sixty-four to seventy-six-inch depths.

The means of the weight-diameter distributions fall into four distinct groups: the surface soil horizons or 0 - 13 inches; the subsoil and the horizon of lime accumulation or 13 - 48 inches; the two parent material horizons from 48 to 76 inches and the lower substratum from 96 to 114 inches. The standard deviations (**r**) indicate that the two horizons between 48 and 76 inches have a somewhat narrower range of sorting than the others.

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RECALCULATED TO		
WEIGHT-DIAMETER DISTRIBUTION IN PHI UNITS BY SOIL HORIZONS AT SITE E	A CLAY-FREE BASIS	

			¢	THAT - LALL	CICAD			
		Medium sand	Fine sand	Very fine sand	Coarse silt	Fine silt	Mø	G
	Dia	imeter Ø units:						
		Range: 0-2 Median: 1.0	2-3 <b>.4</b> 2,7	3. 4-4. 4 3. 9	4, <b>4-5.</b> 7 5, 0	5,7-9.0 7.4		
Horizo	on Depth (inches)							
A	0 - 4	。0068	, 0666	。6831	2.2875	2.4857	5, 53	1.45
A <sub>12</sub>	4-8	, 0039	。0497	, 6974	2.3490	2.4354	5.54	1,43
A <sub>13</sub>	8-13	, 0038	.0497	. 6995	2.3833	2.3814	5, 52	1。40
B <sub>21</sub>	13-21	.0035	. 0562	. 8668	2,3035	2。1667	5.40	1。40
B <sub>22</sub>	21-28	° 0036	。0725	. 9464	2, 2295	2,0794	5, 33	1.42
o C	28-48	, 0023	。(428	. 8726	2,4133	2。0384	5, 37	1,36
c <sup>2</sup>	48-64	, 0014	, 0136	, 4733	2,8368	2。2563	5, 58	1。27
່ິ	64-76	0013	0184	, 4540	2, 7005	2.4527	5,65	1, 30
ົບິ	96-114	.0061	, 0849	. 8163	2.1040	2.4599	5, 47	l <sub>°</sub> 49

As the horizon at 96 to 114 inches is in the lower part of the loess section, the higher sand content and poorer sorting may be due to mixing with the underlying Tertiary sediments. The groupings at 0 - 13, 13 - 48, and 48 - 76 inches may reflect substages in the deposition of the loess, the upper and lower depths being periods of lower local wind velocities as compared to the thirteen to forty-eight-inch depth.

The data for the three depths sampled at Site B and the two depths at Site C do not appear to show differences in the particle-size distributions.

3. <u>Mineralogy</u>: Minerals with densities heavier than 2.95 were separated from the very fine sand (50 to 100 microns) and coarse silt (20 to 50 microns) separates for all samples except the 28- to 48inch depth at Site C and the 36- to 48-inch depth at D. Those with densities of 2.75 to 2.95 were also separated from all samples at Site E. The data are given in Table 5.

Analysis of variance of the samples in the 28- to 48-inch depth-range at Sites A, B, and E and of the samples at the 96- to 114inch depth at Sites B, C, and E show significant variation at the five per cent and one per cent levels, respectively. The least significant difference at the five per cent level for the 28- to 48- inch depth is 1.07, and for the 96- to 114-inch depth is 0.47. Means of all samples below twenty-eight inches at each site give heavy mineral percentages for the very fine sand fraction of 4.25 for Site A, 4.97 for Site B, 6.58 for Site C, and 5.59 for Site E. For the coarse silt, the equivalent

#### TABLE 5

				Den	sity	
	ι		2.9	5 +	2.75 -	2.95
Site	Horizon	Depth (inches)	Very fine sand (percent)	Coarse silt (percent)	Very fine sand (percent)	Coarse silt (percent
A	C2	32 - 40	4. 25%	7.17%		
, B	C2	36-48	4. 35	6.27		
	C4	64-76	5.50	8.20		
	c <sub>6</sub>	96-114	5.06	7.18		
с	c <sub>2</sub>	96-114	6.58	10.09		
E	A <sub>11</sub>	0 - 4	4.46	5.84	. 87%	2.02%
	A <sub>12</sub>	4-8	4.12	4.42	1.16	2.50
	A <sub>13</sub>	8 - 1 3	4.17	5.27 <sup>a</sup>	1.19	3.43
	B <sub>21</sub>	13-21	4.57	5.36	1.15	1.57
	B <sub>22</sub>	21-28	5.64	6.96	1.34	1.66
	C <sub>ca</sub>	28-48	5.74	7.01	1.67	1.94
	c2	48-64	5.89 <sup>a</sup>	6.98	1.89 <sup>a</sup>	2.20
	C3	64-76	5.68	7.45	1.67	2.00
	C <sub>5</sub>	96-114	5.05	6.88	1.26	2.21

#### DENSITY SEPARATIONS OF VERY FINE SAND (.1 - .05 mm) AND COARSE SILT (.05 - .02 mm) FRACTIONS

<sup>a</sup>Single determination, replicate calculated.

percentages are 7.17 for Site A, 7.22 for Site B, 10.09 for Site C, and 7.08 for Site E.

Sites A and B are immediately to the leeward and some ten to twelve miles to the leeward, respectively, of the floodplains of the Madison River. Sites C and E are immediately to the leeward and some eight to ten miles to the leeward, respectively, of the West Gallatin River floodplains. Sites A and B should be influenced only by the Madison River floodplains, and Sites C and E should be influenced by the floodplains of both rivers.

Thus, it would appear that the percentage content of heavy minerals is associated with the source, and the distance from the source, of the loess and the size of the particles involved.

The percentage of heavy minerals in the very fine sand fraction at Site C is sharply higher than any at the other locations. This is apparently due to the influence of the adjacent West Gallatin floodplains which must be of different mineral composition than the Madison floodplains. The drop in the heavy mineral content from Site C to Site E would indicate that the influence of an added local source of dust decreases as its distance from the point of deposition increases. However, comparison of the average heavy mineral content of the 28- to 48-inch, 64- to 76-inch and 96- to 114-inch depths at Site E (5.49%) with the average of the same depths at Site B (4.97%) would indicate that the West Gallatin floodplains are still affecting the composition of the very fine sands at a distance of some eight to ten miles.

In the case of coarse silt, influence of the West Gallatin floodplain is evident at the adjacent Site C. However, comparison of the average at Site E (7. 11%) with the average at Site B (7. 22%) would indicate that it may not have had a significant influence on the composition of coarse silt at Site E. This could indicate that many of the finer particles may have blown in from some distance and would further suggest that the contribution of local sources may be largely to the coarser fractions of the loess.

At Site E, the percentages of heavy minerals in both the very fine sand and the coarse silt to a depth of twenty-one inches are clearly lower than those below that depth. Analysis of variance shows the differences to be significant. It could be presumed that the differences largely are due to differential mineral weathering during the process of soil formation. The percentages of coated and opaque minerals at Site E (Table 6) would indicate that at least some weathering has occurred to a depth of 64 inches.

Examination of both the light and heavy mineral fractions under the polarizing microscope indicated that the mineral suites of all samples are qualitatively alike. Quartz and the feldspars constitute more than ninety per cent; these plus the amphiboles, micas, and chlorite constitute much more than ninety-five per cent. The minerals identified in their approximate decreasing order of abundance are quartz, feldspars (oligoclase, andesine, labradorite, orthoclase, microcline), amphiboles - green hornblende, brown hornblende, actinolite, tremolite),

# TABLE 6

GRAIN FREQUENCY COUNTS OF SELECTED MINERALS
OF THE HEAVY PORTION (DENSITY > 2.95) OF
THE VERY FINE SAND FRACTIONS

Site	Horizon	Depth (inches)	Green hornblende (percent)	Brown hornblende (percent)	Garnet (percent)	Zircon (percent)	Opaque or coated (percent)
A	c <sub>2</sub>	32-40	49.7%	2.5%	2.0%	1.7%	9.7%
в	C <sub>2</sub>	36-48	57.6	3.2	1.7	. 3	6.7
	C4	64-76	56.1	2.1	2.9	. 9	5.7
	c <sub>6</sub>	96-114	54.5	3.8	3.2	. 7	7.5
С	c <sub>2</sub>	96-114	55.1	5.9	2.8	. 9	5.2
E	A <sub>11</sub>	0 - 4	51.7	. 8	3.1	. 7	15.5
	A <sub>12</sub>	4-8	45.6	. 5	3.3	. 3	20.5
	A <sub>13</sub>	8-13	50.6	1.9	2.5	1.2	15.2
	B <sub>21</sub>	13-21	40.0	. 9	2.8	1.5	21.2
	B <sub>22</sub>	21-28	43.3	1.0	2.2	. 2	15.2
	C <sub>ca</sub>	28-48	44.5	. 9	3.0	.9	15.3
	c2	48-64	41.2	1.5	1.5	. 7	12.5
	C3	64-76	52.6	1.4	3.0	1.0	8.7
	C <sub>5</sub>	96-114	54.0	2.3	3.8	1.1	6.7

micas (muscovite, biotite), chlorite, pyroxenes, olivine, epidote, garnet, volcanic glass, zircon and rutile. There appeared to be no differences in the suites between the very fine sand and coarse silt fractions.

The quantitative counts of certain minerals are shown in Table 6. Only green hornblende occurred in sufficient quantity to give data suitable for statistical analysis. It constituted from 40.0 per cent to 57.6 per cent of the heavy minerals. Analysis of variance of all determinations of green hornblende, considering them to be completely random samples, gave significance at the one per cent level with a least significant difference of 8.71 at the five per cent level. However, the differences do not follow a pattern which can be explained either on the basis of location or depth.

#### C. IMPLICATIONS OF THE DATA

Particle-size distribution, heavy mineral percentage, and mineral suites were determined for sixteen samples collected at various depths at five sites (Figure 1) along an east-southeast traverse extending twenty-five miles from the bluffs of the Madison River to the foot of the Gallatin Mountains.

The data indicate that, in the intermountain basin under investigation, local floodplains contributed significantly at least to the silt and sand fractions of the loess. In this study, texture of the loess is shown to be finer as the distance from possible local sources increases. This is in accordance with the findings of various investigators of the extensive Wisconsin loesses of the Great Plains and Central Lowlands provinces. A significant finding of this study is that a local source (the floodplains of the West Gallatin River) markedly affected the percentage of heavy minerals in the very fine sand fraction (50 to 100 microns) as much as ten miles from the source. It also changed the percentage of heavy minerals in the coarse silt fraction (20 to 50 microns) immediately adjacent to the source but apparently had little influence at a distance of ten miles. This would be evidence that local sources may contribute most to the coarser fractions of loess.

With respect to vertical distribution, the data indicate that variations in texture may exist. At Site E, the particle-size distributions adjusted to exclude clay appear to be distinctive for the 0 - 13, 13 - 48, 48 - 76, and 96 - 114-inch depths. Each of these may represent different periods of deposition or different stages in waning cycles of deposition. If the latter were true the  $C_5$  would represent the beginning and the  $C_2$ the ending of one cycle, and the  $C_{ca}$  would represent the beginning and the  $A_1$  the ending of another. In areas where loess of different ages can be separated stratigraphically, further study would be desirable to determine whether particle-size distribution data could be used as a criterion for separating depositional stages in loess.

It is probable that the differences in the percentage of heavy minerals at Site E may be due to weathering. If this were true, the zone of marked weathering during the formation of the soil, a modal northern Chernozem, would be limited to the upper twenty-one inches. Soil profile characteristics would indicate that weathering has occurred at least to the bottom of the lime-free B horizon, or twenty-eight inches.

The percentages of coated and opaque minerals would indicate that some weathering may have extended to a depth of sixty-four inches. Therefore, the variations in heavy mineral percentages may reflect stratification as much as weathering.

## D. INITIAL UNIFORMITY OF THE LOESS AT SITE "E"

A determination of whether the primary (parent) material of the soil was initially uniform or stratified is basic to studies of pedogenesis. Without such a determination it is difficult to distinguish the separate influences of depositional and pedogenetic forces on the anisotropy of the soil profile. The Marshall and Haseman (1942) method of quantitative evaluation of pedogenesis requires the assumption that some underlying stratum is representative of the primary material in which the solum is developed.

In the case of soils developed in Wisconsin loess, an <u>a priori</u> assumption that pedogenetic forces began in essentially uniform primary material has been widely accepted. In gross features vertical sections of Wisconsin loess free from paleosols or intervening strata of glacial till, alluvium or other material, almost invariably do appear to be uniform in character. Very frequently, different strata of loess cannot be distinguished in the field by gross lithology but must be identified by their stratographic relationships to the intervening formations. On field examination, the vertical section of loess at Site E, where the profile of Bozeman silty clay loam was sampled for this study, appears to be uniform.
Nevertheless, although not conclusive, the data presented here strongly suggest that an <u>a priori</u> assumption of uniformity may not be valid in all cases. For the soil profile studied it appears better to assume that at least two cycles of deposition have occurred, and that the material deposited may have become somewhat finer in texture as the cycles waned. For this reason the quantitative pedogenetic changes presented in Chapter V are calculated relative to the  $C_{ca}$  horizon, and it is recognized that the solum may not be entirely free of depositional variation.

#### CHAPTER V

## CHARACTERISTICS OF THE SOIL, BOZEMAN SILTY CLAY LOAM

#### A. FIELD OBSERVATIONS

Bozeman silty clay loam is a well-drained Chernozem developed in calcareous, pale brown, silty loess. It occurs in intermountain valleys or basins in central and western Montana where it is one of a vertical climo-sequence of Brown to Chernozem soils which occur in traverses of less than twenty-five miles. The parent loess is one of a complex of local and general loesses extending westward to the Palouse of Washington and Oregon. It is similar in field observable characteristics to members of the Peorian loess of the Midwestern States.

The climate is cool, subhumid with an average annual precipitation of eighteen inches and an average frost-free period of about 115 days. The mean January and mean July temperatures are 20.8° F. and 64.6° F., respectively. The average snowfall is 70 inches (Climate and Man, 1941). Ordinarily the snow cover disappears in April. Thawing usually alternates with freezing, and rains or wet snows are common during that period. At the beginning of the growing season the soil is usually saturated, but with the beginning of rapid vegetative growth, usually late May, transpiration exceeds precipitation and the soil becomes dry to the depth of significant root penetration. Ordinarily it is dried to fifty or sixty inches by mid-July. The subsoil is rarely re-saturated until the following winter or spring. The surface soil, however, may be rewetted several times during the summer. Thus, the A horizons may undergo several wetting and drying cycles per year, but the B horizons and the  $C_{\rm ca}$  only one. Appreciable percolation of water through the B horizons rarely occurs except during the cool autumn and spring seasons of low vital activity. The substrata below sixty inches is permanently moist indicating that there is some movement of water downward beyond the reach of roots.

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The native vegetation of the area is tall-grass prairie dominated by bluebunch wheatgrass (Agropyron spicatum) and Idaho fescue (Festuca idahoensis). Brush and low deciduous trees grow in ravines and on steep north slopes but never in areas of Bozeman silty clay loam. Conifers are dominant in the adjacent mountains. Growth begins slowly after the disappearance of the snow cover in April. It becomes rapid in late May with the appearance of warm weather, and continues until the grasses mature in July. By that time available moisture may be deficient and the activity of the microflora will also have slowed. The annual period of intensive biological activity is thus short.

The soil profile studied was collected on a broad, gently convex slope of three to five per cent gradient. In terms of fieldobservable morphology, it shows medial development. A, B, C<sub>ca</sub> and C horizons are clearly distinguishable but lack abrupt boundaries. The surface horizons are black and granular. In contrast, the subsoil

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horizons are dark brown and compound, prismatic-blocky. They are slightly finer textured than the surface and parent material horizons, and are leached of free carbonates. The lime-carbonate horizon is characterized by accumulations of lime along a very coarse network of prismaticblocky structure planes and tubular pores. The underlying parent material is massive with only disseminated lime.

The profile was described in the field as follows:

Soil type ..... Bozeman silty clay loam.

Classification .... Medial Chernozem.

Collected ...... August 21, 1955, by W. C. Bourne. The profile was dry to forty-eight inches and moist below.

Location ...... Ft. Ellis Experiment Farm; east of Bozeman, Montana; 225 feet east of lane and 70 feet south of Highway US #10 fence, in pasture north of house. NW 1/4 of NW 1/4, Section 15, T.2S., R.6E.

Parent material.. Calcareous loess of silt loam texture, apparently of uniform character to 100 inches or more.

Physiography .... Moderately dissected, undulating to gently rolling Tertiary Age bench mantled with loess.

Relief at profile .. Four per cent, smooth, northwest slope about 100 yards below the crest of a broad, gentle interfluve.

Vegetation ...... Native sod pasture consisting mostly of blue-bunch wheatgrass, Idaho fescue, bluegrass, western wheatgrass, thread-leaf sedge, vetch, goldenrod, and yellow sweet clover.

Root distribution. Abundant to 21 inches; plentiful to 40 inches; few to 76 inches.

Drainage ...... Well drained. Run off is moderate. The entire profile is permeable and the water table is very deep.  $\operatorname{Profile:}^*$ 

A <sub>11</sub>	(Vhl)	0-4"	Dark grayish yellowish brown (10 YR $3/1.3$ dry to 10 YR $2/1$ moist); silt loam; moderate, fine, platy breaking to moderate, fine, crumb; soft when dry; friable when moist; clear, smooth boundary.
A <sub>12</sub>	(Vh2)	4-8''	Dark grayish yellowish brown (10 YR 3/1.3 dry to 10 YR 2/1 moist); coarse silty clay loam; compound, moderate, medium pris- matic breaking to moderate, medium and fine granular; hard when dry; friable when moist; slightly sticky and slightly plastic when wet; clear, smooth boundary.
A <sub>13</sub>	(Vh3)	8 - 1 3''	Dark yellowish brown (10 YR 3/2.5 dry) to dark grayish yellowish brown (10 YR 2/2 moist); silty clay loam; compound, moderate, medium prismatic breaking to moderate, and strong coarse granular medium blocky; very hard when dry; friable when moist; sticky and plastic when wet; clear, smooth boundary.
B <sub>21</sub>	(Itl)	13-20"	Grayish yellowish brown (10 YR $4/2$ dry) to dark yellowish brown (10 YR $3/3$ moist); silty clay loam; compound, moderate, medium prismatic breaking to very strong fine blocky, very hard when dry; firm when moist; sticky and plastic when wet; diffuse, smooth boundary.
B <sub>22</sub>	(It2)	20-28	Grayish yellowish brown (10 YR 5/3 dry) to moderate yellowish brown (10 YR 4/3 moist); coarse silty clay loam; compound, moderate, medium prismatic breaking to strong medium and fine blocky; very hard when dry, firm when moist, sticky and plastic when wet; clear, wavy boundary.
C <sub>ca</sub>	(Sk)	28-48"	Light grayish yellowish brown (10 YR $6/3$ dry) to grayish yellowish brown (1 Y $4/2$ moist);

<sup>\*</sup> Horizon designations outside the parentheses are according to the Soil Survey Manual (1951). Those within the parentheses are according to the system suggested by Whiteside (1959). Color names are according to ISCC-NBS method of designating colors (Kelly and Judd, 1955).

silt loam with thick, soft white lime coats on all structure planes and fine lines of lime along root channels; compound, weak to moderate, coarse prismatic and medium and coarse blocky; hard when dry; friable when moist, slightly sticky and slightly plastic when wet; effervesces violently with dilute acid; gradual boundary.

- C<sub>2</sub> (II Pul) 48-64" Light grayish yellowish brown (1 Y 6/3 dry) to grayish yellowish brown (1 Y 5/3 moist) silt loam; a few splotches of lime as above, but the lime is mostly disseminated and in fine flecks; generally massive but with some horizontal cleavage planes and a few vertical structure planes; slightly hard when dry; friable when moist; effervesces strongly with dilute acid; very diffuse boundary.
- C<sub>3</sub> (II Pu2) 64-76" Silt loam; not visibly different from horizon above except that lime is all disseminated or in fine flecks and there are virtually no structure planes.

### B. MICROMORPHOLOGY

Examination of the soil under the binocular microscope at magnifications of 9x to 54x revealed striking changes in the microfabric of the soil body from horizon to horizon indicating that, since the deposition of the parent material, marked progress has been made by the soil forming processes. Characteristics of the individual horizons are described below.

1.  $C_2$  and  $C_3$  Horizons: The parent material is a uniform-grained, pale brown, sugary mass. Its general appearance is porous and non-shiny. Simplicity in size, shape and arrangement of the individual grains and pores is the outstanding characteristic. There is no clustering of particles. Rather, the particles form a relatively uniformly spaced lattice interlaced with an equally uniform and simple lattice of pores. Uniformity in size of both the coarse-grain fraction and the pores is striking. Those of 0.02 to 0.05 millimeters diameter predominate. There is no orientation nor stratification of silt and sand particles. The individual grains are not obviously coated. Their color and physical characteristics are readily discernable. They are angular to subangular and equidimensional to flattened or elongated. They are oriented at random, more often than not touching angle to angle or angle to side rather than side to side. The dominant minerals are quartz, feldspars, volcanic glass, calcite, micas and green, brown, and black amphiboles and pyroxenes. They are distributed randomly. Clay is not evident. Apparently, it is mixed with the non-crystalline or micro-crystalline calcium and magnesium carbonate and distributed throughout the mass. There is no hint of colloidal films. However, the scattered fine root channels are visibly lime coated and show some crowding of silt and sand particles in their walls.

2. C<sub>ca</sub> Horizon: The horizon of lime carbonate accumulation has the same generally sugary appearance as the parent material, but lime is more prominent both as coatings in fine root channels and on plane surfaces. These coatings are prominent and brilliantly white as compared to the very pale brown of the mass. Lime also coats individual sand and silt grains and in places forms bridges between them. However, the outlines of the individual grains are never masked. They are as visible as in the parent material. Incipient clustering of the grains and the development of flat ped faces is beginning. Pressure appears to have been involved in their formation, for the grains on the ped faces show orientation to expose a flat surface. However, internal shrinkage also may be an important factor in the clustering of the particles into peds. As a result of ped formation and the clustering of individual particles there is considerable variation in pore size, and the pore lattice is much more complex than in the parent material. Also there are more root channels. There is no evidence of stratification nor redistribution of any fraction.

 $B_{21}$  and  $B_{22}$  Horizons: The principle subsoil horizons are a 3. rich brown. Their structure appears as somewhat irregular blocks which fit together with many angular edges and flat surfaces. Most surfaces appear glazed, indicating the presence of clay films. In contrast to the lower horizons the sand and silt grains are embedded in a ground mass of clays and organic matter and the individual grains appear more rounded. Because of the clay coatings, identification of the minerals is difficult. The silt and sand grains are packed closer together with more faces touching faces than in the C horizons. The fabric appears more dense with fewer observable pores than the horizons below. The scarcity of silt and sand-size pores within the ped is particularly noticeable. Total pore space as calculated from volume weight and density (Table 14) is about the same as the horizon of lime accumulation and less than the parent material horizons. Thus, there must have been a considerable increase in very fine pores, which are not visible at the magnifications used, as

well as in coarse pores between peds. The observable pore space lattice is complex, being dominated by root channels, joints between peds and voids larger than the silt and sand particles.

In viewing the  $B_2$  horizons under the binocular microscope, in comparison with the C horizons, one gains the impression that there has been an inflow of a viscous fluid (the clay and organic colloids) to envelop the coarse particles and fill much of the original simple lattice of medium-sized pores within the peds. The surfaces of the peds are smoothed and the surface silt and sand grains are coated by the clay. However, the development of this clayey ground mass with the varnish-like or sirupy appearance could have resulted from the partial dispersion of the clays formed en situ or already present, following the removal of excess calcium and magnesium carbonates and soluble salts. It would seem that most of the clay leached down from the overlying horizons would be deposited on the walls of the larger pores such as those occurring between ped faces or along old root channels. Clearly the deposition of lime in the  $C_{ca}$  has been almost entirely on the walls of such coarse pores as these. The clay would not be expected to be more mobile than the lime and capable of entering a matrix that the lime could not. Since the clay films on the ped surfaces are seen to be thin, it is concluded from microscopic study of the soil that there has been only a very moderate accumulation of clay in the  $B_2$  horizons by migration from the  $A_1$  horizons.

4. A<sub>11</sub> and A<sub>12</sub> Horizons: In contrast to the subsoil the uppermost surface horizons are more coarsely porous. They, however, show

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marked granular aggregation so that the pore-space lattice is made up complexly of pores of varying size. The general appearance is that of a black, complex, sugary mass. Both organic matter and clay are in evidence, but they do not occur as an apparently continuous, varnishlike mass as in the subsoil. Rather, they appear as discrete coatings around silt and sand grains and as net-like interconnections. The appearance of the peds is light and feathery. There are numerous clean, bleached quartz grains. Either or both expansion in volume and removal of material by leaching could have been important in the fabric development. The greater concentration of fine grass roots, which may play a role in weathering by removing cations such as potassium from mineral lattices, and the more frequent wetting and drying cycles which occur in the surface horizons were likely important controlling forces in the process of fabric development.

5.  $A_{13}$  Horizon: The lower part of the surface horizon is transitional but more like the other surface horizons than the subsoil. It has a complex, weak, blocky structure made up of aggregated granules and some varnish-like clay material. On the basis of micropedological examination this subhorizon would be designated  $A_2$ .

6. <u>Thin Sections</u>: Thin sections were prepared by setting peds from each horizon in castolite, cutting, grinding and mounting on glass slides. The final sections were from 0.08 to 0.12 millimeters in thickness. Their examination under the petrographic microscope shows that the clays and other microscopic constituents envelop the coarser separates. In the case of the  $C_2$  and  $C_3$  horizons, they form discrete films joined only at the points of contact between silt and sand-sized particles, and many voids are clearly visible.

In marked contrast, the clays in the B horizons form a nearly continuous matrix in which the clustered silt and sand grains are embedded. Within peds there are few discernable voids. The relative portion of the total volume occupied by the clays is markedly greater than in the C horizons. However, strongly oriented clay films do not occur. This would indicate that clay accumulations and changes in kind have been largely the result of weathering in place. The small differences in clay content between the subsoil and the surface horizons would tend to support the hypothesis that there has been little movement of clay. Under the microscope the color of the clay in the B horizons is a rich, translucent, reddish brown.

In the A horizons, being masked with organic matter, the clay is black. It occurs as films and bridges around and between the coarser particles. The fabric of these horizons is clustered and porous. Some of the quartz sand is clean and free of films.

### C. PHYSICAL AND CHEMICAL PROPERTIES

Particle-size distribution, specific gravity, volume weight, porosity, organic matter, acid soluble constituents, pH and specific surface of the whole soil, and cation exchange capacity and total K content of the clay fractions were determined by conventional methods as described in the Chapter on Procedures. These data, like the field observable and the micropedological characteristics, would indicate that the soil is a northern Chernozem of medial development.

The particle-size distribution data (Table 2) show the clay content of the solum to be moderately higher than that of the C horizons. The A horizons have specific clay contents 9.3%, 12.3%, and 13.5% higher; the B horizons, 14.1% and 9.3% higher; and the  $C_{ca}$ , 6.9% higher than the specific clay contents of the  $C_2$  and  $C_3$  horizons. As indicated in Chapter IV it is possible that only part of this increase may be due to pedogenetic weathering. If the conclusion reached in the previous chapter that the solum and  $C_{2}$  are in one stratum of the loess and the  $C_{2}$  and  $C_{3}$ horizons in another is accepted, then only the percentage of clay above that of the  $C_{ca}$  horizon may be considered to be due to pedogenetic weathering. These percentage increases are 2.4%, 5.4% and 6.6% for the A horizons and 7.4% and 2.4% for the B horizons. It is believed that these latter figures represent the minimum possible specific clay increases due to pedogenesis and that the former set represents the maximum amounts ascribable to pedogenesis. It would thus appear, even if one accepts the more conservative figures, that at least moderate pedogenetic weathering and clay formation has occurred. The progressively higher specific increases through the A to the B2 horizon coupled with the micromorphological observations of the previous section strongly suggest that some translocation of clay also may have occurred.

The volume weight data (Table 14) show the A horizons to be

relatively lighter and the B horizons relatively heavier than the C horizons or parent material. The lightening of the surface horizons is undoubtedly due to a combination of the expansion of the silt-sand lattice by root and frost action and the loss of material such as lime, salts, oxides and clay by weathering and leaching. The increase in density of the B horizons would be due to contraction of the silt-sand lattice or to illuviation sufficient to more than offset the eluviation of such soluble materials as lime and salts. As mentioned in another section, microscopic examination of the profile suggests that both may have occurred.

Careful heating of crushed fragments of the soil to cherryred heat sufficiently long to remove the organic matter and completely oxidize the free iron but not long enough for fusion to occur did not indicate differences among the soils with respect to oxide accumulation. The carbonate-free horizons were changed to light brown (5 YR 5/4 dry) and the calcareous horizons were changed to light brown (7.5 YR 5/4dry) by this heating.

The percentage of acid-soluble material (Table 11) indicates that slightly more than three per cent of lime has accumulated in the  $C_{ca}$ as compared to an average loss of acid-soluble material of about nine per cent from the A and B horizons. Thus, it would appear that considerable lime has been removed from the profile or deposited in the parent material horizons, or the upper part of the loess was less calcareous when deposited.

The pH data (Table 7) show the soil to be neutral to alkaline. Each horizon is slightly more alkaline than the one above. The reaction

### TABLE 7

# pH, POTASSIUM CONTENT OF THE CLAY, CATION EXCHANGE CAPACITY, AND SPECIFIC SURFACE OF THE WHOLE SOIL

pН	K <sub>2</sub> O content of clay	Meq/10 Clay <sup>1</sup>	C Ogm. Whole soil2	Spe External M <sup>2</sup> /gm.	ecific Surf Internal M <sup>2</sup> /gm.	ace Total M <sup>2</sup> / gm.
7.08	2.88	55.2	12.1			
7.15	2.63	58.5	15.8	60.1	93.5	153.5
7.18	2.46	63.2	17.8			
7.26	2.23	80.3	23.2	62.0	181.4	243.4
7.53	1.63	78.4	21.1			
8.75	2.19	75.3	13.0	57.4	183.1	240.5
8.84	2.28	70.0	7.2			
8.93	2.29	73.4	5.0	54.1	145.7	199.8
	pH 7.08 7.15 7.18 7.26 7.53 8.75 8.84 8.93	pH K2O content of clay   7.08 2.88   7.15 2.63   7.18 2.46   7.26 2.23   7.53 1.63   8.75 2.19   8.84 2.28   8.93 2.29	pH $K_2 O$ content of clay $Meq/10$ Clay17.082.8855.27.152.6358.57.182.4663.27.262.2380.37.531.6378.48.752.1975.38.842.2870.08.932.2973.4	pH $K_2 O$ content of clay $Meq/100 \text{ gm.}$ Clay 1 Whole soil27.082.8855.212.17.152.6358.515.87.182.4663.217.87.262.2380.323.27.531.6378.421.18.752.1975.313.08.842.2870.07.28.932.2973.45.0	pH $K_2O$ content of clay $CEC$ Meq/100 gm. Clay1 Whole soil2 $Spe$ External M2/gm. $M^2/gm.$ 7.082.8855.212.17.152.6358.515.860.17.182.4663.217.862.07.531.6378.421.162.08.752.1975.313.057.48.842.2870.07.25.054.1	pH $K_2 O$ content of clay $CEC$ Meq/100 gm. Clay1 Whole soil2 $Specific Surf.External InternalM^2/gm.7.082.8855.212.17.152.6358.515.860.17.182.4663.217.87.262.2380.323.262.07.531.6378.421.18.752.1975.313.057.48.842.2870.07.28.932.2973.45.054.1$

l By ammunium acetate replacement.

 $^2$  By neutralization to pH 8 of whole soil minus organic matter and carbonates.

of the solum ranges from pH 7.08 in the  $A_{11}$  to pH 7.53 in the  $B_{22}$ . The  $C_{ca}$  to  $C_{3}$  horizons range from pH 8.75 to 8.93. The exchange complex of all horizons is undoubtedly at or near saturation. Certainly intense leaching is not indicated.

The cation exchange capacities of the clay fractions (Table 7) range from 55.2 to 63.2 for the A horizons; whereas they range from 70.0 to 80.3 for the B and C horizons. This would suggest that the composition of clay minerals in the surface horizons is different from those of the subsoil and parent material horizons. This suggestion is substantiated by the x-ray diffraction data discussed in a later section.

An approximation of the cation exchange capacity of the whole soil minus organic matter and carbonates was obtained during the course of mechanical analysis by the neutralization of the acidified soil to a pH of 8.0 with NaOH. The data are given in Table 7. They are relatively lower than those obtained for the clays by ammonium acetate replacement. Equating the latter to the whole soil minus organic matter and carbonates would give exchange capacities higher than those actually measured for the whole soil by about 3 meq. / 100 gms. for the solum and about 7 meq. / 100 gms. for the C horizons. It is possible that the clay separated for the analyses did not include the coarser part which would be of lower exchange capacity. If this is true the C horizons must have had a higher proportion of coarse clays than the horizons of the solum. The coarse fraction would likely be largely clay-size particles of the primary minerals such as quartz, and the feldspars, plus kaolinite and illite. Specific surface determinations were made only for the  $A_{12}$ ,  $B_{21}$ ,  $C_{ca}$  and  $C_{3}$  horizons.<sup>1</sup> These data (Table 7) indicate an increase in the specific external surface of the solum over that of the parent material. They show a marked increase in the specific internal surface of the  $B_{21}$  and  $C_{ca}$  horizons and a decrease in that of the  $A_{12}$  over that of the underlying material ( $C_{3}$ ). As shown graphically in Figure 3, the internal specific surface area is closely correlated with the cation exchange capacity of the clay. On the other hand, the external surface area shows a poor correlation with cation exchange capacity but a good correlation with per cent of clay.

These relationships would support the validity of using cation exchange capacity as a basis for quantitatively estimating clay-mineral composition of a soil as suggested by Ormsby and Sand (1954). It is believed that cation exchange capacity data and total K-content of the clay taken together provide the most reliable indication of the quantitative clay mineralogy of the soil studied The x-ray diffraction patterns, of course, are essential for identifying the minerals present.

### D. MINERALOGY OF THE SILTS AND SANDS

Examination of the sand and the coarse silt under the polarizing microscope indicates that the mineral suites of all horizons are qualitatively alike. Quartz, the feldspars, and volcanic glass constitute

<sup>&</sup>lt;sup>1</sup> Thanks is given to Raymond T. Choriki for assistance in making these determinations.



FIGURE 3: THE RELATION OF EXTERNAL AND INTERNAL SPECIFIC SURFACE TO PERCENT OF CLAY AND CATION EXCHANGE CAPACITY OF THE CLAY FRACTION

more than ninety per cent of the suites; these, plus the amphiboles, micas and chlorite constitute much more than ninety-eight per cent. The minerals identified in their approximate order of abundance are quartz, feldspars (oligoclase, andesine, labradorite, orthoclase, microcline, microperthite), volcanic glass, amphiboles (green hornblende, brown hornblende, actinolite, tremolite), micas (muscovite, biotite), chlorite, pyroxenes, olivine, epidote, garnet, zircon, and rutile. Qualitatively, there appeared to be no differences among the suites in the fine sand, very fine sand and coarse silt, fractions.

The results of heavy mineral separations and mineral counts on the heavy fractions have already been given in Tables 5 and 6.

The percentages of heavy minerals in the horizons above twenty-one inches are less than those in the horizons below this depth. Analysis of variance shows the differences to be significantly greater than the variation between duplicate analyses. It is probable that the differences are largely due to differential weathering during the process of soil formation. The calculations of quantitative changes that have occurred during pedogenesis, which are discussed later, show that the depth of significant clay formation is 21 inches. The percentage of coated and opaque minerals would indicate that at least some weathering has occurred through the  $C_2$  horizon. The frequency counts of specific heavy minerals do not indicate differential weathering. It is possible that the differences in heavy minerals reflect variations in the rate of deposition as much as weathering.

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The relative content of quartz as determined by x-ray diffraction is shown in Table 8. It is relatively higher in the A and B horizons than in the C horizons. Thus, if the quartz content of the entire profile was equal at the starting time of soil formation and quartz has not weathered appreciably, then both the surface soil and subsoil have suffered a net loss of material during the process of soil formation. These losses would presumably be the lime plus soluble salts left over in the transformation of primary minerals to clays. Some of this loss would be offset by the addition of organic matter, which would at least partially account for the smaller net loss of material from the surface soil horizons. Apparently, in the  $C_{ca}$  the accumulation of lime has just offset the loss due to clay formation and the removal of soluble salts by leaching, or there was stratification in the deposition of the loess.

Because of the wide differences of peak intensities of the different feldspars in the 3. 15 Å - 3. 25 Å range the estimates of relative feldspar content (Table 9) are only approximations. They are based on measurements of the total areas under the curves between 27. 0° - 28. 6° two theta (d = 3. 31 Å - 3. 14 Å). Nevertheless, the data appears reasonably consistent. Only the  $C_2$  horizon appears out of line. Based on the percentage of residual constituents the estimate of the feldspar content of the appears to be low for this horizon. If the estimate for the  $C_3$  horizon can be taken as the proper estimate of the feldspar content of the parent material the data would indicate marked weathering of the feldspars during soil formation. It is interesting to note that if the  $C_2$ 

TABLE 8

PERC	CENT (	DF (	QUARTZ	IN WHC	LE SOII	BY X	-RAY	DIFFR	ACTION	JANAL	YSIS
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			X- ray i	Intensities		Per cent
	Depth	Actu	al	Adju:	sted	quartz
Horizon	(inches)	Qtz. std.	Soil	Qtz. std.	Soil	in soil <sup>1</sup>
Α.,	0-4	3616	1136	3840	1206	
11		3648	1184	3840	1246	
		3616	1152	3840	1223	
					1225	25.0%
A,,,	4-8	3776	1264	3840	1285	
12		3840	1328	3840	1328	
		3808	1200	3840	1210	
					1274	26.4
A, 2	8-13	3776	1216	3840	1237	
15		3744	1264	3840	1296	
		3776	1312	3840	1334	
					1289	26.7
B <sub>21</sub>	13-21	3712	1344	3840	1390	
21		3712	1296	3840	1340	
				3840	1130	
					1286	26.6
B <sub>22</sub>	21-28	3904	1328	3840	1306	
66		3872	1280	3840	1269	
		3872	1344	3840	1333	
					1303	27.0
C	28-48	3776	1248	3840	1269	
Ca		3840	1072	3840	1072	
		3808	1136	3840	1146	
					1162	23.5
c,	48-64	3968	1184	3840	1146	
-		3968	1168	3840	1130	
		3968	1248	3840	1208	
					1161	23.5
C2	64-76	3744	1248	3840	1280	
-		3872	1104	3840	1095	
		4000	1152	3840	1106	
					1160	23.5

<sup>1</sup>Determined graphically from Figure 6.

## TABLE 9

	Denth	Adjı	isted peak-a	rea intensi	ties <sup>1</sup>	Estimated
Horizon	(inches)		-Readings		Average	feldspars <sup>2</sup>
A <sub>11</sub>	0-4	0.32	0.40	0.38	0.37	29%
A <sub>12</sub>	4-8	0.32	0.27	0.29	0.30	25
A <sub>13</sub>	8-13	0.31	0.20	0.33	0.28	24
B <sub>21</sub>	13-21	0.31	0.31	0.30	0.31	26
в <sub>22</sub>	21-28	0.38	0.36	0.32	0.35	28
C <sub>ca</sub>	28-48	0.31	0.34	0.39	0.35	27
c <sub>1</sub>	48-64	0.24	0.28	0.34	0.30	25
c <sub>2</sub>	64-76	0.51	0.48	0.34	0.44	33

## ESTIMATED PER CENT OF FELDSPARS IN THE WHOLE SOIL BY X-RAY ANALYSIS

<sup>1</sup>Adjusted to a quartz standard peak-height reading of 3480.

<sup>2</sup>Determined graphically from Figure 7.

horizon is dropped out, there is a good correlation between the contents of feldspars and heavy minerals, and an approximate inverse correlation between the contents of feldspars and quartz.

The x-ray diffraction estimates of calcite and dolomite are shown in Table 10. In comparison with the content of acid soluble material (Table 11) the estimates appear to be somewhat low. This would indicate that the major portion of the calcium and magnesium carbonate is not well crystallized.

The relative mineralogical composition of each horizon is summarized in Table 11. Although it is admitted that they involve some approximations, it is believed that the data reflect the relative differences and similarities between horizons. The unidentified fraction of the silts and sands may be in part such amorphous materials as volcanic glass. However, much of it is likely due to underestimation of the feldspar content. It will be noted in the case of the  $C_2$  horizon, for example, that the marked increase in unidentified minerals is accompanied by a marked decrease in feldspars relative to the  $C_3$  horizon.

In general, the mineralogical data indicate a relative decrease in the weatherable fractions from the C horizons toward the surface of the soil. Inversely, there is an increase in quartz and clays. These trends could be taken to indicate that weathering and possibly some translocation of silicate minerals have occurred. Clearly, of course, marked translocation of acid-soluble material has occurred, and the organic matter has tended to remain at the depth of synthesis.

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#### TABLE 10

#### ESTIMATED PER CENT OF CALCITE AND DOLOMITE IN WHOLE SOIL BY X-RAY DIFFRACTION ANALYSIS

			Adjust	ed peak-	area inte	ensities	
	Depth (inches)	Mineral		Readings		Average	Estimated per cent <sup>2</sup>
A.,	0-4	Calcite				0.00	
11		Dolomite				0.00	
A.,.	4-8	Calcite				0.00	
12		Dolomite				0.00	
A, 2	8-13	Calcite				0.00	
15		Dolomite				0.00	
B.,	13-21	Calcite				0.00	
21		Dolomite				0.00	
B.,	21-28	Calcite				0.00	
22		Dolomite				0.00	
С	28-48	Calcite	0.16	0.19	0.18	0.18	4.5
ca		Dolomite	0.13	0.10	0.10	0.11	3.5
С,	48-64	Calcite	0.12	0.14	0.17	0.14	3.5
1		Dolomite	0.08	0.09	0.12	0.10	3.0
C <sub>2</sub>	64-76	Calcite	0.23	0.16	0.14	0.18	4.5
2		Dolomite	0.11	0.08	0.09	0.09	2.5

 $^{1}\,\mathrm{Adjusted}$  to a quartz standard peak-height reading of 3840.

<sup>2</sup>Determined graphically from Figure 8.

A COLOR

TABLE 11

Relative mineralogical composition of a profile of bozeman silty clay  $\operatorname{Loam}^1$ 

		Aci	id Soluble		1 1 1 1 1		Sands a	nd Silts			
Hor-	Organic matter total per cent	Total per cent	Calcite per cent	Dolomite per cent	Total per cent	Heavy minerals per cent <sup>2</sup>	Inter- mediate minerals per cent <sup>3</sup>	Feld- spars per cent	Quartz per cent	Un- iden- tified per cent	Clays total per cent
A	6.4%	8.4%			61.0%	3.4%	1.1%	29.0%	25.0%	2.5%	24.2%
A12	3.1	9.0	-	1	60.2	2.6	1.3	25.0	26.4	4.9	27.7
Ala	1.8	6.4	-		61.7	3.1	1.8	24.0	26.7	6.1	30.1
B	0.7	5.0	1	{	62.8	3, 2	0.9	26.0	26.6	6.1	31.5
B22	0.4	4.9	-	1	67.5	4.5	1.1	28.0	27.0	6.9	27.2
	0, 1	19.1	4.5	3.5	59.6	4.0	1.1	27.0	23.5	4.0	21.2
c,	0, 1	15.6	3.5	3.0	68.0	4.7	1.5	25.0	23.5	13.3	16.3
	0, 2	16.3	4.5	2.5	67.4	4.9	1.3	33.0	23.5	4.7	16.1
	-										

<sup>1</sup>Oven-dry whole soil considered as 100%.

<sup>2</sup>Density 2.95. Calculation is based on assumption per cent in total sand = per cent in very fine sand and per cent in total silt = per cent in coarse silt. <sup>3</sup>Density 2.75 - 2.95. Calculation is based on assumption per cent in total sand = per cent in very fine sand, and per cent in total silt = per cent in coarse silt. 69





#### FIGURE 4

## X-RAY DIFFRACTION CURVES FOR THE CLAY FRACTION OF EACH HORIZON AFTER SODIUM SATURATION AND GLYCERATION, AND AFTER POTASSIUM SATURATION AND HEATING AT 110° C., 300° C., AND 575° C.





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### E. MINERALOGY AND CHARACTERISTICS OF THE CLAYS

X-ray diffraction intensities together with potassium contents, cation exchange capacities and specific surfaces of the fraction less than two microns in diameter indicate that the clay minerals in Bozeman silty clay loam are illite, montmorillonite, kaolinite, and possibly halloysite. The first two, illite and montmorillonite, are the dominant minerals. They are distributed throughout the profile. Kaolinite is mostly in the parent material and subsoil horizons; whereas, halloysite is mostly in the surface soil. The ratios of intensities between the 7.2 Å, 3.6 Å, and 3.3 Å basal spacings clearly show that the relative proportions of clay minerals are different by master horizons. Indications are that illite and halloysite are increasing in the A horizons and that montmorillonite is accumulating in the  $C_{ca}$ . There is an increase of montmorillonite in the B<sub>2</sub> horizons over the A horizons but not relative to the C horizons. The A horizons exhibit a markedly higher degree of random interstratification than do the B and C horizons.

The possible explanation might be that the new clay being formed is largely illite and that montmorillonite is moving downward into the subsoil. Hydrated halloysite is apparently forming in the surface soil at the expense of kaolinite.

1. <u>X-Ray Identification</u>: Traces of the x-ray diffractometer intensity curves for oriented, clay films from each horizon after various treatments are given in Figure 4 (a-h).



It will be noted that, in the case of the surface soil horizons  $(A_{11}, A_{12}, A_{13})$ , the only clearly identifiable peaks on the traces of the glycerated Na-clays are at 10  $\overset{\circ}{A}$  and 3.3  $\overset{\circ}{A}$ , and these are rather broad. There are distinct rises in the curves at 3.6 Å but no peaks except for the A<sub>13</sub> horizon. The traces are somewhat irregular at less than 8<sup>°</sup> 2  $\theta$ particularly for the  $A_{1,3}$  horizon. The height of the curves in this arc are considerably higher than would be expected by purely background reflections. After saturation with potassium and heating to  $110^{\circ}$  C. for two hours, the traces are very irregular between  $2^{\circ}$  to  $8^{\circ}$  2  $\theta$  with indications of peaks at 14 Å. The 10 Å and 3.3 Å peaks are somewhat more intense, and small but distinct peaks occur at 7.2  $\stackrel{\circ}{A}$  and 3.6  $\stackrel{\circ}{A}$ . After heating for an additional two hours at  $300^{\circ}$  C., the background portions of the curves are much less irregular, and clear, sharp peaks occur at 10  $\stackrel{o}{A}$ , 7.2  $\stackrel{o}{A}$ , 3.6 Å, and 3.3 Å. Also the peak at 5 Å is clearer but still rather broad. The peaks at 10  $\stackrel{\circ}{A}$  and 3.3  $\stackrel{\circ}{A}$  are the most prominent. After heating for two hours more at 575  $^{\circ}$  C., the 7 Å and 3.6 Å peaks disappear but the others remain as before.

In accordance with the interpretations proposed by Johns and Grim (1954) the occurrence of the 10 Å peak in the traces from the glycerated Na-clays would indicate that one of the dominant clay minerals in the surface soil horizons is illite. The lack of resolution of an 18 Å peak coupled with the occurrence of a peak at 3. 6 Å in the traces of the glycerated Na-clays, and the development of a sharp increase in the 10 Å peaks on heat treatment, would indicate that montmorillonite and possibly

other expanding lattice clays, are present but largely in randomly stacked mixtures. The appearance of a 7.2 Å peak coupled with the continued absence of a 14 Å peak after heat treatment, would indicate that a small amount of hydrated halloysite is present.

The subsoil and parent material horizons show markedly different x-ray diffraction intensity curves. The glycerated Na-clays have very strong 18 Å peaks; broad, sometimes double or triple peaks ranging between 9 Å and 10 Å; weak but distinct 7.2 Å peaks; rather broad multiple peaks at 4.5 Å to 5.0 Å; and strong peaks at both 3.6 Å and 3.3 Å. After saturation with potassium and heating for two hours at 110° C., the 18 Å peaks shift to the range 14.3 Å to 13.2 Å. Upon further heating at 300° C., these peaks shift to a sharp 10 Å, and the peaks at 5.0 Å and 3.3 Å become sharper and more intense. The peaks at 7.2 Å and 3.6 Å are somewhat diminished but remain clear and sharp. However, upon heating at 575° C., these latter two peaks disappear completely.

In accordance with the interpretations proposed by Johns and Grim (1954), the occurrence of the strong 18 Å peaks which collapse completely to 10 Å on K-saturation and heating, are interpreted as indicating a significant content of montmorillonite. The presence of broad, multiple peaks at 9 Å to 10 Å together with clearly defined peaks at 3. 3 Å in the glycerated clay are taken to indicate that illite is an important component. The absence of 14 Å peaks and the presence of 7. 2 Å and 3. 6 Å peaks which persist after the 300° C. heat treatment are interpreted to indicate that kaolinite is a minor component. In the absence of clearly defined, persistent peaks at 4.26 Å and in the 3.15 Å - 3.25 Å range it is believed that little if any clay-size quartz or feldspars occur in the fractions x-rayed. Since the peaks for the various basal spacings of the minerals indicated are clear and sharp, it is believed that randomly interstratified lattices are of relatively minor extent compared to the A horizons. This would indicate that greater formation or alteration of clays by pedogenic forces has occurred in the surface soil than in the subsoil.

It is concluded that in the B and C horizons, species of illite and montmorillonite are about equally dominant and that kaolinite occurs in small amounts. Hydrated halloysite appears to be absent.

As is described in Chapter III, Procedures, and discussed in the Appendix, an attempt was made to make quantitative measurements of the clay minerals using the scaler of the x-ray diffractometer. The net intensities measured for the soil clays are given in Table 12 a. Various ratios between intensities are given in Table 12 b. The data tend to substantiate the conclusions drawn from the strip-chart curves. It is interesting to note that many of the ratios indicate similarity in clay mineralogy of the sub-horizons within a master horizon but differences in the clay mineralogy between master horizons.

The following relationships and conclusions are indicated by the x-ray intensity counts:

### TABLE 12

#### X-RAY DIFFRACTION INTENSITIES AND RATIOS OF INTENSITIES FOR THE CLAY FRACTION

	Gly	cerated N	a-clay	K-clay	heated t	o 300° C.
Sample	7.2 Å	3.6 Å	3.3 Å	7.2 Å	3.6 Å	3.3 Å
A <sub>11</sub>	18	360	928	191	500	1,983
A <sub>12</sub>	23	532	958	154	376	2,014
A <sub>13</sub>	37	533	973	117	392	2,110
B <sub>21</sub>	149	661	769	215	347	2,108
B <sub>22</sub>	191	812	682	235	396	2,076
Cca	149	761	563	173	368	2,005
C 2	185	546	584	311	467	1,720
C <sub>3</sub>	171	462	549	307	400	1,714

a. X-Ray Diffraction Intensities

b. Ratios of Intensities

	)	Heated	K-clay		Glyc	erated Na	-clay	)H	eated K-c]	ay
Horizon				3.6 Å	3.6 Å	7.2 Å	3.6 Å	3.6 Å	7.2 Å	3.6 A
	7.2 Å	3.6 Å	3. 3 Å	+ 3, 3 Å	7.2 Å	3, 3 A	3.3 Å	7.2 Å	3, 3 Å.	3.3 Å
A	10.61	1.39	2.14	1.93	20,00	。02	。39	2.62	. 10	. 25
A12	6.70	.71	2.10	1.60	23, 12	。02	。56	2.44	. 08	. 19
A13	3,116	。74	2,17	1.66	14.41	。04	. 55	3, 35	, 06	. 19
B <sub>2.1</sub>	1.44	. 52	2.74	1.72	4.44	. 19	. 86	1,61	.10	. 17
B <sub>22</sub>	1, 123	. 49	3, 04	1.66	4, 25	. 28	1.19	1.69	. 11	。19
Cca	1, 16	。48	3.56	1.79	5.11	. 27	1.35	2.13	. 09	. 18
C <sub>2</sub>	1.68	. 86	2.94	1.94	2.95	。32	。94	1.50	. 18	. 27
°,	1,80	。73	3, 12	2.09	2.70	。31	。84	1.30	。18	. 23

a. The intensities of the 7.2 Å spacing and the ratios of the heated K-clays to the glycerated Na-clays show that the A horizons differ from the B and C horizons. The high ratios of the A horizons differ from the B and C horizons. The high ratios of the A horizons and the low ratios of the B and C horizons tend to substantiate the conclusion that the surface soil contains hydrated halloysite and little kaolinite, whereas the subsoil and parent material contain kaolinite and very little halloysite.

b. The intensities and ratios of the 3.3 Å spacing indicate an increase in illite in the surface soil relative to the subsoil and parent material.

c. The ratios between the different spacings of the glycerated Na-clays suggest differences in clay mineralogy among the three master horizons and similarities among the subhorizons within a master horizon.

2. Total Potassium Content of the Clays: The total  $K_2O$  contents of the clays (Table 7) range from 1.63% in the  $B_{22}$  to 2.88 in the  $A_{11}$ . They would indicate that the surface soils are higher in illite than the subsoil and underlying horizons. If the percentage contents of illite in the clays are calculated on the assumption that illite has a  $K_2O$  content of 6.09% (Grim, 1953), the A horizons are shown to have 47%, 43% and 40% illite; the B horizons, 37% and 27% illite; and the C horizons, 36%, 37% and 38% illite.

These percentages show a significant correlation (r = .8338) with the cation exchange capacities of the clays. This relationship is shown graphically in Figure 5. The equation for the curve shows the



FIGURE 5: RELATION OF PERCENT ILLITE CALCULATED FROM K20 CONTENT (100 K20/6.09) TO CATION EXCHANGE CAPACITY OF THE CLAY

rate of change in the cation exchange capacity to be 131 meg. per 100% change in illite. Since the x-ray diffraction data show that the contents of illite and montmorillonite in the clavs tend to vary inversely and the other components are of minor extent and relatively constant from horizon to horizon, this rate of change can be considered to be the difference between the cation exchange capacities of the illite and montmorillonite components. In other words the cation exchange capacity of montmorillonite would be 131 plus that of illite. If that of illite is 20, it would be 151 and 79% of montmorillonite would satisfy an exchange capacity of 119 which is that indicated for zero per cent illite. Seventy-eight per cent would leave an exchange capacity of 1.5 meq. for the kaolinite component and 22% as the content of kaolinite. The specific cation exchange capacity of this kaolinite would be seven meq. per 100 grams of clay. If the exchange capacity of illite were ten that of montmorillonite would be 141 and 83% would satisfy all but 1.5 meq. of the total exchange capacity of 119 for clay without illite.

Thus, when correlated with cation exchange capacities, a  $6.09\% \text{ K}_2\text{O}$  content for illite appears to give a satisfactory estimate of the relative clay mineralogy of the clay fractions of the various horizons of the profile being studied. If ten per cent were used as the total  $\text{K}_2\text{O}$  content of illite, the change in cation exchange capacity would be at the rate of 217 meq. per 100% change in illite and one would be required to assume an exchange capacity of some 230 to 240 meq. per 100 grams of clay for montmorillonite.

3. <u>Discussion of the Data</u>: On the basis of the illite/cation exchange capacity curve (Figure 5), the x-ray diffraction data (Table 12 and Figures 4), the specific surface data (Table 7), and considering that the clay fractions separated for study may not have included all of the coarse clay, the best approximations of relative clay mineralogy of the profile of Bozeman silty clay loam would be as given in Table 13. This table indicates that the clays of the surface soil horizons are largely in mixedlayer form, and that there has been a relative gain of illite in the A horizons and of montmorillonite in the B horizons as compared to the C horizons.

To summarize, apparently the principal changes which have occurred in the clay fraction of the solum during the course of soil formation have been (1) a relative gain of illite in the A horizons, (2) a relative gain of montmorillonite in the B horizons, (3) a reconstitution of much of the clay of the A horizons to randomly interstratified, mixedlayer forms, and (4) an increase of total clay by formation.

The shifts in relative illite and montmorillonite contents of the A and B horizons are likely the result of differential leaching. The montmorillonite, being finer, would be more mobile and more readily leached.

Differential leaching alone, however, would not account for the marked shift to randomly interstratified clays in the surface horizons. Some of it may be due to the formation of new types of clay minerals, such as halloysite, which have become randomly interstratified with the

#### TABLE 13

Α.	PPR	OXIN	IATE I	RELAT	IVE	CLAY .	MIN	ERALO	GY.	ESTIMAT	ЕD
	ON	THE	BASIS	OF CH	EC/K	RATIC	DS,	X-RAY	DIF	FRACTIC	N
			CURV	ES AN	D SPE	CIFIC	SU	RFACE	DAT	ГA	

Horizon	Illite	Mixed layer <sup>1</sup>	Montmor- illonite	Halloy- site	Kaolinite	Un- known <sup>2</sup>
	per cent	percent	per cent	per cent	per cent	per cent
A <sub>11</sub>		50-25		10		15
A <sub>12</sub>		45-30		10		15
A <sub>13</sub>		45-30		10		15
B <sub>21</sub>	30		45		10	15
B <sub>22</sub>	30		45		10	15
Cca	35		35		10	20
C <sub>1</sub>	35		30		10	25
C2	35		30		10	25

 $^{1}\,{\rm First}$  number indicates illite part and the second number indicates montmorillonite part.

<sup>2</sup>Probably largely quartz, feldspars and other primary minerals of clay size. Allowance is made for the part of the coarse clay lost during separation.

illite and montmorillonite. Much of it must be due to an active cycle of potassium recirculation. This may be in part an equilibrium reaction between the clay minerals. At least it would seem logical to suppose that random stratification could occur by the addition and subtraction of binding potassium between the lattice layers as experiments of Wear and White (1951) and Mortland et al. (1956) indicate is possible. Both addition and subtraction would need to have occurred in this case as the clays of the surface horizons are so completely interstratified that both the original illite and montmorillonite must have suffered modification. For this to be possible there would need to be an active biological potassium cycle. Potassium extracted by plant roots from one interlayer would have to be replaced in another by potassium released by the mineralization of organic matter or by the weathering of primary minerals. The alternate wetting and drying, the moderately high content of potash-rich minerals. the complete permeation by plant roots, and the relatively low rate of leaching which are characteristic of the environment in which the surface horizons have developed, would be conducive to the cyclic transfer of potassium.

The marked changes in the character of the clay minerals of the surface horizons are strong evidence that the higher clay content of the solum in comparison with the underlying horizons is due at least in part to pedogenesis. Were it not for such evidence it would be difficult to show that the variation in texture was not due to changes which occurred during the deposition of the loess.



#### F. QUANTITATIVE PEDOGENETIC CHANGES

Considering that field and microscopic study of the profile indicate that during the course of soil formation marked changes have occurred in both the macro and micro arrangements of the individual mineral particles and that other changes have occurred, resulting in the development of characteristic structure, texture and color profiles, it appears of interest to calculate if possible the extent to which chemical and mineralogical changes have occurred and whether there may have been translocation of materials from horizon to horizon.

This is possible, as suggested by Marshall and Haseman (1942) and described in greater detail by Barshad (1955), if it can be assumed that the material in which the solum has developed was originally similar to the underlying C horizon or varied in a known way. The calculations are based on the ratios in the various horizons between the quantities of the factor being analyzed and some index substance such as quartz which is believed to have resisted transformation or translocation during soil formation and can be related to the C horizon. The calculations are parallel in form to the gravimetric factor of quantitative chemical analysis.

As discussed in Section D of Chapter V, it is believed that the  $C_{ca}$  horizon has a silicate mineral composition similar to that of the primary material of the solum at the beginning of pedogenesis, and may safely be used as the horizon of reference in calculating quantitative pedogenetic changes. In using it to represent the original primary material of the solum, it is assumed that there has been no addition to the clay either by transformation or illuviation. The validity of this assumption is supported by the freshness of the surfaces and the low percentage of coated grains in the uncleaned silt and sand fractions of this horizon.

Calculations of changes based on the C<sub>ca</sub> as representing the original primary material of the solum will give minimum values.

To calculate the percentage content of a constituent originally in a horizon the equation is:

$$w'_{x} = w_{x} \cdot \frac{r_{x}}{r_{p}}$$

where  $w_x^{T}$  is the original percentage content of the constituent in horizon X,  $w_x$  is the present percentage content of the constituent in horizon X,  $r_x$  is the percentage content of the index resistant mineral in horizon X, and  $r_p$  is the percentage content of the index resistant mineral in the C horizons or assumed parent material.

To calculate the weight of the constituent per unit crosssection of horizon the present volume weight and thickness of the horizon is also needed. The equation in this case becomes:

$$W'_{x} = \frac{W_{p}}{100} \cdot V_{x} \cdot T_{x} \cdot \frac{R_{x}}{R_{p}}$$

Where  $W'_x$  is the original weight in grams of the constituent W in horizon X per cm<sup>2</sup> of cross-section,  $W_p$  is the percentage of the constituent W in



the assumed parent material,  $V_x$  is the volume weight of horizon X,  $T_x$ is the thickness of horizon X in centimeters,  $R_x$  is the percentage of the resistant index mineral in horizon X, and  $R_p$  is the percentage of the resistant index mineral in the assumed parent material horizon. In the above equation  $V_x$ <sup>\*</sup>  $T_x$  is equal to the present weight of the soil per horizon per cm.<sup>2</sup> and  $R_x/R_p$  may be considered the weight change factor. Knowing the original weight  $W'_x$  and the present weight  $W_x$  of the constituent W in horizon X, it is a simple matter of subtraction to calculate the change which has occurred in constituent W during the course of soil formation, the equation being:

$$C_x = W_x' - W_x$$

where  $C_{x}$  is the loss or gain of constituent W in horizon X.

In practice the calculations consist of:

- a. Calculating the weight change factor or resistant index mineral ratio  $R_{\rm v}/R_{\rm p}.$
- b. Calculating the present weight of the soil horizon V . T.
- c. Calculating the original weight of the soil horizon by multiplying the weight change factor times the present weight of the soil horizon.
- d. Calculating the original volume of the soil horizon by dividing the original weight of the horizon by the volume weight of the C horizon or assumed parent material.
- Calculating the original weight of specified constituents, such as clay, per horizon per cm.<sup>2</sup> of cross-section by multiplying



its percentage in the horizon times the present weight of the horizon.

- g. Calculating the gains or losses by subtracting the present weights from the original weights.
- Calculating the relative gains or losses in terms of appropriate standards such as 100 grams of original parent material, 100 grams of clay, 100 cc. of parent material, etc.
- i. Calculating the ratio of clay gained to silts and sands lost in each horizon. Since the clay formed is the product of silt and sand weathering a comparison of this last ratio among the various horizons should indicate whether there has been translocation of clay.

In this study quartz has been used as the resistant index mineral. Calculations have been made for the silts and sands, the clays, the acid soluble material, soil weight, and soil volume. The results are given in Table 14.

It should be understood that the calculated volume changes represent only the differences in change relative to the  $C_{ca}$  horizon. This horizon is believed to have settled during pedogenesis. Relative to the  $C_{ca}$  the A horizons have gained and the B horizons have lost in specific volume. It is to be noted that, had there been no loss of material during the course of soil formation, with the same volume weight changes the relative increases in the volumes of the A horizons would have been approximately 64, 22, and 15 per cent instead of 46, 6 and -1; thus

#### TABLE 14

# QUANTITATIVE EVALUATION OF PEDOGENESIS IN A PROFILE OF BOZEMAN SILTY CLAY LOAM

Horizon	Depth (inches)	Volume weight gms./cc	Specific gravity	Porosity per cent	Quartz per cent	Weight change factor <sup>1</sup>
A <sub>11</sub>	0-4	0,83	2,45	66.1	25.0	1.064
A <sub>12</sub>	4-8	1.08	2.50	56.8	26.4	1.123
A <sub>13</sub>	8-13	1.14	2,50	54.4	26.7	1,136
B <sub>21</sub>	13-21	1,31	2,55	48.6	26.6	1.132
B <sub>22</sub>	21-28	1,28	2,55	49.8	27.0	1.149
C <sub>ca</sub>	28-48	1.29	2.55	49.4	23.5	1.000
C <sub>2</sub>	48-64	1.24	2.60	52.3	23.5	1.000
C <sub>3</sub>	64-76	1,25	2,60	51.9	23.5	1.000

a. Physical Constants

 $^1$ Per cent quartz of horizon/per cent quartz of parent material. This ratio equals grams of parent material required to contain the same amount of quartz as one gram of the horizon.

### b. Weight Changes

Hor- izon	Original weight per hor.	Present weight perhor. with OM. gms. / cm	Change in weight per hor. with OM.	Relative change with OM. <sup>1</sup> r	Present weight perhor. ninus OM	Change in weight per hor. . minus O . / cm <sup>2</sup>	Relative change minus M OM <sup>1</sup>
A <sub>11</sub>	9.0	8.4	-0.6	-6.6	7.9	-1.1	-12.7
A <sub>12</sub>	12.3	11.0	-1.3	-10.9	10.7	-1.6	-13.3
A <sub>13</sub>	16.5	14.5	-2.0	-12.0	14.2	-2.3	-13.7
B <sub>21</sub>	30,1	26.6	-3.5	-11.6	26.4	- 3. 7	-12.2
B <sub>22</sub>	26,2	22.8	-3.4	-13.0	22.7	- 3. 5	-13.4
A 1	37.8	33.9	- 3. 9	-10.3	32.8	-5.0	-13.3
B <sub>2</sub>	56.3	49.4	-6.9	-12.3	49.1	-7.2	-12.8
Solum	94.1	83.3	-10.8	-11.5	81.9	-12.2	-13.0

 $^{\rm l}{\rm Grams}$  per 100 grams of original parent material.

## c. Relative Volume Changes

Horizon	Depth (inches)	Original volume per hor. cc/cm <sup>2</sup>	Present volume per hor, cc/cm <sup>2</sup>	Change in volume per hor. cc/cm <sup>2</sup>	Relative change in volume
A <sub>11</sub>	0-4	7.0	10.2	+3.2	+45.7
A <sub>12</sub>	4-8	9.5	10.2	+0.6	+ 6.3
А <sub>13</sub>	8 - 1 3	12.8	12.7	-0.1	- 0.8
B <sub>21</sub>	13-21	23.4	20.3	- 3. 1	-13.2
B <sub>22</sub>	21-28	20.3	17.8	-2.5	-12.3
A <sub>1</sub>	0 - 1 3	29.4	33. 1	+3.7	+12.6
B <sub>2</sub>	13-28	43. 7	38. 1	- 5. 6	-12.8
Solum	0-28	73.1	71.2	-1.9	- 2.6

 $^{1}$  cc/100 cc of the present volume of the C horizon.



# d. Silt and Sand Weathering

Horizon	Depth (inches)	Original silt-sand per hor. gms./cm <sup>2</sup>	Present silt-sand per hor. gms./cm <sup>2</sup>	Loss of silt-sand per hor. gms./cm <sup>2</sup>	Relative loss silt-sand <sup>1</sup>
A <sub>11</sub>	0-4	5.3	5.1	0.2	3.8
A <sub>12</sub>	4-8	7.3	6.6	0.7	9.6
A <sub>13</sub>	8-13	9.8	8.9	0.9	9.2
B <sub>21</sub>	13-21	18.0	16.7	1.3	7.2
B <sub>22</sub>	21-28	15.6	15.4	0.2	1.3
A <sub>1</sub>	0-13	22.4	20.6	1.8	8.0
B <sub>2</sub>	13-28	33.6	32.1	1.5	4.5
Solum	0-28	56.0	52.7	3.3	5.9

<sup>1</sup>Grams per 100 grams of silts and sands originally in the parent material.

# e. Clay Formation

Horizon	Depth (inches)	Original clay per hor. gms./cm <sup>2</sup>	Present clay per hor. gms./cm <sup>2</sup>	Gain of clay per hor. gms./cm <sup>2</sup>	Relative gain of clay <sup>1</sup>	Ratio of clay gain to silt-sand loss
A <sub>11</sub>	0-4	1.9	2.0	0.1	1.9	0.50
A <sub>12</sub>	4-8	2.6	3.0	0.4	5.5	0.57
A <sub>13</sub>	8-13	3. 5	4. 4	0.9	9.2	1.00
в <sub>21</sub>	13-21	6.4	8.4	2.0	11.1	1.54
в <sub>22</sub>	21-28	5.5	6.2	0.7	4. 5	3.46
A <sub>1</sub>	0 -1 3	8.0	9.4	1.4	6.3	0.79
B <sub>2</sub>	13-28	11.9	14.6	2.7	8.0	1.78
Solum	0-28	19.9	24.0	4. l	7.3	1.25

Grams per 100 grams of silts and sands originally in the parent material.

emphasizing the importance of root, frost, wetting and drying, and animal action in loosening and making more porous the surface soil. Loss of weight is largely responsible for the loss of volume of the B<sub>2</sub> horizons. The small increase in volume weight in the B horizons would suggest that a little clay may have been added by illuviation.

The calculated relative weight changes through the losses of mineral matter from the soil column are approximately equal through the B<sub>22</sub> horizon; the average relative loss being 13.0 grams per 100 grams of original parent material.

Of primary interest to a study of the trends in soil formation are the transformations which have occurred with respect to clay. Calculations for the profile of Bozeman silty clay loam being studied indicate, as shown in Table 14e, a gain of clay in the solum of 4. l grams over that of an equivalent amount of  $C_{ca}$  horizon or assumed parent material. This is 1.25 times as great as the decrease in silt and sand. The relative gains in clays per 100 grams of silts and sands in the assumed parent material increase from 1.9 grams in the  $A_{11}$  to 11. l grams in the  $B_{21}$  and then decrease to 4.5 grams in the  $B_{21}$ . These are net gains, the resultants of clay formation and clay movement.

The calculation of the probable amount of clay formed in each horizon based on the assumption that the ratio between clay formation and silt-sand loss should be the same in each horizon, indicates that the relative rate of clay formation is the highest in the lower A horizons. However, the differences through the  $B_{21}$  are small. The rate of formation

definitely drops in the  $B_{22}$  horizon. The ratios of clay gain to silt and sand lost suggest that there has been some clay movement from the A to the B horizons. According to the calculations, about a fourth of the clay gain in the  $B_{21}$  and about two-thirds of the clay gain in the  $B_{22}$  are due to translocation from the A horizons. The clay moved undoubtedly would be from the finer fraction and mostly montmorillonite. This would partially explain the increased cation exchange capacity of the  $B_2$  horizons and the lowered cation exchange capacity of the A horizons as compared to the C horizons.

### CHAPTER VI

### DISCUSSION AND CONCLUSIONS

The Bozeman silty clay loam investigated in this study has developed in an area where environmental factors are within the limits of those characteristic of the extensive Chernozem soil zones. The climate is subhumid, cool temperate and has markedly contrasting seasons. The native vegetation is mixed mid-grass and tall-grass prairie; the topography is undulating to gently rolling, local plains with good surface and internal drainage; and the parent material is deep, permeable, calcareous loess of silt loam texture deposited in late Pleistocene time.

During much of the year, biological, chemical, and physical activity is greatly reduced by cool temperatures or lack of moisture. The period of vigorous plant growth is seldom more than two months. Percolation of moisture beyond the carbonate horizon is very limited. Undoubtedly, the rate of weathering has been comparatively slow. Yet, field and laboratory investigations show clearly that pedogenetic forces have been effective. They have transformed the parent material from its original state at the time of deposition into an organized, anisotropic body clearly related to the environment in which it exists.

Morphologically, the soil is similar to those recognized as medial, Northern Chernozems. The most obvious changes which have occurred, of course, are the accumulation of organic matter in the surface horizons and the leaching of carbonates from the solum to the  $C_{ca}$  horizon. These have long been recognized as fundamental processes in the pedogenesis of Chernozems. It is sufficient to emphasize here that the dark, surface horizons which contain almost all of the organic matter coincide with the zone of abundant roots, and that the carbonates have accumulated mostly on ped surfaces and in tubular pores.

Almost equally obvious is the distinct structure profile, the fabric of which is most strikingly observed under the stereoscopic binocular microscope. Such observation shows that the granular structure of the A horizons is open and porous. Even the blocks which occur in the lower A are porous. In contrast, the ultimate blocks making up the B horizons are devoid of visible pores. Waxy appearing material, the clay, fills the lattice of silt and sand grains. The surfaces of the peds have the appearance of being coated with thin films. The visible pores are restricted to the interconnecting system of voids between adjoining ped faces. Interestingly although 1 ss well developed, the C<sub>ca</sub> also has a blocky fabric similar to the B horizons. The unaltered C horizons have massive, porous fabric.

Complex organic substances and the clays long have been recognized as important binding substances in the formation of soil aggregates. Living organisms, both microbes, and rooting plants, are effective agents in structure formation Perhaps, equally important in the development of the fabric of the Bozeman profile have been the forces associated with the downward percolation of the entering rainfall

and its later removal by evapo-transpiration.

It has been noted in both field and laboratory studies that when soils such as the Bozeman are dry, added water percolates rapidly through the pores between the interfaces of the peds in both the A and B horizons. In the case of the A horizons, the peds themselves also absorb moisture rapidly so that they are usually supersaturated before the excess gravitational water has drained from the horizon. Thus, in its downward movement some of the gravitational water undoubtedly passes through as well as between the peds. This would encourage leaching of mineral and organic constituents and the comminution of the aggregates. It is probably an important force in the development of the characteristic, porous-granular fabric of the A horizons.

In the B horizons, however, moisture is absorbed much more slowly by the peds, and the little water that moves beyond the B horizons very likely passes through the interfacial pore system without entering the peds. Therefore, there is little possibility for the leaching of material from the interior of the peds. Moisture absorbed by the peds is filtered as it enters causing the formation of the clay and organic matter films characteristic of the ped surfaces in the B horizons. Although not thick and strongly oriented in the Bozeman profile, they are quit e apparent under the stereoscopic microscope.

As the moisture is removed by plant roots, forces of surface tension undoubtedly compress the aggregates and make them more stable.



Associated with the development of the characteristic structure profile, there have been changes in volume weights and pore space. In the A horizons the volume weights have decreased while the pore space has increased. In the surface four inches, the present volume weight is only sixty-six per cent of that of the underlying C horizons. Root and frost action, alternate wetting and drying cycles, and increases in organic matter content, as well as the leaching effect of percolating waters, have undoubtedly contributed to the loosening of these horizons.

In contrast the B and the C horizons have somewhat higher volume weights than the underlying C horizons. Both illuviation and settling appear to have occurred. Shrinkage by surface tension and compression by roots have likely been important forces contributing to the latter.

Indications are that the texture of the loess in which the solum and  $C_{ca}$  horizon are developed may have varied at the time of deposition from that of the underlying  $C_2$  and  $C_3$  horizons. However, if one assumes that the greater clay content of the  $C_{ca}$  is due to variation in texture at the time of deposition and uses this horizon to represent the original distrubution of silicate minerals in the parent material of the solum, he still finds that there has been appreciable clay formation during pedogenesis. The relative gain of clay per 100 grams of silts and sands calculated to have been in the horizon originally ranges from 1.9 grams in the  $A_{11}$  to 11.1 grams in the  $B_{21}$ .

The ratios of clay gained to silt and sand lost indicate that
considerable clay from the  $A_{11}$  and  $A_{12}$  and some clay from  $A_{13}$  has moved downward to be deposited in the  $B_{21}$  and  $B_{22}$ . Probably the clay moved has been mostly from the fine fraction and dominantly montmorillonite.

This would account in part for the relative increase of illite in the surface soil but would not explain the very marked increase in randomly interstratified clays. Apparently, forces have been operative which have been conducive to the alternative release and fixation of potassium at random in the internal surfaces of the clays. Both release and fixation must have occurred for the amount of interstratification indicated is so great that both the illite and the montmorillonite must have suffered alteration. In the soil studied the net change has been toward greater fixation. Just what were the forces involved is not clear, but since the phenomenon occurs only in the surface soil, they were likely dominantly biological. Perhaps under conditions of rapid growth fixed potassium is absorbed by the roots of grasses; whereas at other seasons when the rate of mineralization of organic matter and the weathering of primary minerals releases potassium in excess of plant requirements, it is fixed by the clays. The alternate wetting and drying which frequently occurs in summer would encourage fixation.

Perhaps the most useful information in characterizing the soil and tracing the course of pedogenesis was obtained by the micropedological studies. The microscopic study of undisturbed structural aggregates reveals much about fabric changes that have occurred during soil formation that escape the eye in field studies. It permits the relation of many phenomena to specific parts of the soil body in a manner that cannot be accomplished by the usual techniques of physical and chemical analysis. The variation of the environment in which reactions must proceed, from point to point in the soil mass is clearly illustrated. Useful information about many phenomena important to the use and management of soils also can be obtained by such study. The routine use of stereoscopic microscopes in the study of soils in connection with programs of soil classification and survey is to be recommended.

The x-ray diffraction analysis appeared to give satisfactory quantitative measurement of the dominant minerals in the silt and sand fractions. It gave a very satisfactory analysis of the kinds of clay minerals in the less than two-micron silicate fraction of the soil and indicated their relative importance. However, it was not possible to estimate the percentage composition with any degree of certainty solely on the basis of the x-ray data. Judging from their close correlation, cation exchange capacity, specific surface and total potassium content of the clay appeared to be the most useful measurements for estimating the quantitative composition of clay minerals.

The calculation of quantitative differences between horizons was useful in indicating the probably extent to which pedogenetic processes have been active. However, it is evident from the texture and mineralogical studies of the loess that even materials which to all appearances should be uniform, can vary in composition. Care should be exercised in selecting the horizon to represent the primary material of the soil before pedogenesis began.

#### APPENDIX

### A. QUANTITATIVE MEASUREMENT OF CLAY MINERAL COMPOSITION BY X-RAY ANALYSIS

During the course of the study an attempt was made to develop a method of quantitatively estimating the percentage composition of the mineral species making up the clay-size fraction of soils by counting the intensities of the reflections of the basal (001) spacings with the scaler and calculating various ratios between the reflections. The calculations were based on the 7.2 Å, 3.6 Å and 3.3 Å reflections of glycerated Na-clays and heated K-clays. Results were not encouraging. They gave somewhat better indications of clay-mineral composition than visual inspection of strip-chart intensity curves but were far from quantitative.

The procedures used are described in Chapter III. The methods of calculation and results are described below.

 <u>Results</u>: Net x-ray diffraction intensities for the soil clays and pure clay mineral mixtures as measured by the scaler are given in Table 15. These are averages of duplicate counts. Various ratios between peaks are given in Table 16.

The following relationships and conclusions are indicated:

a. The intensities at the 7.2 Å spacing and the ratios of the heated K-clays to the glycerated Na-clays show (1) that the A horizons differ from the B and C horizons and from the pure clays, and (2) that the B and C horizons have about the same



	Gly	ycerated Na	a-clay	K-clay	heated to a	300 <sup>°</sup> C.
Sample	7.2Å	3.6 Å	3.3Å	7.2 Å	3.6 Å	3.3 Å
		Bozema	n Silty Clay	y Loam		
A <sub>11</sub>	18	360	928	191	500	1,983
A <sub>12</sub>	23	532	958	154	376	2,014
A <sub>13</sub>	37	533	973	117	392	2,110
B <sub>21</sub>	149	661	769	215	347	2,108
B <sub>22</sub>	191	812	682	235	396	2,076
C <sub>ca</sub>	149	761	563	173	368	2,005
C2	185	546	584	311	467	1,720
C <sub>3</sub>	171	462	549	307	400	1,714
		Artifici	al Clay Miz	<u>ktures</u>		
100-0-0 <sup>a</sup>	33	227	2,563	91	140	4, 582
0-100-0	41	1,357	33	-62	115	2,455
0-0-100	7,004	4,986	160	9,115	6,339	335
90-5-5	598	748	2,604 <sup>b</sup>	770	635	3, 819
50-25-25	2,138	2,248	994	3, 362	2,914	3, 234
25-50-25	1.873	2,544	444	2,759	2,458	3, 394

## X-RAY DIFFRACTION INTENSITIES FOR THE CLAY FRACTION OF BOZEMAN SILTY CLAY LOAM AND ARTIFICIAL MIXTURES OF CLAYS

<sup>a</sup>Relative proportion of illite, montmorillonite and kaolinite.

<sup>b</sup>The two replicates differed by 800, which is excessive.

### RATIOS OF X-RAY DIFFRACTION INTENSITIES FOR THE SOIL CLAYS AND ARTIFICIAL MIXTURES OF CLAY MINERALS

		-Heated	K-clay		Glyce	rated 1	Na-clay	Hea	ted K-c	lay
	Gl	ycerate	d Na-cl	- ay						
				3.6 Å	36Å	7. 2 Å	3.6 Å	36Å	7.2 Å	3.6 Å
Hor- izon	7. 2 Å	36 Å	33Å	+ 7. 2 Å	33Å	3.3Å	3.3Å	7.2 Å	3.3Å	3.3Å
			Bozem	an Silty	Clay L	oam				
A	10.61	1.39	2.14	1.93	20.00	. 02	. 39	2.62	. 10	. 25
A12	6.70	. 71	2.10	1.60	23, 12	. 02	. 56	2.44	.08	. 19
A <sub>13</sub>	3.16	. 74	2.17	1.66	14.41	.04	. 55	3.35	. 06	. 19
B <sub>21</sub>	1.44	. 52	2.74	1.72	4.44	. 19	. 86	1.61	.10	.17
B <sub>22</sub>	1.23	. 49	3.04	1.66	4.25	. 28	1.19	1.69	. 11	. 19
C	1.16	. 48	3, 56	1.79	5.11	. 27	1.35	2.13	. 09	. 18
C <sub>2</sub>	1.68	. 86	2.94	1.94	2.95	. 32	.94	1.50	. 18	. 27
C <sub>3</sub>	1.80	. 73	3.12	2.09	2.70	. 31	.84	1.30	. 18	. 23
			Ave	erage (S	Soil Clay	<u>y)</u>				
				1.80						
			Artific	ial Cla	y Mixtu	res				
100-0-0			1.79	1.69						
0 - 100 - 0				1.85						
0-0-100	1. 30	1.27		1.30	. 71			. 70		
90-5-5	1.29			1.36				. 83		
50-25-2	5 1. 57			1.90				. 87		
25-50-2	5 1.47			1.96				. 89		
Average (art.	e 1.41 mix.)			1.68				. 82		

ratios as the pure clays which contain kaolinite, thus substantiating the conclusion that the surface soil contains hydrated halloysite and little kaolinite, whereas the subsoil and parent material contain kaolinite and very little halloysite.

- b. The intensities and ratios at the 3.3 Å spacing indicate an increase in illite in the surface soil relative to the subsoil and parent material.
- c. The ratios of the combined intensities of the 3, 6 Å and 3, 3 Å spacings of the heated K-clays indicate that the increase in intensities of montmorillonite and illite after potassium saturation and heating is on the order of 1, 8. The ratios for the pure clays indicate that the increase in intensity for illite is as great as that for montmorillonite.
- d. The ratios between the different spacings of the glycerated Na-clays suggest differences among the three master horizons and similarities among the subhorizons within a master horizon.
- e. The ratios of the 3.6 Å spacing to the 7.2 Å spacing of the heated K-clays of the soil are different from those of the pure clays containing kaolinite. This would indicate the presence of kaolinite group minerals in the soil which have a different structure factor from the pure kaolinite used, or that the bases of the 3.3 Å peaks for the soil clays are sufficiently broad to cause some intensity increase in the 23.3°

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25.  $7^{\circ}$  2  $\theta$  arc. Considering the shape of the strip chart curves, the latter probably accounts for much of the difference. However, it does not account for the difference between the surface soil and the lower horizons.

Table 17 shows the calculated percentage distribution of the clay minerals based upon the assumptions listed at the bottom of the table. Table 18 shows the calculated percentage distribution of the minerals in the pure clay mixtures as compared to the actual. The fit is reasonably good except for the 90-5-5 mixture. In this case, there is a difference of 800 in the intensities at 3.3 <sup>o</sup>A between the replicates of the glycerated Na-clay. Were the calculations based on the intensity of the lower replicate, the montmorillonite ratio would be -148. An increase factor of 2.0 would fit the pure clay data equally well, but this would result in virtually no montmorillonite being calculated for the A horizons. The approximate cation exchange capacities for these horizons would indicate that this is not possible.

In Tables 19 and 20, the percentage distributions are calculated with additional assumptions which permit the calculation of the montmorillonite ratios directly. This has the effect of distributing the error equally between montmorillonite and illite. It should be noted that the combined calculated illite-montmorillonite ratios are generally high as compared to the actual 3.3 Å intensities for the heated K-clays. This again illustrates that ratios from pure clays are not directly applicable to soil clays.

## B. QUANTITATIVE MEASUREMENT OF QUARTZ, FELDSPARS AND CARBONATES BY X-RAY DIFFRACTION ANALYSIS

Various artificially prepared mixtures of pure minerals were analyzed with the x-ray diffractometer as described in the chapter on Procedures. The data are presented in Tables 21 and 22.

The average intensities adjusted to a 100%-quartz standard reading of 3840 were used to plot the curves in Figures 6, 7, and 8. These curves were used for scaling the percentage composition of quartz, feldspars, calcite and dolomite in the whole soil.

Hor- izon	Illite		Montmor- illonite		Kaolinite Halloysite				Total	
	Ratio	%	Ratio	%	Ratio	%	Ratio	%	Ratio	%
A <sub>11</sub>	1, 578	78	405	20	8	0	56	2	2,047	100
A <sub>12</sub>	1,629	78	385	19	11	1	40	2	2,075	100
A <sub>13</sub>	1,654	76	456	22	17	1	22	1	2,149	100
в <sub>21</sub>	1,307	60	801	37	69	3	3	0	2,180	100
B <sub>22</sub>	1,159	54	919	42	89	4	0	0	2,167	100
C	957	46	1,048	51	70	3	0	0	2,075	100
c <sub>2</sub>	993	53	727	41	86	5	18	1	1,824	100
$c_3$	933	51	781	43	80	5	22	1	1,816	100

# PERCENTAGE DISTRIBUTION OF IDENTIFIABLE CLAY MINERALS IN SOIL PROFILE CALCULATED ACCORDING TO ALTERNATIVE ASSUMPTIONS LISTED

## Assumptions:

1. The 3.3 Å peak of glycerated Na-clay is due to illite. It times 1.7 equals the intensity for illite in the 3.3 Å heated K-clay peak.

2. The 3.3 Å spacing of heated K-clay is due to illite and montmorillonite combined.

3. The intensity of the 7.2 Å peak for the glycerated Na-clay is due to kaolinite. This peak for heated K-clay is due to kaolinite and halloysite combined. Intensity of kaolinite in the 7.2 Å peak after K-heat treatment is 1.4 times that in glycerated Na-clay.

4. After potassium saturation and heat treatment, the relative intensity of the 7.2 Å peak for kaolinite is three times that of the 3.3 Å peak for montmorillonite and illite.

## Calculation of Ratios:

- a. Illite = 3.3 Å glyc. X 1.7.
- b. Montmorillonite = 3.3 Å heated (a).
- c. Kaolinite + halloysite = 7.2 Å heated X 0.33.
- d. Kaolinite = 7.2 Å glyc. X 1. 4 X 0. 33.
- e. Halloysite = (c) (d).



Actual	Calculated Composition									
sition	Illite		Montmori	illonite	Kaol	inite	Total			
	Ratio	%	Ratio	%	Ratio	%	Ratio	%		
100-0-0	4. 357	95	225	5	15	0	4, 597	100		
0-100-0	56	2	2,399	97	19	1	2,474	100		
0-0-100	272	7	63	2	3, 269	91	3,604	100		
90-5-5	4. 427	95	-608	?	279	5	4, 706	100		
50-25-25	1,690	41	1,544	37	997	22	4,236	100		
25-50-25	755	18	2,639	61	907	21	4,301	100		

CALCULATED AND ACTUAL PERCENTAGE DISTRIBUTION OF CLAY MINERALS IN ARTIFICIAL MIXTURES OF PURE CLAYS ACCORDING TO ASSUMPTIONS LISTED BELOW TABLE 17

\* In the order illite, montmorillonite, kaolinite.

Hor- izon	Illite		Montmor-		Kaolinite		Halloysite		Total	
	Ratio	%	Ratio	%	Ratio	%	Ratio	%	Ratio	%
A11	1, 578	69	588	26	8	0	56	3	2,230	100
A12	1,629	64	874	34	11	0	40	2	2,554	100
A13	1,654	65	855	33	17	1	22	1	2,548	100
B <sub>21</sub>	1, 307	57	921	40	69	3	3	0	2,300	100
B22	1,159	48	1,120	48	89	4	0	0	2,368	100
C_	957	46	1,091	51	70	3	0	0	2,118	100
c,	993	57	677	38	86	4	18	1	1,774	100
C <sub>3</sub>	933	59	553	35	80	5	22	1	1, 588	100

#### PERCENTAGE DISTRIBUTION OF IDENTIFIABLE CLAY MINERALS IN A PROFILE OF BOZEMAN SILTY CLAY LOAM CALCULATED ACCORDING TO ASSUMPTIONS LISTED

#### Assumptions:

1. The 3.3 Å peak of glycerated Na-clay is due to illite. It times 1.7 equals the intensity for illite in the 3.3 Å heated K-clay peak.

2. The 3.6 Å peak of glycerated Na-clay is due to montmorillonite and kaolinite.

3. The intensity of the 7.2 Å peak for the glycerated Na-clay is due to kaolinite. This peak for heated H-clay is due to kaolinite and halloysite combined. Intensity of kaolinite in the 7.2 Å peak after K-heat treatment is 1.4 times that in glycerated Na-clay.

4. After K saturation and heat treatment, the relative intensity of the 7.2 Å peak for kaolinite is three times that of the 3.3 Å peak for montmorillonite and illite.

5. Remainder of 7.2 Å peak after K-heat treatment is due to halloysite with the same intensity ratio to 3.3 Å peak as kaolinite.

Calculation of Ratios:

a. Illite = 3.3 Å glyc. X 1.7

b. Kaolinite = 7.2 Å glyc. X 0.33 X 1.4

c. Kaolinite + halloysite = 7.2 Å heated X 0.33.

d. Halloysite = (c) - (b)

e. Montmorillonite = (3.6 Å glyc. - (7.2 Å glyc. X 0.8)) X 1.7

# CALCULATED AND ACTUAL PERCENTAGE DISTRIBUTION OF CLAY MINERALS IN ARTIFICIAL MIXTURES OF PURE CLAYS ACCORDING TO THE ASSUMPTIONS LISTED BELOW TABLE 19

Actual	Calculated Composition										
compo- sition <sup>*</sup>	Illite		Montmor	illoni	te Kaol	inite	Total				
	Ratio	%	Ratio	%	Ratio	%	Ratio	%			
100-0-0	4, 357	92	362	8	15	0	4, 755	100			
0-100-0	56	2	2,251	98	19	0	2,353	100			
0-0-100	272	8	-612	?	3, 269	.92	8,116	100			
90-5-5	4, 427	90	270	5	279	5	5, 366	100			
50-25-25	1,690	52	538	19	997	31	4, 622	100			
25-50-25	755	28	1,046	39	907	33	3, 898	100			

\* Percentage in the order illite, montmorillonite, kaolinite.



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	-Peak height i			
Percent quartz	100% quartz standard	Mineral mixture	Adjusted average intensity <sup>1</sup>	
100			3840 <sup>2</sup>	
50	1000	22.40	2150	
50	4000	2240	2150	
	4064	2208	2093	
			and the second	
			2150	
40	3936	1952	1904	
	3984	1920	1849	
	3968	1760	1703	
			1819	
20	3984	1008	970	
	4000	1136	1090	
	3968	960	929	
			996	
10	3600	576	608	
	3616	640	680	
	3648	608	640	
			643	

# X-RAY PEAK-HEIGHT INTENSITIES OF QUARTZ IN PURE MINERAL MIXTURES

 $^{\rm l}$  Readings adjusted to a 100% quarts standard of 3840.

<sup>2</sup>Average of 31 readings ranging from 3616 to 4126.

		Adjusted peak-area intensities <sup>1</sup>					
Mineral	Per cent of mixture		Average				
Orthoclase	100	2.10	2,21	2.15	2.15		
Microcline	100	2.47	2.52	2.49	2.50		
Albite	100 90 30	2.12 2.53 0.25	1.82 2.75 0.31	1.99 2.56 0.31	1.97 2.61 0.29		
Andesine	100	1.45	1.89	1.51	1.62		
Labordorite	100 30	1.63 0.46	1.52 0.36	1.56 1.58	1.57 0.47		
Anorthite	100	1.22	1.09	1.00	1.10		
Albite and labordorite	60	0.77	0.95	0.91	0 <b>. 88</b>		
Calcite	100 30 20	1.97 1.12 0.57	1.65 0.96 0.61	2.11 0.93 0.67	1.91 1.00 0.62		
Dolomite	100 20	3.74 0.61	3.43 0.63	3.26 0.73	3.48 0.65		

# X-RAY DIFFRACTION PEAK-AREA INTENSITIES FOR FELDSPARS, CALCITE AND DOLOMITE IN PURE MINERAL MIXTURES

TABLE 22

Adjusted to a quartz standard reading of 3840.





FIGURE 6: RELATION OF PERCENT QUARTZ IN STANDARD WIXTURES TO PEAK-HEIGHT INTENSITY AT THE 3.35 Å PEAK



FIGURE 7: RELATION OF TOTAL FELDSPARS IN STANDARD MIXTURES TO PEAK-AREA INTENSITIES IN THE ARC 27.0° to 28.6° TWO THETA



PIGURE 8: RELATION OF CALCITE AND DOLOMITE IN STANDARD MIXTURES TO PEAK-AREA INTENSITIES AT THE 2.88 Å and 3.04 Å PEAKS

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